Proceedings of the ICAR5/GCTE-SEN Joint Meeting

Joint meeting of the Fifth International Conference on Aeolian Research and
The Global Change & Terrestrial Ecosystem-Soil Erosion Network

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Editors:

Jeffrey A. Lee
Dept. of Economics & Geography
Texas Tech University
Lubbock, TX, USA  79409-1014
jeff.lee@ttu.edu

Ted M. Zobeck
Wind Erosion and Water Conservation Unit
USDA, Agricultural Research Service
3810 4th Street
Lubbock, TX USA 79415
tzobeck@lbk.ars.usda.gov

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Cover Photo: West Texas Dust Storm, 1930s. Photo courtesy Texas Tech University Southwest Collections.

Special thanks to Linda Jones, Amy Bownds and Melissa Nance for editorial assistance in the preparation of this volume.
Dedicated to the Memory of  
Dr. Harold E. Dregne  
Distinguished Professor – Emeritus

Harold E. Dregne was born on September 25, 1916 in Ladysmith, Wisconsin. He received his B.S. degree in Chemistry and Mathematics from the University of Wisconsin at Stevens Point (1938), and M.S. degree in Soil Chemistry from the University of Wisconsin in Madison (1940), and a Ph.D. from Oregon State University in Soil Chemistry (1942). After graduating he was a soil scientist with the Soil Conservation Service and served in the U.S. Navy as a Radio-Radar Officer from 1942-46.

His academic career included work at the University of Idaho, Washington State University, and New Mexico State University. He came to Texas Tech University as a professor and Chairman of the Department of Agronomy in 1969. In 1976, he was named Director of International Center for Arid and Semiarid Land Studies. Since 1985 he served as a distinguished Horn Professor Emeritus and Special Consultant to the ICASALS. Dr. Dregne’s major academic interests were arid land development and degradation on a global basis. He authored 10 books and over 170 scientific articles and he received many awards during his distinguished career.

Dr. Dregne was internationally known for his work on desertification. He loved working in arid and semiarid environments and was always traveling and promoting and performing research in these areas. His favorite places to visit were Vienna and all of China. He consulted extensively across the globe for individual countries and for several major organizations including the Food and Agriculture Organization of the UN (FAO) and the United Nations Environment Program (UNEP). Dr. Dregne was respected and loved by all who knew him. His genuine interest in students, his compassion and professionalism have left an indelible mark on Texas Tech University and the international arid and semiarid research community. We are all honored to have shared the life of such a wonderful man and distinguished scientist and dedicate this meeting and Proceedings in his memory.
The meeting has been sponsored by:

USDA, Agricultural Research Service
Texas Tech University, Office of International Affairs, International Center for Arid and Semiarid Land Studies
The Soil and Water Conservation Society, Golden Spread Chapter
United States Geological Survey

WELCOME

Welcome to the Proceedings of the ICAR5/GCTE-SEN, a joint meeting of the Fifth International Conference on Aeolian Research (ICAR 5) and the Global Change and Terrestrial Ecosystems, Soil Erosion Network (GCTE-SEN). Previous ICAR meetings have been held in Aarhus, Denmark (1985), Sandbjerg, Denmark (1990), Zyzxx, CA, USA (1994), and Oxford, U.K. (1998). The ICAR conferences attract aeolian geomorphologists, physical scientists, soil scientists, and erosion specialists from around the world to discuss the latest challenges and discoveries of aeolian research.

The GCTE-SEN is a core project of the International Geosphere-Biosphere Programme. The thrust of the GCTE-SEN is twofold: 1) To design and undertake experimental and monitoring programs to provide a predictive understanding of the impacts of changes in climate and land-use on soil erosion and 2) To refine and adapt current erosion models (for use in global change studies) from plot to regional scales. The network was developed to encourage international collaborations and has been very successful. There have been several meetings in support of GCTE-SEN evaluating, testing and comparing water erosion models but very little has been done in wind erosion and aeolian processes. More information on the network can be found on their web page at (http://mwnta.nmw.ac.uk/GCTEFocus3/networks/erosion.htm).

MEETING ORGANIZERS:

Meeting Conveners: Ted M. Zobeck, Jeffrey A. Lee, Thomas E. Gill, John E. Stout

Program and Planning Committee:
Ted Zobeck*, Committee Co-Chair, USDA, Agricultural Research Service, tzobeck@lbk.ars.usda.gov
Jeff Lee, Committee Co-Chair, Texas Tech University, USA, jeff.lee@ttu.edu
Tom Gill*, Texas Tech University, USA, tom.gill@ttu.edu
John Stout, USDA, Agricultural Research Service, jstout@lbk.ars.usda.gov
Cheryl McKenna-Neuman*, Trent University, Canada, cmckneuman@trentu.ca
Andrew Warren, University College, UK, andrew.warren@micl.lu.se
Yaping Shao*, City University of Hong Kong, apyshao@cityu.edu.hk
Giles Wiggs, Univ. of Sheffield, UK, G.Wiggs@sheffield.ac.uk
PROGRAM HIGHLIGHTS

The program included a mix of poster and oral presentations. A list of the session topics and session coordinator follows.

Session 1. Fundamental Aeolian Processes - Dr. Bill Nickling, University of Guelph, Canada nickling@uoguelph.ca, and Dr. Jack Gillies, Desert Research Inst., Reno, NV jackg@dri.edu

Session 2. Instrumentation/Measurement in the Field and Lab – Dr. Cheryl McKenna-Neuman, Trent University, Canada, cmckneuman@trentu.ca

Session 3. Field Studies – Dr. Ted M. Zobeck, Committee Co-Chair, USDA, Agricultural Research Service, USA, tzobeck@Lbk.ars.usda.gov

Session 4. Modeling - Dr. Tom Gill, Texas Tech University, USA, tom.gill@ttu.edu, and Dr. Yaping Shao, City University of Hong Kong, apyshao@cityu.edu.hk

Session 5. Environmental Impacts and Erosion Control – Dr. Larry Hagen, USDA, Agricultural Research Service, HAGEN@weru.ksu.edu

Session 6. Landforms and Aeolian Paleoenvironments – Dr. Ian Livingstone, University College Northampton, UK, ian.livingstone@northampton.ac.uk and Dr. Nicholas Lancaster, Desert Research Inst., Reno, NV nick@dri.edu

Session 7. Dunes and Related Landforms: Issues and New Approaches - Dr. Nicholas Lancaster, Desert Research Inst., Reno, NV nick@dri.edu
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Scaling Effects of Standing Crop Residues on Aerodynamic Transfer Processes

R. M. Aiken, Kansas State University, Colby, Kansas 67701, raiken@oznet.ksu.edu
D. C. Nielsen, USDA-ARS, Akron, Colorado, dnielsen@lamar.colostate.edu
L. R. Ahuja, USDA-ARS Fort Collins, Colorado 80926, ahuja@gpsr.colostate.edu

Introduction

Standing senescent stems increase the aerodynamic roughness of the surface, reducing wind energy available for momentum transfer at the soil surface, such as for wind erosion, and also the soil-atmosphere convective exchanges of heat, water vapor, and trace gases. These roughness elements alter convective exchanges and near-surface (<0.05 m) wind velocities by absorbing kinetic energy and modifying aerodynamic roughness. These effects are readily quantified as a log-linear decrease in wind velocity relative to distance above the land surface. The slope of this relationship reflects the friction velocity, while the intercept can be interpreted as the aerodynamic roughness of the surface, or roughness length. Vertical stems tend to raise, or displace, the level of near-zero wind velocity; while increasing aerodynamic roughness and altering friction velocity (Pereira, and Shaw, 1980). Though displacement height and aerodynamic roughness are phenomenological coefficients, they tend to scale with crop canopy characteristics (height: Campbell, 1973; leaf area: Choudhury and Monteith, 1988). Analogous relationships exist between residue architecture (horizontal projected stem area) and threshold velocities required to initiate soil erosion (Hagen, 1996).

Our research objective was to derive a modified algorithm, which quantifies effects of standing stems on wind profiles above and within sparse canopies and to conduct field measurements in standing residues of wheat, corn, millet, and sunflower for wind profiles and geometries to validate the modified algorithm.

Method and Materials

Extending wind profile theory to sparse canopies of standing crop stems requires a procedure to quantify the aerodynamic parameters \(d\) and \(z_p\). We hypothesize that in sparse canopies, these effects are proportional to silhouette area index (SAI), the horizontal projected area of roughness elements per unit of land area (Nielsen and Aiken, 1998) analogous to similar relations derived for crop canopies. Specifically, we extend the algorithm of Choudhury and Monteith (1988) to standing stems, specifying \(d/h\), relative displacement height, as a function of aerodynamic drag \((C_f, \text{ dimensionless})\) and SAI.

\[
\frac{d}{h} = 1.1 * \ln \left( 1 + (C_f \cdot SAI)^{0.25} \right)
\]  \hspace{1cm} (1)

Following Shuttleworth and Gurney (1990), we compute \(z_p\) as the sum of roughness lengths for standing stems \((z_{s(s)})\) and surface \((z_{s(s)})\) layers, where \(z_{s(s)}\) is represented, according to Choudhury
and Monteith (1988), as
\[ \frac{z_{e(n)}}{h} = a \cdot \left(C_{fa} \cdot SAI \right)^{0.5} \quad \left(C_{fa} \cdot SAI \right) < 0.2 \]
\[ \frac{z_{e(n)}}{h} = a \left(1 - \frac{d}{h} \right) \quad \left(C_{fa} \cdot SAI \right) > 0.2 \]

with the value of \( a \) set to 0.3. Here the aerodynamic drag coefficient \( C_{fa} \) represents form drag of individual residue elements, perpendicular to fluid flow, distinguished from skin drag, tangential to fluid flow. We compute SAI from
\[ SAI = d_s \cdot h \cdot N \]

where \( d_s \) is stem diameter (m), \( h \) is stem height (m), and \( N \) is number of stems per square meter.

We conducted field studies to determine the predictive accuracy of an algorithm derived for plant canopies to scale effects of standing crop residues on the wind profile. We used this algorithm to calculate displacement height and roughness length of standing crop residues related to the log wind profile equation. We also calculated apparent roughness length from wind profiles measured under neutral stability conditions over stems of wheat (Triticum aestivum), corn (Zea mays), millet (Panicum miliaceum), and sunflower (Helianthus annuus) using calibrated single-needle and cup anemometers at up to 10 heights ranging from 0.07 m to 2.40 m above the soil surface on fields with fetch:height ratios exceeding 200:1. We compared roughness length and wind profiles computed using Eqs. 1 and 2, and measured standing stems, with those computed from field observations of wind profiles under neutral stability conditions.

**Results and Discussion**

Residues at the selected experimental sites were typical of those found in semi-arid cropping systems. Sunflower stubble represented the simplest system, with roughness elements approximating the shape of thin vertical cylinders. Corn stubble, comprised of husks, leaves, and broken stems, added complexity to the roughness elements. The tillering growth habit of wheat added to row orientation effects, resulting in a stiff hedge structure. The millet field was planted on ridges (height of 30 mm and spacing of 0.21 m) into standing wheat stubble.

A least-squares fit of roughness length calculated by an algorithm derived for crop canopies indicated a systematic, positive bias when it was applied to standing stems (Fig. 1). After adjusting for bias, calculated windspeeds generally were contained in 80% confidence intervals for observations above and within the crop stubble (Fig. 2). Predictive root mean square errors (RMSE) within profiles ranged from 0.5 to 4.6% of reference wind speed. The adequate fit is expected, because the revised coefficient, \( a \), for Eq. 2 was derived from wind profiles observed above the roughness elements. The nonlinear forms of the scaling algorithms are consistent with theory and wind tunnel observations, representing an advance over schemes assuming a linear relation with residue height.

The scaling approach represented by Eqs. 1 and 2 is adequate to quantify effects of standing stems on wind speed profiles above and within these roughness elements. Further evaluation of the coefficient \( a \) used in Eq. 2 is warranted, because we used the same profile data
to derive the coefficient and to evaluate subsequent wind speeds. Further work also is required to evaluate the adequacy of Eqs. 4 and 5 for drag partitioning and to investigate aerodynamic properties of complex surfaces containing ridges and standing stems. The similarity between our results and those of Raupach (1992) and Hagen (1996) indicates that the algorithms may be suitable for process-level wind erosion and drag partitioning, though further work is warranted. Application to momentum transfer problems requires further investigation of drag partitioning.

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Figure 1. Roughness length, scaled by height of standing stems and depicted in relation to canopy drag. The continuous function was calculated from Eq. 2 using suggested (Choudhury and Monteith, 1988) and fitted values for the coefficient ‘a’. Observed roughness length and 80% confidence intervals constructed from standard error about the means were calculated from wind profiles over standing crop residues.

Figure 2. Relative wind speed, scaled to wind speed at reference height (2.4 m) above and within standing stems of wheat, sunflower, corn, and millet. Height is presented on the vertical axis; arrows indicate height of standing stems. The continuous function was calculated from the wind profile equation parameterized by Eqs. 1 and 2 using a fitted value of 0.24 for the coefficient ‘a’. Observed wind speeds and direction relative to row orientation are depicted with 80% confidence intervals constructed from standard errors about the means.
Estimation of PM20 Emissions by Wind Erosion:
Main Sources of Uncertainties

Stephane C. Alfaro, LISA, UMR-CNRS 7583, Universite de Paris 12 (E-mail: alfaro@lisa.univ-paris12.fr)
Jean Louis Rajot, IRD LISA, UMR-CNRS 7583, Universite de Paris 12 (E-mail: rajot@lisa.univ-paris12.fr)

Introduction

Particles that can be observed a few centimeters above a surface undergoing wind erosion cover a wide range of diameters - from about 0.1 µm to several hundreds of µm. After they have been lifted from the surface, their fate mostly depends on their weight. In usual conditions, air flow turbulence can only maintain particles larger than 20 µm in suspension for a short time (less than a few hours) and they rapidly fall in the vicinity of the place they were lifted from. In consequence, the corresponding mass redistribution is of some importance at local scale only. On the contrary, particles finer than 20 µm (PM20) can be transported over long distances - hundreds or thousands of kilometers in the case of the smallest whose residence time in the lower troposphere can reach a week. This has several effects: 1) fine particles are the richest in soil-nutrient, and their departure from semi-arid areas can further deplete stocks of already poor soils, 2) fine particles being easily inhaled in the respiratory tract they have an impact on human or animal health, and 3), while they are suspended in the atmosphere PM20 affect transfer of solar and terrestrial radiation. This direct effect is presently one of the major source of uncertainties in climate modeling. Quantifying effects 1 and 3 requires estimations of PM20 mass fluxes at field scale. Relatively numerous studies have been dedicated to this problem, but their results have not yet allowed to ascertain which parameters, among the numerous ones apparently involved (soil texture, soil composition, soil-aggregate size distribution, soil roughness, wind friction velocity, humidity ...), are the more relevant. The aim of this work is to show that it is possible to pinpoint key parameters, and to explain much of the apparent variability of field measurements, by using a dust production model (DPM) based on a physically explicit parameterization of the aeolian processes leading to fine dust emissions.

Present understanding of the physics of aeolian processes

Saltation, splashing and sandblasting processes

Chatenet et al. (1996) have shown that the loose wind-erodible fraction of arid soils can usually be considered as a mixture of at most three lognormally distributed soil-aggregate populations. The geometric mean diameter (gmd) and geometric standard deviation (gsd) of the smallest of these modes are 125 µm and 1.6, respectively. This shows that the mass of PM20 present in arid soils in a free state is insignificant. These fine particles, that indeed exist within the soil, may be contained in two types of aggregates. Indeed, they can either be glued to the surface of sand-sized grains or imbedded in aggregates of fine material. When aggregates are set into motion by wind strength, their movement (saltation) remains essentially horizontal because of their important weight. At the downwind end of their trajectories their kinetic energy is partly transformed into heat in inelastic shocks, partly used to eject other aggregates from the soil (splashing), and the rest is used to release PM 20 from the aggregates or from the
soil surface (sandblasting). There is much experimental evidence (e.g., Gillette, 1977; Shao et al., 1993; Houser and Nickling, 2001) to prove that direct mobilization by aerodynamic forces plays a minimal role in PM20 emissions and that these emissions can be considered as a direct consequence of saltation. Thus, from a formal point a view, it is natural to think of disconnecting the study of sandblasting from the one of saltation. This idea was first developed by Gillette (1977) who compared sandblasting efficiencies ( ) of various natural soils in the southwestern part of the USA. These sandblasting efficiencies were defined as ratios of measured PM20 vertical fluxes to measured horizontal saltation fluxes (Fh). Of course, the same idea of uncoupling saltation and sandblasting can be applied to modeling.

Physics of saltation

After Bagnold (1941), the basic knowledge of the saltation process was derived from wind tunnel simulations performed in ideal conditions. Indeed, experiments were carried out in dry conditions, and at controlled wind speed, over sand beds made of grains having all the same size and deprived of non-erodible elements. Various expressions accounting for the dependence of the horizontal saltation flux to the wind friction velocity were proposed (see Greeley and Iversen, 1985, for a review). These expressions involve a parameter of crucial importance for saltation, the threshold friction velocity’ (u*) under which wind stress is too low to set aggregates into motion. Well above this threshold the horizontal flux is proportional to u*3. In the case of loose beds, that is to say when all inter-particle bonds but electrical ones can be neglected, u* depends on particle characteristics such as size (D) and density ( ), but it also depends on the degree of protection brought to the soil by non erodible elements (pebbles, boulders, vegetation, …). The effect of these non erodible elements is to increase the soil roughness length (Z0), and consequently u*t for each soil aggregate size class. Z0 can be used as a proxy to model the influence of non erodible elements on u*t (Marticorena and Bergametti, 1995; Alfaro and Gomes, 1995). Humidity has also several important effects. First, it can increase u*t indirectly by promoting vegetation growth. This effect can, as before, be taken into account by the means of Z0. Secondly, it can enhance the strength of inter-particle bonds in two ways: by promoting development of a humid film between grains (Fecan et al., 1999), or by favoring soil crusting. In the latter case, soil texture and composition are key parameters. For example, the type of physical crust that develops on soils with a very low content in fine particles (sandy soils) does not affect saltation (Rajot et al., in press), while stronger crusts that form on clayey or loamy soils are much more efficient at limiting the availability of soil-aggregates for saltation (Gillette, 1988; Sterk et al., 1999; Gomes et al., in press).

Physics of sandblasting

Wind tunnel simulations of wind erosion realized with different natural soils collected in source areas have shown that the PM 20 that are ejected from soil aggregates by sandblasting can always be considered as a mixture, in various proportions, of only three lognormally distributed populations (Alfaro et al., 1998). In first approximation, the size characteristics of these populations seem to be independent of the soil texture and mineral composition. Moreover, the experiments showed that the largest particles could be released even at slow wind speeds, but that it took increasingly larger energies to produce the second finest, and finest types of particles. In other words, the finer the particles, the higher their binding energy (e;) within the soil aggregates.

Modeling

The Dust Production Model
A physically explicit saltation model incorporating the effect of non-erodible elements, and of the humidity films was first developed (Marticorena and Bergametti, 1995). It allows computation of the size resolved saltation mass flux from input parameters that are the following: soil roughness length, dry size distribution of the loose soil aggregates, humidity, and wind friction speed. To this date, the model does not yet account for crusting of fine textured soils after wetting. In a second step, a sandblasting model based on a scheme describing the partition of the soil aggregates kinetic energies between the binding energies of the 3 PM20 populations was proposed (Alfaro et al., 1997). The $e_i$ values were also derived from the experiments. Then, a Dust Production Model (DPM) was obtained (Alfaro and Gomes, 2001) by combining the saltation model to the sandblasting one. In the DPM, the energy partition scheme is applied to each size class of the saltation flux. Integration of the results over the full size range of saltating aggregates, then yields the vertical number (and mass) flux of PM20 ($F_{v}$) and its size distribution. The saltation part of the model also provides $F_{h}$. Thus, the model also yields $F_{h}/F_{v}$ as the ratio of these two fluxes. It has to be noted that, since the characteristics of the three PM20 populations are provisionally considered as fixed, the input data required by the DPM are the same than those necessary to run the saltation model (see above).

The DPM has been validated by comparing predictions of its two (saltation and sandblasting) sub-models to direct measurements performed on a sandy soil in Niger and in northeast Spain (Gomes et al. b, in press).

Implications of the model: sensibility of the PM20 flux to key variables

Efforts have been made by numerous authors to express PM20 vertical flux as a function of $u^*$. It is generally considered that $F_{v}$ is a power function of $u^*$. Due to important scatter of experimental data collected on the field, it has not been possible to find a unique value for the exponent ($n$) of this power law. Thus, attempts have been made to sort experimental data according to soil texture in order to provide $n$ values for various textural groups. For example, Nickling and Gillies (1989) have found that $n = 3.03$ for soils with a silt and clay content lower than 15%, but that $n = 4.27$ for soils with more than 25% silt and clay. But even so, an important data scatter remains and some authors have been led to doubt that $u^*$ is a relevant parameter to estimate dust emissions from a given field (Houser and Nickling, 2001). We propose to reexamine the problem with the insight provided by the physical descriptions of saltation and sandblasting that constitute the backbone of the DPM.

By definition of $F_{v}$, $F_{v}$ can be expressed as:

$$F_{v} = (D, u^*, Z_0, e_i) F_{h}(D, u^*, Z_0, h, C) \quad (1),$$

where $F_{v}$ depends on 1) the size ($D$) of the saltating soil aggregates, and hence the size distribution of the parent soil, 2) the wind friction velocity, and 3) the soil roughness length. It also depends on the PM20 binding energies that have been considered provisionally as constant in a first step. As seen above, $F_{h}$ also depends on $u^*$, $D$, $Z_0$, humidity ($h$), and $C$ that represents the degree of limitation due to crusting ($C = 1$ for no crusting effect).

Tests meant to determine the sensibility of $F_{v}$ to its parameters have shown (Alfaro and Gomes, 2001) that the effect of $Z_0$ is to increase saltation, and hence sandblasting, thresholds. The size distribution of the soil aggregates has the most important effect on the $F_{v}$ behavior. When the soil loose fraction contains either one of the two smaller fine sand populations (identified by Chatenet et al. (1996)), the influence of this mode is predominant and, after a sharp increase just above sandblasting threshold, $F_{v}$ becomes approximately independent of $u^*$. As a consequence, well above threshold, $F_{v}$ becomes more or less proportional to $F_{h}$, that is to say proportional to $u^*^3$ (Fig. 1). When the erodible fraction of the soil contains neither of the
fine sand modes and is only made of the coarsest ones (520 or/and 690 µm modes of Chatenet et al.), does not stabilize after the fast initial increase but regularly decreases with $u^*$. Equation (1) then implies that well above threshold $F_v$ increases less rapidly than in the previous case. It is found (Fig. 1) that $F_v$ approximately increases as $u^{*2}$.

![Graph showing PM20 fluxes computed by the DPM for two soils made of fine aggregates (125 µm, bold line), or coarse aggregates (690 µm, thin line) populations. Soils are considered as deprived of nonerodible elements.](image)

Fig. 1: PM20 fluxes computed by the DPM for two soils made of fine aggregates (125 µm, bold line), or coarse aggregates (690 µm, thin line) populations. Soils are considered as deprived of nonerodible elements.

**Summary and conclusion**

Above results indicate that, contrary to a common assumption, the aptitude of a soil to release PM20 does not depend directly on its texture but rather on 1) the roughness length that conditions the emission threshold, and 2) the size distribution of the loose soil aggregates available for saltation. More precisely, when a fine sand component (soil aggregates smaller than 210 µm) is available for saltation, PM20 production is enhanced by at least an order of magnitude relative to cases where only coarser aggregates are present in the topsoil. Moreover, the soil dry size distribution also conditions the way $F_v$ increases with $u^*$. In all cases, increases quite rapidly with $u^*$ just above threshold and equation (1) implies that $F_v$ then increases more rapidly than $F_h$, that is much more rapidly than $u^{*3}$. At higher friction speeds the rate of increase of $F_v$ with $u^*$ progressively goes down. Though this is not mathematically true, $F_v$ can be considered in first approximation as tending towards a power function of $u^*$. Depending on the absence, or presence, of fine sand aggregates in the topsoil, the exponent of this function varies respectively from 2 to 3. The fact that, even for a given soil of fixed roughness and dry size distribution, $F_v$ cannot be considered as a power function over a wide $u^*$ range explains in part the large scatter obtained when trying to plot $F_v$ versus $u^*$ in a log-log scale. Indeed, field measurements of PM20 fluxes made in natural conditions are usually performed over the widest possible range of wind speeds. Another source of scatter is the
duration of field campaign. Many factors (rain, vegetation growth, changes in microtopography, in wind direction, …) that lead to changes in soil roughness or/and in the degree of crusting could also explain data scatter. When comparing PM20 fluxes measured at different places, grouping of the sites should be made according to the dry size distribution of the topsoil rather than according to soil texture that is not directly a relevant parameter for dust emissions.

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Analysis of velocity profiles from wind tunnel experiments with saltation

B.O Bauer, Department of Geography, University of Southern California, Los Angeles, California, USA 90089-0255 (E-mail: bbauer@usc.edu)

C. A. Houser, Scarborough College Coastal Research Group, University of Toronto, Scarborough, Ontario, Canada M1C 1A4 (E-mail: houser@scar.utoronto.ca)

W.G. Nickling, Department of Geography, University of Guelph, Guelph, Ontario, Canada N1G 2W1 (E-mail: nickling@uoguelph.ca)

Introduction

In the past two decades, increasing effort has been directed at probing and modeling the internal dynamics of saltation layers. Wind tunnels have proven indispensable in this regard, especially for testing the legitimacy of various ideas about sediment-fluid interaction under tightly controlled experimental conditions. Unfortunately, it is not always apparent how to analyze and interpret even relatively simple measurements such as the vertical profile of time-averaged wind speed. Complete descriptions of the mean streamwise velocity distribution, \( U(z) \), in a turbulent boundary layer are generally derived by a classical asymptotic matching procedure that requires the inner layer profile (Law of the Wall) to be matched to the outer layer profile (Velocity Defect Law) in an overlap region or ‘inertial sublayer’ (Raupach et al., 1991), which leads to the well-known logarithmic law,

\[
\frac{U(z)}{u_*} = \frac{1}{\kappa} \ln \left( \frac{z}{\nu} \right) + B
\]

Equation (1)

where \( u_* \) is shear (or friction) velocity, \( \kappa \) is the von Karman constant (taken as 0.41), \( z \) is height above the bed, \( \nu \) is kinematic viscosity, and \( B \) is an integration constant that depends on surface roughness and shear velocity. In most earth science applications, the logarithmic law takes an alternative, more general form that employs \( z_0 \), the roughness length or height above the bed at which flow velocity tends to zero.

\[
\frac{U(z) - U(z_0)}{u_*} = \frac{1}{\kappa} \left[ - \ln \left( \frac{z}{\delta} \right) + 2 \Pi + 2 \Pi \sin^2 \left( \frac{\pi z}{2 \delta} \right) \right]
\]

Equation (2)

where \( U_\infty \) is free stream velocity, \( \delta \) is boundary layer depth, and \( \Pi \) is a profile parameter that depends on the distribution of stress in the boundary layer. This expression applies to the entire boundary layer above \( z_0/\nu = 30 \) presuming that \( \Pi \) is constant (Spies et al., 1995). A value of \( \Pi = 0.55 \) for clean air flows has been widely adopted, but Spies et al. (1995) suggest that a value of \( \Pi = 0.6 \) may be more appropriate when aeolian sand transport is active.
It is widely appreciated that the presence of moving sediment in the near-surface region of the boundary layer alters the fluid dynamics of the inner layer. Nevertheless, the logarithmic law continues to be applied widely, either in the form of the equations presented above or those proposed by Bagnold or Owen, which were derived specifically for aeolian saltation systems. The challenge facing the aeolian geomorphologist, and the subject of this paper, is to determine which of these many alternative expressions provides the most realistic and complete description of a measured velocity profile whether from the field or wind tunnel.

**Methods**

A series of runs was conducted in the University of Guelph recirculating wind tunnel with fine-grained (D$_{50} = 0.19$ mm) and coarse-grained (D$_{50} = 0.25$ mm) quartz sand. For each run, the sand bed was flattened by running a straight-edged bar along the tops of two metal side rails fixed to either side of the tunnel. The fan motor was set to a constant frequency, the sediment-feed system was then set to a supply rate that was sufficient to preclude sediment build-up or erosion beneath the hopper, and the transport system was given ample time for an equilibrium surface to establish itself (with low-amplitude ripples in most cases). Sediment transport rate was measured continuously using a wedge-type trap and high-precision electronic balance connected to a computer-controlled data-acquisition system.

Wind speed was measured using a high-speed thermal anemometry system (TSI 300) and a stainless-steel hot-film probe (TSI Model 1266). The probe was fixed to a precision rack-and-pinion mount that was located in the center of the tunnel immediately in front of the wedge trap. The probe was lowered toward the sand surface and then raised back up to the free-stream core by reversing the steps. After a vertical wind-speed profile was measured in its entirety, the fan and sediment-feed systems were turned off, the wind tunnel windows were opened, the entire sand surface was sprayed with water mist to 'fix' the surface, and an identical experiment was conducted on the stationary surface, absent sediment transport. Such paired runs were conducted across a range of free-stream velocities for both the fine-grained and coarse-grained sediment mixtures, yielding a total of 28 runs.

**Results**

Figure 1 shows several representative velocity profiles, with and without saltation, for the fine-grained and coarse-grained cases. Several features of these profiles are noteworthy. The uppermost portions of all the profiles show a pronounced deviation from the expected log-linear trend. This is believed to be an artifact of the constrained dimensions of the wind tunnel rather than a velocity-defect phenomenon as envisioned by Coles. Detailed inspection of the wind-speed time series showed that the turbulent signatures in these upper locations were distinctly different from those lower in the profile, and on this basis, the boundary-layer depth was assessed at 0.24 m for virtually all runs. The middle sections of the profiles (roughly between 0.06 m and 0.18 m) are consistently log-linear even when saltation was active. However, the slopes of the profiles are considerably different between the no-transport and with-saltation cases, especially as free stream velocity increases. Shear velocities derived from log-linear regressions through these middle sections are consistently greater for the with-saltation case in any given profile pair, reaffirming that the presence of the saltation layer does indeed have the effect of contributing an enhanced roughness to which the overlying wind field must adapt. Interestingly, these with-saltation regression lines appear to converge roughly near Bagnold's focal point (i.e., at a the height of ~ 0.003 m and an average wind speed of ~ 2.5 ms$^{-1}$). Nevertheless, such 'focusing' of the profiles never truly occurs because
the lowermost portions of the profiles (less than ~ 0.05 m) begin to deviate significantly from
the log-linear trend of the middle profile section. This is most evident in the high-speed
profiles and especially when the saltation layer is well-developed. Shear velocities derived
from the near-surface section of the profile are invariably smaller than those derived from the
middle section. The question arises as to which is the most robust shear velocity estimate to
use in transport modeling.

Figure 1. Representative wind-speed profiles with and without saltation for fine-grained
(solid symbols) and coarse-grained (open symbols) sand. 'Hz' refers to fan speed.

A series of tests was conducted to fit six alternative velocity profile parameterizations
to the data in order to determine what the implications might be for estimates of shear velocity
and roughness length. Figure 2 shows the results of the shear velocity analysis for the fine-
grained case only. Using a constant value of $\Pi = 0.55$ in Equation (2) yields suspiciously small
values for shear velocity, and this is exacerbated if a larger constant value for $\Pi$ is employed.
When the value of $\Pi$ is derived independently from an analysis of the innermost profile
segment and then applied to the outer flow region, realistic values of shear velocity are
produced. These are surprisingly similar to the shear velocity estimates derived from a simple
regression through only the innermost measurements (< 0.045 m). Shear velocity estimates
based on two versions of the logarithmic law (e.g., Equation (1)) are similar to each other, and
when there is no transport, these estimates are virtually the same as those using Equation (2).
However, when there is active transport, these methods produce diverging estimates for shear
velocity. A simple regression applied to the middle segment of the profile (between 0.06m
and 0.18 m) produces unrealistically large estimates of shear velocity when there is a well
developed saltation layer.
Figure 2. Shear velocity estimates from six alternative velocity-profile parameterizations for the fine sand case only.

Conclusions

Detailed analysis of wind-speed profiles from a series of wind-tunnel experiments shows that estimates of shear velocity can vary by almost two-fold depending on which profile parameterization is adopted. The prospects for sediment-transport prediction are therefore not encouraging. Employing a constant value of $\theta=0.55$ (or greater) in Coles' Law of the Wake is not recommended. Calculating $\theta$ based on near-surface measurements is preferred, but this is time-consuming and requires very closely-spaced velocity measurements within the saltation layer—this is unlikely in field experiments. Restricting attention only to the boundary-layer segment immediately above the saltation layer, as is conventionally done, appears to produce over-estimates of shear velocity. Using wind-speed information from the entire boundary layer appears to provide the most robust estimates of shear velocity, although it is not known how these relate to sediment flux.

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Aeolian abrasion and fine particle production from red sands: an experimental study.

J.E. Bullard, Department of Geography, Loughborough University, Loughborough, Leicestershire LE11 3TU UK. (E-mail: j.e.bullard@lboro.ac.uk)

G.H. McTainsh, Australian School of Environmental Studies, Griffith University, Brisbane, Queensland 4111, Australia. (E-mail: g.mctainsh@mailbox.gu.edu.au)

C. Pudmenzky, Australian School of Environmental Studies, Griffith University, Brisbane, Queensland 4111, Australia. (E-mail: c.pudmenzky@mailbox.gu.edu.au)

Introduction

The production of dust-sized particles (<100 μm) has been discussed at length in the literature, particularly with reference to the origins of loess deposits. The main mechanisms proposed for fine particle production are weathering, glacial grinding, fluvial comminution and aeolian abrasion and all have been identified as effective processes for producing loess-sized particles (20-60 μm; Wright et al. 1998). Finer particles (<20 μm) can be produced by similar mechanisms. The size, shape and composition of the sediments from which fine particles are derived has an impact upon the nature and rate of dust production. During aeolian abrasion experiments on freshly-crushed quartz, angular sand-sized particles initially yield high amounts of dust-sized material as corners and protruberances are removed. This is followed by a decrease in the production of fine particles as the grains become increasingly more rounded (Whalley et al. 1987; Wright et al. 1998).

Most natural aeolian sand grains are sub-angular to sub-rounded (Goudie & Watson, 1991) suggesting that lower quantities of fines will be yielded during abrasion compared with angular quartz. In addition, many natural dunefield sediments do not comprise ‘clean’ quartz grains. In particular, clay coatings on grains have been reported from a number of dunefields (e.g. Walden & White, 1997; Walker, 1979; Wasson, 1983). There has been some debate in the literature about the extent to which these clay coatings can be removed by aeolian abrasion (Walker, 1979, Wopfner & Twidale, 2001). If the clay coatings can be removed during saltation then they could provide a source of dust-sized material additional to that produced by the rounding of sand grains.

This paper investigates fine particle production by aeolian abrasion of natural dune sands featuring a clay coating. This provides three possible sources for the dust-sized material: (1) fine particles initially present in the natural sand population and released during saltation; (2) fine particles produced by spalling and chipping of larger particles; (3) fine particles derived from the removal of the clay coating from grain surfaces.

Method

A series of experiments was conducted using a ‘test-tube’ chamber similar to that described by Whalley et al. (1987) and Wright et al. (1998) to simulate the aeolian abrasion process
The sample is placed in the bottom of the glass chamber and the grains agitated by an air stream. Fine particles raised into suspension within the chamber are trapped in an electrostatic precipitator operating at 5 kV. The efficiency of this technique for trapping fine particles was estimated at 95% by Whalley et al. (1987).

Two main experimental runs were conducted each using 10 g of sand collected from a linear dune in western Queensland, Australia (modal grain size 172 μm). Clay coatings were clearly visible on the quartz grains under a light microscope and sand colour can be described as 7.5 YR 5/6 using the Munsell colour chart. For the first test 10 g of dune sand was agitated for a total of 120 hours with fine particles collected after 1, 2, 4, 8, 16, 32, 48, 72, 96 and 120 hours of abrasion. For the second test, the dune sand was acid washed using 10% hydrochloric acid followed by stannous chloride solution to remove the grain coatings (Newsome & Ladd, 1999) and then subject to the same abrasion procedure. The amount of material collected in the electrostatic precipitator was weighed after each time period. A Coulter Multisizer was used to determine the particle size distribution of the samples and the products of abrasion.

Results and Discussion

Figure 2 shows cumulative dust production over a period of 120 hours for the natural and acid washed sand samples. The natural sand yielded a total of 0.2519 g of dust from an initial sample weight of 10.8352 g. In comparison, the acid washed sample yielded less than half that amount of fines with 0.1253 g of dust coming from 10.5304 g of sand.
A detailed analysis of the particle size characteristics of the fine sediments captured in the electrostatic precipitator shows that the particle size characteristics of the dusts varies during the abrasion period. Figure 3a shows the initial natural sand particle size distribution with a modal particle diameter of 172 μm. Fines collected during the second hour of abrasion show a distinct mode at approximately 60 μm (Figure 3b) which can be recognised in the original sand sample and probably represents small particles present in the natural sample rather than the products of abrasion. Fine particles yielded in hours 9-16 of the abrasion period show a very different particle size distribution with modes at approximately 3 μm, 17 μm and 40 μm (Figure 3c). The 3 μm and 17 μm modes can also be recognised in the fines collected in hours 49-72 of the abrasion period (Figure 3d).

The larger two of these modes may represent spalling and chipping of the sand grains. The finest mode may be produced from removal of the clay coatings from the outside of the grains.
however more detailed analyses of the fine particle products are necessary to investigate these hypotheses further. Visual observations of the unwashed sand grains before and after abrasion did indicate a change in colour of the grains possibly caused by removal of the red clay coating. Following 120 hours of abrasion, the particle size mode of the sand sample was found to have increased to 218 μm.

Conclusions

Preliminary analysis of the products of abrasion in this instance suggest first that sands which are clay-coated yield greater quantities of fine particles than those from which the coatings have been removed. The particle size distribution of the fine particles produced varies through time and may be related to the abrasion process. Further investigations, including the implications for dust production and the interpretation of sediment colour are ongoing.

References


A Physical-Conceptual Model to Predict the Threshold Shear Velocity of Wet Sediment

W.M. Cornelis, Dep. Soil Management and Soil Care, Ghent University, Coupure links 653, Gent, Belgium (E-mail: wim.cornelis@rug.ac.be)

D. Gabriels, Dep. Soil Management and Soil Care, Ghent University, Coupure links 653, Gent, Belgium (E-mail: donald.gabriels@rug.ac.be)

R. Hartmann, Dep. Soil Management and Soil Care, Ghent University, Coupure links 653, Gent, Belgium (E-mail: roger.hartmann@rug.ac.be)

Introduction

A crucial parameter in predicting wind erosion is the threshold shear velocity. This is the minimal shear velocity required to initiate deflation of soil particles. Amongst the several factors that govern threshold conditions, surface moisture is one of the most significant. Through adhesion and capillary effects it strongly contributes to the binding forces keeping particles together (McKenna-Neuman and Nickling, 1989). Albeit the several studies that were conducted to determine the influence of moisture on entrainment of soil or sand particles by wind, its effect is still not well understood (Namikas and Sherman, 1995, Shao, 2000).

A new model to predict the threshold shear velocity for deflation of wet particles is presented in this paper. It is based on the balance of moments acting on wet particles at the instant of particle motion. The model includes a term for the aerodynamic forces, including the drag force, the lift force and the aerodynamic moment force, and a term for the interparticle forces. The effect of gravitation is incorporated in both terms. Rather then using an implicit function for the effect of the aerodynamic forces as in the model of Iversen and White (1982), a constant aerodynamic coefficient was introduced. The term for the interparticle force was deduced from consideration of the electrostatic force, the van der Waals force, and forces due to liquid-bridge bonding (capillary forces) and adsorbed-layer bonding (adhesion forces). The finally obtained model can be written as:

\[ u_s = \sqrt{\frac{A_1 \left[ 1 + w + A_2 \left( \frac{\sigma^2}{\rho_f - \rho_s} \right) g d^3 \right]}{d + A_3 \left( \frac{w}{w_{1.5}} \right)}} \frac{\rho_f - \rho_s}{\rho_f} \frac{g d^2}{\mu_{md} e^{-6.5 \frac{w}{w_{1.5}}}} \]

where \( u_s \) is the threshold shear velocity (m s\(^{-1}\)), \( A_1, A_2 \) and \( A_3 \) are regression coefficients (resp. dimensionless, in N m\(^{-1}\), and in kg m\(^{-2}\)), \( w \) is gravimetric moisture content (kg kg\(^{-1}\)), \( w_{1.5} \) is gravimetric moisture content at –1.5 MPa (kg kg\(^{-1}\)), \( \rho_s \) is particle density (kg m\(^{-3}\)), \( \rho_f \) is fluid density (kg m\(^{-3}\)), \( g \) is gravitational acceleration (m s\(^{-2}\)), \( d \) is particle diameter (m), \( \sigma \) is surface tension (N m\(^{-1}\)), and \( \mu_{md} \) is matric potential at oven dryness (≈ -10\(^6\) kPa). For detailed information about the model development, we refer here to Cornelis (2002).
Materials and Methods

The values for the parameters $A_1$ and $A_2$, were determined from non-linear least-squares analysis on the wind-tunnel data as reported by Iversen and White (1982) and validated on data from own wind-tunnel experiments on six fractions of dry dune sand and silt loam aggregates, with particle densities of resp. 2.65 Mg m$^{-3}$ and 1.47 Mg m$^{-3}$. Note that the latter value corresponds to bulk density of the silt loam aggregates. The value for the parameter $A_3$ was derived from non-linear least-squares analysis on data from wind-tunnel experiments with the same but prewetted sediment, and the final model was validated against simulations with the model of Chepil (1956), which was, though controversial (Gregory and Darwish, 1990; Namikas and Sherman, 1995; Fécan et al., 1999), chosen for reasons as mentioned in Cornelis (2002).

The own experiments were performed in the closed-circuit blowing-type wind tunnel of the International Centre for Eremology, Ghent University, Belgium (Gabriels et al., 1997). The test section of the wind tunnel was 12 m long, 1.2 m wide and 2.5 m high, and the boundary layer was set at a height of about 0.60 m, by using a combination of spires and roughness elements (Cornelis, 2002).

A sample tray 0.95 m long, 0.40 m wide and 0.02 m deep was located at a distance of 6.0 m downwind from the entrance of the wind-tunnel test section. The tray was filled with air-dried or prewetted (by spraying a fine mist) sediment. The sample trays were then exposed to wind with different shear velocities and were subjected to evaporation in the case of the wet sediment.

A saltiphone, which is an acoustic sediment sensor that measures the number of saltating particles that bounce against a microphone (Spaan and van den Abeele, 1991), was continuously monitoring any particle deflating from the sample tray. It can hence be used to determine the instant of particle motion. The impact energy of the soil aggregates smaller then 200 $\mu$m was, however, too low to be recorded by the saltiphone. In those cases, particle entrainment was determined visually using a Neon-Helium laser beam (Logie, 1982). The detailed procedures to determine the deflation threshold for dry sediment and for wet sediment are both described in Cornelis (2002).

The wind velocity was measured at a 1-Hz frequency with 16-mm vane probes (Testo, Lenzkirch, Germany$^1$) mounted at five different heights, and was used to compute the shear velocity using the well-known Prandt-von Kármán logarithmic law.

Moisture content was determined gravimetrically on three small samples taken down to 1 mm, from the instant of continuous particle entrainment, i.e. from the moment that three impacts were recorded on the saltiphone within one minute. Since we could visually observe differences in moisture content over the tray, two samples were taken at a dry section and one sample at a wetter section. Continuous deflation only occurred once the first dry sections appeared. The moisture content at $-1.5$ MPa $w_{1.5}$ was determined using a pressure chamber (Soilmoisture Equiment, Santa Barbara CA, USA).

The relative humidity, evaporation rate and temperature were constant within each test run.

Results and Discussion

$^1$ Mention of company names is for the convenience of the reader and does not constitute any endorsement in whatever sense from the authors.
The values for the parameters $A_1$ and $A_2$ as determined from non-linear least-squares analysis against the Iversen and White dry-sediment data (1982) were $0.013$ and $1.7 \times 10^{-4}$ N m$^{-1}$ respectively. Our simple model (with $w = 0$) showed a similar performance with the model of Iversen and White (1982) and $R^2$ was resp. $0.997$ and $0.992$, when omitting the extremely high particle density data ($8.9$ Mg m$^{-3}$ and $11.4$ Mg m$^{-3}$). When validating our model against our own wind-tunnel data, $R^2$ was $0.995$ and $0.972$ for dune sand grains and silt loam aggregates resp.. This illustrates that it is acceptable to replace the implicit function in the Iversen and White model (1982) for the aerodynamic term with a constant, and that soil aggregates can be treated as individual particles with a density equal to their air-dry bulk density. Prediction of mass transport for a given soil should therefore be based on minimally-dispersed particle-size distributions, rather then fully-dispersed ones.

The third model parameter $A_3$, that expresses the effect of surface moisture, was determined from non-linear least-squares analysis against wind-tunnel experiments on prewetted sediment, and was found to be linearly proportional to $d^2$. In that respect, it can also be considered as a kind of compensation factor that keeps the term for the wet-bonding force scaled with $u_\tau$ which to a large extent depends on particle diameter $d$. In Fig. 1, the observed $u_\tau$ values are plotted as a function of $w$. Also plotted is the proposed model with $A_3 = 4 \times 10^{14} d^2$ kg$^{-1}$ m$^{-2}$ s$^2$. Although the data show a lot of scatter and $R^2$ was in all cases lower then 0.75, it can be seen that the model follows the dry-section data, where deflation was observed, rather well. At low moisture contents, the increase in $u_\tau$ with $w$ is very gradual. But at a given moisture content, a steep increase in $u_\tau$ becomes apparent. The threshold shear velocity soon reaches a very high value and a critical moisture content, above which there is no wind erosion, can be observed. This is in line with observations of Chepil (1956), Bisal and Hsieh (1966) and Saleh and Fryrear (1995). Therefore, there is no need to quantify the upper boundary conditions of our model.

As a validation, in Fig. 2, the threshold shear velocities $u_\tau$ as predicted with the model of Chepil (1956) is plotted against the threshold shear velocities $u_\tau$ as predicted with the model presented in this study. Figure 2 clearly illustrates that both models agree very well for
threshold shear velocities not exceeding 0.5 m s\(^{-1}\), which was about the maximum shear velocity generated in Chepil’s study. This value corresponds with a \(w/w_{1.5}\) ratio of 0.75 and is close to the critical moisture content above which no wind erosion will take place. This means that the soil surface has to dry to a rather low value before deflation will occur.

**Conclusions**

The presented model expresses the threshold shear velocity as a function of the soil-surface moisture content, particle diameter, and particle (or bulk) and fluid density. It is indexed to the moisture content at \(-1.5\) MPa, a unique value for a given soil. Furthermore, it contains only three model coefficients. Although the model was derived from theoretical considerations it is rather simple. Calibration and validation of the model on dry and wet sediment showed that it performs very well.

\[
\text{Predicted } u^*_{t}, \text{ Chepil (1956) (m s}^{-1})
\]

\[
\begin{array}{cccccc}
0.0 & 0.2 & 0.4 & 0.6 & 0.8 & 1.0 \\
\end{array}
\]

\[
\text{Predicted } u^*_{t}, \text{ this study (m s}^{-1})
\]

\[
\begin{array}{cccccc}
0.0 & 0.2 & 0.4 & 0.6 & 0.8 & 1.0 \\
\end{array}
\]

![Figure 2. Threshold shear velocities \(u^*_t\) as predicted with the model of Chepil (1956) vs. threshold shear velocities \(u^*_t\) as predicted with the model presented in this study.](image)

**References**


Aerodynamic Roughness and Displacement Height Prediction for Waves

M. M. Darwish, Ph.D., Engineering Technology Department, Texas Tech University, Lubbock, TX 79409 (mukaddes.darwish@coe.ttu.edu)

J. M. Gregory, Ph.D., College of Engineering, Texas Tech University, Lubbock, TX 79409 (james.gregory@coe.ttu.edu)

Abstract

Aerodynamic roughness and displacement heights are important parameters needed to accurately estimate evaporation from free water surfaces. It is evident that evaporation rate is influenced by the mean and turbulent flow properties near the surface of a given body of water. If the aerodynamic roughness can be estimated for the water surface, then the accuracy of predicting evaporation from the free-water surface will be greatly improved.

This study presents an equation to predict displacement height and aerodynamic roughness over water bodies. Wind is the main force, which causes the waves on water surfaces. Surface roughness of water is directly influenced by the velocity of the wind. The equation presented in this study predicts wave heights as a function of wind speed and displacement height and then the aerodynamic roughness as a function of the wave height. This model expands the overall accuracy of evaporation prediction from free-water surfaces.

There are contradictory discussions in the literature about roughness of the water surfaces and between surface roughness and wind speeds. A need exists to clarify the physics and mechanics of the process of evaporation and apply sensible understanding through model verification to better predict evaporation. The model presented in this study is comprised of a set of interactive physically based equations, which relate aerodynamic roughness to surface waves. This Equation when applied to Borelli -Sharif Equation Model was validated by comparing output determined by the model with observed data. Since there are similarities in the process of predicting erosion from beaches, the same equation should be valid for predicting wind erosion of beaches from wind coming off water bodies.

Introduction and development

Soil erosion and evaporation are both surface phenomenon and greatly effected by the wind profile. In order to accurately predict soil erosion and evaporation from water surfaces, the wind velocity profile parameters of displacement height D, and aerodynamic roughness $z_0$, must be accurately determined. Objective of this paper is the overview of the development and evaluation of equations for displacement height and aerodynamic roughness for water surfaces.

Overview and discussion

(Abte et all., 1989) derived equations for displacement height and aerodynamic roughness for solid objects such as sand grains, soil aggregates, residue or vegetative cover such as trees. Over land surface displacement height and aerodynamic roughness directly
related to surface characteristics but over water surfaces there is relation between wind velocity and displacement height and aerodynamic roughness because the wind causes surface roughness to develop as expressed as waves. Darwish 1998, using concept of potential and kinetic energy derived a relationship for wave height and wind velocity. Building on this relationship and by application of (Abtew et al., 1989) relationship for displacement height and aerodynamic roughness a system of equation was developed to predict aerodynamic roughness. Wave height was treated as ridge height in (Abtew et al., 1989)’s equation. Because waves move with wind and the relative velocity between wind and waves is less than between the relative velocity over a solid object (Abtew et al., 1989)’s equation was modified by slight change adding coefficient.

Results

The wave equations were tested involving variation in wind and fetch length with seven data sets (Darwish, 1998) five of the seven sets with and R² of values 0.90 or better. The other two comparisons had values of 0.77 and 0.80. Based on this evidence the equation was accepted as the valid predictor wave height as a function of wind speed. Aerodynamic roughness equation was indirectly verified through prediction of evaporation, R² of 0.61 was obtained using Sharif-Borelli model for data collected by Sharif, 1989. Although, R² of 0.61 does not sound like a good fit. It was highly significant α = 0.001. It improved the prediction of evaporation from R² of near 0 to 0.61. Another words it represents a displacement height, and aerodynamic roughness equations derived accounted for 61 % increase in the observed variance. It should be noted that the measured data was for a daily bases which is difficult to measure. Therefore the observed R² of 0.61 represents significant improvement in prediction of evaporation. Based on this evidence the aerodynamic roughness equation was also accepted as a valid predictor of aerodynamic roughness for water bodies. It should be noted based on personal communications with Dr.Wossenu Abtew, he too has learned that his aerodynamic roughness equation was applied for ridges as reasonable prediction for aerodynamic roughness over water surfaces.

Conclusions

Aerodynamic roughness of the water surface can be determined from the wave heights, if the wave heights can be predicted from the wind velocity measurements. Equations tested in this study provides a relation between wind velocity and wave heights therefore can be used for more accurate prediction of evaporation over water bodies.

References


Wind-driven Rainsplash Erosion


L. D. Norton USDA - ARS National Soil Erosion Research Lab., 1196 SOIL Bldg., Purdue Univ., West Lafayette, IN, 47907-1196, USA.

D. Gabriels Department of Soil Management and Soil Care, Ghent University, Coupure Links 653, B 9000 Ghent, Belgium.

Introduction

The erosion process involves detachment of soil particles from a soil surface and transport of these particles from their first location. The main agents that loosen, break down, and carry the soil particles are wind and water.

Wind and water erosion processes have traditionally been separately studied, and independent models were developed to predict soil erosion by either wind or rain. In nature, erosive rainstorms are usually associated with high winds. Therefore, a quantification of wind and rain interactions and the effects of wind on detachment and transport processes provides a great opportunity for a given technology to improve the estimation of erosion.

Soil detachment and transport under wind-driven rain differs from that under windless rain (Lyles et al., 1969, 1974; Disrud and Krauss, 1971; Moeyersons, 1983; De Lima et al., 1992). Usually, if a raindrop falls at an angle, only the component of velocity normal to the soil surface gives rise to an impact pressure (Ellison, 1947; Springer, 1976; Gilley and Finkner, 1985). If we assume that the effect of the wind shear stress on the detachment is insignificant when compared to the effects of the impacting raindrops, the detachment rate at which soil particles are supplied into the air is a function of the normal component of raindrop impact velocity. Additionally, wind, as well as overland flow, is another possible factor capable of transporting the detached particles. Consequently, our approach to the rainsplash transport process under wind-driven rain is based on the concept that once lifted off by the raindrop impact, the soil particles entrained into the splash droplets travel some distance, which varies directly with the wind shear velocity. The raindrop impacts induce the process that wind would otherwise be incapable of transporting.

This paper presents experimental data obtained on the wind-driven rainsplash erosion and aims to provide a better basis for modeling the process.

Materials and Methods

The study was conducted in a wind tunnel rainfall simulator facility at Ghent University, Belgium (Gabriels et al., 1997). A continuous spraying system of downward-oriented nozzles was used at 1.5 bar operating pressure. A detailed description of the raindrop size distribution for the simulated rainfalls of the wind tunnel is given by Erpul et al. (1998, 2000).

Three loess derived agricultural soils, Kemmel1 sandy loam (57.6% sand, 31.1% silt, and 11.3% clay) and Kemmel2 loam (37.8% sand, 44.5% silt, and 17.7% clay) from the Kemmelbeek watershed (Heuvelland, West Flanders, Belgium) and Nukerke silt loam (32.1%...
sand, 52.3% silt, and 15.6% clay) from the Maarkebeek watershed (Flemish Ardennes, East Flanders, Belgium) were used in this study. The soil samples were collected from the Ap horizon and air-dried prior to the experiment. Soil was sieved into three aggregate fractions: 1.00 - 2.75, 2.75 - 4.80, and 4.80 - 8.00 mm, and the weighing factors assigned to each fraction were 28, 32, and 40%, respectively to reconstitute the packing soil. A 5-kg soil sample was then packed loosely into a 55-cm-long and 20-cm-wide pan after the three fractions of aggregates were evenly mixed.

Rains driven by horizontal wind velocities of 6, 10, and 14 m s\(^{-1}\) were applied to the soil pan placed at both windward and leeward slopes of 7, 15, and 20%. There were three replicates for each soil and slope aspect.

Wind velocity profiles were measured up to 2-m-nozzle height with a vane type anemometer and associated recording equipment, and the wind velocity profiles above the soil pan were characterized by the following logarithmic equation:

\[
\frac{u(z)}{u_*} = \frac{0.41}{\kappa} \ln \left( \frac{z}{z_o} \right) \quad \text{for} \quad z > z_o
\]

where, \(u(z)\) is the wind velocity at height \(z\), \(z_o\) is the roughness height, \(u_*\) is the wind shear velocity, and \(\kappa\) is von Karman's constant. The boundary layer was set at 0.30 m above the soil pan, subsequently, the reference shear velocities were derived from the logarithmic wind profiles, assuming a fixed roughness height of 0.0001 m for a bare and smoothed soil surface from the relation \(b = a_e z^*\), where, \(a = z_o\) and \(b = \kappa/u_*\).

The energy of simulated rainfalls was measured by a piezoelectric ceramic kinetic energy sensor (Sensit\textsuperscript{TM}, 2000). The kinetic energy sensor is a 5-cm ceramic disk and works on the piezoelectric effect of a ceramic disc, which produces electric charges proportional to the kinetic energy of impacting raindrops. The intensity of simulated rains (I) was directly measured with 5 small collectors on the inclined plane with respect to the prevailing wind direction.

In the present study, we assumed rainsplash detachment rate under inclined rain was related to the normal component of raindrop impact velocity. Accordingly, the fluxes of rain energy (KE) based on the normal velocity of raindrop impact was used as a rainfall parameter:

\[
KE = \Xi_a \left( \frac{1}{2} mV^2 \right) \phi^2
\]

where, KE is in Wm\(^{-2}\) and \(m\) is the raindrop mass in kg. \(\Xi_a\) is the number of raindrops in \# m\(^{-2}\) s\(^{-1}\) and calculated by \(\Xi_a = 1/\forall\), where, \(I\) is in ms\(^{-1}\), and \(\forall\) is the raindrop volume in m\(^3\). \(\phi\) is cosine of the angle of rain incidence between the wind vector and the plane of the surface and calculated by \(\phi = \cos(\alpha + \theta)\), where, \(\alpha\) is the raindrop inclination from vertical (degree), and \(\theta\) is the slope gradient (degree).

Rainsplash transport was evaluated by the amount of the splashed particles trapped at set distances on a 7-m uniform slope segment. Troughs were placed in the upslope and downslope directions, respectively, for windward and leeward slopes. The soil particles trapped in the collecting troughs were washed, oven-dried, and weighed. Mass distribution curves were then determined, from which rainsplash transport rates were calculated based on:

\[
q_s = \frac{1}{At_r} \int_{x_i} m_i \, dx
\]

where, \(q_s\) is in g m\(^{-1}\) min\(^{-1}\), \(A\) is the collecting trough area \((1.20\, m \times 0.14\, m = 0.168\, m^2)\), and \(t_r\) is the time (min) during which rainsplash process occurred. \(m_i\) is the mass of soil (g) splashed over the distance \(x_i\) (m).

Since soil particles left the surface with different initial lift-off speeds and angles, prediction of particle trajectories was considered as an average in this study. We used the first
moment of the mass distribution curves (Van Heerden, 1967), which is the center of gravity of the curves, to approximate the mean rainsplash distance:

\[ \sum_{i=1}^{n} (x_i - X) m_i = 0 \]  \[ \text{[4]} \]

and

\[ X = \frac{\sum_{i=1}^{n} x_i m_i}{\sum_{i=1}^{n} m_i} \]  \[ \text{[5]} \]

where \( X \) is mean rainsplash distance (m).

**Results and Discussion**

Measured rainsplash rates varied in close relationship to the energy flux of rain and wind shear velocity. Similar results were obtained for all three soils. The statistical fit of the power law model for the combined data from three soils is:

\[ q_s = 119.95 KE^{0.81} u_s^{0.09} \]  \[ \text{[6]} \]

The model of Eq. [6] performed equally well and provided similar \( R^2 \) values, which were ≥ 0.94 for three soils. The analysis of variance also showed that the model coefficients were significant at \( P = 0.0001 \) level of significance. The form of the model developed above features an integration of wind effects on the physical raindrop impact, and hence detachment, and on the transport process. In this experimental study, wind increased the raindrop resultant velocity and altered the angle of raindrop incidence, which resulted in a variable raindrop impact frequency and impact angle. Therefore, differential delivery rates occurred depending on the variations in raindrop trajectory and frequency with wind velocity and direction. More significantly, the wind had a greater effect on soil particle transport.

For the description of the average path of raindrop-induced splash droplet, a statistical analysis was conducted with non-linear regression model of:

\[ X = C_1 (u^2/g) \]  \[ \text{[7]} \]

where \( X \) is expressed in unit of m, \( u^* \) in units of m s\(^{-1}\), and \( g \) is the gravitational acceleration and in units of m s\(^{-2}\). \( C_1 \) is a model parameter. Eq. [7] shows the fit of the data collected in this study, and for all data \( C_1 \) is equal to 32.7 ± 1.9. An important point here is that the average trajectory of raindrop-induced particle movement is approximately 3 times greater than the trajectory of a typical sand particle (White and Schulz, 1977). Longer particle trajectory might result from a change in the ejection velocity of droplets and lower density of soil aggregates. The greater lift-off speeds are probably caused by the raindrop impact than that by hitting sand grains. Therefore, raindrop-induced particles could attain greater heights and travel longer distances.

**Conclusions**

A wind-tunnel study under wind-driven rains was conducted to determine the combined effect of rain and wind on the rainsplash transport process. Transport by this process for the three soils studied was adequately described \( (R^2 \geq 0.94) \) using log-linear regression technique by Eq. [6] relating transport rate to the rain energy flux and wind shear velocity. Equation [6] reflects the combined effect of wind: one is on the detachment by changing the raindrop impact parameters, and the other is on transport by carrying the detached and lifted soil particles. Therefore, a model of
this form could provide the basis for modeling interrill rainsplash transport under wind-driven rains, a common phenomenon in erosion events.

Average trajectory of a raindrop-induced and wind-driven particle was also adequately predicted by 32.7 \((u^2/g)\), and the travel distance is found three times greater in raindrop-induced process than the path of a typical saltating sand grain. We ascribed this to the greater lift-off speeds possibly caused by the raindrop impact and the lower densities of soil aggregates.

References


Introduction

The influence of valley topography on boundary layer winds has been investigated in relation to wind velocity and turbulence (Sierputowski et al., 1995) and the dispersal of pollutants (Beniston, et al., 1989; Kalthoff et al., 2000). These studies, however, find their focus on valleys in hilly or mountainous terrain. Research has not adequately examined the effect of isolated valleys in flat, plateau terrain on boundary layer winds and no investigations to link consequent changes in wind patterns to sediment transport have been undertaken.

In the drylands of the world, where wind is a major transporting mechanism of sediment and organic matter, the interaction of ephemeral rivers and airflow has implications for the accumulation, transport and deposition of airborne material. This paper presents the first attempt to quantify the impact of incised valley topography on wind direction, velocity and turbulence using measurements from both field and wind-tunnel investigations.

Methodology and Procedures

Data are presented from wind tunnel measurements, using hot-wire anemometry and particle image velocimetry (PIV), made across an idealised 1:1000 scale model of a typical 200 m wide and 20 m deep incised valley with wind flowing normal to the long axis of the valley. Data are normalised to the free stream velocity in the wind tunnel (\(U_r\)).

The extent of flow modification caused by the valley and shown in the wind tunnel measurements is compared to field data collected from a tributary of the Gaub River in central Namibia, southern Africa. A site with a valley long axis of 300m in length, incising flat, homogenous terrain, was chosen for the experiments. Within this section the valley width, or short axis, varies between 150 m and 175 m at the valley edges. The depth of the valley is 20-22 m with and the slopes have gradients of 20-25\(^\circ\).

Six stations, each with vertical arrays of rotating cup anemometers and a potentiometer windvane measured mean velocity and direction across the valley under a range of wind directions and thermal conditions. The portable stations were always aligned in a transect parallel to the dominant wind and measurements, made at a range of spatial scales in the vicinity of the valley, were compared to those made at the reference station.

In the co-ordinate system used for the data below, \(x\) refers to the horizontal distance and \(z\) to the vertical height above the ground surface. The leading edge of the valley encountered
by the oncoming wind is always referred to as \( x = 0 \) for the field data and \( x/L = 0 \) (where \( L \) = length of short axis of the valley) for the wind tunnel data. Measurements upwind of this position are given a negative value. In all proceeding figures wind is from left to right.

Values of mean velocity calculated at a given height \( (u_z) \) were normalised with measurements at the corresponding height at the reference station \( (U_z) \). The changes in velocity across the transects are displayed below in the form of the fractional speed-up ratio \( (\delta \xi) \) which presents wind acceleration as a fraction and is an effective way of showing changes in velocity across a topography (Jackson and Hunt, 1975; Lancaster, 1985; Mulligan, 1988; Wiggs et al., 1996): \[
\delta \xi = (u_z - U_z)/U_z
\]
The changes in wind direction as wind passes the valley are presented by showing the deviation of direction at each point of measurement from the mean direction recorded at the reference station:

Deviation = direction measured at point ‘x’ - direction at reference

**Experimental observations**

Field data
Figure 1 shows the fractional speed up measured at \( z = 0.4 \) m and \( z = 2.8 \) m above the surface for a westerly wind of 263\(^\circ\) blowing normal to the valley axis.

There is an acceleration in wind velocity towards the upwind edge of the valley with \( \delta \xi = 0.3 \) at \( z = 2.78 \) m and \( 0.2 \) at \( z = 0.4 \) m. This acceleration occurs from \( x = -150 \) m. There is then a dramatic reduction in wind velocity into the centre of the valley where \( \delta \xi = -0.44 \). Following this is an acceleration up the downwind valley side and velocity reaches a maximum at the downwind edge of the valley, where \( \delta \xi = 0.81 \) at \( 0.4 \) m in height and \( 0.53 \) at \( 2.78 \) m in height. Velocity reduces in the wake of the downwind edge and begins to resemble upwind values beyond \( x = 300 \) m.

Visualisation of the flow using smoke canisters confirmed that flow separates off the leading edge and a recirculation zone extends for at least 50 m into the valley with flow reversing up the upwind slope. Smoke experiments also indicated a smaller recirculating zone in the wake of the downwind edge. They also indicated that the incident angle of the approaching wind to the axis of the valley may limit the extent of the separation zone.

Figure 2 compares the deviation of wind within the valley channel for a perpendicular wind and a wind blowing at an acute angle to the long axis of the valley. The maximum deflection of a 93\(^\circ\) wind is only some 25\(^\circ\), occurring at the downwind edge (\( x=200 \) m). This is contrasted with a deflection of
Figure 2  Deviation of wind from the reference station across the valley 160° in the centre of the valley (x = 100 m) when wind approaches the valley at 36°. It appears that the role of the valley in deflecting wind along its geometry is more pronounced when the angle of wind approach is less than 45°.

Wind tunnel data
Results from the wind tunnel hot-wire anemometry experiments show a considerable increase in both velocity and turbulence (Figure 3) from x/L=−1 to the leading edge of the valley (x/L=0). There is a slight increase in turbulence towards the leading edge, followed by a tremendous increase over the valley as the shear layer is encountered. At the downwind edge (x/L=1) the flow has passed the shear layer and, although turbulence is depressed, values remain significantly greater than those upwind of the valley.

Figure 3  Horizontal profiles of streamwise (u^2), vertical (w^2) turbulence and Reynolds shear stresses (-uw) measured at z/H= 0.2.

Vertical turbulence increases slightly to the leading edge, shows much greater values over the shear layer and, unlike the other profiles, greatest values at the downstream edge beyond which values decrease rapidly. The increased vertical component of the flow as flow is forced over the downwind edge indicates that separation is likely and is in agreement with observations from field smoke visualisation experiments.

Reynolds shear stress increases in intensity to the leading edge, shows greatest values over the centre of the valley and a decrease to the downwind edge, where intensities equate to those at the leading edge before decreasing at x/L = 1.5.

PIV experiments indicate that on passing the leading edge flow decelerates and separates just below the edge. The vertical and horizontal extent of the separation zone is shown by the streamlines in Figure 4 and flow reattaches by x/L = 0.7. Wind then begins to accelerate streamlines converge on encountering the downwind slope, consistent with the findings from the hot-wire and field data.
Data presented here have shown the significant effect valley topographies have on wind velocity, turbulence and direction. An increase in velocity and turbulence to the leading edge of the valley is followed by rapid deceleration as flow separates off the leading edge and zone of recirculation extends to almost two thirds of the valley length. This recirculation zone is limited if flow approaches at an angle, whereby streamlines are funnelled along the long axis of the valley. High turbulent intensities are apparent above the shear layer of this zone.

Following reattachment, streamlines converge and velocity begins to accelerate up the downwind slope reaching a maximum at the downwind edge. There is evidence of flow separation at this edge as high turbulence and low velocities are recorded in its wake. Flow begins to recover between one and two valley lengths downwind of the valley.

These findings have important implications for predicting the distribution of wind borne material. Sediment erosion might be expected from $x/L = -1$ of the valley as windflow accelerates and shear stress increases towards the valley edge. Deposition of material in transport may result from flow separation and deceleration into the valley. Sediment eroded as flow accelerates up the downwind valley slope will potentially be deposited as the flow decelerates from a maximum velocity at the valley edge to $x/L = 1.5$ beyond which mean and turbulent velocities been to recover. The empirical evidence presented forms an important backdrop for the future prediction of wind patterns and transport of particulate organic and inorganic matter in relation to valley topographies.

### References


The Long-distance “Transportable Fraction” of the Vertical Flux of wind-transported dust

Dale Gillette, Air Resources Laboratory, NOAA, Research Triangle Park, NC 27711 (Email: gillette.dale@epa.gov)

Introduction

A “lumped control volume” approach is used here to gain understanding of fugitive dust sources including wind erosion and road dust emission. This case considers dust generated from a road surface. By letting the road be directed into and out of the page while wind is directed from right to left we may invoke two-dimensionality or symmetry in the direction into the page (fluxes are equal into and out of the direction into the page). Figure 1 shows the geometry of our control volume. A dirt road exists at the right side of the control volume. To the left of the road is a surface that is grass or shrub-covered and does not emit particles. The ceiling of the control volume is the surface of primary interest as to vertical flux.

Figure 1. Control Volume for Fugitive Dust Model Depicting Vertical and Horizontal Fluxes. The quantities shown in the figure are as follows:
M is the mass of particles in the control volume (CV),
\( \frac{dm_{up}}{dt} \) is the mass per unit time passing out vertically at the top of the CV,
\( \frac{dm_{depos}}{dt} \) is the mass per unit time depositing to the floor of the CV,
\( \frac{dm_{ambout}}{dt} \) is the mass per unit time passing out of the CV through the left wall,
\( \frac{dm_{ambin}}{dt} \) is the mass per unit time passing into the CV from the right wall,
\( \frac{dm_{road}}{dt} \) is the mass per unit time emitted from the road, and
\( \frac{dm_{ceil}}{dt} \) is the mass per unit time passing out of the ceiling directly above the road.

A relation between the mass per unit time emitted from the road and the mass per unit time emitted from the road that can be considered to be regionally transported developed in the appendix is:

\[
\Phi = \frac{\frac{dm_{up}}{dt}}{\frac{dm_{road}}{dt}} = \left[ 1 - \frac{V_d}{(V_d + K)} \right] = \frac{K}{(V_d + K)}
\]

(1)

We can evaluate the ratio \( \Phi \) given by Equation 1 by making an approximation based on the data presented by Gillette (1974) that \( K = 0.08u^* \) where \( u^* \) is the friction velocity. Using Equation 1 and the above expression for \( K \), we can derive values of \( \Phi \) for typical values of \( V_d \) where the values for \( u^* \) and \( V_d \). Figure 2 shows families of \( \Phi \)'s for differing \( V_d \) values. Values of \( V_d \) versus size and environmental conditions are given by Slinn (1982).

**Discussion**

The expression given in Equation 1 explains a large part of why observed concentrations are smaller than that predicted by regional scale models that do not correct for the large scale transportable vertical flux of dust but rather use the entire amount of dust emitted by roads. Because dust is deposited to the surface close to the source, uncorrected dust vertical fluxes lead to overestimates of dust concentrations downwind of the source. Other effects that are expected based on equation (1) are given below:

- For low wind speeds, little dust is input above the surface level of a roadway. This agrees with observations by Johnson, et al. (1992).
- For fugitive dust like downloading of sediment, Equation 1 overestimates \( \Phi \) since horizontal diffusion would make more dust available for deposition to the ground. A correction for this would be to reduce the ratio \( \Phi \).
- For wind erosion, the threshold friction velocities are larger than 19 cm/s. Therefore, it would be expected that regional scale vertical flux of dust would be a very large fraction of field-scale flux of wind erosion dust.
- Dust devils, another class of fugitive dust sources, occur when the overall friction velocity is fairly low, but the height of the initial dust input is very high. For these conditions, the vertical flux \( \frac{dm_{up}}{dt} \) is virtually the same as for the input dust. Consequently, dust devils should be considered to be effective sources of dust. See Gillette and Sinclair (1989) for a discussion of dust devil dust fluxes.
**Conclusion**

A semiempirical model expresses in a simple expression the “transportable fraction of dust generated at ground level and carried by the wind.

**Appendix: Derivation of Equation (1)**

A conservation of mass equation can be written for the control volume shown in Figure 1 as:

\[
\frac{dM}{dt} + \frac{dm_{up}}{dt} + \frac{dm_{depos}}{dt} + \frac{dm_{ambout}}{dt} - \frac{dm_{ambin}}{dt} - \frac{dm_{road}}{dt} + \frac{dm_{ceil}}{dt} = 0
\]

where, \( M \) is the mass of particles in the control volume (CV),
\( \frac{dm_{up}}{dt} \) is the mass per unit time passing out vertically at the top of the CV,
\( \frac{dm_{depos}}{dt} \) is the mass per unit time depositing to the floor of the CV,
\( \frac{dm_{ambout}}{dt} \) is the mass per unit time passing out of the CV through the left wall,
\( \frac{dm_{ambin}}{dt} \) is the mass per unit time passing into the CV from the right wall,
\( \frac{dm_{road}}{dt} \) is the mass per unit time emitted from the road, and
\( \frac{dm_{ceil}}{dt} \) is the mass per unit time passing out of the ceiling directly above the road.

A simplifying assumption is that of steady state emissions, that is, \( \frac{dM}{dt} = 0 \). For this case:

\[
\frac{dm_{up}}{dt} = -\frac{dm_{depos}}{dt} - \frac{dm_{ambout}}{dt} + \frac{dm_{ambin}}{dt} + \frac{dm_{road}}{dt} - \frac{dm_{ceil}}{dt}
\]

\[\text{(A-2)}\]

**Relationship of mass per unit time to the horizontal mass flux from the road**

A subvolume of the control volume is shown in Figure 3-1 as that part that contains the road but no area downwind of the road. The left wall of this part of the control volume is the surface through which all of the dust emitted from the road passes. That is, we specify that none of the road dust is part of the ceiling flux (\( \frac{dm_{ceil}}{dt} \)). For the condition of steady state, a description of the conservation of mass for this part of the control volume is:

\[
\frac{dm_{road}}{dt} = \frac{dm_{out}}{dt} - \frac{dm_{ambin}}{dt} + \frac{dm_{ceil}}{dt}
\]

\[\text{(A-3)}\]

where \( \frac{dm_{road}}{dt} \) is the horizontal mass per unit time passing out through an imaginary wall just left of the left edge of the road reaching from the surface to the top of the control volume, and \( \frac{dm_{ceil}}{dt} \) is the mass per unit time passing through the ceiling of the control volume. Because the road dust is usually the overwhelming source of dust, that is, \( \frac{dm_{road}}{dt} \gg \frac{dm_{ambin}}{dt} + \frac{dm_{ceil}}{dt} \) the horizontal flux \( \frac{dm_{out}}{dt} \) for these conditions is approximately:

\[
\frac{dm_{road}}{dt} = \frac{dm_{out}}{dt}
\]

\[\text{(A-4)}\]

which is the horizontal mass flux of dust from the road.

**Deposition at the floor and vertical flux of dust through the ceiling of the control volume downwind of the road**

The total loss of material diffusing vertically through the ceiling and depositing on the floor of the control volume to the left of the road may be calculated by first calculating the effective concentration at position \( x \) to the left of \( x_0 \) (i.e., the left edge of the road). We use the equation:

\[
v \frac{dC(x)}{dx} = \frac{-C(x)[V_d + K]}{\Delta z}
\]

\[\text{(A-5)}\]

where, \( V \) is the wind speed that carries the dust through the control volume (CV),
\( C(x) \) is the concentration of dust mass at position \( x \) in the CV,
\( x \) is the horizontal position in the CV, increasing to the left,
\( z \) is the vertical position in the CV, increasing from the floor to the ceiling,
\( V_d \) is the deposition velocity, and
K is a coefficient having the dimensions of velocity.

A solution to Equation (3-5) is:

\[ C(x) = \frac{\frac{dm_{\text{road}}}{dt}}{V \Delta z \Delta L} \exp\left[ \frac{V_d + K}{V \Delta z} (x - x_0) \right] \]  

(A-6)

Multiplying \( C(x) \) by \( V_d \Delta L \) (i.e., the deposition velocity times the unit length of the road) and integrating with respect to \( x \) from \( x = x_0 \) to \( x = \infty \) gives:

\[ \frac{dm_{\text{depos}}}{dt} = \frac{V_d \frac{dm_{\text{road}}}{dt}}{(V_d + K)} \]  

(A-7)

By making the approximation that \( \frac{dm_{\text{amb}}}{dt} - \frac{dm_{\text{amb}}}{dt} = 0 \) and ignoring \( \frac{dm_{\text{ceil}}}{dt} \) we may rewrite equation (3-2) as:

\[ \frac{dm_{\text{up}}}{dt} = \frac{dm_{\text{road}}}{dt} - \frac{dm_{\text{depos}}}{dt} \]  

(A-8)

Watson (personal communication, 2000) stated that the condition \( \frac{dm_{\text{amb}}}{dt} - \frac{dm_{\text{amb}}}{dt} = 0 \) was observed in the field for a distance \( x \) of about 200 meters in the absence of any intervening dust sources.

Ratio of vertical flux of road dust into the atmosphere to the horizontal flux of road dust This ratio, expressed by the symbol \( \Phi \), is given by:

\[ \Phi = \frac{\frac{dm_{\text{up}}}{dt}}{\frac{dm_{\text{road}}}{dt}} = \left[ 1 - \frac{V_d}{(V_d + K)} \right] = \frac{K}{(V_d + K)} \]  

(A-9)

References


Evidence for Direct Suspension of Loessial Soils

Jim Kjelgaard, Biological Systems Engineering, Washington State University, Pullman, WA 99164-6120, jkkelgaard@wsu.edu

David Chandler, Plant, Soils, and Biometeorology Dept., Utah State University, 4820 University Blvd, Logan, UT, 84322-4820, dchandle@mendle.usu.edu

Keith Saxton, USDA-ARS, Biological Systems Engineering, Washington State University, Pullman, WA 99164-6120, ksaxton@wsu.edu

Abstract

Dust emissions by high wind events have traditionally been modeled with saltation-based wind erosion processes. This approach gives generally good results when saltation-sized soil particles, 60 μm to 2 mm mean diameter, are present on the exposed soil surface. The Columbia Plateau, located in north-central Oregon and south-central Washington, is a region with extensive loess deposits where 90% of particles (by weight) have diameters less than 60 μm. During high wind events, large amounts of particulate matter are suspended. However, field surfaces may show little evidence of surface scouring or saltation, e.g. soil drifts, covered furrows, etc.. Wind profile analysis of two large, regional high wind events and additional data from a third event show evidence of direct suspension process where saltation is not a major mechanism for generating dust emissions. More intensive studies of the wind erosion process are underway to better understand and quantify the direct suspension mechanism.
Shrub Spacing and Surface Shear Stress Distributions: A Wind Tunnel Study

Lee, Jeffrey A., Dept. of Economics & Geography, Texas Tech University, Lubbock, Texas, USA 79409-1014 (Email: jeff.lee@ttu.edu)

Greeley, Ronald, Dept. of Geology, Arizona State University, Tempe, Arizona, USA 85287 (Email: ron.greeley@asu.edu)

Abstract

This study attempts to quantify the role of desert shrub spacing on the spatial pattern of shear stress on the soil surface. In a wind tunnel, the naphthalene sublimation technique was used to map shear stress under varying situations of shrub spacing. The technique is based on the amount of sublimation of naphthalene during a wind tunnel run; the amount is a function of surface temperature and shear stress (Lee, Chyu and Greeley, submitted, *Earth Surface Processes and Landforms*). The results suggest that flow acceleration around shrubs enhances shear stress, with maximum reduction in shear stress at intermediate spacings. At high spacings, the shrubs have little effect on the ground surface away from them and at low spacings, a high fraction of the ground surface is affected by flow acceleration around individual shrubs. Field studies are needed to see if these laboratory findings are applicable to real world conditions.
Wind tunnel simulation experiment on the erodibility of the fixed aeolian sandy soil by wind

Xiao-Yan Li, Jia-Hua Wang, Cold and Arid Regions Environmental and Engineering Research Institute, Chinese Academy of Sciences, Lanzhou 730000, P.R. China (Email: xxyyli@fm365.com)
Lian-You Liu, Institute of Resources Science, Beijing Normal University, Beijing 100875, P.R. China (Email: lianyou@public.lz.gs.cn)

Introduction

Approximately 20% of the world’s arid zones are covered by sand seas, sand sheet, and by dune fields (Pye and Tsoar, 1991). In the case of China, aeolian sandy soils are widely distributed in the arid regions of northwest China, humid regions of southeast China and the Tibet Plateau. The fixed aeolian sandy soils cover extensive areas of the southeastern edge of the Tengger Desert, where sand dunes bordering the railway have been vegetated by means of straw checkerboard barriers since 1956. The fixed aeolian sandy soil has been effective in stabilizing soil surface and preventing desert railway from migrating sand dune (Li et al., 2001). During the past 44 years, an 8-59 mm thick microbiotic crust has been formed in the surface of the fixed aeolian sandy soil. In order to get a better understanding of the effectiveness of the fixed aeolian sandy soils in controlling soil erosion by wind and to identify its certain underlying principles for its strong resistance against wind erosion, the objectives of this paper were to investigate the natural wind erodibility of the fixed sandy soil and the accelerating effect of human disturbance on wind erosion by means of wind tunnel simulation and discuss the effects of microbiotic crusts and vegetation covers on wind erosion.

Materials and methods

The wind erosion experiment was conducted in a push-type wind tunnel of the Laboratory of Blown Sand Physics and Desert Environments at the Cold and Arid Regions Environmental and Engineering Research Institute, Chinese Academy of Sciences. The undisturbed fixed sandy soil samples were taken from Shapotou at the southeastern edge of the Tengger Desert in the arid region of north China. The soil sampling and wind tunnel experiment were described by Li et al. in detail (2001). First, three soil samples subjected to free-stream wind velocity from 10 to 26 m s⁻¹, with equal intervals of 2 m s⁻¹ to measure natural erodibility by wind. The tests lasted from 2 to 15 min depending upon wind speed. Then the soil samples in the tray were “plowed” by a hand chisel to a depth of 10 cm, and larger clod were broken into small pieces, which represented cultivation. The “plowed” soils also subjected to the above mentioned wind velocity. Vegetations in other three soil samples were clipped to a cover density of 30, 15 and 0% to simulate the effects of vegetation cover on soil erosion. Then surface crust was broken with the point-grid method to the percentage of 20, 40, 60, 80 and 100% to simulate the effects of crust on soil erosion. The soil samples were blown under the wind velocity of 10, 18, 26 m s⁻¹ for each treatment.

Results and discussions

The experimental results indicated that there was a close exponential relationship between wind velocity and soil erosion rate (Figure 1). In the case of the undisturbed (natural) fixed sandy soil, there was virtually no wind erosion at wind speeds less than 8 m s⁻¹, while wind
erosion began at velocity 6 m s\(^{-1}\) for the cultivated sandy soil. Erosion rates of the cultivated treatments were 3.6, 51.8, 102.6, 311.6, 426.7 times of the undisturbed fixed sandy soil at the wind velocity of 10, 16, 20, 24, 26 m s\(^{-1}\), respectively. The soil loss rate (SLR = soil loss from undisturbed soil/ soil loss from cultivated soil) versus wind velocity is presented in Figure 2, indicating that the relationship follows negative exponential function. The ratio between total soil loss from the undisturbed soil and the cultivated soil was about 0.004. The results suggest that cultivation or trample can substantially accelerate erodibility of the fixed sandy soil by wind.

Soil erodibility by wind reflects the fragility of soils suffering from wind deflation and abrasion. The natural wind erodibility of the fixed sandy soil is affected appreciably by surface vegetation and crust. Figure 3 shows that wind erosion rate increases with decreasing vegetation cover at the wind velocity of 10, 18 and 26 m s\(^{-1}\), suggesting that vegetation is one of the factors responsible for the natural wind erodibility of the fixed sandy soil, which would reduce soil erosion by wind.

![Figure 1](image1.png)

**Figure 1** Relationship between wind erosion rate and wind velocity for the undisturbed fixed sandy soil and cultivated fixed sandy soil

![Figure 2](image2.png)

**Figure 2** The relationship between soil loss ration (soil loss from undisturbed soil/ soil loss from cultivated soil) and wind velocity
The importance of microbiotic crusts for surface stabilization and in particular for dune stabilization has long been noted (Belnap and Gillette, 1997). Biological soil crust, which are formed by communities of several types of microphytes including mosses, lichens, fungi, green and blue-green algae as well as bacteria, are a common and widespread phenomenon in sandy soils. By creating a multi-layered net of sheaths and filaments and by exopolysaccharide and slime excretion, the filamentous cyanobacteria, together with other microorganisms, bind the upper surface, forming an intricate webbing of fibers in the soil. In this way, loose soil particles are joined together, and otherwise unstable and highly erosion-prone surfaces become resistant to both wind and water (Belnap and Gardner, 1993). Figure 4 indicates that surface microbiotic crusts on fixed sandy soils have great effects on wind erosion rates, and wind erosion increases with decreasing percent of crust cover, following negative exponential functions. Wind erosion rate for sandy soil with 0% crust cover was about 1.8, 9.5 and 9.4 times of the soil with 100% crust cover at the wind velocity of 8, 18 and 26 m s⁻¹ respectively. The results demonstrate that soil crust has a high effectiveness in controlling wind erosion.
range <0.05 mm for the cultivated sandy soil than for the undisturbed fixed sandy soil (Table 1). Also, Table 1 indicates that the contents of >0.25 mm fractions increase with the increase of wind velocity for the cultivated sandy soil, while <0.05 mm fraction shows an opposite trend. There are no obvious changes of size distribution for the undisturbed fixed sandy soil.

Table 1. Grain size distribution of eroded soil samples at the wind velocity of 10 m s\(^{-1}\), 18 m s\(^{-1}\), and 26 m s\(^{-1}\) from the wind tunnel experiments

<table>
<thead>
<tr>
<th>Velocity</th>
<th>Treatments</th>
<th>Percent distribution of soil grains (mm)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>&gt;2.0</td>
<td>2.0-1.0</td>
</tr>
<tr>
<td>10 m s(^{-1})</td>
<td>Undisturbed</td>
<td>0</td>
</tr>
<tr>
<td></td>
<td>Cultivated</td>
<td>0.5</td>
</tr>
<tr>
<td>18 m s(^{-1})</td>
<td>Undisturbed</td>
<td>0</td>
</tr>
<tr>
<td></td>
<td>Cultivated</td>
<td>0.3</td>
</tr>
<tr>
<td>26 m s(^{-1})</td>
<td>Undisturbed</td>
<td>0.4</td>
</tr>
<tr>
<td></td>
<td>Cultivated</td>
<td>3.1</td>
</tr>
</tbody>
</table>

**Conclusions**

The main results of the experiment can be summarized as follows:

1. Human disturbance such as cultivation or trample can accelerate the erodibility of the fixed sandy soil. The ratio between total soil loss from the undisturbed soil and the cultivated soil was about 0.004.
2. Surface vegetation and microbiotic crust are the main factors responsible for the natural wind erodibility of the fixed sandy soil. Wind erosion rate increases with decreasing percent of the vegetation and crust cover.
3. The grain size distribution showed a higher percentage of particles in the range >1.0 mm and a lower percentage in the range <0.05 mm for the cultivated sandy soil than for the undisturbed fixed sandy soil.

**Acknowledgments**

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**Reference**


Sidewall Effects of Wind Tunnel on Aeolian Sediment Transport

Z. S. Li, Center for Environmental Sciences, Peking University, Beijing 100871 China
The Key Laboratory of Water and Sediment Sciences, Ministry of Education, China
(E-mail: lizhenshan@iee.pku.edu.cn)

J. R. Ni, Center for Environmental Sciences, Peking University, Beijing 100871 China
The Key Laboratory of Water and Sediment Sciences, Ministry of Education, China
(E-mail: nijinren@iee.pku.edu.cn)

Introduction

Wind tunnels are deployed for investigating aeolian sediment transport. But the sidewalls of wind tunnels may distort the measurements of aeolian sediment transport. The width, or more general parameter--aspect ratio (width/height), of a wind tunnel is a major factor in determining if sidewalls effect will be important. Rasmussen and Mikkelsen (1991) presumed that sidewalls did influence wind measurements of Kawamura (1951) experiments conducted in a 5 cm wide wind tunnel. Horikawa and Shen (1960) and Belly (1964) suggested the sidewall effect on the velocity distribution could be ignored compared with a relative wide wind tunnel. Obviously, there needs to be more experimental data to explore the relation between the sidewall effect and the aspect ratio.

The aim of this short paper is to describe experimentally the sidewall effects of wind tunnel on wind velocity distributions and sand mass flux. The experiments were conducted in a wind tunnel, the aspect ratios of which can vary from 0.33 to 1.00.

Methods

The experiments were carried out in a straight-line blowing wind tunnel at the Shapotou Desert Research Station, Lanzhou Institute of Desert Research (Chinese Academy of Sciences). The tunnel body rests on the lab floor. The test section is 21 m long, 1.2 m high, and 1.2 m wide. The floor of the test section consists of seven panels, each one of which is 3 m long and can be removed to meet specific experiment needs. These panels were used as the added sidewalls of wind tunnel in the experiments (Fig.1). The widths of added sidewalls are 1 m, 0.8 m, 0.6 m and 0.4 m, respectively. The wind tunnel in the experiments had five aspect ratios (width/height) of 1.00, 0.83, 0.67, 0.50 and 0.33, respectively.

Wind velocities were measured with a single hack tube connected to a digital pressure meter that, in turn, was controlled by an IBM PC. Figure 2 shows the measurement positions of wind velocity. In Fig.2, $y$ is the distance from the left wall, $B$ the width of wind tunnel, $z$ the height above the bed, and $H$ the height of wind tunnel. A Liu’s type passive vertical array sand trap (Liu, 1995), used to measure the vertical sand mass flux profiles, is 30 cm high and has 30 collection chambers. The aperture of each chamber is 1 cm wide and 1 cm high. The test sand was obtained from Tengger Desert in China. The mean grain diameter is 0.35 mm. To guarantee the same flux of wind flow in the experiments with different aspect ratios, the rotate speed of electric machine kept at 375 r.p.m. in all the runs.
Results and Discussion

The results of velocity for five aspect ratios are plotted in Fig. 3. The sidewall affects: (i) wind velocity field; (ii) upper-edge heights and wind velocity distribution of the inner saltation layer; (iii) effective roughness and upper-edge heights in the outer saltation layer; (iv) wind friction velocity in the cross sections; (v) free stream wind velocities; (vi) sand transport rate; and (vii) profiles of sand mass flux.

With the decrease of aspect ratio, the region where sidewall effects on wind velocity increase compared with the width of wind tunnel. The pattern of wind velocity isoline near the bed surface vary from W shape into beeline with the decrease of aspects ratio. The region of maximum wind velocity moves down with the decreasing aspects ratio.

In a cross section, the effective roughness corresponding to the central vertical profile is greater than those in other vertical profiles. The wind friction velocity as well as the upper-edge heights for velocity profiles in either inner or outer saltation regions increase from the sidewall to the central line of wind tunnel. Both the ratio of friction velocity to free stream wind velocity and the thickness of airflow are subject to the sidewall effect, showing the increase trend with the decreasing aspect ratios.
Figure 3. Cross sections of wind velocity at x = 13.5 for five aspect ratios of the wind tunnel.

The measurements of sand mass flux for four aspect ratios are shown in Fig. 4. The sand transport rates increase with the decreasing aspect ratio. Gradient of sand mass flux profiles in central line are less than ones in other lines in the wind tunnel.

Figure 4. Measured sand mass fluxes in four wind tunnels.
Conclusions

The sidewall exerts significant effects on the parameters of vertical profiles of wind velocity and sand mass flux. The parameters are effective roughness, wind friction velocity, free stream wind velocity, sand transport rate, profiles gradient of sand mass flux etc. Sidewall effects should be taken into account especially for the wind tunnels with small aspect ratio.

References


Mechanics of Crust Rupture and Erosion

Cheryl McKenna Neuman, Department of Geography, Trent University, Peterborough Ontario Canada, K9J 7B8 (cmckneuman@trentu.ca)

Ann Rice, Dept. of Engineering, University of Aberdeen, Kings College, Aberdeen Scotland, AB32 6PP (A.Rice@eng.abdn.ac.uk)

Despite their recognized importance worldwide, very little work has addressed the physical characteristics of crusts in relation to the threshold for particle entrainment by wind, their response to impact, and the mechanisms by which they disintegrate. Over the last several years parallel and complimentary studies of the mechanics of crust rupture and erosion have been carried at the University of Aberdeen and Trent University. The work at Aberdeen has specialized in crusts formed from clay minerals, as well as the development of physical models designed to compliment the measurement of crust strength using a needle penetrometer. Research at Trent has focused upon biological crusts, which are thinner and weaker, but more elastic than clay-set crusts. These surfaces rupture initially under grain impact, though eventually undercutting of the thin crust leads to large-scale mass loss as wind drag on the surface pries off large flakes. Measurement of crust failure in flexure testing appears to be appropriate in the analysis of crust breakdown at this advanced stage of wind erosion.

A number of important research questions have emerged from this work which are now being addressed in collaborative experiments. These experiments focus upon characterising the susceptibility of crusted surfaces to the abrasion that occurs during saltation. Penetrometry seems to be a reliable method for evaluating the relative strength of varied crust types, but in fact we still don’t know much about what the absolute values actually mean. Most test plots show an early maximum peak strength and then a series of minor peaks until the penetrometer reaches the unconsolidated sand below the crust. Does this mean that the breaking of the adjacent interparticle bonds by the initial penetration leads to cracks that extend to the surrounding material? As a consequence, the applied load would not need to be as high during the rest of the penetration. An important question to address is whether or not saltating particles have the same effect. Recent wind tunnel experiments at Trent seem to suggest that some form of fatigue is important, since the duration of impact seems to outweigh the importance of impact velocity as a control of mass loss. Ongoing experiments also address effects associated with the rate of loading, the diameter of the needle relative to grain size, and the angle and spacing of the crust penetrations.
Scaling of surface wind speed

Thomas M. Over, Department of Geology/Geography, Eastern Illinois University, Charleston, IL 61920 (Email: tmover@eiu.edu)

Paolo D’Odorico, Department of Environmental Sciences, University of Virginia, Charlottesville, VA 22904 (Email: paolo@virginia.edu)

Introduction

Studies of the scaling of surface wind speed have mainly considered power spectra at small time scales where specialized instrumentation such as hot-wire and sonic anemometers are required, i.e., in the turbulent inertial (microscale) range where a power law with exponent $\beta = -5/3$ is expected and has been observed (e.g., Kaimal et al., 1972), and in the turbulent energy production (mesoscale) range where a power law with exponent $\beta = -1$ power law is predicted and observed (e.g., Katul and Chu, 1998, and references therein). In this study, however, we are interested in what can be seen at time scales observable by standard cup anemometers (i.e., at time resolutions of one minute and above, at the upper end of the mesoscale and above). In addition, recently more general analysis and modeling tools for scaling processes have been developed that go beyond the second-order statistics of power spectra and thus give a fuller picture of the wind structure. These have been applied to micro- and mesoscale winds by Lauren et al., 2001. Here these tools are applied to anemometer data with time resolutions of one to five minutes from Lubbock, Texas and Dodge City, Kansas.

Methods

Several two to four week periods of anemometer data from Lubbock, Texas, and Dodge City, Kansas were analyzed. For some selected series, the periods considered and some basic information along with some results are given in Table I.

A multiscaling analysis following Davis et al. (1994) was performed. This consists of the following steps: (1) spectral analysis to identify scaling ranges, (2) singular measure analysis, and (3) structure function analysis. Power law behavior in the power spectrum gives the initial indication of scaling behavior, and the spectral exponent $\beta$ gives the domain of the process, which is needed for the rest of the analysis.

A singular measure analysis, as we use the term here, consists of computing the moment scaling function $K(q)$, which is given by

$$\langle \varepsilon(\lambda, t)^q \rangle \sim \lambda^{-K(q)}, \ \lambda_{\text{min}} \leq \lambda \leq \lambda_{\text{max}},$$

where $\varepsilon(\lambda, t)$ is the singular measure at time $t$ and coarse-grained to resolution $\lambda$, $\langle \rangle$ indicates averaging over all times $t$, and $\lambda_{\text{min}}$ and $\lambda_{\text{max}}$ are the minimum and maximum resolutions, respectively, over which the scaling holds. In practice, $K(q)$ is found by regressing $\log \langle \varepsilon(\lambda, t)^q \rangle$ vs. $-\log \lambda$ for a set of moment orders $q$, the slope of the regression giving $K(q)$. Note that in
the case of several scaling ranges as are observed here, different $\lambda_{\text{min}}$, $\lambda_{\text{max}}$, and $K(q)$ apply to each. A special value of $K(q)$ is its derivative at $q = 1$, called $C_1$, which gives the co-dimension of the (measure-theoretic) support of the singular measure. Qualitatively, the bigger $C_1$ is, the more intermittent is the measure.

The moment scaling function $K(q)$ may be used to estimate the parameters of a process called a random cascade which may be used to simulate a singular measure with the desired $K(q)$ function (see, e.g., Tessier et al., 1993; Over and Gupta, 1994).

Singular measures have spectral power law exponents $\beta = -d + K(2)$, thus $-1 < \beta < 0$. Thus if the spectral exponent of some series has the property $\beta < -1$, then the absolute or squared gradients of the series should be analyzed via a singular measure analysis (if $-3 < \beta < -1$), or gradients of gradients for smaller values of $\beta$.

The structure function is designed, as the discussion above suggests, to study the fluctuations of a process $\mathcal{S}(t)$. The structure function $\zeta(q)$ is defined as

$$\left\langle |\Delta \mathcal{S}(\lambda, t)|^q \right\rangle - \lambda^{\zeta(q)}, \quad \lambda_{\text{min}} \leq \lambda \leq \lambda_{\text{max}},$$

(2)

where $\Delta \mathcal{S}(\lambda, t) = \mathcal{S}(t + \lambda) - \mathcal{S}(t)$. So here the resolution $\lambda$ indicates the separation distance of two values of the process $\mathcal{S}(t)$ rather than a coarse-grained version of it as it does in singular measure analysis. Similar to the case of the computation of $K(q)$, in practice, $\zeta(q)$ is computed by regressing $\log(|\Delta \mathcal{S}(\lambda, t)|^q)$ vs. $\log \lambda$. A special value of $\zeta(q)$ is $H = \zeta(1)$ which gives the Hurst exponent $H$ of fractional Brownian motion (fBm). In fact, for fBm with exponent $H (0 < H < 1)$, $\zeta(q) = Hq$. For fBm, the power spectrum follows a power law with $\beta = -1 - 2H$, so $0 < H < 1$ implies $-3 < \beta < -1$, thus a Gaussian process with $\beta$ in this range may be fBm, while if $-1 < \beta < 1$, it could be modeled as the increments of fBm, i.e., fractional Gaussian noise (fGn). Thus if a process has $-1 < \beta < 1$, before computing its structure function, it should be cumulated.

**Results**

Results of the are given below in Table I. The most significant finding was that most of the series showed a set of three power law scaling ranges in their power spectra, approximately from 1-10, 10-100, and 100-1000 minutes, with $\beta$ near $-1$ in the first, $-1 < \beta < 0$ in the second (suggesting the fGn or singular measure domains), and $-3 < \beta < -1$ in the third (suggesting fBm). An example spectrum from the ARS 7/01 series is given in Figure 1. It should be noted further that the ASOS data was originally in knots and were rounded to the nearest knot, which explains the large value of $C_1$ for the absolute gradient in the highest frequency range.

<p>| Table I. | Basic information and results regarding wind data and simulations used in this study. All data is from the 10-m anemometer. The ASOS and ARS data are from Lubbock. The TTU-Mesonet site used is at Reese Center. |</p>
<table>
<thead>
<tr>
<th>Data Source, Month</th>
<th>Time Res. (min.)</th>
<th>Duration (min.)</th>
<th>Average Speed (m/s)</th>
<th>Scale Range (minutes)</th>
<th>$-\beta$</th>
<th>series $H_1$</th>
<th>cum. $H_1$</th>
<th>series $C_1$</th>
<th>grad. $C_1$</th>
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<td>2</td>
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<td>5.04</td>
<td>2-8</td>
<td>0.98</td>
<td>0.22</td>
<td>0.98</td>
<td>0.004</td>
<td>0.26</td>
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<td></td>
<td></td>
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<td></td>
<td>8-128</td>
<td>0.71</td>
<td>0.26</td>
<td>0.98</td>
<td>0.004</td>
<td>0.051</td>
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<td></td>
<td></td>
<td></td>
<td></td>
<td>128-1024</td>
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<td>0.86</td>
<td>0.017</td>
<td>0.066</td>
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<tr>
<td>ARS, 7/01</td>
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<td>1-8</td>
<td>1.24</td>
<td>0.20</td>
<td>0.96</td>
<td>0.007</td>
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<td>4.75</td>
<td>5-80</td>
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<td>0.97</td>
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<td></td>
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<td></td>
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<td>160-640</td>
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<td>0.24</td>
<td>0.76</td>
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<td>160-640</td>
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<td>0.32</td>
<td>0.85</td>
<td>0.017</td>
<td>0.066</td>
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</tbody>
</table>

Results for simulated series:

- $fBm$, $H = 0.3$
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<th>Scale Range (minutes)</th>
<th>$-\beta$</th>
<th>series $H_1$</th>
<th>cum. $H_1$</th>
<th>series $C_1$</th>
<th>grad. $C_1$</th>
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</thead>
<tbody>
<tr>
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<td>0.20</td>
<td>4-2048</td>
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</table>

- $fGn$, $H = 0.8$
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<thead>
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<th>Time Res. (min.)</th>
<th>Duration (min.)</th>
<th>Average Speed (m/s)</th>
<th>Scale Range (minutes)</th>
<th>$-\beta$</th>
<th>series $H_1$</th>
<th>cum. $H_1$</th>
<th>series $C_1$</th>
<th>grad. $C_1$</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>65,536</td>
<td>0.00</td>
<td>4-2048</td>
<td>0.58</td>
<td>0.022</td>
<td>0.75</td>
<td>0.002</td>
<td>0.012</td>
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</tbody>
</table>

**Figure 1.** Power spectrum of ARS-Lubbock 10-m wind data for July, 2001.

It generally appears that the singular measure analysis is showing weak intermittency ($C_1$ near zero), especially if other values of the gradient $C_1$ are affected by rounding. This would suggest that $fGn$ or $fBm$ (and not a cascade) might be appropriate modeling/downscaling tools, and moment analyses (not shown) suggest the densities (though positively skewed) are not far
from Gaussian, and the complete structure functions (also not shown) are indeed nearly linear. However, the expected behavior of the series and cumulative $H_i$ values, as may be seen by comparing the simulated fBm and fGn results, is not obtained for the data series. A basic conceptual issue in the use of fBm to model these series would be that they probably should be, on a physical basis, stationary. We note that the value of $\beta$ near –1 for the highest frequency range matches previous results for the energy production range, though in our case, this value may be affected by the inertia of the instrument. Also, the behavior at the largest scale includes some effects of the diurnal cycle. In summary, while many questions remain to be answered, this analysis has a valuable window on the structure and possible modeling approaches to low frequency surface wind data.

Acknowledgements

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References


Preliminary laboratory studies of shear stress partitioning

Keld R. Rasmussen, Department of Earth Sciences, University of Aarhus, Block 520, Ny Munkegade, DK8000 Aarhus C, Denmark. Email: geolkrr@geo.au.dk.

Introduction

In the saltation layer the moving grains carry part of the momentum while in the flow above the horizontal momentum is carried entirely by the fluid. As explained by Owen (1964) the saltation process thus leads to a partitioning of the shear stress in the saltation layer between the wind flow and the saltating particles. Ultimately this leads to an upward convexity of the velocity profile near the bed. The influence from moving grains shear on the vertical wind speed profile in the saltation layer has been deduced from numerical models of the saltation process (e.g. Sorensen, 1985; McEwan and Willetts 1991). Calculations have indicated that for moderate friction speeds there should be a noticeable increase in the wind speed (expressed as a deviation from the logarithmic wind law) up to a few centimeters above the bed (e.g. Sorensen, 1985).

The experimental verification of this influence has been debated earlier a matter for debate over the years (Butterfield, 1999). The reasons are several including the obvious fact that it is difficult to measure precisely the air speed shortly above a grain bed, which systematically deforms because of moving ripples. Furthermore, most (if not all) aeolian laboratory studies have been made in wind tunnels where a horizontal pressure gradient influences the wind profile during saltation. Thus the upper part of the wind profile deviates from its ideal log-linear shape because of the influence from a slight wake at the top of the boundary layer (White, 1991). Also Froude Number effects may have influenced some results obtained in wind tunnels with a small cross section (White and Mounla, 1985). Finally, results from atmospheric investigations and wind tunnels show that the shear stress will only attain its surface value near the bottom of the boundary layer (e.g. Garrett, 1994).

In order to overcome some of the problems outlined above the long aeolian wind tunnel at the University of Aarhus was modified so that the boundary layer thickness could be increased. Furthermore, a technique based on measuring the shear stress in the saltation layer using hot-wire anemometry was introduced. Previous use of this sensor (Rasmussen and Sørensen, 1998) indicated that it survives several hours of use in the saltation layer. In the following the new experimental set-up will be described including some of the rather cumbersome hot-wire calibration experiments. Finally initial measurements made over a stationary grain bed will be presented.

Experimental set up

The 20 m long, horizontal wind tunnel at the University of Aarhus is a low speed, open circuit suck-down tunnel. It is driven by an axial fan at the flow outlet. Its width and height are 0.6 m by 0.9 m, and there is a small bell-mouth and filters at the entry while moving particles are collected in a large settling chamber (with filters) just in front of the fan. The fan can be set at fixed RPM with a resolution of about 1 %. Enhanced growth of the boundary layer is made immediately downstream from the entry using turbulence spires and roughness arrays.

For each of five nominal values of friction speed $u_*$ (0.20, 0.30, 0.40, 0.60, and 0.75 m/s), corresponding sets of turbulence spires and roughness block arrays (Irwin, 1981) are placed in
the first 2.5 m downwind of the entry. The spires produce a boundary layer thickness of 15 cm approximately 2.5 m downwind of the entry while the roughness block arrays are designed to match values of aerodynamic roughness length (z0) typical for aeolian grain beds. In the experiments reported here, a 0.5 cm layer of almost uniformly sized quartz sand or gravel was spread on the bed between the roughness array and the sand collector.

While the wind tunnel was being modified calibration experiments were made with the split-fiber sensor. In principle the sensor records the velocity (speed and direction) of a two-dimensional flow using signals from the two electrically separated half parts of a platinum coated cylindrical ceramic fiber (D=200 µm). Thus the friction speed (stress) can be obtained as the time average of the product \( u^* = \sqrt{-u'w'} \) where \( u' \) and \( w' \) are the fluctuating components of the horizontal and vertical velocities, respectively. However, the velocity signal depends on flow direction and vice versa (Rasmussen and Sorensen, 1998), but existing information on experimental error and calibration procedure for this sensor is scarce. Therefore several sensors were calibrated with an automatic calibration system (DANTEC, 90H10) using different calibration procedures and different settings of the electronics. In the calibrations reported here the sensor was first placed perpendicular to the flow (zero pitch) and the output was measured for 15-20 velocity readings. Next, the sensor was placed at a fixed pitch of -60˚ and the output was recorded once more for 10-15 velocity readings. This was repeated for every 10˚ increase in pitch angle in the interval -60˚ to +60˚.

In the wind tunnel vertical profiles of the wind speed and the shear stress were recorded above a quiescent grain bed. At each height data were recorded at a rate of 100 Hz continuously during 400 seconds. The profiles were obtained for different wind speeds below the saltation threshold at the centerline of the tunnel approximately 15 downstream from the entry.

Results and Discussion

Results from a simple velocity calibration of a split fiber sensor which is placed with the fiber perpendicular to the flow at zero pitch angle is presented in figure 1. The bridge signals from the lower and upper sensor halves, is denoted by V1 and V2, respectively, and the total output (V=V1 + V2) is shown versus the velocity of the calibration jet. When a 4-th order polynomial is fitted to the calibration points the standard error on the deviation between observations and fit is only 0.04 m/s.

The directional response of the sensor is investigated using the voltages obtained for different velocities of the jet (U) and pitch-angles (θ). The voltage at zero velocity recorded from the sensor halves is denoted V_{0,1} and V_{0,2}, respectively. Thus the logarithm of the increase in signal relative to the zero-value for the two sensor is \( \log(V_1-V_{0,1}) \) and \( \log(V_2-V_{0,2}) \), respectively which is plotted in figure 2.

![Figure 1. Calibration characteristic of sensor 5.](image)
It is seen that the data are almost uniformly distributed for the range of angles and speeds tested here. At very low speeds there are small irregularities in the signals. However, the Reynold’s at such low speeds is small (Re≈100) indicating that cooling is influenced in a complex manner by the flow around the fiber. For aeolian work, however, such low speeds (< 1 m/s) are far below the saltation threshold and will not be discussed in further detail.

With the calibration device used in the set-up reported here, very consistent readings were obtained. Setting a similar overheat temperature for the two sensor halves appears to produce almost identical calibration runs. Therefore it was decided to convert data via interpolation within the calibration matrix (conversion of low speed-readings is slightly different, but will not be discussed here). Initially is found the four calibration values surrounding a given set of reduced bridge voltages. Then interpolation is made between these four points in order to calculate the weighing factors for conversion of the corresponding sets of speed and angle measurements. The standard error between the calculated and measured velocity readings is only 49 mm/s for velocities up to 12 m/s when this procedure is used with the calibration data given in figure 1. At higher velocities there is somewhat larger deviation so that for the entire set of values in the range 1-30 m/s the standard error is 19 cm/sec.

Two vertical profiles of the horizontal speed (u) measured above flat beds of different grain size are presented in figure 3. Profile A was measured above a bed of uniform sand with diameter 0.54 mm while profile B was measured above a bed of gravel with median diameter 5 mm. A line representing the logarithmic wind profile \( u = \frac{u_*}{\kappa} \log \frac{z}{z_0} \) was fitted to each set of data, but only data points up to 120 mm above the bed were used in the fit. In both cases the observed velocities follow the logarithmic law quite well. The values of aerodynamic roughness length \( z_0 \) and friction speed \( u_* \) for the two profiles are \( z_{0,A}=6.8 \times 10^{-5} \) m, \( z_{0,B}=2.6 \times 10^{-4} \) m, \( u_{*,A}=0.21 \) m/s, and \( u_{*,B}=0.37 \) m/s.

At the same heights where the horizontal wind speed was measured the shear stress and hence the friction speed can be calculated, too. For the two beds the \( u_* - \)values normalised with the value found at 4 mm height are listed in Table 1. The values are found from averaging over four runs at each height.
The two profiles both have a fairly uniform stress near the bottom. However, for the profile above the smoother surface and at lower friction speed the stress decreases from about 2 cm above the bed, while for the rougher bed and higher speed the decrease starts at about 6 cm height. Also above this bed there is more scatter in the values. At the top of the profiles the stress decreases rapidly above approximately 10 cm height.

Conclusions

The measurements presented here clearly points in the difficulties in investigating experimentally in the laboratory the influence from stress partitioning in the saltation layer. The decrease of the stress with height even when there is no saltation makes it difficult to establish a proper reference for the measurements in the densest cloud of particles. More experiments are needed in order to determine how the height of the constant stress zone depends on bed texture and friction speed. In addition it may seem necessary to compare stresses made in the saltation layer to values obtained above quiescent beds of similar roughness.

References


On the Rate of Aeolian Sand Transport

Michael Sørensen, Department of Statistics and Operations Research, University of Copenhagen, Universitetsparken 5, DK-2100 Copenhagen Ø, Denmark (E-mail: michael@math.ku.dk)

Introduction

During the last 15 years a number of numerical models of aeolian sand transport have been published (Anderson & Haff, 1988; Werner, 1990; McEwan & Willetts, 1991; Spies & McEwan, 2000) that incorporate the present understanding of the physics of sand transport by wind, see Anderson, Sørensen, & Willetts (1991). An analytic model of aeolian saltation based on essentially the same physics (Sørensen, 1991) resulted in a formula for the transport rate that seems to compare well with transport rate data (Rasmussen & Mikkelsen, 1991); see also the discussion in McEwan & Willetts (1994).

Here the results in Sørensen (1991) are improved significantly in two directions. By introducing into the theory a hypothesis by Owen (1964), a more satisfactory and simpler model is obtained. Moreover, the derivation of the formula for the transport rate is far simpler and more lucid. In particular, the constants in the formula have a physical interpretation. The new formula is calibrated to wind-tunnel data by Iversen and Rasmussen (1999).

The wind profile during saltation

The following theory of how the cloud of saltating grains modifies the wind profile is based on the concepts of eddy viscosity and grain borne shear stress. The latter was introduced by Owen (1964). The grain borne shear stress at height \( y \), \( T(y) \), is the part of the shear stress that at height \( y \) is carried by the saltating grains. It can be calculated as (Sørensen, 1985)

\[
T(y) = \Phi \nu(y),
\]

where \( \nu(y) \) is the average increase of the horizontal velocity component of a saltating grain while it is above the level \( y \), and where \( \Phi \) is the mass of grains that on average leaves one unit of surface area per time unit. The air borne shear stress at height \( y \) is equal to \( \rho U^2 - T(y) \), where \( U_* \) denotes the friction speed and \( \rho \) is the density of air. This is the difference between the shear stress in the grain free wind above the saltation cloud and the grain borne shear stress at height \( y \). Owen (1964) argued that at all friction speeds, the air borne shear stress at the surface is equal to its value at the impact threshold \( U_{c*} \), i.e. it is equal to \( \rho U^2_{c*} \). If this is true, we find that

\[
\Phi = \rho U^2_{c*} \left( 1 - V^{-2} \right) / \nu(0),
\]

where \( V = U_*/U_{c*} \).

Next we calculate how the wind profile is modified by the saltating grains based on eddy viscosity theory. We assume that the eddy viscosity is given by

\[
\nu(y) = \kappa y U^2_{c*} - T(y)/\rho = \kappa y U_* \sqrt{1 - (1 - V^{-2}) \nu(y)/\nu(0)},
\]

where \( \kappa \) is von Kármán’s constant. This is analogous to the usual derivation of the logarithmic wind profile and was first proposed by Anderson (1986). The quantity \( \sqrt{U^2_{c*} - T(y)/\rho} \) is
an equivalent friction speed corresponding to the air borne shear stress at height \( y \). The expression (2) implies that the wind profile \( U(y) \) satisfies

\[
\frac{dU}{dy} = \frac{U_*}{k y} \sqrt{1 - (1 - V^{-2}) u(y)/v(0)}.
\]  

(3)

This equation does not yield an explicit expression for the wind profile, but if we approximate \( \sqrt{1 - x} \) by \( 1 - x \) for \( x \) between zero and one, we find that

\[
U(y) = \kappa^{-1} U_* \left[ \ln(y/y_0) - (1 - V^{-2}) b(y) \right],
\]  

(4)

where \( y_0 \) is the height at which the wind speed is zero, and where the dimensionless function \( b \) is given by

\[
b(y) = \int_{y_0}^{y} z^{-1} \frac{v(z)}{v(0)} dz.
\]  

(5)

An alterative derivation of the expression (4) for the wind profile in the saltation layer was made in Sørensen (1985) by assuming that the eddy viscosity is \( \nu(z) = \kappa z U_* \).

![Figure 1: The dimensionless transport rate \( Q g/(\rho U_*^3) \) plotted agains the dimensionless friction speed \( V = U_*/U_* \) for the experiment with homogeneous sand of size 170 \( \mu \)m (indicated by \( \circ \)). The curve is the transport rate formula (7) with \( \alpha = 0, \beta = 2.98, \) and \( \gamma = 2.06. \)](image)

**The transport rate**

The transport rate \( Q \) is given by

\[
Q = \Phi \bar{\ell},
\]  

(6)

where \( \bar{\ell} \) denotes the mean jump length of a saltating grain, see Sørensen (1985). By combining the analytic model for the saltation trajectories in Sørensen (1991) (based on Owen’s (1964)
linear drag approximation) with the wind profile (4), an expression for \( \tilde{e} \) can be obtained, and using (1), we find that the dimensionless transport rate is given by

\[
\frac{Qg}{\rho U^3_*} = \left(1 - V^{-2}\right) \left[\alpha + \beta V^{-2} + \gamma V^{-1}\right]
\]

where \( g \) denotes the acceleration of gravity. The quantities \( \alpha, \beta \) and \( \gamma \) have the following interpretation: \( \alpha v(0)/g \) is the average jump length of a grain that starts vertically in the wind profile \( \kappa^{-1} \left[ \ln(y/y_0) - b(y) \right] \), \( \beta v(0)/g \) is the average jump length of a grain that starts vertically in the wind profile \( \kappa^{-1} b(y) \), while \( \gamma v(0) U_{ec}/g \) is the average jump length of a grain if it were not accelerated by the wind. A simple formula for the transport rate is thus obtained if the quantities \( \alpha, \beta \) and \( \gamma \) do not depend on the friction velocity. They will obviously depend on the type of sand being transported, e.g., on the grain size. Note that \( v(0)/g \) is of the order of the typical duration of a saltation jump.

**Calibration to wind-tunnel data**

The transport rate formula (7) was fitted to data obtained in wind tunnel experiments by Iversen and Rasmussen (1999). Figure 1 shows a plot of the data from an experiment using very homogeneous sand with a typical size of 170 µm and the fitted curve. The formula fits the data well. A slightly better fit can be obtained if \( \alpha \) is given a negative value. Data from an experiment using natural sand with a typical grain size of 230 µm are plotted in Figure 2. Here a negative value of \( \alpha \) has been allowed since the fit for the largest friction speed was otherwise less satisfactory.

Figure 2: The dimensionless transport rate \( Qg/(\rho U^3_* ) \) plotted against the dimensionless friction speed \( V = U_*/U_{ec} \) for the experiment using natural sand with a typical grain size of 230 µm (indicated by o). The curve is the transport rate formula (7) with \( \alpha = -1.78, \beta = 15.9, \) and \( \gamma = 0. \)
Conclusions

A simple explicit formula for the transport rate has been derived based on physical reasoning. The formula contains three parameters with a physical interpretation that can be estimated with measurements of transport rates. The formula fits data from wind tunnel experiments well.

Acknowledgement

I am grateful to Jim Iversen and Keld Rømer Rasmussen for allowing me to use the original data from their wind-tunnel experiments.

References


Flattened residue effects on wind speed and sediment transport

Geert Sterk  Erosion and Soil & Water Conservation Group, Department of Environmental Sciences, Wageningen University, Nieuwe Kanaal 11, 6709 PA Wageningen, Netherlands

Abstract

Mulching with flattened crop residues is widely used to protect soils from wind erosion. Several wind tunnel and field experiments have shown decreased protection of some soil covers with increasing wind speed. In some studies sediment transport was enhanced with flattened residue as compared with the bare soil condition. The purpose of this article was to determine the behavior of wind speed and sediment transport when the soil surface is covered with randomly applied, flattened crop residues. A literature review was conducted to evaluate recent insights in turbulent flow properties and related sediment transport. A conceptual model was then developed to explain decreasing soil protection of a certain residue quantity when the free stream wind speed increases. The main reason is the change in turbulent flow properties of the near-surface wind when non-erodible roughness elements are added to an otherwise smooth surface. The average wind speed is reduced by the roughness but the probability distribution of instantaneous wind speed becomes wider and positively skewed. If free stream wind speed increases, at a certain moment the changed turbulence will cause more wind gusts that exceed the threshold wind speed for soil particle movement than would occur over a bare surface. But it is likely that this only happens when soil cover is less than 10%. For higher soil covers, increasing wind speed will also cause a decrease in soil protection, but natural wind speeds are normally not sufficient to cause enhanced sediment transport, because average wind speed is sufficiently reduced.
Interactions between turbulent wind flow and saltation sand transport

Geert Sterk,a* John van Boxelb Rosaline Zuurbierva

a Erosion and Soil & Water Conservation Group, Department of Environmental Sciences, Wageningen University, Nieuwe Kanaal 11, 6709 PA Wageningen, Netherlands
b Institute for Biodiversity and Ecosystem Dynamics, University of Amsterdam, Nieuwe Achtergracht 166, 1018 WV Amsterdam, Netherlands

Introduction

Aeolian transport of fine to medium sized sand particles is usually occurring as saltation, the jumping movement of grains over the surface. Saltating grains receive momentum from the near-surface wind, which causes particle lift, entrainment, acceleration, and when the particle rebounds with the surface, previously stationary particles will be splashed up (Rice et al., 1999). Saltation transport is highly intermittent under natural conditions (Bauer et al., 1998; Sterk et al., 1998). The gustiness of the natural, unsteady wind causes pulses of transport that are followed by relatively quiet periods without transport. Most predictive mass flux models relate saltation transport to the cube of shear velocity u* without incorporating time-dependence (Bauer et al., 1998). Three recent papers (Bauer et al., 1998; Butterfield, 1998; Sterk et al., 1998) have raised some skepticism about the use of u* or shear stress for modeling of saltation transport at time scales in the order of one second.

The aim of the present study was to determine the driving force of saltation transport. Specific objectives were: (i) to determine instantaneous wind speed and shear stress near the saltation layer, (ii) to relate saltation transport to fluctuations in wind speed and shear stress, (iii) to analyze the effects of turbulent flow structures on saltation transport.

Materials and methods

A field experiment was conducted during the months June and July of 2001. The location was the nature reserve Kootwijkerzand, a drift sand area in the central part of the Netherlands. Two 3-D sonic anemometers (R.M. Young Co., model 81000) were used to measure the three orthogonal components of the wind vector. The anemometers were mounted on a tower, with one anemometer fixed at 2.0 m height. The second anemometer was on an arm of 0.50 m and adjustable in height from 0 to 1.7 m. The sampling frequency of both sonics was 32 Hz. The output was low pass filtered with a 3.33 Hz first order filter and stored at 8 Hz in a CR10 (Campbell Scientific Ltd.) data logger.

Saltation transport was measured with two saltiphones, which is an acoustic sensor that records particle impacts with a microphone (Spaan and Van den Abeele, 1991).

During erosion, part of the saltating sand grains moving through the tube hit the microphone and create pulses. The centre of the microphones was positioned at 0.10 m height. The created pulses were sampled at 8 Hz and stored in the CR10 data logger at the same frequency.

The wind speed data from the sonic anemometers were used to determine shear stress and turbulent flow structures. Reynolds decomposition was applied to separate the instantaneous wind speed vectors into average and turbulent fluctuating parts: \( u = \bar{u} + u' \),
\( v = \bar{v} + v' \), and \( w = \bar{w} + w' \), where overbars denote average values and primes denote the fluctuating turbulent part. The fluctuating velocity components \( u' \) (horizontal streamwise direction) and \( w' \) (vertical direction) were used for kinematic shear stress \((-u'w' \text{ [m}^2 \text{s}^{-2}]\)) calculations, and to determine the turbulent flow structures ejection \((u' < 0, w' > 0)\), sweep \((u' > 0, w' < 0)\), inward interaction \((u' < 0, w' < 0)\), and outward interaction \((u' > 0, w' > 0)\). Finally, saltation transport was correlated with 1) instantaneous horizontal streamwise and vertical wind speeds, 2) kinematic shear stress, and 3) turbulent flow structures.

Results and discussion

During the experiment five days with saltation transport occurred. Two events with a duration of 30 minutes and intense saltation transport were selected for further analysis. The events were almost similar in average wind speed conditions. During the first event, wind speed was measured at 2.0 and 0.4 m height. The average wind speeds at these two heights were 6.21 and 5.38 m s\(^{-1}\), respectively. During the second event, the sonic anemometer at 2.0 m was not working, and wind speed was only measured at 0.3 m. The average wind speed was 5.55 m s\(^{-1}\). Wind directions were exactly opposite, with a westerly wind during the first event and an easterly wind during the second event.

Instantaneous values of horizontal wind speed correlated much better with saltation flux than instantaneous shear stress. Correlation coefficients between \( u' \) and saltation for the first event were 0.54 at 0.4 m height and 0.51 at 2.0 m, and for the second event a correlation of 0.57 at 0.3 m height was obtained. The correlation coefficients between \(-u'w'\) and saltation for the first event were 0.24 at 0.4 m height and 0.15 at 2.0 m, and for the second event a correlation of 0.22 at 0.3 m height was obtained. All correlation coefficients gradually improved when the wind speed data were averaged and the saltiphone data summed for time periods of 0.25, 0.50 and 1.0 sec. The correlation between \( u' \) and saltation improved to 0.69 for the second event (measurement height = 0.3 m), and the correlation between \(-u'w'\) and saltation reached a maximum value of 0.47 for the first event (measurement height = 0.4 m). In short, the horizontal wind speed fluctuations have the greatest impact on saltation transport, while shear stress seems to play a less important but still significant role.

The analysis of turbulent flow structures showed that these structures occur only 16% (first event) to 20% (second event) of the time, but created 62% and 64%, respectively of the average shear stress. The main structures occurring in the flow were sweeps and outward interactions. These structures have a positive \( u' \) and result in opposite contributions to the average shear stress. However, they both result in significant saltation transport, while ejections (positive shear stress contribution) and inward interactions (negative shear stress contribution) did not result in saltation transport. This leads to the conclusion that fluctuations in horizontal streamwise wind speed (\( u' \)) are of more importance for saltation transport than the instantaneous fluctuation in shear stress.

References

The dynamic effects of moisture on the entrainment and transport of sand by wind

Wiggs, G.F.S. Department of Geography, University of Sheffield, Western Bank, Sheffield, S10 2TN, UK (E-mail: g.wiggs@sheffield.ac.uk)

Atherton, R.J. Department of Geography, University of Sheffield, Western Bank, Sheffield, S10 2TN, UK

Baird, A.J. Department of Geography, University of Sheffield, Western Bank, Sheffield, S10 2TN, UK (E-mail: a.baird@sheffield.ac.uk)

Introduction

Understanding the role of moisture in the aeolian sand transport system is crucial in accurately predicting the mass flux of wind-driven sand. The majority of investigations into the role of moisture in aeolian transport have been conducted in laboratory wind tunnels (McKenna Neuman and Nickling 1989). The findings from these works confirm the conceptual idea of a moisture-dependent threshold and the existence of a critical moisture content above which entrainment is difficult and sediment transport is suppressed. However, comparison of these wind tunnel studies has demonstrated that the findings vary widely (Namikas and Sherman 1995). The paucity of field-based enquiries means that results from wind tunnel studies cannot easily be contextualised in terms of the natural system. Furthermore, only a few studies have explored the time-dependent behaviour of aeolian sand transport in relation to natural moisture levels (see Jackson and Nordstrom 1997). The aim of this research was to investigate the influence of moisture on the temporal dynamics of two aspects of the aeolian sand transport system in the field; (i) entrainment thresholds and (ii) mass flux.

Methods

Experiments were carried out in September 1999 and 2000 at Aberffraw beach on the southwest coast of Anglesey, UK. Three near-surface hot wire anemometers on the upper part of the beach were used to record wind speed fluctuations at 1 Hz. Sediment transport was measured using a Sensit grain impact sensor logged at a frequency of 1 Hz. Sediment flux was measured over periods of 10 minutes using a sand trap and gravimetric surface moisture content to a depth of 2 mm was also measured at 10 minute intervals. Two experiments were conducted. The first involved a series of measurements taken during a surface drying episode and was used to determine the effect of moisture on grain entrainment. The second was undertaken during a period of high wind velocity and included a period of surface wetting by rainfall and subsequent drying. This experiment was used to determine the effect of moisture on sediment flux.

Results and Discussion

Establishing a representative threshold

A modification of the time fraction equivalence method (TFEM, Stout & Zobeck, 1996) was used on the data from the first experiment to determine the change in critical threshold for entrainment as the beach surface dried. The technique assumes that the fraction of time in
which sand transport events occur is equivalent to the fraction of time that the wind speed is equal to or in excess of the threshold value. By converting the Sensit and wind speed data to two binary series the threshold value that gave rise to this equivalence was determined from a cumulative frequency plot of wind speed data.

The data were analysed in 20 minute periods with results expressed as the number of sand transport events occurring above the threshold as a percentage of the total number of sand transport events recorded. Overall, the results indicated a poor relationship between the calculated wind speed threshold and simultaneously measured sand transport. In most cases fewer than 50% of the sand transport events were explained by a corresponding threshold. In an attempt to improve the relationships the data were re-analysed taking into account a time lag between wind velocity and sediment transport, velocity measurement height, undetected transport events, sampling period, shear velocity and turbulence intensity. Results indicated that most sand transport events (67-91%) were explained by a critical threshold analysed using the TFEM based on a 40 second time-averaged wind velocity measured at height of 0.25 m.

Dynamic effects of moisture on entrainment

Figure 1 shows data for sand transport (expressed as the intermittency value) and wind speed during a surface drying episode. Figure 2 shows corresponding critical thresholds calculated from the TFEM analysis.

Figure 1: Wind speed (closed symbols) and intermittency value (open symbols)

The data presented in Figures 1 and 2 indicate that from 12.52 onwards the intermittency value responded directly to changes in wind velocity \((p = 0.001)\). Between 12.00 and 12.40 the calculated value of critical threshold reduces, but this is not statistically significant. This lack of a significant change in the threshold over time suggests that once saltation was initiated, the surface conditions did not change sufficiently for any effects to be reflected in the calculated critical values. However, the similarity in wind speeds recorded prior to and during saltation activity suggests that the comparatively high moisture content measured in the early part of the experiment restricted the availability of entrainable material.
These results provide evidence of a switch from a system controlled dominantly by moisture, prior to 12:20, to a largely wind regulated system thereafter in which moisture assumes a less significant role. This switch between dominant controlling variables suggests the existence of a wind speed specific moisture threshold for the initiation of saltation activity. Measured data suggest that this threshold lies within the region of 4 - 6 % which is in excess of critical moisture contents specified in previous wind tunnel investigations.

Dynamic effects of moisture on transport

Data in Figure 3 suggest that prior to rainfall sediment flux is responding to changes in wind speed ($p < 0.001$). After rainfall flux is reduced by an order of magnitude but steadily increases to approach pre-rainfall levels. Step-wise regression identified moisture content as the most significant control on transport in this period ($R^2 = 69.3\%$), whilst wind speed was insignificant. This experiment illustrates a switch from mass flux being controlled by windspeed prior to rainfall and regulated by sediment moisture content (for a period of 1 hour) after rainfall. The system appears to be very sensitive to subtle changes in moisture content although it is clear that moisture levels of up to 1.68% are not a barrier to aeolian transport.

Conclusion

The experiments have demonstrated the sensitivity of aeolian sand transport to small changes in moisture content. There is good evidence for the existence of a wind speed specific and moisture threshold, with a decrease in moisture content of the order of 1 - 2 % resulting in the initiation of entrainment. The sensitivity of mass flux to moisture changes of the order of 1 % was also shown. Recovery in the mass flux record, following light rainfall, accord with a decrease in moisture content and not with fluctuations in wind speed, as found prior to rainfall. The aeolian sand transport system is conceptualised as responding to time-dependent shifts or ‘switching’ between wind speed and surface moisture as the main controls on saltation activity.
Figure 3: Records of measured system parameters before and after rainfall: (a) wind speed, (b) trap moisture content, (c) mass flux rate

References


The formation and stability of surface armoring by coarse sand particles

Yu Yi, Faculty of Agriculture, Tottori University. (yuyi@phanes.muses.tottori-u.ac.jp)

Okumura Takenobu, Faculty of Agriculture, Tottori University. (okumura@muses.tottori-u.ac.jp)

Sueyasu Katsumi, Faculty of Agriculture, Tottori University.

Kamichika Makio, ALRC, Tottori University. (kamichi@alrc.tottori-u.ac.jp)

Introduction

Considerable knowledge about the influence of coarse grains on sorting processes has been gained during past years. Bagnold (1941) indicated that sand removal must cause a progressive coarsening of the bed surface, when the wind strength was below the ultimate threshold, movement must cease after a certain interval of time, because the surface becomes stabilized by the formation of a protective layer of grains sufficiently large to be immovable. Rasmussen and Mikkelsen (1991) also pointed out, that when grains were gradually exposed to the air flow, the average speed of creeping grains was less than that of saltating grains and the latter could progressively be blown away from the bed leaving the creeping grains as a residual armouring. As a result of sorting processes on an erodible surface, Yu and Okumura (2002) found that the positioning of coarser particles on erodible surfaces changed after an initial period, and an alignment of the coarser particles with wind direction was observed. The assembled coarser sand particles were more effective in protecting erodible sand than individual coarser particles. Even if the coarser particles on the erodible surface were not large enough to be immovable, the erodible surface would also become stabilized through positioning of assembled coarser particles (referred to as oriented assemble). Yu and Okumura (2002) suggested that the movement and stability of the oriented assemble would be related to interspaces of coarser particles on the surface in different wind condition.

The purpose of this study was to further rarefy the formation and stability of oriented assemble influenced under interspaces of topmost grains through a serial of wind tunnel experiments.

Materials and Methods

Experiments were conducted in an indoor wind tunnel with a 50cm×40cm (spanwise) in cross-section and 17m in length at the Arid Land Research Center of Tottori University, Japan. The dimension of the experimental section is presented in Figure 1. The sampling area was 10cm perpendicular to the stream and 20cm streamwise. A coarser sand layer (2.00>d>0.84mm, dyed black), with thickness of approximately one grain and 5 varieties of interspaces (Table 1), were placed on a top of the fine sand glued onto the tunnel bed. Interspaces sizes of coarser sands were estimated by the ratio of empty area (REA) which was obtained by calculating the ratio of the empty area to the total sampling area from photographs obtained with a digital camera system mounted on the tunnel roof. Coarser sands layers were intermittently exposed to 3 kinds of wind friction velocities (VL: 24, VM: 32 and VH: 39 cm/s; Willetts and Rice, 1988). Wind conditions and elapsed erosion time are
presented in Table 2. Before starting of each experiment and at 5 minutes intervals the sand trays were weighted and REA were measured.

![Figure 1. Dimensions of experimental section in wind tunnel.](image)

**REA:** ratio of empty area (=Empty area (strip)/Total sampling area).

<table>
<thead>
<tr>
<th>Sample</th>
<th>S5</th>
<th>S10</th>
<th>S15</th>
<th>S20</th>
<th>S25</th>
</tr>
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<tbody>
<tr>
<td>Weight(g)</td>
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<td>10.000</td>
<td>15.000</td>
<td>20.000</td>
<td>25.000</td>
</tr>
<tr>
<td>Weight per unit area(g/cm²)</td>
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<td>0.050</td>
<td>0.075</td>
<td>0.100</td>
<td>0.125</td>
</tr>
<tr>
<td>REA(%)</td>
<td>75.800</td>
<td>64.000</td>
<td>32.700</td>
<td>14.600</td>
<td>8.400</td>
</tr>
</tbody>
</table>

REA: ratio of empty area (=Empty area/Total sampling area).

<table>
<thead>
<tr>
<th>Pattern</th>
<th>Time elapsed (min.)</th>
<th>0-5</th>
<th>5-10</th>
<th>10-15</th>
<th>15-20</th>
<th>20-25</th>
<th>25-50</th>
<th>50-55</th>
<th>55-60</th>
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<tbody>
<tr>
<td>VL</td>
<td>Friction wind velocity</td>
<td>VL</td>
<td>VL</td>
<td>VL</td>
<td>VL</td>
<td>VL</td>
<td>VL</td>
<td>VL</td>
<td>VL</td>
</tr>
<tr>
<td>VM</td>
<td>Friction wind velocity</td>
<td>VM</td>
<td>VM</td>
<td>VM</td>
<td>VM</td>
<td>VM</td>
<td>VM</td>
<td>VM</td>
<td>VM</td>
</tr>
<tr>
<td>VH</td>
<td>Friction wind velocity</td>
<td>VH</td>
<td>VH</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

**Results and Discussions**

The effect of interspaces sizes on the oriented assemble

With the reduction of friction velocity and REA, more coarser sand particles were remained on the surface (Figure 1). After 50 minutes, most of the remaining coarser sand particles appeared on oriented assemble condition (Photo. 1) except S25 in the case of VL.

The residual coarser sands on the surface presented in the cases of VL and VM, which REA was as smaller as that of 33% and 15% respectively. However, in the cases of VH, the coarser sands layer was eroded completely in 2 minutes, even though the REA was below 9%. The phenomenon of oriented assemble can occur only for certain interspaces sizes of coarser sands. If the interspaces size is too large, all the sand grains will be blown away. On the other
hand, if the interspaces size is too small, the sand grains cannot adjust their positions and the oriented assemble cannot form. The friction velocity is also an effective factor.

Figure 2. The weight loss of coarser sands in 3 kinds of friction velocity.

\begin{figure}
\centering
\includegraphics[width=\textwidth]{friction_velocity_weight_loss}
\caption{The weight loss of coarser sands in 3 kinds of friction velocity.}
\end{figure}

\begin{figure}
\centering
\includegraphics[width=\textwidth]{positioning_of_sands}
\caption{The positioning of residual coarser sands after blowing.}
\end{figure}

The stability of oriented coarser sand

In a past research, it appeared that the surface of sorting process formed under some certain friction velocity, remained relatively stable as long as the friction velocity and
direction were not changed. In the present study demonstrated the same result was obtained, indicating that with the initial formation of oriented assemble, even if erosion time increased to 50 minutes, the weight loss of coarser sands was less (Figure 2), and the oriented assemble remained stable (Photo.1). However, natural wind is changing all the time, and therefore the stability of oriented assemble under wind blowing from the opposite direction (counter-wind) was studied.

Figure 3 shows, the weight loss after 50 minutes of counter-wind most cases of VL and VM was decreased to an extreme low level. However, only with relative weak VL counter-wind of 5 minutes (from 50 to 55 minutes), coarser sands were removed, so was in the cases of VM. And then during the period of another 5 minutes from 55 to 60 min, with the wind of original direction and velocity, the weight loss was increased again in much higher level than that of volume in the end of 50 minutes. The oriented assemble, formed under a strong long-term wind, could be destroyed by a relative weak wind from the opposite direction. Moreover, inference is made that wind from the side would be destroy the oriented assemble too. The oriented assemble gets formed only when the wind blows from the original direction.

Conclusions

As a result of sorting processes by wind erosion, the phenomenon of oriented assemble (armoring) can occur only for certain sized interspaces and function in original wind direction to stabilize the erodible surface.

References


KISR Long Atmospheric Boundary Layer Wind Tunnel

W. Al-Nassar, S. Alhajraf, M. Al-Sudirawi and G. Joseph, Kuwait Institute for Scientific Research (KISR), KUWAIT. (shajraf@kisr.edu.kw)
Keld R. Rasmussen, Department of Earth Sciences, University of Aarhus, DENMARK.

Introduction

Wind tunnel studies are used in many engineering and environmental applications as a key tool to understand problems associated with aerodynamics and transport phenomena. Wind blown sand, sand accumulation on structures, wind load on civil installations and the dispersion of pollutants over industrial and residential areas are examples where wind tunnel simulations can be used to understand and then control related problems.

This paper introduces a new long wind tunnel designed for atmospheric and aeolian studies. Turbulence spires are used to assist the formation of a thick, uniform, steady boundary layer, and initial data on the flow in the test section are presented. A special problem in aeolian tunnels is the sand feeding at the entry, which will often cause instabilities in the momentum budget. Therefore, it is also discussed how arrays of immobile blocks are applied at the upwind edge of the tunnel. It is attempted to design the arrays with an aerodynamic roughness length that will match the dynamic roughness of saltating particles in the test section thereby simulating an infinite upwind fetch.

The KISR Wind Tunnel

The wind tunnel is a low speed, open circuit wind tunnel driven by a centrifugal air fan of 75 kW maximum power at the flow inlet with maximum air speed of 30 m/s, Fig. 1. The intake and the centrifugal fan itself, a wide-angle diffuser, settling chamber, and contraction are all placed in a separate room attached to the main laboratory with the working section. Air can enter the fan from the laboratory or from the outdoor environment through a large gate covered by a filter.

Figure 1: KISR wind tunnel components.
The wide-angle diffuser is connected to the fan via a short, straight transition section with an anti-vibration device to avoid vibrations from the fan to progress to the test section. Screens are placed before the wide-angle diffuser and within the settling chamber where a honeycomb is placed, too. A contraction leads to the working section, which ends in a low angle exit diffuser. The 22 m long working section has a 0.95 m by 1.20 m cross section. There are smooth glass walls on both sides of the tunnel and access is possible through the plywood ceiling and floor. The working area for model studies is situated 15 - 20 m downstream from the entry. In the working area there is also access through the walls via a 0.5 m section of stainless steel sheet fitted into the glass panes. Finally, a turntable in the floor is placed within the working area for modelling of different wind direction.

**Experimental set up**

The flow in the tunnel was investigated for fan frequency settings of F=6, 11, 20 Hz which correspond to inlet free stream velocities of 2.8 m/s, 6.3 m/s, and 12.5 m/s at the center of the tunnel inlet cross section, P0. Wind speed measurements were made for at three positions in the test section: P0, P1 and P2 at 0.0m, 3.4 m and 17.0 m, respectively. Turbulence was measured using a 1-D hot-wire probe (DANTEC 55P11), while Dwyer Pitot-static tubes connected to an electronic micro-manometer (Flow Master) were used for measuring the mean velocity field. Four levels of boundary layer control scenarios have been tested. Run 1 was made with an empty tunnel primarily to check the wind tunnel specifications. The test is of limited relevance to aeolian conditions since a laminar sub-layer is present above the smooth plywood floor except at high speeds. Therefore only turbulence data measured at P0 will be discussed. Run 2 was conducted with a thin gravel bed covering the tunnel floor. This bed was chosen because its static roughness is comparable to the dynamic roughness of a saltating sand bed (Rasmussen et al., 1994). The mean flow field at cross sections P1 and P2 was mapped. Measurements were taken along 9 vertical lines with 13 reading points at each line.

In Run 3 the gravel bed was replaced by non-erodible roughness elements (roughness array) for the first 3.5 m of the test section, Fig. 2. The roughness array was designed to match the roughness of the gravel bed, and the flow patterns at P1 and P2 were mapped again. During Run 4 turbulence spires (Irwin, 1981) were fixed at P0 in front of the roughness array Fig. 2. The spires produce a thick boundary layer, which makes estimation of the surface friction speed and roughness length more reliable and also makes saltation dynamics closer to its natural state. The flow was also mapped at P1 and P2.
Results and Discussions

The traversing data made during Run 1 (not shown here) indicate that the wind profile is very uniform across the entry and that the wall layer is about 15 mm thick at P0. The turbulence measurements made at P0 at the center of the entry are shown in Fig. 3 for F=6, 11, 20 Hz. The flow is very steady with a coefficient of variation of 1 % or less, which is relatively closed to 0.1 % obtained in many aeronautical tunnels (Bradshaw and Pankhurst, 1963).

For gravel the velocity field at P1 and P2 is shown in Fig. 4a,b and a vertical profile at P1 is depicted in Fig. 5a. In the free stream the flow is fairly uniform, and at P1 the boundary layer is about 10 cm high while at P2 it is about 15 cm high. Along the glass panes the boundary layer thickness increases from less than 10 cm at P1 to about 30 cm at P2.
The vertical profiles recorded above gravel at P1, Fig. 5a, as well as at the downstream end when using the roughness array at the same position P1, Fig. 5b, are fairly similar. This indicates that the flow pattern at the upwind end of the tunnel is reasonably stable and that the design of the roughness array as given by Irwin (1981) works well in the KISR tunnel. Measurements were repeated with spires added to the previous setting (Run 4). Vertical profiles are shown for different fan speeds at P1, Fig. 6a, and P2, Fig. 6b. These plots show that the boundary layer thickness increases considerably when the turbulence spires are added. At P1 the boundary layer is now about 30 cm high while at P2 is a few centimeters higher. Near the top of the boundary layer the presence of a slight wake (White, 1991) is clearly noticed. The effect of the slight wake is insignificant in the lower part of the boundary layer, say below 10 cm, where good estimates of $u^*$ and $z_0$ can be obtained directly from the profile.

Table 1 presents the friction speed and aerodynamic roughness of the gravel bed for different fan speeds during the final run where both roughness arrays and turbulence spires were inserted in the tunnel. It is observed that the friction speeds $u^*$ at P1 is smaller than that at P2 except at the lowest fan speed where they are very close to each other given the experimental uncertainty. Similarly, the aerodynamic roughness increases as the distance increases from the tunnel entry. Variation of shear stress between up and down stream sections is a common modeling problem especially for long wind tunnels. However, in the KISR wind tunnel a moderately thick boundary layer is created using control measures at the inlet section of the tunnel. The boundary layer
grows slowly between P1 and P2. There is a change of about 10% at all fan settings which means that at the central part of the test section, say between 13m and 18m, the net change of $u^*$ is less than 5%.

Table 1: Friction speed ($u^*$) and aerodynamic roughness length ($z_o$) for Run4.

<table>
<thead>
<tr>
<th>Run 4: gravels + arrays + spires</th>
<th>$F = 6$ Hz</th>
<th>$F = 11$ Hz</th>
<th>$F = 21$ Hz</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>$u_m/s$</td>
<td>$z_o m$</td>
<td>$u_m/s$</td>
</tr>
<tr>
<td>P1 (3.4m from inlet)</td>
<td>0.16</td>
<td>0.00042</td>
<td>0.29</td>
</tr>
<tr>
<td>P2 (17m from inlet)</td>
<td>0.14</td>
<td>0.00009</td>
<td>0.34</td>
</tr>
</tbody>
</table>

Conclusion

Preliminary measurements of the long atmospheric boundary layer wind tunnel established at KISR are presented with various inlet and bed surface settings. Turbulence spires, in addition to gravel bed and rough arrays, show that a good and stable boundary layer can be produced at the test section. In order to control the boundary layer development, different settings of turbulence spires and rough elements will be tested in the next stage and before serious sand drift experiments may take place.

References

Clay and Carbonate Effect on Fine Dust Emissions Measured in a Rotating-Tube Dust Generation System

A. Amante-Orozco, Colegio de Postgraduados, Salinas de Hgo., SLP 78600, MEXICO (E-mail: aamanteo@yahoo.com.mx)

T. M. Zobeck, USDA-ARS, Lubbock, Texas 79415 (E-mail: tzobeck@lbk.ars.usda.gov)

Introduction

Dust emissions from wind erosion are a significant component of the atmospheric aerosol in regions of highly erodible soils (Gatz, 1995). Frequently, human activities and the force of the wind give rise to suspended dust in the atmosphere. Dust created in this way is called “fugitive dust” Because field evaluation of fugitive dust presents serious difficulties (Nickling and Gillies, 1989), considerable effort is now directed toward developing equipment and techniques to generate and analyze aerosol PM$_{10}$ emissions in the laboratory. In this study, we used the Lubbock Dust Generation, Analysis and Sampling System (LDGASS) (Gill et al., 1999; Singh, 1994) developed by the USDA-ARS Wind Erosion and Water Conservation Research Unit in Lubbock, Texas. The LDGASS includes a dust generator module that applies kinetic energy to a dust source sample to simulate aerosol emissions by wind erosion. This system is capable of generating fugitive dust, measuring particle characteristics of the generated dust in situ, and collecting PM$_{10}$ and PM$_{2.5}$ particulate aerosol samples. PM$_{10}$ and PM$_{2.5}$ are particulate matter having aerodynamic diameters $\leq 10$ and $\leq 2.5$ [$\mu$m], respectively. We will present an evaluation of the effect of soil clay and calcium carbonate (CaCO$_3$) content on aerosol PM$_{10}$ and PM$_{2.5}$ production as determined by the LDGASS using two types of aerosol samplers for eight soils from the Southern High Plains near Lubbock, Texas.

Materials and Methods

The LDGASS consists of a dust aerosol generating tube, laser particle size analyzer, and dust settling chamber containing other dust aerosol monitoring and sampling devices (Figure 1). Source samples are placed in a 1-m long, 7 cm square tube (Figure 1-A). The tube is oscillated with the long axis perpendicular to the floor and inverted 27 times per minute, dropping the sample and generating dust by the impact. The dust is conveyed through a laser particle analysis system (Figure 1-E) and then to a settling chamber (Figure 1-F). In the settling chamber, the dust is sampled by a MiniVol sampler (Figure 1-I) to determine PM$_{10}$ concentration, and DataRAM nephelometers (Figure 1-J) to monitor and sample in situ aerosol PM$_{10}$ and PM$_{2.5}$.

Eight agricultural soils from the Southern High Plains near Lubbock, Texas, with combinations of carbonate and clay content were selected from a preliminary study. Two levels (low and high) of soil clay content and two levels (low and high) of soil CaCO$_3$ content were evaluated. Levels of clay content were $< 20\%$ for low and $> 20\%$ for high. For CaCO$_3$ content the low level was $< 3\%$ and the high level was $> 3\%$. Selected soils series were: Amarillo “Amr” (low CaCO$_3$, low clay); Acuff “Acf1 & Acf2” and Olton “Olt” (low CaCO$_3$, high clay); Gomez...
“Gmz” (high CaCO$_3$, low clay); and Drake “Drk1, Drk2 & Drk3” (high CaCO$_3$, high clay). Soils clay and carbonate content ranged from 12.7 to 32.6% and from 0.2 to 13.0%, respectively.

Figure 1. The Lubbock Dust Generation, Analysis and Sampling System (LDGASS).

A mass of 25 g of soil aggregates from 2 to 19 mm were placed in the dust generator tube and agitated for 30 min to create the aerosol. Five replications for each soil were made. An airflow of 200 l/min (generated by a vacuum) was used to transport the dust through the LDGASS.

Results and Discussion

For all soils, notable but proportional differences were observed in PM$_{10}$ concentration (average in 30 min) as measured with a MiniVol and a DataRAM instruments (Figure 2), which showed a correlation coefficient ($r$) of 0.97. The average proportion of PM$_{10}$ as measured with a DataRAM in relation to PM$_{10}$ as measured with the MiniVol was 66%. Paired $t$-test results showed significant differences ($\alpha = 0.05$) among overall means for DataRAM and MiniVol as well as by soil means. Also PM$_{2.5}$ as measured by a DataRAM was proportional to PM$_{10}$ as measured by a DataRAM and MiniVol in all soils (Figure 2). Correlation coefficients among PM$_{2.5}$ and PM$_{10}$ as measured by DataRAM and MiniVol were 0.92 and 0.87, respectively. The overall average proportion of PM$_{2.5}$ in relation to PM$_{10}$ as measured with DataRAM was 27%.

A general trend was observed for aerosol PM$_{10}$ and PM$_{2.5}$ to increase as soil clay and CaCO$_3$ content increased. The effect of soil CaCO$_3$ content on aerosol PM$_{10}$ (as measured with both DataRAM and MiniVol) and PM$_{2.5}$ (as measured with DataRAM) was significant as indicated in Table 1. Soil clay content, on the other hand, produced significant effect for PM$_{10}$ data set for MiniVol but no significant effects for PM$_{10}$ and PM$_{2.5}$ data sets from DataRAMs (Table 1).

As Table 1 indicates, differences among levels of soil CaCO$_3$ content produced greater differences in fine dust generation than those differences for levels of soil clay content. Gill et al. (1999) reported that a calcareous Drake soil produced more PM$_{10}$ than an Amarillo soil of the same soil texture. They suggest that a lower binding energy in highly calcareous aggregates in the silt fraction of the Drake soil resulted in much less stable aggregates than those of the Amarillo soil, and that this fact might have favored their easier disaggregation into fine dust. Also, the increase in aerosol PM$_{10}$ concentration with increases in soil clay content has been previously
reported. Zobeck et al. (1999) observed a general increase in aerosol PM$_{10}$ production as the clay content in agricultural soils of the Southern High Plains increased. Stetler et al. (1994) and Stetler and Saxton (1995) also have pointed out the higher potential of fine textured soils to generate dust when disturbed or eroded by wind.

Figure 2. Aerosol PM$_{10}$ and PM$_{2.5}$ concentrations as measured by the MiniVol and DataRAMs dust monitors.

Table 1. Means comparison of CaCO$_3$ and clay levels.

<table>
<thead>
<tr>
<th>Level of factor</th>
<th>PM$_{10}$ (MiniVol)</th>
<th>PM$_{10}$ (DataRAM)</th>
<th>PM$_{2.5}$ (DataRAM)</th>
</tr>
</thead>
<tbody>
<tr>
<td>High clay level</td>
<td>39.04 a</td>
<td>25.61 a</td>
<td>6.87 a</td>
</tr>
<tr>
<td>Low clay level</td>
<td>34.24 b</td>
<td>23.35 a</td>
<td>6.56 a</td>
</tr>
<tr>
<td>High CaCO$_3$ level</td>
<td>44.85 a</td>
<td>29.97 a</td>
<td>8.19 a</td>
</tr>
<tr>
<td>Low CaCO$_3$ level</td>
<td>31.01 b</td>
<td>20.12 b</td>
<td>5.42 b</td>
</tr>
</tbody>
</table>

Means with the same letter by columns within clay and CaCO$_3$ levels are not significantly different at $\alpha = 0.05$.

Conclusions

Significant differences were found among measurements of PM$_{10}$ concentration made by DataRAM and MiniVol instruments. DataRAM PM$_{10}$ measurements were on average 66% of measurements made by the MiniVol. Measurements between instruments, however, were proportional with a correlation coefficient of 0.97. PM$_{2.5}$ concentrations obtained with a DataRAM were also proportional to PM$_{10}$ production. The overall average proportion of PM$_{2.5}$ in
relation to PM$_{10}$ as measured by a DataRAM was 27%. Correlation coefficients of 0.92 and 0.87 were found among PM$_{2.5}$ and PM$_{10}$ as measured by a DataRAM and a MiniVol, respectively.

The increase of aerosol PM$_{10}$ and PM$_{2.5}$ concentrations with the increase of soil CaCO$_3$ content was significant for the three datasets. The increase of particulate matter concentrations with the increase of soil clay content was significant for the PM$_{10}$ data set from the MiniVol dust monitor, but was not significant for PM$_{10}$ and PM$_{2.5}$ as measured with DataRAMs. Differences in soil CaCO$_3$ content produced greater differences in aerosol PM$_{10}$ and PM$_{2.5}$ concentrations than the differences in soil clay content.

Disclaimer: Names of commercial products and/or their manufacturers are necessary to describe the equipment, processes and products in this study. Colegio de Postgraduados and USDA-ARS imply no approval of these products to the exclusion of others that may also be suitable.

Acknowledgements: The authors are grateful to Dean Holder, USDA-ARS for assistance with data collection and compilation.

References


Clay and Carbonate Effect on Fine Dust Emissions as Generated in a Wind Tunnel

A. Amante-Orozco, Colegio de Postgraduados, Salinas de Hgo., SLP 78600, MEXICO (E-mail: aamanteo@yahoo.com.mx)

T. M. Zobeck, USDA-ARS, Lubbock, Texas 79415 (E-mail: tzobeck@lbk.ars.usda.gov)

Introduction

A growing concern for health effects and climatic impact of airborne dust has motivated a number of studies focusing on sediment properties and meteorological conditions that influence dust emissions. Determination of the capacity of different soil types to produce PM$_{10}$ emissions and identification of the causes for variations in PM$_{10}$ production among them, is critical to improve the estimation of PM$_{10}$ emission by wind erosion (Hagen et al., 1996). PM$_{10}$ is particulate matter having aerodynamic diameter $\leq 10$ $\mu$m. Wind erosion is the main source of these types of dust emissions in the Southern High Plains of Texas (Gill et al., 1999). In this study, we used a suction-type wind tunnel where wind erosion was reproduced to evaluate the effect of soil clay and calcium carbonate (CaCO$_3$) content on aerosol PM$_{10}$ production from eight agricultural soils of the Southern High Plains near Lubbock, Texas.

Materials and Methods

Eight agricultural soils from the Southern High Plains near Lubbock, Texas, with different combinations of carbonate and clay content were selected from a preliminary study. Two levels (low and high) of soil clay content and two levels (low and high) of soil CaCO$_3$ content were evaluated. Levels of clay content were $< 20\%$ for low and $> 20\%$ for high. For CaCO$_3$ content the low level was $< 3\%$ and the high level was $> 3\%$. Selected soils series were: Amarillo (low CaCO$_3$, low clay); Acuff (two soils) and Olton (low CaCO$_3$, high clay); Gomez (high CaCO$_3$, low clay); and Drake (three soils) (high CaCO$_3$, high clay). Soils clay and carbonate content ranged from 12.7 to 32.6 $\%$ and from 0.2 to 13.0 $\%$, respectively.

A suction-type wind tunnel, 10 m in length and 0.5 m wide by 1 m high cross section, was used to conduct the experiment. The wind profile was developed on the first 7.8 m of the wind tunnel over a fixed roughness, and the test section occupied the next 2.2 m. A fan at the end of the tunnel draws air from the work section of the tunnel and generates the air stream. An array of 10 pitot tubes mounted on a vertical bracket was used to measure velocity gradients in the wind tunnel, just upwind of the test section. Pitot tubes were connected to a scanivalve pressure transducer. A 21X-micrologger program controlled the valve sequence to make the pressure readings, and provided the wind velocities at the different heights. The wind profile parameters were then determined according to the Prandtl-von Karman equation (Nickling, 1994).

A wind profile with a similar aerodynamic roughness length ($Z_o$) to that measured at a highly erodible agricultural field near Lubbock, Texas (Stout and Zobeck, 1996), was developed in the wind tunnel. To develop the desired wind profile in the wind tunnel, it was necessary to manipulate diverse roughness elements. The wind profile friction velocity ($U_*$) was 0.79 m/s.
Before each test, a layer of 2.01 to 19.0 mm soil aggregates was placed in the test section of the wind tunnel. An effective area of 217 cm by 47.6 cm of the wind tunnel floor was totally covered by soil aggregates. Abrader sand (0.65 mm diameter) was released into the wind tunnel during a test, 9 cm above the tunnel floor near the tunnel entrance at a feed rate of 0.277 g/cm/s. Abrader sand had to travel 4.6 m before impacting the soil aggregates in the test section to generate PM$_{10}$ aerosol.

An isokinetic vertical dust sampler was mounted at the end of the test section in the wind tunnel. A blower connected to the top of the sampler (the outlet) produced a flow of air by drawing air from the sampler. Soil being eroded was sampled through a 3 mm wide vertical inlet that extended the entire height of the wind tunnel on the centerline of its width. PM$_{10}$ aerosol was sampled at the center of a 10.16-cm diameter duct that connected the blower to the outlet of the vertical soil sampler. A DataRAM dust monitor was used to measure generated PM$_{10}$ aerosol with an isokinetic sampling probe to sample the dusty air from the center of the pipe. A running time of 10 min was given to measure PM$_{10}$ concentrations. A background of the PM$_{10}$ concentration with the saltating sand alone was taken for each test. Five replications for each soil were made.

**Results and Discussion**

PM$_{10}$ tended to decrease as the soil clay content increased and airborne PM$_{10}$ concentration for soils with low clay content was significantly larger ($P = 0.034$) than that for soils with high clay content (Figure 1). Average PM$_{10}$ concentration for the low and high soil clay content levels were 118.9 and 91.2 $\mu$g/m$^3$, respectively. Conversely, PM$_{10}$ tended to increase as soil CaCO$_3$ content increased and significantly larger ($P < 0.0001$) aerosol PM$_{10}$ concentrations were produced by soils with high CaCO$_3$ content than those with low CaCO$_3$ content (Figure 1). Average PM$_{10}$ concentrations of 69.1 and 127.1 $\mu$g/m$^3$ were observed for the low and high soil CaCO$_3$ content levels, respectively. As can be seen in Figure 1, differences in soil CaCO$_3$ content levels produced greater differences in aerosol PM$_{10}$ concentrations than those produced by differences in soil clay content levels.

In a wind tunnel study, Mirzamostafa (1996) and Mirzamostafa et al. (1998) reported significant differences in the suspension fraction (of the eroded soil) generated by the abrasion of aggregates from soils with different clay content. The suspension fraction ($< 0.106$ mm) decreased as the soil clay content increased up to 20%, but increased when the soil clay content was >20%. On the other hand, the PM$_{10}$ fraction of suspension increased as the soil clay content increased (Mirzamostafa, 1996; Hagen et al., 1996). Gillette (1978) found that fine particle ($< 25$ $\mu$m diameter) fluxes in a wind tunnel were relatively independent of the soil texture. Gillette attributed the lack of effect to the short length (21.7 cm) of the soil bed in the test section of the wind tunnel over which abrasion from saltating sand took place. Fryrear et al. (1994) and Usman (1995) reported that soil aggregates $> 0.84$ mm increased as the soil clay content increased and, therefore, the soil erodibility decreased. Since PM$_{10}$ production is directly related to soil erodibility, wind tunnel results are in agreement with these findings. The increase in aggregate stability with increases in soil clay content has also been observed by a number of researchers (e.g. Skidmore and Layton, 1992; Usman, 1995).
Figure 1. Means comparison of soil clay and CaCO₃ content levels. Means with the same letter within clay and CaCO₃ factors are not significantly different.

Fryrear et al. (1994) and Usman (1995) also observed that the soil aggregation increased as soil CaCO₃ content increased up to about 25%, which in turn decreased the soil erodibility. Fryrear et al. (1994) also mention that, depending on soil clay content and type, soil CaCO₃ content of about 40% or greater tend to increase soil erodibility. However, PM₁₀ concentration increased significantly in this study for soils in the high level of CaCO₃ content, which varied from 3.1 to 13.0%. This result suggests that even if the soil aggregation increases, soil CaCO₃ content of about 3% and more significantly weakens the stability of the soil aggregates. As a consequence, soil aggregates were more easily abraded and the aerosol PM₁₀ increased.

The variation in the ability to produce aerosol PM₁₀ among the tested soils was highly significant (\( P < 0.0001 \)). The Gomez soil (the only soil with low clay content among soils with high CaCO₃ content) produced the largest average aerosol PM₁₀ concentration (174.2 \( \mu g/m^3 \)). On the other side, an Acuff soil (the soil with the highest clay content) produced the lowest average aerosol PM₁₀ concentration (42.6 \( \mu g/m^3 \)).

Conclusions

Soil clay and CaCO₃ content significantly affected airborne PM₁₀ emissions from soil aggregates. Airborne PM₁₀ concentrations increased as soil clay content decreased and soil CaCO₃ content increased, with CaCO₃ having the greatest impact. Consequently, soils from agricultural fields in the Southern High Plains of Texas that would produce the highest PM₁₀ concentrations have more than 3% by weight CaCO₃ content and less then 20% clay. Conversely, the lowest PM₁₀ concentrations were observed in soils with low CaCO₃ content (< 3%) and high clay content (> 20%). PM₁₀ production from abraded soil aggregates was found to be significantly different among soils.
Disclaimer: Names of commercial products and/or their manufacturers are necessary to describe the equipment, processes and products in this study. Colegio de Postgraduados and USDA-ARS imply no approval of these products to the exclusion of others that may also be suitable.

Acknowledgements: The authors are grateful to Dean Holder, USDA-ARS for assistance with data collection and compilation and Texas Tech Univ., Dept. of Civil Engineering for the use of their wind tunnel.

References


Evaluation of Saltation Flux Impact Responders (Safires) for measuring instantaneous aeolian sand transport rates

Andreas C.W. Baas, Department of Geography, University of Southern California, Los Angeles, California 90089-0255 (E-mail: baas@usc.edu)

Abstract

The assessment of aeolian sand transport rates in the field on small temporal and spatial scales is of primary interest for the development and testing of more detailed models of sand movement by wind in general and of interactions between the flow field and the saltation layer in particular. One basic method of assessing detailed transport rates employs a responder that registers and counts the impacts of saltating grains hitting the probe through time. The Sensit and the Saltiphone are two examples of this type of instrument that have been used in various previous studies. This paper reports on laboratory calibrations and preliminary field tests of a new design of piëzo-electric impact responders capable of measuring saltation impacts at a frequency of 20 Hz, called ‘Safires’.

Laboratory calibrations were performed with a vertical gravity flume generating known sand grain fluxes using regular sand and specific individual size-fractions. The fall velocities of grains as a function of their size and fall distance along the vertical tube were calculated using a numerical solution of the Basset-Boussinesq-Oseen equations. This enabled the determination of the momentum flux as a function of fall distance along the tube. Initial tests of 35 Safires involved investigation of three fundamental characteristics: correspondence between digital and analogue signals generated by the instrument, relative response as a function of azimuth angle around the probe (directional response), and linearity in response as a function of mass flux in the fall flume.

Calibrations of the instruments involved two approaches. The first is an investigation of direct correspondence between mass flux and signal response. The second involves determination of the minimum momentum threshold required for the instrument to register a grain impact. Using this lower limit and the known distribution of grain size and speed at different fall depths a prediction can be made as to the sand grain flux the Safire ought to measure, which is then compared with the signal response. The latter approach essentially determines the instrument’s measuring efficiency.

These Safires were also deployed in the field as part of a larger experiment designed to investigate aeolian streamers. During this experiment sand traps were deployed along-side Safires on time scales of 10 minutes and this provided the opportunity to compare the instrument’s recordings with traditional sand trap yields.

The advantages of the Safire are: 1) that they provide high-frequency measurements, 2) that they present a very minimal obstruction to the wind flow (no scour observed in the field), and 3) that they are of a (relatively) low-cost. However, the sensor shares the limitation inherent in all impact responder-type instruments, namely that measurements are based on grain momentum rather than mass alone and that results are therefore difficult to translate to traditional mass fluxes.
Measurement PM$_{2.5}$ Emission Potential from Soil Using the UC Davis Resuspension Test Chamber.

O. F. Carvacho, C.N.L- A.Q.G. One Shield Ave. University of California, Davis, Ca. 95616-8569 (carvacho@crocker.ucdavis.edu)

L. L. Ashbaugh, C.N.L- A.Q.G. One Shields Ave. University of California, Davis, Ca. 95616-8569 (ashbaugh@crocker.ucdavis.edu)

M.S. Brown, C.N.L- A.Q.G. One Shield Ave. University of California, Davis, Ca. 95616-8569 (brown@crocker.ucdavis.edu)

R.G. Flocchini, C.N.L- A.Q.G. One Shields Ave. University of California, Davis, Ca. 95616-8569 (flocchini@crocker.ucdavis.edu)

Introduction

A new National Ambient Air Quality Standard (NAAQS) for particulate matter less than or equal to 2.5 microns in aerodynamic diameter is being considered by the United States Environmental Protection Agency (U.S. EPA). Particulate matter of this size is commonly referred to as PM$_{2.5}$ or “fine” particulate matter.

We have constructed a dust resuspension chamber to identify different soil texture to generate fugitive geological source profiles and to investigate the potential of soil to emit dust in the PM$_{10}$ size range (Carvacho et al., 2002). Using the same protocol we also investigate PM$_{2.5}$ size range. The PM$_{2.5}$ is then separated from the dust cloud using an AIHL-design PM$_{2.5}$ cyclone and collected on Teflon filters for gravimetric and elemental analysis. The dust generated in the chamber can be modeled by a decaying exponential function. The model parameters are related to the inherent PM$_{2.5}$ emission potential of the soil and the energy input necessary to separate the PM$_{2.5}$ from the parent material.

We have optimized the chamber operating parameters to produce results that can be related to underlying soil properties. We have tested the procedure on 44 soils spanning a range of soil textures. The chamber gives consistent results when used with the optimized operating parameters and will describe the potential of geological material to emit PM$_{2.5}$ based on the 44 soils tested. It will also compare these results to the earlier results for PM$_{10}$.

Materials and Methods

Since we are primarily concerned with soil particles that remain suspended in ambient air, only dried soil is used to measure the maximum potential to emit PM$_{2.5}$ Index of the soil. Approximately 1.0 g of sieved soil material with size fraction of 75 to 0 µm is placed in the dust resuspension chamber, which is then sealed. An aluminum tube of 1.0 cm diameter connects the end of the dust suspension chamber to the inside of the dust collection chamber.
A measured volume of air (3.5 lpm for 15 seconds) is forced through the soil sample at the base of the fluidizing bed. This is sufficient to suspend dust particles of ~50 µm in aerodynamic diameter, these particles are carried out of the resuspension chamber and into the collection chamber as shown in Figure 1. The particles are then collected on a 47 mm Teflon filter after passing through an AIHL-design PM$_{2.5}$ cyclone. We sample each 15 seconds “puff” of dust for 15 minutes on a single Teflon membrane filter. We then repeat this procedure using the same sample of the soil until the soil sample is depleted of PM$_{2.5}$ material.

Figure 1. Schematic of the C.N.L resuspension and collection chamber.

For this study, we collected 44 soil samples from agricultural fields, unpaved roads, paved roads, disturbed land areas, construction sites, and equipment staging areas in California’s San Joaquin Valley. These soils spanned a wide range of texture, as shown in Figure 2. Some of the agricultural soils were replicates from different parts of the same field. Generally, the unpaved road sample was collected from agricultural roads adjacent to the field where crop soil sample was collected.

Both the PM$_{10}$ and PM$_{2.5}$ Index are calculated by fitting the cumulative mass CM as a function of time $t$ to the equation $CM = A(1 - e^{-Bt})$ as shown in Figure 3. The time parameter is the cumulative sampling time of soil suspension in the collection chamber for each filter. The parameter $A$ is the asymptote of the decaying exponential curve and represents the PM$_{10}$ or PM$_{2.5}$
Index. This represents the maximum amount of PM$_{10}$ or PM$_{2.5}$ that would be released by repeated “puffs” if disaggregation did not occur.

Figure 2. San Joaquin Valley soil distribution in the soil texture triangle.

Figure 3. Curve fit for PM$_{10}$ or PM$_{2.5}$ Index.

**Results and Discussions**

Figure 4. Show the relationship between PM$_{2.5}$ Index and the standard soil texture parameters sand, silt, and clay. The PM$_{2.5}$ Index is plotted for 0 to 75 µm fraction of dry-sieved soil, recall that the index is the maximum amount of PM$_{2.5}$ dust that is generated from one gram of soil material.
The sand, silt, and clay were measured by wet sieving and gravimetric pipette suspension, this represents the soil particle size distribution for completely disaggregated soils. There is a good correlation between the PM_{2.5} Index with clay and sand fractions.

![Graphs showing the relationship between PM_{2.5} Index and soil texture parameters.](image)

Figure 4. The relationship between the PM_{2.5} Index and soil texture parameters. Also included is the relationship between the PM_{2.5} Index and the PM\textsubscript{10} Index.

Our results show that both the PM\textsubscript{10} and PM\textsubscript{2.5} Index have a better correlation to the soil texture than to the dry silt content. Furthermore, the soil texture is readily available, while the dry silt content is not. For these, we expect the PM\textsubscript{10} and PM\textsubscript{2.5} Indexes to be more useful parameters to use in emission calculations.

**References**

Characterising soil surface susceptibility to wind erosion using bi-directional reflectance: a preliminary assessment.

A. Chappell, School of Environment & Life Sciences, University of Salford, Manchester, M5 4WT, UK. (E-mail: a.chappell@salford.ac.uk)

G. Sheridan, School of Environment & Life Sciences, University of Salford, Manchester, M5 4WT, UK. (E-mail: g.sheridan2@pgr.salford.ac.uk)

Introduction

Recent developments in models for the dust cycle (e.g. Sokolik and Toon, 1996) and for wind erosion (Shao and Leslie, 1997) have highlighted the need for information on the spatial and temporal variation of soil surface conditions. This is because soil surface conditions such as structure, aggregation and roughness control the surface susceptibility to wind erosion (erodibility) and the emission of dust. The model requirements are difficult to meet over large areas because surface characterisation is traditionally conducted using labour-intensive in situ soil measurements or soil samples. Arguably, these field-based approaches are even inadequate at the field scale to account for the dynamic nature and evolution of surface conditions. Consequently, there is a requirement for an approach that can be used to rapidly assess changes in the compositional and structural nature of a soil surface in both time and space. This would improve model predictions of wind erosion over different spatial and temporal scales and would ensure that soil surface characteristics are considered on a continuum (Geeves et al., 2000).

Reflectance of the Earth’s surface is directly related to the fundamental vibrational bands determined by the biophysical and geometric characteristics of a surface in the visible and infrared region. The Bi-directional Reflectance Distribution Function (BRDF) of a particular target represents the reflectance at all possible illumination and sensor view angles. It has the potential to provide consistent estimates of the biophysical and three-dimensional structural properties of a soil surface simultaneously over many spatial scales and rapidly over time. This can be estimated by sampling the spectral reflectance of the target (soil surface) at multiple view angles (MVA) (Barnsley et al., 1997) on the ground or from air or space-borne platforms. The BRDF can be subsequently estimated following integration of the MVA using established physically-based soil reflectance models, (Ciemiewski 1987. Jaquemoud et al., 1992). Soil BRDF models utilise radiative transfer theory and the geometric optics principles of anisotropic surface scatter.

The aim here is to present preliminary results of a larger project into the use of the BRDF to characterise soil surface changes due to wind erosion. The results are of initial experiments to (a) determine the resolution of soil surface composition that can be detected using spectral reflectance data; (b) quantify difference in the structure (roughness) of prepared soil surfaces and (c) characterise soil surface change during controlled wetting and drying conditions.
Methods

Analytical laboratory spectroscopy, using an ASD field spec spectroradiometer, a 1000w halogen lamp, and a calibrated spectralon panel was performed on four dryland soil types from around the world. A goniometer was constructed to allow measurements of multi-angular reflectance (including nadir) of the soil surface including 10 degree increments in the sensor zenith from 0 – 50 degrees in the backward scatter, and 0-50 degrees in the forward scatter, and a user defined viewing azimuth plane was used along the principle plane and at two other angles.

Soil moisture, organic matter content, soil texture or mineralogy, and iron oxide content are identified as the central soil compositional features altering the spectral reflectance of dryland surfaces in the visible and near-infrared wavelength spectrum (Heute and Escadafel, 1991). Soil moisture, organic matter content, iron oxide and particle size were artificially altered in the four soils to provide a range of sub-samples with known variation in the key properties controlling reflectance. Quantitative analysis of the samples provided validation of composition using established soil laboratory methods. Nadir reflectance was measured (350 nm – 2500 nm) in a controlled laboratory setting on the sub-samples from each soil types. Angular spectral measurements were also made on all sub-samples for all soils.

The third objective was achieved by loading the four soils into soil trays with perforated bases that allowed drainage. The trays were subjected to a series of treatment cycles consisting of flooding, through a fine noose spray and drying at 40 degrees centigrade for a minimum of twelve hours. Bi-directional reflectance was recorded after each drying stage to allow investigation into the biophysical and micro-topographical changes in the soil surface. Reflectance, using a probe attachment to the ASD field-spec pro radiometer was also recorded at each stage, readings being taken from 15mm below the surface, to quantify any sub-surface changes in the soil composition as derived from the post spectral reflectance analysis. Digital images were taken following every treatment stage to derive accurate, detailed measurements of the 3D geometric structure of the soil surface using digital soft-copy photogrammetry. This information was used to provide validation for investigation into the ability to detect using the BRDF structural changes induced by the environmental processes.

Analysis

A qualitative analysis of the nadir spectra from controlled experiment was undertaken to investigate the composition of the soil samples and to determine the coarse (primary) information content. This subjective analysis rapidly supported many of the findings of soil reflectance evident in the literature. For example, it enabled the measured spectra to be placed within the context of the findings of Heute and Escadafal (1991): variation in spectral reflectance is explained by (and in order or importance) brightness, organic matter and iron oxide.

A semi-quantitative analysis was conducted in order to investigate the more subtle (secondary) information content. For example, figure 1 shows the first order derivative of soil spectral reflectance. This analysis exaggerates the subtle variation in spectra of six samples of soil samples from south-west Niger. Using the findings from other literature it is possible to identify the central wavebands responding to changes in iron oxide, organic carbon and water molecule bindings following extreme heating of the soil sub-samples. Magnitude changes in moisture content and organic carbon content can be identified simultaneously from a single spectrum. Variation in iron oxide type can be distinguished and it is perhaps possible to detect its
magnitude once the obliteration affects of organic content are removed as in the case of the 8-hour sample (Galvao et al., 1998).

Quantitative analysis was used in an attempt to determine the resolution at which spectral reflectance data can detect individual magnitude changes in the central biophysical soil properties. Principle Component Analysis (PCA) and Canonical Ordination were used to provide and indirect and direct gradient analysis of the selected wavelengths and soil properties. The analysis initiated the construction of calibration tables of absolute changes in soil properties and corresponding changes in reflectance in specific wavebands.

Difference in the magnitude of reflectance as a function of view angle can be used to quantify the magnitude of shadowing at the soil surface. The shadowing in a soil surface is a direct consequence of illumination and the soil surface micro-topography. Figure 2 shows the results of bi-directional reflectance experiments designed to investigate the ability to detect difference in particle size; a simplified simulation of surface roughness in this case. Changes in reflectance with view zenith angle are indicted by variation in forward (negative view angles) and backward (positive view angles) scatter reflectance following standardisation to the nadir view. Polynomials are fitted as a guide only to the pattern of scattered reflectance for each soil surface. In general the forward scatter direction is better able to separate the shadowing of the surfaces. In this direction at the lowest view angle the surfaces separate according to their respective particle sizes. The results suggest that in this experiment we are able to distinguish the roughness of surfaces upto ca. 250 µm (0.25 mm).

Figure 2. Reflectance in the solar principal plane of material sieved into four separate fractions to provide controlled surface roughness simulations.

The resolution at which roughness can be detected using bi-directional reflectance was validated using photogrammetry-derived digital elevation models of the soil surface. These analyses were repeated for angular spectral reflectance collected during the process-based wetting and drying experiments (including crusting, cracking and slaking processes). Cierniewski’s (1987) soil roughness model was used to investigate the ability to retrieve information on surface roughness. This model is essential for quantifying roughness under less controlled (field) situations.

Conclusion

Properties investigated in these experiments play a direct role in the susceptibility of a soil surface to wind erosion and collectively determine erodibility of an area under investigation at any one time. Preliminary results of this study provide the first indication that these properties may be replaced by spectral reflectance information. Hence there appears considerable potential for this information to be gathered frequently over many scales simultaneously. Continued process-based investigations using the calibrations determined in this study and validation in field settings will enable application of this method directly to the study of wind erosion processes.

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Use of Landsat Thematic Mapper multi-temporal images to detect and study vegetation dynamics in the Mojave Desert: Applications to studies of dust emission

Pat S. Chavez, Jr., Miguel G. Velasco, David J. Mackinnon, Kathryn A. Thomas, and Robert Fulton; United States Geological Survey and California State University Desert Studies Center

Abstract
Remotely sensed satellite, airborne, and ground-based digital images are being used to investigate landscape vulnerability to wind erosion in the Mojave Desert, with particular interest in the detection and mapping of surface change in the Mojave Desert. Land surface parameters critical in determining the vulnerability of the landscape to wind erosion include vegetation type (annual or perennial) and cover, topography, and surface soil properties. Of these the amount of vegetation cover is the most dynamic over relatively short periods of time due to its sensitivity to seasonal and yearly rainfall. Digital change-image maps generated using satellite images collected under different vegetated surface conditions in the Mojave Desert are being used to 1) extract information directly related to the presence of fine-grained sediment (fines) available to wind erosion and 2) map the amount of sheltering of the surface provided by both annual and perennial vegetation. The temporal characteristics of the image data used as input for change detection and analysis are critical, with selection dependent on the main objective of a given study (i.e., short-term seasonal or long-term multi-year change detection).

In our study, Landsat Thematic Mapper images collected during very contrasting precipitation conditions (April 1992/wet El Nino year and June 1997/dry year, as well as spring 2000/dry and spring 2001/wet) are being analyzed to detect and map vegetation dynamics in the Mojave National Preserve. Resulting digital change images are used to help identify and map dust source areas, as well as extract information related to vegetation types and amounts. By analyzing vegetation change affected by both seasonal and highly contrasting climate/rainfall conditions, we extract information that is mostly related to annual vegetation. Dust sources composed of sand intermixed with clay (loamy sand) or clay intermixed with sand (sandy loam) are ideal soils for the growth of annuals, that is controlled primarily by annual rainfall. Using change detection procedures we are detecting and mapping information related to changes in the spatial distribution and cover of annual, as well as some perennial, vegetation in the Mojave National Preserve, and thereby, changes in availability of dust. A direct result of this research is the development of tools and capabilities to characterize surface features and detect surface changes, including the capability to evaluate and monitor the vulnerability of land and vegetation to environmental and climate-induced changes on a regional scale, especially with respect to potential dust emission.
Use of satellite and ground-based images to monitor
dust storms and map landscape vulnerability to wind
erosion

Pat S. Chavez, Jr., David J. Mackinnon, Richard L. Reynolds, and Miguel G. Velasco;
U. S. Geological Survey

Abstract

Wind-induced dust emission in the Southwestern United States is important regionally because of
its impact on human health and safety and its influence on ecosystem dynamics. Wind velocity,
sediment availability, and surface conditions are important factors that determine landscape
vulnerability to wind erosion. We are investigating remotely sensed satellite, airborne, and
ground-based image data to detect and monitor active dust storms, as well as to map areas
vulnerable to wind erosion in the Mojave Desert of the Southwest United States. Data collected
by various satellite imaging systems, a ground-based digital camera station, and several in-the-
field instruments during several dust storms are being used to correlate landscape characteristics
to wind erosion vulnerability. Multispectral satellite images were used to generate a wind
erosion vulnerability image that represents a first-order vulnerability image map, with field data
and observations being used to validate the results. The GOES satellite imaging system is the
only one available that has the required temporal resolution to detect and monitor active dust
storms; however, its spatial and spectral resolutions are low, and only very large dust storms can
be detected. A satellite imaging system with three to five spectral bands and approximately 100
m spatial and 15 to 30 minutes temporal resolutions is needed to effectively monitor short-lived
events.
Mapping surface armoring of arid and semi-arid environments using remotely sensed images

Pat S. Chavez, Jr., David J. Mackinnon, and Miguel G. Velasco; USGS

Landsat Thematic Mapper (TM) image data were used to develop an algorithm to generate a digital image map that represents the amount of protective covering (armoring) that exist over soils in arid and semi-arid environments.
WITSEGG Sampler:  
A Segmented Sand Sampler for Wind Tunnel Test

Zhibao Dong, Laboratory of Blown Sand Physics and Desert Environments, Cold & Arid Regions Environmental and Engineering Research Institute, Chinese Academy of Sciences, No.260, West Donggang Road, Lanzhou, Gansu Province 730000, People’s Republic of China (E-mail:zbdong@ns.lzb.ac.cn)

Hongyi Sun, Laboratory of Blown Sand Physics and Desert Environments, Cold & Arid Regions Environmental and Engineering Research Institute, Chinese Academy of Sciences, No.260, West Donggang Road, Lanzhou, Gansu Province 730000, People’s Republic of China (E-mail:zbdong@ns.lzb.ac.cn)

Introduction

Seeking reliable sand traps or samplers for direct sediment transport measurement have been of continuing significance for decades and many types of sand samplers have been made for different purposes (Shao et al., 1993; Nickling and McKenna Neuman, 1997; Goosens and Offer, 2000; Goosens et al., 2000). Here we report a newly designed segmented sampler for wind tunnel test (WITSEGG sampler).

Design of the sampler

The WITSEGG sampler is of vertically integrating, passive type that follows earlier design by Bagnold (1941). So it is a modified Bagnold sampler. WITSEGG sampler is designed to measure the flux profile of a blowing sand cloud in the sand wind tunnel of the Laboratory of the Blown Sand Physics and Desert Environments, Cold and Arid Regions Environmental and Engineering Research Institute, the Chinese Academy of Sciences. The cross-sectional area of the wind tunnel is 1.2m × 1.2m.

The sampler is constructed of 0.5mm stainless steel and has four main components (Fig.1): A removable side cover, a wedge-shaped leading edge, a support and 60 sand chambers. Fig.2 shows the technical scheme. The gross height of the sampler is 700mm and the gross width 160mm. The gross thickness of the sampler is designed to be 25mm so that it has no significant interference with the airflow while maintaining enough sampling volume of the sand chambers. The wedge-shaped leading edge has 60 nozzle orifices connecting to 60 sand chambers. Each orifice is 10mm high and 5mm wide. The wedge-shaped leading edge is chosen to reduce the interference of the sampler with the airflow at the inlet on one hand and avoid the too short sampling time of the sand chambers by reducing the sampling width of the orifices. The spacers between the orifices are milled very sharp to reduce the particle rebound on them and errors in the measurement of total flux. The leading edge and a side cover are removable so that the sand chambers can be removed and the collected sand weighed. The size of the sand chambers is 140mm × 15mm × 6mm. Each chamber has a full sand capacity of about 18g and is inclined 30° with respect to the horizontal.

A key design in the leading edge of the sampler is the vent. The function of the vent is to ensure the isokinnetic sampling. To avoid the problem of stagnation and the related errors in measuring the sediment flux, each chamber of the WITSEG sampler is vented. There are vent holes with diameter of 2mm on both sides of the orifice, which are open to a sand chamber and connected to the vertical common vent. The airflow goes out of the vent and sand particles in the sand-laden wind entering an inlet fall into the sand chamber by gravity. To prevent the blown sand particles from going out with the wind the vent holes are covered with the same fine stainless steel wire mesh (200 mesh, 62.5 μm openings, 60% porosity) as that used in Nickling and McKenna Neuman’s (1997) wedge-shaped sand traps.

Wind tunnel evaluation of the sampling efficiency
The efficiency of the sampler was evaluated in the sand wind tunnel of the Laboratory of Blown Sand Physics and Desert Environments, Cold and Arid Regions Environmental and Engineering Research Institute, Chinese Academy of Sciences. The sand used for the evaluation test was typical dune sand in the Shapotou area, southeast of Tengger Desert of China.

**Figure 2.** Technical scheme of the WITSEG sampler (A. General outline, B. Sand chamber)

The efficiency is defined as the ratio of the collected total flux by the sampler to that considered being a correct value of the total flux rate:
$E = \frac{Q_s}{Q_c}$  \hspace{1em} (1)

where, $E$ is the sampling efficiency, $Q_s$ is the collected total flux by the sampler, $Q_c$ is the correct value of the total sediment flux. So the key to obtaining the sampling efficiency is to define $Q_s$ and $Q_c$. In our calibration, $Q_c$ was obtained by:

$$Q_c = \frac{(W_o - W_e)}{(L \times T)} \hspace{1em} (2)$$

where, $W_o$ is the total weight of the sand sample before test, $W_e$ is the total weight of the sand sample after test, $L$ is the width of the sand tray, in cm, and $T$ is the sample time, in second.

The sampling efficiency at different wind speed is obtained by the data of $Q_s$ and $Q_c$. The sampling efficiency ranges from 0.87 to 0.96, with an average of 0.91.

**Conclusions**

A segmented aeolian sampler for studying the flux profile of a blowing sand cloud in a wind tunnel (WITSEG sampler) has been developed. The sampler can measure the sediment flux at 60 heights with one-centimeter intervals and provide the detailed data for establishing the function of blown sand flux profile.

The WITSEG sampler is passive type but each sand chamber is well vented. It has been evaluated in a wind tunnel using typical aeolian sand. The overall efficiency is 0.91.

**References**


A 200-m Tall Instrumented Tower for Atmospheric Measurements in Wind Events

Thomas E. Gill, Department of Civil Engineering and Department of Geosciences, Texas Tech University, P.O. Box 42101, Lubbock, TX 79409-2101: (email Tom.Gill@ttu.edu)

Douglas A. Smith, Department of Civil Engineering, Texas Tech University, P.O. Box 41023, Lubbock, TX 79409-1023: (email Doug.Smith@wind.ttu.edu)

Russell R. Carter, Department of Civil Engineering, Texas Tech University, P.O. Box 41023, Lubbock, TX 79409-1023: (email Russell.Carter@wind.ttu.edu)

Richard E. Peterson, Department of Geosciences, Texas Tech University, P.O. Box 41053, Lubbock, TX 79409-1053 (email Richard.Peterson@ttu.edu)

DESCRIPTION

A 200-m tall instrumented tower dedicated to atmospheric measurements has been constructed at Texas Tech University’s Wind Science and Engineering Research Center field facilities at Reese Technology Center (formerly Reese Air Force Base), in Lubbock County, Texas, approximately 15 km west of the city of Lubbock in the flat, aeolian-active landscape of the Southern High Plains (Figure 1). The tower includes instrumented boom arms (Figures 2, 3) at 10 levels to continuously collect high-resolution full-scale data on meteorological variables: $u$, $v$, $w$ winds; pressure; relative humidity; temperature; and aerosol concentrations. Aerosols in the PM$_{10}$ range are sensed via a nephelometer-type laser scattering device specially designed for operation during dust storms (Stopenhagen and Pottberg, 2000). Particulate concentrations are typically recorded as 1-minute averages; the other quantities are sampled and reported at tens of Hertz. Data collected on the tower is transmitted to a computer in a central building on the ground via a fiber optic system. The facility is expected to be operational in 2002. Schematic layouts of the boom arm and tower are given in figures 3 and 4.

Figure 1. 200-meter tower (slender tower in foreground) at Reese site.

Figure 2. Close-up of tower section showing boom arm prior to deployment of instrumentation.
POTENTIAL APPLICATIONS TO AEOLIAN RESEARCH

Until now, vertical profiles and fluxes of aeolian dust emissions have only been measured at heights up to a few meters above the surface (e.g., Fryrear and Saleh, 1993). Standard meteorological towers are rarely greater than 10 meters tall; above this height, aerosols are characterized by sampling from buildings or other tall structures (usually in urban areas), tethered balloons or aircraft sampling, or remotely sensed by lidar. Other tall towers dedicated to meteorological research (Boulder, Colorado; Beijing, China; Cabauw, Netherlands) either are no longer in operation (Boulder) or not routinely used or designed for aerosol measurements.

The primary application of the tall tower at Reese Technology Center for research in aeolian processes, wind erosion and airborne dust will be to document concentrations, profiles and fluxes of particulate matter on a vertical scale of hundreds of meters. The Southern High Plains of Texas is an active mineral-aerosol-producing region, with blowing dust having been regularly reported in the vicinity of the present tower many days each year (Wigner and Peterson, 1987). The site’s location at Reese Technology Center is > 10 km away from the urban influence of the city of Lubbock, and surrounded in all directions by agricultural fields vulnerable to wind erosion. However, land surfaces within the former Air Force base immediately surrounding the tower are not dust-emissive; they are typically paved (runways or roads) or vegetated (landscaped or prairie), with no significant local aerosol sources or other tall structures nearby, making it an ideal receptor site for particulate matter measurements.

The ten instrumented boom arms on the 200-m tall Lubbock tower will allow for simultaneous measurement of aerosol concentration along a vertical scale extending upward from the surface several orders of magnitude larger than past studies of aeolian emission and transport. The collocated meteorological instruments and fast-response aerosol sensors will allow for correlation of short-term aerosol concentrations with fluctuations in boundary-layer temperature, pressure, moisture, and wind fields. On longer temporal scales, dust concentrations and their variations with height could be correlated with different meteorological patterns and dust-storm forcing mechanisms characteristic of the Southern High Plains (Wigner and Peterson, 1987) and seasonal to interannual variations in land use, soil, and climatic factors.

There are many other potential applications of this facility and its data (which will be made available in the future to qualified investigators) for research in lower atmospheric boundary layer characterization as well as aeolian processes.

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Figure 3. Instrumentation layout along the tower boom arms. Dimensions of instruments on diagram not necessarily to scale.
Figure 4. Schematic of tower levels. All heights given are for the anemometer level. The boom arm is .71m below the given level (see below).
Temperature Sensitivity of a Piezo-Electric Sensor used for Wind Erosion Measurements

S. K. Heidenreich, DLWC, Gunnedah, NSW, Australia (E-mail: sheidenreich@dlwc.nsw.gov.au)
J. F. Leys, DLWC, Gunnedah, NSW, Australia (E-mail: jleys@dlwc.nsw.gov.au)
G. H. McTainsh, Griffith University, Brisbane, QLD, Australia (G.McTainsh@mailbox.gu.edu.au)
F. J. Larney, Agriculture and Agri-Food, Lethbridge, Alberta, Canada (larney@EM.AGR.CA)

Introduction

The use of piezo-electric elements for the determination of moving particles is well documented in wind erosion literature (Arens 1996, Fryrear et al. 1991, Gillette et al. 1996, Gillette and Stockton 1986, Larney et al. 1995, Stout 1997). As part of another study (Heidenreich et al., 1999) we have calibrated an instrument, based on a piezo-electric quartz crystal (SENSIT™) with data derived from instrumentation with a different underlying measuring principle (passive dust catcher) (Fryrear 1986). SENSIT™ measurements are expressed as particle flux (impacts per square meter per second) and kinetic energy count flux (erg per square meter per second. The kinetic energy output displays a particle impact independent background noise. To minimize measurement errors of the kinetic energy, we first need to subtract the background kinetic energy.

Methods

Method 1: Average constant

Assuming a constant background noise level a numeric method was chosen to determine the value. The constant was calculated by averaging the kinetic energy (KE) per logging interval [LI] for particle count \( n(t) = 0 \) for each sampling period \( t=0 \) to \( t=T \).

\[
\text{Equation 1: } KE_{\text{Background}} = \frac{\sum_{t=0}^{T} KE_{t(n(t)=0)}}{\sum_{t=0}^{T} LI_{n(t)=0}}
\]

We dismissed this assumption of a constant background noise because of the large standard deviation about the mean.

Method 2: Temperature dependent background noise determination

One physical feature of piezo-electric cells is their sensitivity towards changes in temperature. The cells are a lead zirconate titanate composition [Pb(Zr,Ti)O₃] which have a pyroelectric sensitivity of \( k_q=4.2 \times 10^{-4} \text{ K}^{-1} \text{ m}^{-1} \). We analyzed approximately 10,000 logging intervals each one minute long and without impacting particles. The correlation coefficient \( \rho = 0.118 \) of the kinetic energy as a function of temperature change in 10 minutes indicated that the velocity of the temperature changes (dT/dt) is too slow to have a significant impact on the signal (Figure 1). We found a strong correlation between kinetic energy background noise and ambient air temperature (Figure 2). The correlation factor between ambient temperature and kinetic
background energy was $\rho = 0.85$. Other factors suspected to have an influence on the background kinetic energy, such as wind speed and barometric pressure, had no correlation.

A visual inspection of the time series data indicates a time lag between the temperature and the reaction of the kinetic energy signal. This can be explained with the experimental set up of the SENSIT™. The instrument was buried underground with the sensing ring 4-5 cm above ground. The insulating effects of the soil would explain the time difference between changes in ambient air temperature and instrument temperature. To find the appropriate time lag to apply to the correction factor we repeated the correlation analysis whilst shifting the Kinetic Energy data in 20 minute intervals from 0 to 240 minutes (4 hours).

We found the optimum correlation factor $r = 0.89$ at $T=112$ min (Figure 3).

In order to keep the required adjustment of the background kinetic energy level simple, we chose a linear fit of the form $y = a+bx$. The result for the examined instrument is

$$KE_{Background}(t) = 10.2 + 0.493 * T_{(t-112 \text{ min})}$$

**Discussion**

The examined piezo electric instrument clearly has a dependency of its kinetic energy output channel on the ambient air temperature, which in turn influences the soil / instrument
temperature. The maximum correlation between kinetic energy and temperature for determining
the background kinetic energy signal was found at a time lag of $t=112$ min.
Possible causes of this temperature interaction could be either the ceramic cell itself or the signal
conditioning electronics (i.e. a temperature sensitive amplifier transistor). The data logging
system is unlikely to be the cause of the temperature dependency because the signal transfer is
digital and therefore rather secure in its accuracy. Further experiments are required to determine
which of the above mentioned possibilities causes the oscillation. This could, for example, be
done by cooling the instruments body, and thereby the electronic components, to a constant
temperature and then changing the ceramic cells temperature.

**Conclusions**

Care should be taken when utilizing the kinetic energy channel of the SENSIT™ to calculate
particle mass flux because the accuracy when determining the background noise level will govern
the accuracy of the calculated flux. This applies mainly to small events where the noise to signal
ratio will remain large compared to times of high saltation activity. Field experiments undertaken
with the described instrument are often event based and therefore do not necessarily include two
hours before the event. This should be taken into account when setting up recording parameters /
intervals in future experiments.

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Simulated Atmospheric Vortex Threshold

James D. Iversen, Professor Emeritus, Iowa State University, 4105 Stone Brooke Road, Ames IA 50010 <iversen@iastate.edu>

A number of people have studied experimentally the threshold of small particles in a boundary layer wind tunnel (see, e.g., Bagnold 1941, Chepil 1945, Zingg 1953,). The writer has also studied boundary layer threshold including the effects of small particle cohesion (Iversen et al 1976, Iversen & White 1982), fluid density (Iversen et al 1987), and surface slope (Iversen & Rasmussen 1994, 1999). Much less work has been done to date on the threshold of particles being lifted from the surface by an atmospheric vortex such as a “dust devil”. Recent data returned from probes to the planet Mars have stirred new interest in the dust devil phenomenon. The writer first performed vortex threshold experiments in 1975 in a “tornado simulator” designed by C.T. Hsu (Hsu & Fattahi 1976, threshold experiments reported in Greeley et al 1981).

The mechanism of entrainment of sand and dust particles into a dust devil seems to be somewhat different from the normal boundary layer entrainment. The surface shear stress most likely reaches a maximum directly under the vortex radius at which the tangential wind speed is a maximum. In addition to the shear stress acting to lift particles from the surface, however, there is also a decrease in pressure from the ambient at large radius towards the vortex center. This pressure difference may also be effective in lifting small particles from the surface. The minimum pressure point, however, is not directly under the maximum tangential wind speed but at the center of the vortex.

Since the dust devil moves across the surface over which it has formed, the flow at a given point on the surface is unsteady. Thus as the vortex moves over a given point, the pressure on the surface at that point decreases with time. If the decrease is sufficiently fast, it is conceivable that the pressure difference, as well as the surface shear stress, can act to lift particles from the surface. Stresses acting to oppose particle motion include the particle layer weight per unit area and a cohesive stress due to interparticle forces. Thus an equation of equilibrium at threshold can be written by equating effective stresses due to pressure difference and shear stress to the sum of particle layer weight and cohesive stress.

A recent improved version of C.T. Hsu’s vortex generator has been constructed at Arizona State University. New threshold experiments have been conducted with this device at ASU and in an altitude chamber at the NASA Ames Research Laboratory in California (Balme et al 2002). The results of these experiments and the earlier experiments at ISU show that the vortex strength required for threshold is a function of particle layer weight, as expected, but there are indications that the effects of vortex Reynolds number and cohesive stress are also important, and the resultant threshold figures look quite different from those for boundary layer threshold.

References


Simulating the Martian Environment

K.M.Kinch, (e-mail: kjartan@ifa.au.dk), J.Jensen, (e-mail: jjensen@ifa.au.dk)
J.P.Merrison, (e-mail: merrison@ifa.au.dk), R.Mugford, M.N.Zadeh, A.B.Christensen,
D.M.Grant, T.L.Jacobsen
All are: Institute of Physics and Astronomy, Aarhus University, Building 520, Ny Munkegade, DK-8000C, Denmark
K. R. Rasmussen, Department of Earth Sciences, Aarhus University, Building 520, Ny Munkegade, DK-8000C, Denmark (e-mail: geolkrr@geo.aau.dk)

Introduction

The Mars Simulation Laboratory at Aarhus University, Denmark is a newly established facility for laboratory simulation of the Martian environment. The main instrument is a recirculating wind tunnel, which runs at Martian surface temperature and pressure. Small particles of a Mars analogue soil are added to the wind stream. Several ongoing experiments use this facility for studying biological, chemical and aerodynamical aspects of the Martian environment. One example is a recently conducted study on triboelectric charging of dust suspended in the atmosphere, which showed some interesting preliminary results.

Design of The Wind Tunnel

The Mars simulation wind tunnel at Aarhus University, is a re-circulating wind tunnel enclosed in a vacuum chamber. The vacuum chamber is cylindrical with a length of ~ 3 m and an inner diameter of 75cm. Inside the tank, the wind tunnel itself is a cylinder measuring 1.5 m in length and with a diameter of 40cm. The actual experiment is situated inside the wind tunnel. A fan placed at one end of the tunnel sets up a circulating wind. Gas runs through the tunnel, passes the fan, and returns on the outside of the tunnel in the space between the outer wall of the tunnel and the inner wall of the vacuum tank. Wind speeds may be varied from 0-20 m/s, which covers the most likely wind speeds on Mars. Good stability at very low wind speeds is an especially nice
Figure 1. Design of the wind tunnel.

feature of the instrument. The chamber can be evacuated to $3 \times 10^{-2}$ mbar, and then filled with any desired gas, normally CO$_2$. It is usually operated around 5-10 mbar. Tubes in the outer part of the tank allow cooling with liquid nitrogen down to a minimal temperature of ~ -120°C. Dust is injected in the outer tank close to the fan and carried with the gas into the actual tunnel. It may be blown in by opening a valve, letting the pressure from outside push dust into the tank, this has the advantage that a lot of dust can be injected in a short time, but lets air into the tank, which increases the pressure and disturbs the composition of the gas in the tunnel. Alternatively dust may be injected from a vibrating container with a sieve in the bottom. The dust used is a Mars analogue soil from Salten Skov, in central Jutland, Denmark. This analogue is believed to be close to Martian dust in terms of optical and magnetic properties. The soil is sieved to size <63 µm and is dried at 110°C before applying.

Wind speed and dust density is monitored with a laser Doppler anemometer (LDA). Here two coherent laser beams intersect to make an interference pattern. As a dust grain passes the stationary interference pattern light reflects off the grain in a series of pulses. The time between each pulse determines the speed of the grain (which is taken as a measure of wind speed). Dust density is estimated from the number of grains passing in a given time.

Triboelectric charging of dust

Charging and discharging of airborne dust on Mars has attracted some attention\textsuperscript{2,3} as a possible hazard for a Mars lander mission, manned or unmanned. Dust deposition on solar panels and camera lenses may degrade these instruments, while charged dust may interfere with electronics and communication. As an example of recent activity in the wind tunnel we describe a preliminary experiment on the effect of charging on dust deposition.

The experimental set up consisted of a glass plate 30 x 16.5 x 0.2 cm which was placed on top of two pieces of conducting foil 7x14 cm, these were separated by ~ 2cm. The one was grounded, as is the wind tunnel, and the other electrode was held at 200 volts. The plate was placed in the wind tunnel, holding it at an angle of 45° facing in the direction of the wind, so that the foils were on the leeward side. The wind tunnel was run at a wind speed of 3 m/s for a period equivalent to
20 sols (Martian days) in terms of the amount of dust passing the experiment (the equivalent of 20 sols is reached in about an hour of actual time). After deposition the difference in deposition between the two electrodes was clearly visible to the eye. The absorbance at 532nm of the glass plate with deposited dust was measured at a series of points. The resulting curve is shown in Figure 2. There is a clear effect as the absorbance jumps from 0.35 to 0.5 at the edge of the charged plate.

A later series of measurements at a wind speed of 4.5 m/s investigated the dust deposition at different voltages from –200V to 200V. For all negative voltages the charged plate showed greater deposition than the grounded plate, this was also true for positive voltages above ~75V. Surprisingly however at positive voltages less than 75V the charged plate showed less deposition than the grounded one. A tentative explanation of this phenomenon suggests that there is a difference in the mean charge, size and/or number density between negatively and positively charged dust grains. If the negative grains react easier to an applied field, then the first effect of applying a positive voltage would be to repel negative grains, resulting in less dust deposition. Only at larger voltages would positive grains be attracted in large numbers.

**Figure 2.** Preliminary measurement of dust deposition on charged and grounded areas. The grounded area begins at 12mm and ends at 80mm, while the charged area begins at 97mm and ends at 165mm.
Perspectives

While we stress that what is presented above are preliminary results, it would seem that this technique can be used to quantify the charge state of suspended dust and there is an interesting effect, which merits further investigation. Several other studies are planned or ongoing at the wind tunnel facility among them an investigation into magnetic attraction of airborne dust, investigation of oxidation processes under Martian conditions and investigation into characteristics of aerodynamics on Mars.

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Relative Dust Emission Estimated From A Mini-Wind Tunnel

J. F. Leys, Centre for Natural Resources, Dept Land Water Conservation, Gunnedah, NSW 2380, Australia (Email: jleys@dlwc.nsw.gov.au)

C. Strong, Australian School of Environmental Studies, Griffith University, Brisbane Queensland, 4111, Australia (E-mail: craig.strong@mailbox.gu.edu.au)

G. H. McTainsh, Australian School of Environmental Studies, Griffith University, Brisbane Queensland, 4111, Australia (E-mail: g.mctainsh@mailbox.gu.edu.au)

S. Heidenreich, Centre for Natural Resources, Dept Land Water Conservation, Gunnedah, NSW 2380, Australia (Email: sheidenreich@dlwc.nsw.gov.au)

O. Pitts, Sinclair, Knight, Merz, 263 Adelaide Terrace, Perth, WA, 6001, Australia (Email: OPitts@skm.com.au)

P. French, Sinclair, Knight, Merz, 263 Adelaide Terrace, Perth, WA, 6001, Australia (Email: PFrench@skm.com.au)

Introduction

Measurement and modelling of fugitive dust emissions has become an increasing area of interest because of the effects of fugitive dust on visibility, air quality and the potential effect on human health (Pietersma et al. 1996). Direct measurements of dust emission can be achieved in various ways including using high volume air samplers down wind of a source or at source. Alternatively, wind tunnels have been used to measure the emission rate from agricultural fields and coal stockpiles. The portable wind tunnels used in most studies are similar in design to a number of “big” wind tunnels used in aeolian research. These big wind tunnels (BWT) have duct cross-sections in the order of 1m x 1m and working section lengths from 4 to 10m. BWT can have fully developed turbulent boundary layer from which measurements of surface roughness ($z_0$) and friction velocity ($u^*$) can be calculated using Pitot tubes (Raupach and Leys 1990). In comparison to these BWT, there are a few smaller tunnels. The most widely reported being that of Gillette (1978), which has a 0.15m x 0.15m cross-section and a 3m length and is reported to have a turbulent boundary layer.

Big wind tunnels have the disadvantage of being large and difficult to transport and generally require at least two people to operate. To overcome these issues, a mini wind tunnel (MWT) with 0.1m x 0.5m cross-section and a 1m working section was developed and used to estimate dust emissions (Zegelin et al. 1997). The flow in the MWT does not have a fully developed turbulent boundary layer, and as such, the wind profile can not be used to measure $z_0$ or $u^*$ and subsequently calculate the equivalent wind velocity at 10 m height ($u_{10}$). To over come this, the ratio of dust flux in the MWT was compared to that in the DLWC big wind tunnel (Raupach and Leys 1990) at a range of wind speeds and where $z_0$, $u^*$ and $u_{10}$ were measured.
This paper presents estimates of relative dust emission for a range of iron ores and road surfaces using a mini-wind tunnel.

Materials and Methods

The design of the MWT is shown in Figure 1. The saltation introduction system (saltation silo) was not used in this study. The details of the DLWC BWT are reported in Raupach et al. (1990) and the sampling methods similar to the MWT and detailed in Leys et al. (1996).

![Figure 1. Schematic of mini wind tunnel (MWT)](image)

In the MWT, five ores and three road surfaces were exposed to three to five wind speeds (each replicated three times for one minute), the resultant wind velocities were measured at three heights (15, 25, 35 mm) with Pitots then averaged (u, m/s) and erosion rates were measured with an integrating trap that sampled 10% of the tunnel air flow (E, g/m²/s). Eroded sediment was captured on glass fibre filter papers (0.1μm pore size) in a sediment filter box. The sediment trap was quasi-isokinetic; ie the average speed of the tunnel was matched to the average speed of the inlet of the sediment trap. The particle-size distribution (PSD) was determined for the eroded sediment with a Coulter Multisizer using the methods of McTainsh et al. (1997). ORE C, was tested in the BWT and MWT. Erosion rates in both tunnels were calculated using equation 1.

\[
E = \frac{m}{x(yT)}
\]

Where \( E = \) erosion rate [g/m²/s], \( m = \) mass collected in trap, \( x = \) upwind fetch of 1 m for MWT and 4.2 m for BWT, \( y = \) trap width of 0.01 m for MWT and 0.005 m for BWT, and \( T = \) time 60 s

Dust fluxes were calculated using equation 2.

\[
DF = E \times CFD
\]

Where \( DF = \) dust flux [g/m²/s] and \( CFD = \) critical fraction of dust.
The CFD is the fraction of the eroded sediment that can be held in suspension by the wind for a particular \( u^* \). This fraction changes with \( u^* \), sediment density and air density and was calculated using the subroutines within the Wind Erosion Assessment Model (Shao et al. 1996) to determine the critical size of the suspension material and ranged from 30 and 37 \( \mu m \).

**Results and Discussion**

The particle-size analysis of the eroded sediments indicated that the CFD ranged from 1.11 – 0.35 depending on wind speed and ore / surface. A comparison of the dust flux from the same ore (ORE C) is given in Figure 2.

![Figure 2. Dust flux (DF) from MWT and BWT for a range of wind speeds](image)

The ratio of dust flux between the BWT and the MWT for the ORE C is described in equation 3.

\[
DF_{BWT} = DF_{MWT} \times 0.516738 - \frac{135740}{(u^*)^2}
\]  

(3)

Where \( DF_{BWT} \) = dust flux equivalent [g/m²/s] in DLWC wind tunnel, \( DF_{MWT} \) = dust flux [g/m²/s] in mini wind tunnel, and \( u = \) wind speed. The MWT overestimates the DF compared to the BWT.

It was then possible to calculate an estimate of the dust flux from a wind speed measured at 10 m height with the following assumptions:

- That the BWT / MWT DF relationship in Figure 2 established for ORE C can be used for all sites and ores. This is a fair assumption for this study as all the ores and surfaces were levelled before testing.
- That the wind speed correction from 10m height to wind tunnel free-stream is valid for all surfaces although only one correction factor been established for ORE C
- That the equations are limited to the wind speed range \((u_{10})\) of 8 to 20 m/s.
- That the results are for first minute of the specified wind speed, after which the DF would expect to decline, especially if the source of erodible material was limited.

Accepting these assumptions, estimates of the BWT equivalent dust flux \((DF_{BWT})\) can be made from the MWT results for a wind measured at 10m height by applying the following equations.
Site | Equation
---|---
Road - bulldust | $DF_{BWT} = 0.000306 u_{10}^3 - 0.410156$
Ore A fines | $DF_{BWT} = 0.000058 u_{10}^3 - 0.145065$
Ore B Fines | $DF_{BWT} = 0.000025 u_{10}^3 - 0.029948$
Ore B sub fines | $DF_{BWT} = 0.000503 u_{10}^3 - 0.757249$
Ore C | $DF_{BWT} = 0.000027 u_{10}^3 - 0.042815$
Ore C sub fines | $DF_{BWT} = 0.001470 u_{10}^3 - 2.602224$
Road - deposition material | $DF_{BWT} = 0.003367 u_{10}^3 - 4.462010$
Road - gravel | $DF_{BWT} = 0.000464 u_{10}^3 - 0.759889$

Where: $DF_{BWT} =$ dust flux equivalent $[\text{g/m}^2/\text{s}]$ in DLWC wind tunnel, and $u_{10} =$ wind speed at 10m height

### Conclusions

The use of a mini-wind tunnel to characterise the erosion rate of a surface in conjunction with particle-size analysis of the eroded sediment can be successfully used to determine the relative dust emission for a range of wind speeds and surfaces. By undertaking similar work with the large DLWC wind tunnel and deriving a ratio of dust emission between the two tunnels, it is possible to calculate indicative dust emissions for a range of ores and surfaces.

Dust emissions vary for the range of iron ores and road surfaces. Deposition material from conveyors that falls on roads (Road - deposition material) is extremely dusty and easily mobilised ($DF_{BWT} = 15.18 \text{ g/m}^2/\text{s}$ at a wind speed of 18 m/s when measured at 10 m height). The Ore C sub fines (ie Ore C with no fraction greater than 1 mm) is also very dusty ($5.97 \text{ g/m}^2/\text{s}$ at a wind speed of 18 m/s). The Ore B sub fines, Road - gravel, the Road - bulldust are moderately dusty ($2.18$ to $1.37 \text{ g/m}^2/\text{s}$), with the remainder of the ores being less than $1.9 \text{ g/m}^2/\text{s}$. These dust emission rates would not be expected to persist for long periods because the sediment supply diminishes with time.

### References

Measured and modeled roughness heights \((Z_0)\) over diverse roughness elements, central Mojave Desert, California, USA

D.J. MacKinnon, U.S. Geological Survey, Flagstaff (dmackinnon@usgs.gov)

G. D. Clow, U.S. Geological Survey, Denver (clow@usgs.gov)

R. K. Tigges, U.S. Geological Survey, Denver (rtigges@usgs.gov)

R.L. Reynolds, U.S. Geological Survey, Denver (rreynolds@usgs.gov)

P.S. Chavez, Jr., U.S. Geological Survey, Flagstaff (pchavez@usgs.gov)

Introduction

We are developing a wind-erosion model to investigate the sensitivity of dust-emission rates in the Southwestern U.S to climatic variability and land-use changes. Dust-emission rates largely depend on the amount that wind shear or “friction” velocity \(u_\ast\) within the boundary layer at the surface exceeds the “threshold” friction velocity of the surface \(u_{\ast s}\) ([3]). To calculate \(u_\ast\), we use an atmospheric boundary layer (ABL) model designed for terrestrial-type planets ([1],[2]). This model includes effects of “free-stream” wind velocity, atmospheric stability, both aerodynamically smooth and aerodynamically rough airflow, and the roughness lengths of momentum, sensible-heat, and water-vapor transfer \((Z_{0m}, Z_{0h}, Z_{0v})\). The ABL model is currently being linked with the non-hydrostatic Penn State/NCAR mesoscale climate model MM5 ([5]) to allow coupling with regional atmospheric dynamics and local topography.

Methods

The momentum roughness height (length), \(Z_0\) (m subscript dropped) is the roughness height for a bare soil surface “covered” by non-erodible roughness elements, such as plants, clasts, and small-scale topography. As cover is removed, \(Z_0\) approaches the roughness height of the bare surface, here noted as \(Z_{0b}\). \(Z_0\) also represents a measure of the rate at which the regional wind flow is dissipated through the boundary layer to overcome surface friction. The roughness height \((Z_0)\), whether the surface is covered or bare ([6], affects both the model-derived, atmospheric friction velocity in the boundary layer \((u_\ast)\) and the surface threshold friction velocity for dust-emission \((u_{\ast s})\).

Many natural surfaces consist of a mixture of non-erodible roughness elements (e.g. plants, clasts, small-scale topography) overlying a bare erodible surface. To account for the effect of non-erodible roughness elements on the threshold friction velocity, Raupach et al. ([7]) define a term called, \(R_e = (u_{\ast s} / u_\ast)\) which is ratio of the threshold friction velocity for a bare erodible surface to that for a surface covered with non-erodible roughness elements. Further, Raupach et
al. ([7]) were able to parameterize $R_t$ in terms of the physical dimensions and drag coefficients of the individual surface roughness species. $R_t$ is equivalent to the (efficient) friction velocity ratio $f_{eff}(Z_0)$ defined by Marticorena et al. (6), which is a function of the surface roughness height. Thus, $Z_0$, is directly connected to measurable physical dimensions and drag coefficients of the surface roughness elements.

The terms $R_t$ and $f_{eff}$, however, do not account for surfaces composed of multiple, coexisting types of roughness elements (e.g., plants, clasts, and small-scale topography). To address this problem, we examine the way momentum is partitioned among arrays of roughness species, and we show mathematically (see appendix) that the total threshold friction velocity ratio, $R_t$, is related to the independent friction velocities of “n” species according to:

$$1/R_t^2 \approx (1/R_1^2) + (1/R_2^2) + \ldots + (1/R_n^2) - (n-1)$$

As the total threshold friction velocity ratio, $R_t$, approaches values of 0.1 to 0.2 ([6]), the lee wakes behind a dominant or combination of species are packed so closely together as to completely shelter the surface against wind erosion. Aerodynamically, the wind profile characterizing this regime is displaced upward from one where bare portions of the surface still experience wind-shear stress to one whose profile represents shear stress only from the overlying roughness elements. Marticorena et al. ([6]) found that when values of $Z_0$ approached 0.6 cm, almost complete sheltering of the ground surface occurred. In contrast, Wolfe ([8], p.162) found that values of $Z_0$ approached 4 cm before complete sheltering occurred. Our measurements showed values of $Z_0$ up to 7.1 cm without indication of a wind-profile displacement height, although we did not place anemometers at a sufficiently low height to verify wind-profile displacement. We used this information to adjust Marticorena et al.’s ([6]) (efficient) friction velocity ratio, $f_{eff}$, to accommodate larger $Z_0$ values than 0.6 cm for complete sheltering.

This theoretical development provides a means to calculate the total roughness height of a complex surface composed of multiple roughness species and is compared below to field measurements.

**Data Analysis**

During one eight-day period in April 2001, we measured $Z_0$ at 12 diverse sites located around the perimeter of Soda (Dry) Lake, in the central Mojave Desert, California. All sites had evidence of past sand saltation, based on scouring of the surface. Three collapsible 9-meter-high wind towers were operated simultaneously at separate sites, each with three or four anemometers mounted in a logarithmic spacing with the lowest anemometer placed above 0.9 meters to protect internal bearings. One-minute averaged wind speeds were recorded from each anemometer during a 24-hour period to include the early-morning and late-afternoon periods of neutral buoyancy. We measured thermal stability from data continuously acquired at two nearby meteorological stations. Every morning the wind and thermal data were processed on-site to determine $Z_0$ and its associated error criteria. If the data proved sufficiently robust, a tower was moved to a new site; otherwise, the tower remained operating at the same site for another 24-hour period.
Results

While acquiring these wind data, we measured (in some cases estimated) a large number of plants for their geometrical widths, heights, and mutual separations for each surface roughness species using digital camera, tape measure, and calibrated rods. The mean and variance from these physical dimension data and estimated drag coefficients were incorporated into equation (1) as part of the Raupach parameterization of the friction velocity ratio for a particular species. Because methods are lacking to accurately measure the drag coefficients for naturally occurring solid and porous roughness elements in terms of physical parameters, estimates and uncertainties (standard deviations) for the drag coefficients were made based on previous reported field and lab studies in the literature (e.g., [4], [9]). These data were placed into equation (1) for each site and converted to calculated mean and standard deviation values of $Z_0$. The results, comparing the measured and calculated values of $Z_0$ for each site, are shown in Table 1 and plotted in Figure 1. Considering the inherent uncertainties in measuring the properties of natural systems, the results (correlation coefficient is 0.845 for mean values) show the measured and calculated values are generally within each other’s standard deviation about the mean ($-\sigma, +\sigma$), which is generally considered to be a strong correspondence. Results that showed the greatest disparity could be ascribed to inadequate documented of the physical dimensions and distributions of non-erodible roughness elements, incorrect estimates of species drag coefficients, or to inadequacies in the model. More work needs to be done to test the functional form of the model (Eq. 1).

Conclusions

Our most important conclusions are: (1) the actual sheltering caused by multiple, coexisting types of roughness elements (e.g. multiple plant species, desert pavement, and small-scale topography) needs to be unambiguously defined to yield an accurate measure of the local surface roughness, $Z_0$; (2) the effects of transitions from non-sheltered to sheltered surfaces may be abrupt and have a strong influence on accurate assessments of roughness element properties; and (3) the sensitivity of our $Z_0$ development to roughness properties suggests that during drought, surfaces largely sheltered by plants can become vulnerable to wind erosion resulting from very small losses in vegetative properties, not easily detected by the eye (e.g., substantially lower drag coefficients as a result of a relatively minor leaf loss or minor bending over of tall grasses).

Because our $Z_0$ development shows such a sensitive response to measured roughness geometry and drag coefficients, it would be useful to make more accurate field measurements of these properties than we achieved during our field experiments in April 2001.
Table 1. Comparison of measured and calculated mean $Z_0$ with added standard deviations ($-s$, $+s$) for a given site. The measured and calculated standard deviation values show skewness, because they depend on a basically logarithmic relationship.

<table>
<thead>
<tr>
<th>Site #</th>
<th>Measured $Z_0$ (cm)</th>
<th>Calculated $Z_0$ (cm)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>-s</td>
<td>Mean</td>
</tr>
<tr>
<td>200-201</td>
<td>1.66</td>
<td>2.79</td>
</tr>
<tr>
<td>202</td>
<td>.89</td>
<td>1.94</td>
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<td>203</td>
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<td>7.1</td>
</tr>
<tr>
<td></td>
<td>.194</td>
<td>.29</td>
</tr>
</tbody>
</table>

Figure 1. Graph of measured and calculated data in Table 1 using a logarithmic scale. Correlation coefficient $r=0.845$ explains 71% of variance; w/site #203 removed $r=0.87$ explains 76% var.

References

Extraction of the geometry of surface clasts from ground-based digital images: Application to studies of wind erosion

D.J. MacKinnon, U.S. Geological Survey, Flagstaff (dmackinnon@usgs.gov)

P.S. Chavez, Jr., U.S. Geological Survey, Flagstaff (pchavez@usgs.gov)

Introduction

We are developing a regional dust-emission model for the Southwestern U.S. This model is strongly controlled by the natural suppression of dust-emission by surface roughness and armoring of the surface by clasts and crusts. In our dust-emission study area located in the central Mojave Desert, California, USA, many wind-erosion surfaces are covered by clasts of varying size and coverage density. These clasts suppress wind erosion and dust emission by both physically covering a portion of the erodible surface and by aerodynamically reducing wind stress on portions of the bare surface downwind from each clast. Although the role of clasts and pavement in sheltering a surface against wind erosion is well documented in the literature, little work quantifying the degree of suppression has been done. A quantitative model of this suppression requires not only knowledge about clast number and size, but also clast height above the protected surface. Removing and sieving a surface sample of these clasts destroys the required height information as well as the amount of surface the clasts covered. A method is presented here to extract these critical clast properties using ground-based digital imagery without disturbing the clasts. Final demonstration that the method works will be achieved by comparing image and hand counts.

An on-going study by Dave Miller (pers. comm.) in the central Mojave seeks to examine the effects of a variety of erosion processes on desert slopes. Freeze-thaw and other processes break down native rock into clasts, and eventually long-term creep and sheet-wash transport these clasts to lower elevations. Miller’s study requires the removal and sieving of a large number of surface samples. We have recently joined this study to investigate methods of extracting surface clast and particle size information from digital surface images at Miller’s field sites. In a parallel effort, Pat Chavez is using spatial variability analysis of the digital images to extract clast size statistically. The two studies discussed above would benefit from the automatic extraction of clast properties using digital image processing techniques discussed here.

Methods and Results

We have acquired a large number of digital photographs from a wide variety of central Mojave sites. These images were taken using a 3-megapixel, 3-color digital camera with JPEG compression invoked. The latter digitally smooths the radiance values of the image pixels so that images generally occupy far less memory than the 9 megabytes normally required. We have yet to determine whether such compression reduces the ability of our algorithms to discriminate clasts from their background. Figure 1 shows a black-and-white image of the red band from a 3-color set used to develop the algorithm; note the scale bar in the lower right.
The eye, similar to our algorithm, uses two methods to discriminate an object from its background: color and texture. Our color algorithm generates a processed image by first identifying the 3-color range occupied by the fine-grained, typically uniform background of erodible sediment on which the clasts lie and sets all such pixels to a DN of 255; all other pixels are set to a DN of 0. If the clast colors are clearly different from the background our algorithm does very well at associating a black polygon (as small as a few pixels in size) with the area occupied by each clast in the image (see Figure 2). Discrimination of smaller-sized particles requires another image be taken closer to the surface.

Careful examination of Figure 2 shows that several large clasts remain undetected by the color algorithm because their DN range is similar to the background, however, the eye can easily discriminate these clasts by texture differences. To duplicate the texture discrimination, we created a resultant image by processing each pixel in the original 3-color image with a 9-sample by 9-line standard-deviation box filter. The standard deviation filter measures the degree of DN variability within the box centered on each pixel in the image. In general, background pixels composed of fine-grained material show much higher standard deviation values than do the clasts, which are composed generally of smoother textured clasts. The exception is when a clast is partially covered by fine-grained material or is composed of finely textured (contrasting) materials itself. Our texture-discrimination algorithm was applied to the red band of our test image (Figure 1) and the results are shown in Figure 3. The algorithm clearly shows the red-colored clasts are detected, but detection of the other clasts by this algorithm is poor. Using the color and texture algorithm together promises to yield better overall discrimination than using each one separately.

To optimize the results of either the color or texture algorithm it is important that the images be fully illuminated and contain no shadows; shadows will be falsely classified as a different color and texture than the background surface and counted as part of the clast dimension, and shadows obscure the true size of the clasts.

In the color-processed image (Figure 2) some of the clast locations have “holes” (255 DN) within the 0 DN area associated with a clast or “bridges” between clasts. Further image processing is required to remove these artifacts. The “cleaned-up” image can then be processed by a cluster algorithm to count all pixels associated with a clast. This cluster counting will contain all size and shape information associated with each clast from which a range of statistics can be derived.

Finally, to obtain the clast-height information required for aerodynamic roughness measurements, the clasts need to be illuminated at a low enough elevation angle to display shadows next to the clasts; the low angle illumination can be achieved artificially with a flash unit or using low-sun conditions. From the length of the shadows the height of the clasts can be inferred. The measurement and detection of these shadows can also be automated either by discriminating DN changes between a fully illuminated image and one containing shadows. Another approach to the shadow detection is to use the 3-color images of the scene containing the clast shadows and create a ratio image between the blue and red bands. Because blue light is scattered by the atmosphere more strongly than is red, a blue-to-red ratio image will appear very bright within the location of the shadows and can be readily discriminated. Using the illumination azimuth and elevation angles in the image, the height is extracted by multiplying the largest shadow pixel length for each clast by the tangent of the elevation angle.
All the methods discussed above can be automated to yield the geometry of surface clasts from ground-based images, which greatly simplify and improve the accuracy of surface clasts analysis for all erosion processes.

Fig. 1 -- Red band image of ground clasts

Fig. 2 -- Color algorithm applied to red, green, blue bands

Fig. 3 — Texture algorithm applied to red band

Preliminary results of the ADEC first IOP in April 2002

Masao, Dr. Mikami (Meteorological Research Institute, Japan) Nagamini 1-1, Tsukuba, Ibaraki 305-0052 Japan

A five-year experiment to investigate the impact of aeolian dust on climate (ADEC) started in April 2000. This Japanese-Seno cooperative project aims to estimate the total supply of mineral dust to the atmosphere and to measure the degree of radiative forcing direct effect by aeolian dust particles. The first IOP was put into practice from April 8 to 21, 2002. One of the main objectives of this IOP is to observe the wind erosion processes and parameterize them for the GCM dust model. To achieve this, direct observation of the wind erosion process and characterization of the dust is required. For this purpose, in situ observation was done at seven sites around the Tarim basin and Dunfuan. We will introduce here the preliminary result of Aksu (lat. 40°37′07″N, long. 80°49′43″E) and Qira (lat. 37°00′56″N, long. 80°43′45″E) in the north and south of the Taklimakan Desert. In these sites, observation was made for roughness length, friction velocity, stability, soil particle size distribution and water content on the ground surface, infrared radiation, solar radiation, and stream wise dust flux. Equipment to be used includes automatic weather stations, visibility meters (Mikami, 2000), dust particle measuring systems (Yamada et al., 2001), and Doppler Sodar systems.

For understanding the saltation process, we developed the Dust Particle Measuring System (DPMS). This system enable us to get the total amount of the drifting sand particle, its size class distribution, and time variation of dust particle size distribution. The system consists of dust catcher and dust particle counter. A cyclone type dust catcher makes use of centrifugal force effect by the drifting sand particle is used in order to collect the drifting dust particles. Dust particle counter is designed for the sequential measurement of dust particle size drifting on the ground surface. For this purpose, we use a semiconductor laser of 670 nm in wavelength and 1 mm in diameter. This instrument is originally from snow particle counter first developed by Schmidt (1977). When a dust particle passes through the laser beam, the detector output drop in proportion to its cross-sectional area. From the time sequence of the detector output, dust particle size, from 40 micron meter to 400 micron meter, and its number of individual bins can be obtained. Stream wise dust flux can be estimated from these data. For long-term use in desert, the instrument is designed for integral-type including sensor, data logging system, and data storage memory (Compact Flash Memory). In addition, solar panel with 42 W of maximum output power is used for the power supply. As is the same as dust catcher, detector slit is designed to be always turning to the windward direction by use of wind plate.

During the First IOP, we came across two weak dust event. Here we summarized the floating dust particle size distribution just before and after the dust storm event, soil particle size information on the gobi surface, and verification of the Gillette scheme used in the MRI GCM dust model. Threshold wind velocity at the gobi site was decided at 7.4 m/s and the saltation flux at each particle size is also discussed.
The role of instrumentation in aeolian research; recent advances and future challenges

Cheryl McKenna Neuman, Department of Geography, Trent University, Peterborough Ontario, Canada K9J 7B8 (cmckneuman@trentu.ca)

William G. Nickling, Wind Erosion Laboratory, Department of Geography, University of Guelph, Guelph Ontario, Canada N1E 2W1 (nickling@uoguelph.ca)

This paper represents a compilation of our collective experience in attempting to measure accurately aeolian transport processes in the field and laboratory. All measurements of aeolian processes can be generalized into at least one of the following categories: 1) a property of the airstream, 2) a mass transport rate (either horizontal or vertical), and 4) a gravitational or inter-particle force moment at the surface that opposes motion. In association with each of these categories, we address the following questions: What do we need to know? What do we presently measure? What problems exist? What improvements need to be made? What spatial and temporal scales need to be addressed? In essence, the purpose of this paper is an attempt to define, review and stimulate discussion regarding the many challenges facing aeolian researchers engaged in empirical work; it is not to provide the answers. Specific topics addressed will include measurement of: 1) wind flow and determination of surface shear stress; 2) tractive stress on rough surfaces; 3) particle transport rate; and 4) crust strength and capillary force. The various techniques currently used to measure these processes and surface characteristics will be discussed in the context of time scales varying from seconds to years, and spatial scales varying from sub-millimeter to meters.
Sonic anemometers in aeolian sediment transport research

J.H. van Boxel: Institute for Biodiversity and Ecosystem Dynamics (IBED), Universiteit van Amsterdam, Nieuwe Achtergracht 166, 1018 WV Amsterdam, The Netherlands (Email: J.H.Boxel@science.uva.nl)

G. Sterk: Erosion, Soil and Water Conservation, Wageningen University, Nieuwe Kanaal 11, 6709 PA Wageningen, The Netherlands (Email: Geert.Sterk@users.tct.wau.nl)

S.M. Arens: Bureau for Beach and Dune Research, Iwan Kantemanplein 30, 1060 RM Amsterdam, The Netherlands (E-mail: Arens@duinonderzoek.nl)

Fast response wind and turbulence instruments, including sonic anemometers are used more and more in the research of aeolian sediment transport. These instruments provide data on mean wind, but also on friction velocity, wind speed fluctuations and turbulence statistics, such as the U-W and W-T covariance (C_{UW} and C_{WT}), which are a measure for the momentum flux and the sensible heat flux.

This short paper will examine two problems that may arise when using sonic anemometers, namely the low/high frequency losses and the interpretation of the sonic anemometer measurements over sloping or non-homogeneous terrain.

The sonic anemometer

The sonic anemometer uses the speed of sound to measure temperature and wind speed. The wind speed is measured in three orthogonal directions, so the complete three-dimensional wind vector is known. Since the instrument has no moving parts the measurement is almost instantaneous. Many sonic anemometers have a path length of 15-20 cm and an output frequency of around 20 Hz. They can be used to obtain high frequency wind measurements, mean wind speeds and standard deviations of wind speed in the stream-wise or longitudinal (U), lateral (V) and vertical (W) direction. The C_{UW} covariance, multiplied with the density of the air yields the vertical flux of momentum and the square root of -C_{UW} is the friction velocity (U\_f \equiv \sqrt{-C_{UW}}).

Generally a good estimate of the heat flux can be obtained by multiplying the C_{WT} covariance with the density and heat capacity of the air. Since the Obukhov length is mainly a function of friction velocity, heat flux and temperature, which are all measured by the sonic anemometer, a sonic anemometer will also provide the Obukhov length, which is a measure for atmospheric stability.
The turbulence spectrum

Let us picture turbulence as a collection of vortices, big and small, superimposed on the mean wind and on each other. These vortices transport air with its characteristics (e.g. momentum, temperature, humidity) up and down and will thereby create the fluxes of momentum, heat, water vapor, carbon dioxide, etc.

Large vortices transport momentum over large distances, but are rare. There are many small vortices, but these are inefficient in transporting momentum. Most of the transport is done by intermediate vortices (compare with frequency-magnitude principle). The shaded area in Figure 1 indicates the frequencies that cause 90% of the total momentum transport.

Low frequency losses of more than 1% occur when normalized frequencies \( N = f \times z / U \) of more than 0.0016 are not resolved any more (table 1). When using for example a measuring height \( z \) of 2 m and a wind speed \( U \) of 5 m/s, under neutral conditions, that would mean that we would have to resolve frequencies \( f = N \times U / z \) down to 0.004 Hz, requiring our period of measurement to be at least 250 s. To be on the safe side, we want to have at least a few of these large eddies, so 750 s would be a good measurement period.

High frequency losses can occur due to the limited measurement frequency of the sonic anemometer or due to the measurement volume of the instrument. In order to keep the high frequency loss below 1% we should be able to resolve normalized frequencies \( N \) up to 2.37 (table 1). In the above example \( z = 2 \text{ m}, U = 5 \text{ m/s} \) that corresponds to a frequency \( f = N \times U / z \) of 6 Hz, so that the sampling frequency of the sonic should be at least 12 Hz.

If the highest frequency that must be resolved is 6 Hz and the wind speed is 5 m/s, that means that vortices as small as 0.8 m \( (\lambda = U / f) \) should be measurable. Kaimal & Finnigan (1994) state that vortices can measured down to a size of \( \lambda = 2 \pi d \). So our sonic with a path length \( d \) of 15 cm should be able to resolve vortices of 0.94 m. In the above example we would have a high frequency loss of slightly more than 1%.

<table>
<thead>
<tr>
<th>Loss</th>
<th>1%</th>
<th>2%</th>
<th>5%</th>
<th>10%</th>
</tr>
</thead>
<tbody>
<tr>
<td>Low freq.</td>
<td>0.0016</td>
<td>0.0023</td>
<td>0.0045</td>
<td>0.0087</td>
</tr>
<tr>
<td>High freq.</td>
<td>2.37</td>
<td>1.53</td>
<td>0.77</td>
<td>0.45</td>
</tr>
</tbody>
</table>

Table 1: Normalized frequencies \( N = f \times z / U \) that must be resolved in order to keep high or low frequency.

Measurements over sloping terrain

Sonic anemometer measurements are easiest to interpret if they are performed over flat and homogeneous terrain, preferably under neutral conditions. However often our objects of study
often are neither flat nor homogeneous. Measuring on a slope or near a roughness transition will cause the streamlines to slope.

When the streamlines are sloping part of the stream-wise fluctuations will be measured as vertical fluctuations. If the slope is upward this will give a positive contribution to $C_{UW}$ (the U-W covariance). Since $C_{UW}$ is negative this will decrease the measured friction velocity. Therefore the sonic coordinate system has to be rotated according to the streamlines, which are not necessarily parallel to the surface.

Kaimal and Finnigan (1994) already discussed the three rotations required, which we will refer to as yaw, pitch and roll. The yaw rotation (along a vertical axis) is required to orientate the longitudinal velocity $U$ along the streamlines, so that the mean of $V$ becomes zero. The second rotation (pitch) will orient the $Z$-axis perpendicular to the streamlines (mean $W = 0$). The roll rotation will set $CV_W$ to zero. Wilczak et al. (2001) applied these rotations to sonic tilt errors and derived equations for simplified cases. If both yaw and roll can be ignored, they argue that the pitch-error in $C_{UW}$ is approximately 6.4% per degree of tilt.

To test their theory we used 293 runs of 15 minutes each of sonic anemometer measurements, taken on a beach near Faro (Portugal) on 17 days between 14 January 1999 and 21 March 1999. Part of the beach was steeply sloping (about 8°). During the measurements almost all wind directions occurred and wind speeds varied from 1 to 10 m/s. The sonic anemometer was placed at different locations on the beach, usually at an elevation of approximately 0.9 m. The high frequency loss in $C_{UW}$ resulting from the sonic path length (15 cm) was estimated to be circa 4% under near neutral conditions and less during unstable conditions. For wind speeds below 10 m/s the high frequency cut off resulting from the sample frequency (21 Hz) will be at a higher frequency than that resulting from the sonic path length. So in our case the path length is the factor determining the high frequency losses in $C_{UW}$.

We have calculated the error resulting from the sloping streamlines, by correcting individual data points as well as by using turbulence statistics and the equations suggested by Wilczak et al. The results were almost identical, except for two runs in which there was a strong change in wind direction.

From figure 2 it can be seen that the error in the $C_{UW}$ covariance can be considerable when the slope corrections are not applied. In three cases the uncorrected $C_{UW}$ was positive, meaning that the friction velocity was not defined anymore. Our data suggest that the error is 9.5% per degree, which is larger than the 6.4% per degree predicted by Wilczak et al. (2001) for neutral cases. If the regression line is not forced through the origin, there is an offset (-1% error for zero slopes).

Wilczak et al. predict a stronger response at unstable conditions. Although the worst outliers in figure 2 (in gray) are very unstable ($z/L < -0.1$), the majority of the unstable points do not show a trend that differs very much from the measurements during near neutral conditions.

![Figure 2: Error when the $C_{UW}$ covariance is not corrected for the slope](image)
Since stability is not the only source of the difference between theory and the results in figure 2 we will examine the high frequency cut off as a possible cause. At first approximation the relative error in $-C_{UW}$ is given by: 
$$RE_{UW} = 100\% \times \frac{\sigma_U^2 - \sigma_W^2}{U^2} \sin(slope)$$,

where $\sigma_U$ and $\sigma_W$ denote the standard deviation of the wind speed component in the X- (stream-wise) and Z-direction (near vertical, but perpendicular to the streamlines) (for a more accurate formula see Wilczak et al.). Since $\sigma_U$ exceeds $\sigma_W$ the relative error is positive for positive slope angles.

The W-spectrum peaks at higher frequencies than the UW-cospectrum and the U-spectrum peaks at lower frequencies. Therefore the high frequency losses in $\sigma_U^2$ become small, but those in $\sigma_W^2$ will be considerably larger than 4%. Panofsky et al. (1977) argue that at near neutral conditions $\sigma_W/U*$ should be near 1.25. In our measurements the ratio was 1.09 for near neutral conditions. The high frequency loss in $C_{UW}$ was estimated 4%, so that our $U*$ is underestimated by 2% and the ratio of $\sigma_W/U*$ becomes 1.07. If we compare this with the value of 1.25 suggested by Panofsky et al. we conclude that our $\sigma_W$ values are about 15% too low, probably because of high frequency losses. If we correct all our $\sigma_W$ values for this high frequency loss and recalculate the slope error, the −1% offset disappears, but the slope error still is 9.0% per degree, which is yet considerably higher than the slope error proposed by Wilczak et al. 2001. So although the high frequency losses for $\sigma_W$ are considerable this does not seem to affect the correction very much. For $\sigma_U/U*$ Wilczak et al. suggest a value of 2.29 for near neutral conditions. Our data show a large scatter for $\sigma_U/U*$, but the mean value in our measurements at near neutral conditions was about 2.65, which is 16% larger than the value suggested by Wilczak et al. If we take this value into account and consider the high frequency loss in $\sigma_W/U*$ that explains the difference in slope sensitivity between our measurements and the theoretical calculations by Wilczak et al. According to our estimates only 2% of the difference between our $\sigma_U/U*$ and the value proposed by Wilczak et al. can be attributed to high frequency losses in our $U*$. The reason for the rest of the difference is not known. It is very well possible that the general relations for $\sigma_U/U*$ and $\sigma_W/U*$ that Wilczak et al. derived from Panofsky et al. (1977) do not hold for sloping or non-homogeneous terrain.

**Conclusions**

- Sonic anemometers can provide very useful information for aeolian research.
- High and low frequency losses can be estimated using the turbulence spectra.
- When measuring over sloping or non-homogeneous terrain the measurements must be corrected for the slope of the streamlines.
- Our best estimate from the measurements on a Portuguese beach is that the slope error in $C_{UW}$ is 9.0% per degree of slope, which is considerably more that the 6.4% suggested by Wilczak et al. (2001) on the basis of theoretical considerations.

**References**

A wind tunnel study of the collection efficiency of an aerodynamically improved ‘Frisbee’ dust trap

G.F.S. Wiggs, Department of Geography, University of Sheffield, Western Bank, Sheffield, S10 2TN, UK (E-mail: g.wiggs@sheffield.ac.uk)

J. Leys, Department of Land & Water Conservation, Gunnedah, NSW, Australia (E-mail: jleys@dlwc.nsw.gov.au)

G. H. McTainsh, Department of Environmental Sciences, Griffith University, Brisbane, Australia

S. Heidenreich, Department of Land & Water Conservation, Gunnedah, NSW, Australia (E-mail: sheidenrich@dlwc.nsw.gov.au)

C. Strong, Department of Environmental Sciences, Griffith University, Brisbane, Australia (E-mail: craig.strong@mailbox.gu.edu.au)

Introduction

Airborne dust is commonplace and dust storms are a well-known natural hazard in dryland regions. Efforts aimed at understanding the dust erosion and transport system are hindered by a lack of reliable data on dust sources, the timing of dust events and the link between land management practices and dust generation. Such information can most easily be gained from direct measurements of dust flux and deposition. However, the successful measurement of such processes remains elusive and is one of the most problematic procedures in aeolian geomorphology (Goosens & Offer 2000). The most important characteristic of a dust trap is its collection efficiency. This efficiency is controlled by the degree to which the trap represents an obstacle to the windflow, and so an aerodynamic design is most efficient. One of the most important components of the dust system which still needs to be understood is the rate of deposition. Existing deposition trap designs have efficiencies of commonly between 20% and 75% (Goosens & Offer, 2000). A novel trap design has been described by Hall et al (1994) which is similar in design to modern “shallow bowl” traps but incorporates an additional deflector ring to improve aerodynamic behaviour. The deflector ring aims to reduce the acceleration in wind speed over the trap opening. Such flow acceleration is largely responsible for reducing the efficiency of standard trap designs (Hall et al, 1994).

The aim of this research was to test the aerodynamic behaviour of the deflector ring and to measure any resulting improvement in the collection efficiency of a ‘frisbee’-type deposition trap.

Methods

All the experiments were carried out in the portable field wind tunnel at Gunnedah Research Station, Australia. The tunnel was of the blowing fan type with a working section 9.50 m long and cross-section 1.15 m wide and 1.00 m in height. Turbulence was induced at the entrance to
the working section using a flow contraction resulting in a turbulent intensity at the trap site of approximately 6%. The trap was placed on the centre-line of the tunnel, 8.50 m along the working section and with the opening at a height of 0.55 m. The design of the trap and surrounding deflector was similar to that of Hall et al. (1994) and is summarised in Table 1.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Length scale (mm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Depth (H)</td>
<td>93</td>
</tr>
<tr>
<td>Internal trap diameter</td>
<td>297</td>
</tr>
<tr>
<td>External trap diameter (D)</td>
<td>354</td>
</tr>
<tr>
<td>H/D</td>
<td>0.263</td>
</tr>
<tr>
<td>Internal deflector diameter</td>
<td>424</td>
</tr>
<tr>
<td>External deflector diameter</td>
<td>707</td>
</tr>
<tr>
<td>Upper deflector ellipse</td>
<td>20</td>
</tr>
<tr>
<td>Lower deflector ellipse</td>
<td>73</td>
</tr>
</tbody>
</table>

Wind velocity profiles and turbulence intensity were measured across the trap opening both with and without the deflector ring using an array of six fast-response pitot-tubes referenced to a freestream velocity measured at 0.25 m from the tunnel roof. These measurements were carried out at a sampling frequency of 1 second and averaged over 3 minute periods. Measurements were also undertaken in the empty tunnel for comparison.

The collection efficiency of the trap was tested both with and without the deflector ring by injecting a known quantity of silica flour dust (mean grain size of 14.6 μm) over a 60 second period at the upwind tunnel contraction. The amount of dust collected in the trap was then compared to the expected catch using the following equation:

\[ E = \chi \cdot A \cdot V_f \cdot t \]

Where: \( E \) = collection efficiency; \( \chi \) = dust concentration; \( A \) = trap area; \( V_f \) = particle falling speed; \( t \) = time.

The dust concentration (\( \chi \)) was determined from the volume of air pushed through the tunnel during the experiment and the measured quantity of dust injected into the airstream. Particle falling speed (\( V_f \)) was calculated from data in Davies (1945). Each test measurement was repeated 3 times.

**Results**

Velocity profiles above the centre of the trap opening are presented in Figure 1. Here, the velocity data are expressed as a fractional speed-up ratio (\( \delta s \), Jackson & Hunt, 1975) in relation to the measured velocity at a point (\( u \)) and the reference velocity measured in the free-stream (\( U_r \)):

\[ \delta s = \frac{(u-U_r)}{U_r} \]
A fractional speed-up ratio ($\delta s$) value of +0.05 therefore represents a 5% increase in wind velocity when compared to that measured at the same point in the absence of the dust trap.

Figure 1. Velocity profiles measured above the trap opening

The data in Figure 1 clearly show that the deflector has a significant impact on the airflow above the trap. Without the deflector the measured velocity profile shows a 66% deceleration immediately above the trap opening (at 0.55 m height) and an 8.5% acceleration at 0.62 m height (7 cm above the trap opening). In contrast, the deflector has the effect of reducing the intensity of both the deceleration immediately above the trap opening (to 53%) and the acceleration higher in the profile (to 6.5%). In addition, the height of the point of maximum velocity is reduced to 0.59 m (4 cm above the trap opening) and the zone of flow acceleration above the trap is consequently much smaller. The deflector appears to have had the desired effect of diverting the flow downwards around the gauge, hence flattening the flow over the gauge and reducing the acceleration in wind speed (Hall et al., 1994).

The reduced impact on the flow dynamics above the trap opening when employing the deflector had a follow-on effect on the efficiency of the collector in terms of dust deposition. The data in Figure 2 show the collection efficiency of the trap, both with and without the deflector, with varying free-stream velocity. At all wind speeds the deflector had the effect of increasing the efficiency of the deposition trap. At lower wind speeds (4 – 8 ms$^{-1}$) the efficiency of both collectors is very high, the deflector giving a collecting efficiency of over 100% at 4 ms$^{-1}$. At 10 ms$^{-1}$ the trap with no deflector experienced a decline in efficiency from 83% to 61%, falling further to 46% at 12.5 ms$^{-1}$. In contrast, the trap with the deflector maintained a very high collecting efficiency of 97% at 10 ms$^{-1}$ which reduced to only 69% at 12.5 ms$^{-1}$. At lower windspeeds the deflector had the impact of increasing the collection efficiency of the trap by approximately 13% to 14%. At windspeeds over 10ms$^{-1}$ this was increased to 33% to 37%.
**Conclusion**

The addition of a deflector ring had a significant impact on both the aerodynamics and collection efficiency of a ‘frisbee’-type dust deposition trap. The deflector successfully partitioned the flow around the dust trap reducing above-trap flow acceleration and helping to maintain velocity immediately above the trap opening. In wind speeds up to 10 ms\(^{-1}\) the deflector improved collection efficiency by between 13% and 37% though some reduction was evident at higher wind speeds. Further testing is required, specifically for a greater range of sediment size, but the wind tunnel results presented here suggest that improved aerodynamic design could have a significant effect on results from field dust collectors.

**References**


Application of Caesium-137 Technique on Wind Erosion in Gonghe Basin, Qinghai Province, China

P. Yan, China Center of Desert Research at Beijing Normal University, Beijing 100875, China (E-mail: yping@bnu.edu.cn)

P. J. Shi, China Center of Desert Research at Beijing Normal University, Beijing 100875, China (E-mail: spj@bnu.edu.cn)

G. R. Dong, Cold and Arid Regions Environmental and Engineering Research Institute, Chinese Academy of Sciences, Lanzhou 730000, China

Introduction

The worldwide fallout $^{137}$Cs, associated with the atmospheric testing of nuclear weapons during 1950s and 1960s, has provided a valuable man-made tracer for studies of soil erosion and sediment delivery (Ritchie & McHenry, 1990). Since its introduction in the 1970s, the $^{137}$Cs technique has widely applied in water erosion leading to profound accomplishments (Ritchie & Ritchie, 1996). Until the 1990s a few attempts was made to estimate the wind erosion rate using $^{137}$Cs technique, and it is still in an exploring stage (Chappell, 1999).

Here we report the results of $^{137}$Cs sampling, measurements to investigate wind erosion in Gonghe Basin of Qinghai Province during 1998-1999. The objectives were to analysis $^{137}$Cs distribution in the area and its depth profile of different soils, determine $^{137}$Cs reference inventory of the area, and estimate wind erosion of sampling plots and the whole area, and discuss the input and output ways of aeolian sediment in the Basin.

Materials and Methods

Study area

The Gonghe Basin of Qinghai Province is located between the Qilian Mountains to Kunlun Mountains at the northeastern edge of the Qinghai-Tibet Plateau and the Qinling Mountains, at latitude 35°20′-36°51′N, longitude 98°24′-101°22′E, and elevation between 2 400-3 500 m above sea level. The basin has 210 km length, 30-90 km width and a total area 13 787 km$^2$. The climate of the area belongs to plateau temperate semiarid type, with a mean annual temperature 1.0-3.3°C. The mean annual precipitation 250-400 mm and annual evaporation is 1 528-1 937 mm. Prevailing wind are northwest wind and southeast, with mean annual wind velocity is 2.1-2.7 m/s and gales (>17.0 m/s) occur on a mean of 15.5-50.6 days/a. Dust storms (visibility < 1000 m) occur on a mean of 6.5-20.7 days/a. Due to dry and cold climate, frequent and strong wind, spare vegetation, plentiful surface sand material, and irrational economic activities in recent decades, the basin becomes one of the regions suffering from severe desertification in the steppe and

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desert steppe zones in China. The total desertified land area in the basin is 4 926.12 km², accounting for 35.73% of the total area (Dong et al., 1993).

**137Cs sampling and measurement**

Total 380 137Cs samples were taken from seven general sampling sites (RS1-7) involving main land types of the area and four typical sampling plots (SAM1-4) represented most severe desertified land (SAM1), severe desertified land (SAM4), on-going desertified land (SAM2) and latent desertified land (SAM2) respectively. At the general sites, the sampling arranged at specific soil with random 2-3 repeatedly sampling points. At typical plots, according to topography, three sampling forms were adopted, i.e. grid (SAM1), parallel transects (SAM2 and SAM3) and vertical transects (SAM4) (Walling & Quine, 1993). The SAM1 plot has 60 × 40 m, SAM2 and SAM3 100 × 100 m and SAM4 100 × 25 m, with a same 10 m spacing between sampling point. 137Cs samples include bulk sample and incremental sample, collected respectively by spiral drill with 7.4 cm internal diameter and 15 cm depth, and scraper-plate with 20 cm length, 10 cm width and 2 cm depth (increment). Meanwhile, at some specific points, grain-size and bulk density samples were collected.

137Cs samples were air-dried, ground to pass through a 1-mm mesh sieve. All samples weighting about 400 g each were test by 7-ray detector (EG&G ORTEC LOAX HPGe) connected to an ORTEC amplifier and multi-channel analyzer in the in the Nuclear Physics Laboratory of Sichuan Union University, Chengdu, China (Yan et al., 2001).

**Computational methods**

The 137Cs point inventory, CPR (Bq/m²) was calculated by Eq.(1) (Walling & Quine, 1993)

\[ CPI = \sum_{i=1}^{n} C_i \cdot B_{di} \cdot D_i \]  

(1)

where \( i \) is the sampling depth and \( n \) is total number of samples with detectable 137Cs, \( C_i \) is 137Cs activity for sample \( i \) (Bq/kg), \( B_{di} \) is the bulk density (t/m³) for depth \( i \), and \( D_i \) is the depth (m) for sample \( i \).

And 137Cs redistribution was calculated as (Walling & Quine, 1993):

\[ CPR = (CPI-CRI) \cdot 100/CRI \]  

(2)

where \( CPR \) is 137Cs percentage residual at a sampling point in the field relative to the native control area(%), \( CPI \) is 137Cs point inventory (Bq/m²) and \( CRI \) represents 137Cs reference inventory (Bq/m²).

**Results and Discussion**

**137Cs inventory and its depth distribution**

The mean 137Cs activity and the total 137Cs inventory of the region were determined to be 4.84±0.34 Bq/kg and 1513.83±108.37 Bq/m² respectively. The 137Cs activity of different land types was in the order: woodland > playa > cold grassland > dry farmland, steppe > fixed sand dune > desert steppe > shift sand dune > interdune (Table 1). The 137Cs depth distribution can be divided into normal profile (NP), aggrading profile (AP), eroding profile (EP) and disturbed profile (DP). The playa RS3 is typical AP (Fig.1a), woodland RS2 DP-EP (Fig.1b), shift dune SAM1and SAM4 AP-EP (Fig.1b), fixed dune RS2 AP (Fig.1g), interdune SAM1 and SAM4 typical EP (Fig.1c), cold grassland RS7 and SAM3 typical NP (Fig.1d), steppe RS4, SAM2 and SAM4 EP-DP (Fig.1e), desert steppe RS5 and SAM1 EP (Fig.1i), dry farmland SAM2 and SAM3 typical DP (Fig.1f).


**137Cs reference inventory**

The 137Cs distribution of cold grassland was uniform over the region and its depth distribution approached a negative exponential curve. From the 137Cs determination of samples collected at RS7 in cold grassland, the 137Cs reference inventory of the Gonghe Basin was 2691.78±196.08 Bq/m² which is essentially equal to the mean 137Cs reference inventory of the Northern Hemisphere. Regression analysis shows that the 137Cs depth profile of RS7 can follow the function of peak distribution as following (Yang et al., 1998):

\[
Cs = 98.576(1 - (0.989 - z/19.208)^3.689)(0.989 - z/19.208)^2.689
\]

Comparing with the exponential distribution and typical peak distribution, it appears a downward shift of 137Cs peak and flattening trend in the curve. This phenomenon which was frequently seen in reference materials may be ascribed to the steadily downward migration of 137Cs in recent several decades (Yang et al., 1998).

**Wind erosion estimation by 137Cs model**

The model of 137Cs profile distribution to estimate wind erosion rate of natural soil profiles (sand dune, interdune and grassland) was developed as Eq.(4) in accordance with 137Cs reference inventory of the region.

\[
ER = -100 \cdot Bd \cdot \ln(1 + 0.9951CPR/100) / 0.2658T
\]

<table>
<thead>
<tr>
<th>Land type</th>
<th>Number of sampling sites/plate</th>
<th>Number of samples</th>
<th>Mean 137Cs activity (Bq/kg)</th>
<th>Mean 137Cs inventory (Bq/m²)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Fixed sand</td>
<td>Fixed sand</td>
<td>1×2</td>
<td>2.48±0.21</td>
<td>881.63±72.04</td>
</tr>
<tr>
<td>Woodland</td>
<td>Woodland</td>
<td>1×2</td>
<td>12.59±0.79</td>
<td>4733.05±293.99</td>
</tr>
<tr>
<td>Playa</td>
<td>Playa</td>
<td>1×2</td>
<td>10.57±0.67</td>
<td>3977.39±240.12</td>
</tr>
<tr>
<td>Shift sand dune</td>
<td>Shift sand dune</td>
<td>1×2</td>
<td>1.75±0.15</td>
<td>2218.36±195.04</td>
</tr>
<tr>
<td>Interdune</td>
<td>Interdune</td>
<td>1×2</td>
<td>0.51±0.05</td>
<td>148.88±14.91</td>
</tr>
<tr>
<td>Grassland</td>
<td>Grassland</td>
<td>1×2</td>
<td>7.67±0.50</td>
<td>1796.54±117.12</td>
</tr>
<tr>
<td>Farmland</td>
<td>Farmland</td>
<td>1×2</td>
<td>0.76±0.08</td>
<td>366.29±39.91</td>
</tr>
<tr>
<td>Total or mean</td>
<td>380</td>
<td></td>
<td>4.84±0.34</td>
<td>1513.53±108.37</td>
</tr>
</tbody>
</table>

![Figure 1](image-url)
where $E_R$ is the net wind erosion rate of the sample site (t/ha·a), $CPR$ is the loss of $^{137}$Cs (%), $Bd$ refers to the bulk density of sampled soil (t/m³), $T$ is the time period between the year of maximum $^{137}$Cs fallout (1963) and the sampling year, in this study, $T=35$ a. For the cultivated soil (farmland SAM2 and SAM3), the wind erosion rate was calculated using the $^{137}$Cs proportional model (Sutherland et al., 1991):

$$E_R = 10^4 \cdot Bd \cdot Pd \cdot \left(\frac{CPR}{100}\right) / T \quad (6)$$

where $Pd$ is the plough depth (m), and $T$ is cultivation time, for SAM2 $T=9$ a (1989-1998) and SAM3 $T=20$ a (1978-1998).

By these models, the wind erosion rate of four sampling plots was calculated to be $24.910\pm1.088$ t/ha·a averagely, reaching the moderate to severe grade (Zachar, 1982). The sand dune plot SAM1 with the largest rate $43.679\pm1.491$ t/ha·a was one experienced severe wind erosion continuously; The farmland/grassland plots SAM2 and SAM3 were in degrees of moderate and slight of wind erosion, $14.901\pm0.974$ t/ha·a and $7.484\pm0.533$ t/ha·a, but land reclamation had such a great influence on wind erosion that the accelerated wind erosion came to 5.74-8.80 times than normal wind erosion. The dune plot SAM4 was in severe wind erosion, $33.576\pm1.353$ t/ha·a, but it was relatively slighter than SAM1 plot because of counteraction by steady accumulation of aeolian sand.

**Regional wind erosion assessment and erosion-deposition equilibrium**

According to the desertified land classification (Dong et al., 1993), the wind erosion rates of four plots were converted into the regional wind erosion assessment by erosion-deposition equilibrium model. The wind erosion rate of the Gonghe Basin was $12.556$ t/ha·a which passed test of erosion-deposition equilibrium at a relatively small error less than 10%. Among the eroded sediments, more than 57% were preserved in the form of sand dunes, which was the main sand source of development of land desertification; 17% were transported into the Yellow River and its tributaries, which were the main sediment source of the upper reach of the Yellow River; 14% were deposited on the surface as dust; less than 1% of the sediments deposited in the inland lakes and rivers; and a very small portion was transported great distance as suspension dust.

**References**


Natural sources of Aeolian dust in Amman, and selected Areas of Jordan

Hadeel Al Dwaikat, PhD student; School of Science and Environment, Coventry University, UK
Ian Foster, School of Science and Environment, Coventry University, UK.
Nasfat Hunjul, Natural Resources Authority, Amman, Jordan.
Adrian Wood, School of Science and Environment, Coventry University, UK
Joan Lees, School of Science and Environment, Coventry University, UK

Abstract

The overall aim of this research programme is to characterise potential sources of atmospheric dusts deposited in the Amman (urban), Mafraq (rural) and Azraq (desert) areas of Jordan. These dusts derive from a number of sources, some of which may be remote from the point of deposition and will reflect long range transport during dust storms. The potential dust sources will be characterised by their mineral magnetic, radionuclide and geochemical signatures which will be quantitatively compared, using an un-mixing model, with the same signatures of deposited dusts in each area.

The aim of this paper is to identify the potential natural and anthropogenic sources of dust and the impact that human activity may have in increasing the availability of dust from natural sources for transport.

Geologically it is important to identify those rocks and minerals that could weather by natural processes to provide material fine enough to be entrained by the wind. Part of this study therefore focuses on an analysis of the geological materials capable of producing such fine sediments. The major natural process increasing air pollution in the research areas is the dust storm (sandstorm) which affects the desert areas more than the highlands. However, in north-eastern Jordan over the last decade, changes in agricultural practices, especially the clearance of surface stone layers in the basalt regions for cultivation, may have increased the availability of dust in these regions for aeolian transport. Soil erosion in the hilly areas causes landslides while tree cutting and overgrazing by large numbers of livestock may also increase sediment availability.

Small localised anthropogenic sources of contaminated dust may also be released in rural and desert areas but these pollutants will be rapidly dispersed and diluted by wind, washout by rain or through deposition with large quantities of natural dusts.

Direct contributions to air pollution primarily affects urban areas where the density of building, industry and vehicles prevents pollutants from being dispersed. Urban air pollutants include particulate matter, heavy metals and acidifying gases such as \( \text{SO}_4 \), \( \text{NO}_x \). The city of greater Amman is suffering from serious air pollution problems from nearby industries.

Other activities, especially quarrying, mining and building construction sites in and around Amman may also make additional natural geological materials available for transport. These include limestone extraction for building construction and cement production in the south of the city. Another major source of dust to the west of Amman derives from the Al Fuhays cement factory because most of the prevailing winds are from the west. The most important source of dust in the north is from the sandstone quarries of the Safout area. The eastern areas of Amman and Ruseifa have been greatly affected by emissions from major industrial plants, phosphate mining and phosphate processing. These phosphate deposits, and processed phosphates, are characterised by elevated levels of U-235.
Man and Dust--A Unique Perspective from Southeastern New Mexico/West Texas

Richard Arimoto Carlsbad Environmental Monitoring & Research Center, New Mexico State University (CEMRC/NMSU), Carlsbad, NM  88220 (E-mail: arimoto@cemrc.org)

Sondra Sage, CEMRC/NMSU (E-mail: sage@cemrc.org)

Cheryl Schloesslin, CEMRC/NMSU (E-mail: schloess@cemrc.org)

Thomas Kirchner, CEMRC/NMSU (E-mail: kirchner@cemrc.org)

Joel Webb, CEMRC/NMSU (E-mail: jwebb@cemrc.org)

Barry Stewart, CEMRC/NMSU (E-mail: bstewart@cemrc.org)

David Schoep, CEMRC/NMSU (E-mail: dschoep@cemrc.org)

Mark Walthall, CEMRC/NMSU (E-mail: mwalthall@cemrc.org)

Introduction

Results of two studies involving humans’ influences on dust in the New Mexico/West Texas region will be presented. The first study deals with aerosol particles from the vicinity of the Waste Isolation Pilot Plant (WIPP), which is located near Carlsbad, NM and is the first permitted deep underground repository for defense-related nuclear waste in the U.S.A. This study was funded by a grant from the U.S. Department of Energy, and it was designed to provide an objective and independent means for evaluating the potential effects of the WIPP on the environment. The second study focused on emissions from a cattle feedyard located near Friona, TX. That study was sponsored by a seed grant from the U.S. Department of Agriculture, and it was a collaborative effort involving scientists from Texas A & M Agricultural Extension Service, the Carlsbad Environmental Monitoring & Research Center (CEMRC), the Department of Biology, New Mexico State University, Las Cruces and Texas A & M University. The objectives of the feedyard study were to begin an evaluation of the physical, chemical and microbial characteristics of fugitive particulate matter from cattle feedyards in the southern High Plains.

The WIPP Environmental Monitoring Project: Aerosols

Methods As part of an environmental monitoring program focusing on the WIPP, aerosol samples were collected and analyzed to characterize the spatial and temporal variations in the activities and concentrations of selected radionuclides and inorganic substances in the atmosphere around the WIPP. High-volume aerosol sampling was conducted at three sites which were named (1) On Site, (2) Near Field, and (3) Cactus Flats. $^{239,240}$Pu was determined by alpha spectrometry following chemical separations.
A separate set of low-volume aerosol samples was collected and analyzed for major ions and trace elements, using ion chromatography and inductively-coupled emission spectrometry and mass spectrometry for the analyses, respectively. Gravimetric determinations were made only for the high-volume radionuclide filters.

**Results** The average $^{239,240}$Pu activity concentrations in total suspended particles (TSP) samples (1.2 to $1.8 \times 10^4$ nBq m$^{-3}$) were similar to those previously reported (Lee et al., 1998), but they varied substantially with season, with the highest values generally in spring. Further, the $^{239,240}$Pu activity concentrations were comparable among the three sites, and therefore there was no evidence for elevated $^{239,240}$Pu activities due to WIPP operations. The partitioning of $^{239,240}$Pu activity concentration in PM$_{10}$ (particles less than 10 μm diameter) samples relative to TSP (50-55% of the $^{239,240}$Pu was in the PM$_{10}$ fraction) was slightly lower than the corresponding PM$_{10}$/TSP ratios of high-volume mass or several inorganics (e.g., sulfate, aluminum or lead), indicating that $^{239,240}$Pu tends to be on relatively large particles. The aerosol mass loadings (g m$^{-3}$) and the $^{239,240}$Pu activity concentrations were correlated for all sets of samples, but there was a clear difference at On Site where the TSP samples showed a higher mass to $^{239,240}$Pu concentration ratio than the other sites. This indicates that activities or processes occurring at or near the WIPP site produced aerosols that contributed to the mass loadings but contained less $^{239,240}$Pu than ambient aerosols. Regression models showed that about 63% of the variability in $^{239,240}$Pu activity concentrations could be explained by wind travel, sampling location, length of the sampling interval and aerosol mass. The $^{239,240}$Pu activity concentrations also were correlated with aluminum (an indicator of mineral dust), implicating the resuspension of contaminated soils as an important determinant of $^{239,240}$Pu concentrations in the aerosols. The $^{239,240}$Pu/Al ratios for the aerosols were substantially higher than in soils, and this could be explained by the preferential binding of $^{239,240}$Pu to small soil particles that have large surface area to mass ratios and also have higher aluminum contents than larger particles.

**Ambient Air Quality Issues Related to Confined Animal Operations**

**Methods** Three sampling trips were made to the Paco Feedyard, near Friona, TX, and on each trip, several types of samplers were used for different purposes. FRM PM$_{10}$ samplers were used to collect aerosols and Whatman 41® filters were used as sampling substrates for the studies of trace elements and major ions presented here. Inductively-coupled plasma mass spectrometry (ICP-MS) and graphite-furnace atomic absorption spectroscopy (AAS) produced data for more than thirty elements (Ag, Al, Ba, Be, Ca, Cd, Ce, Co, Cr, Cu, Dy, Er, Eu, Fe, Gd, Hg, K, La, Li, Mg, Mn, Mo, Na, Ni, Pb, Pr, Sb, Sc, Si, Sm, Sn, Sr, Th, Ti, Tl, U, V and Zn) in acid digests of the filters. The detection limits for the ICP-MS and AAS analyses are in the low parts per billion range and uncertainties are typically 10 to 20%. The concentrations of major ions (chloride, nitrate, phosphate, sulfate, sodium, ammonium, potassium, magnesium, and calcium) in aqueous extracts of the samples were determined by ion chromatography (IC).

Samples of feedyard soils also were collected, and they were analyzed by AAS for As and by ICP-MS for the same suite of elements as in the aerosol samples. Aqueous
extracts of the these soil samples were filtered through a 0.45 m syringe filter and then analyzed for the same suites of anions and cations as the aerosol samples using the same IC methods.

Results  The concentrations of thirty-four elements and nine ions were determined in the majority of the aerosol samples. Of the trace elements readily determined by ICP-MS, only Sn was below detection in all of the aerosol samples, but the data for As, Be, Cd, Hg, La, Se either were sparse or below detection in the soil samples (see below), and therefore those results will not be presented here. Most of the elements (Ba, Ca, Ce, Dy, Er, Eu, Fe, Gd, K, Li, Mg, Mn, Mo, Na, Nd, Pr, Sc, Si, Sm, Sr, Th, Ti, Tl, U, and V) in the aerosol samples were associated with mineral dust particles, i.e., they were correlated with Al, an indicator of dust, and they exhibited ratios to Al similar to those in crustal material. In contrast, several elements Ag, Cr, Cu, Ni, Sb, Zn, and to a lesser extent Co and Pb, were enriched in the feedlot atmosphere over the levels expected from mineral dust, suggesting there are important non-crustal sources for these elements.

In general, the concentrations for the trace elements in the upwind and the downwind aerosol samples were comparable, with differences in the arithmetic mean concentrations typically less than 30%. None of the elements analyzed had concentrations that were more than 50% lower in the downwind samples relative to the upwind samples. However, K, Mo and Na had mean concentrations that were 2-fold higher in the samples collected downwind of the feedyard, and Ni and Mg were 40% to 70% higher in the downwind samples. While these differences are not statistically significant at probabilities for chance occurrence of 5% or less, they are generally consistent with trends in the soil data.

The major ion data for aqueous extracts of the aerosol samples also show some differences related to their orientation relative to the feedlot. For nitrate, sulfate, ammonium, and calcium, the differences between the upwind and downwind were small, less than 30%, but chloride, phosphate, sodium, potassium, and magnesium were 2- to nearly 5-fold higher in the downwind samples relative to the upwind ones. As was true of the elemental data for the acid digests of the aerosol samples, these downwind/upwind differences in aerosol ion composition are generally consistent with patterns evident in soils.

The concentrations of most elements (expressed as mass of the element per unit mass of soil) in the upwind soil sample were either about same or higher than those in the downwind samples. This is precisely what one would expect if there were a substantial amount organic material in the downwind samples, a situation that would lead to an effective "dilution" of the elements of interest on a per unit mass soil basis. There are several notable exceptions to the downwind/upwind depletion of trace elements however: K, Mg, Na, Ni and Zn exhibited concentrations that were ~2-fold higher in the downwind samples. Although the available sample set is small, the concentrations of these elements were quite comparable in the three independently collected downwind samples, suggesting a reasonable degree of homogeneity among those soil samples. Calcium was higher in acid digests of the downwind soils by about 50% compared with the upwind sample, and Mo was higher by ~25%. Ca, K, Mg and Zn are all essential nutrients that are either naturally elevated in feed components for the cattle or are routinely included as dietary supplements in feedyard rations. Zinc, in particular, is added to feedyard diets as a micronutrient owing to its involvement in protein and carbohydrate metabolism as well as immune system function. The slight enrichment of Mo is likely not be significant while the elevated concentrations of Ni
in the downwind samples may be due either to some mechanical aspects of the feeding operations in which metal particles are shed, or to some unknown cause.

Interestingly, the major ions showed some even larger differences than the trace elements between upwind versus downwind soils. All of the ions except calcium and nitrate in the aqueous soil extracts (this includes chloride, phosphate, sulfate, sodium, ammonium, potassium, and magnesium) were enriched in the downwind soils compared with the one from upwind of the feedyard; these major ion enrichments ranged from ~7X to more than 100X, but most were around 10 to 20X. Soluble calcium also was enriched in the downwind samples, but the difference was less than 2-fold and thus probably not significant. Nitrate was measurable in the upwind sample but below quantitation limits in the downwind samples--this is equivalent to a difference of more than 1000X, and at present is difficult to explain.

Conclusions

The WIPP EM studies indicate that the resuspension of mineral particles contaminated with radioactive fallout from nuclear weapons tests is a major, if not controlling, influence on the concentrations of \(^{239,240}\)Pu in the atmosphere. No effects from the WIPP itself could be detected, but continued monitoring serves a useful purpose, i.e., maintaining public confidence that the WIPP is operating safely. In contrast to the WIPP, the effects of confined animal operations on visibility are readily observable even to the untrained eye, and the odors emanating from feedyards often are considered a nuisance. The preliminary study of the Paco cattle feedyard is a first step in developing a scientific understanding of the chemical and health effects and socioeconomic implications of airborne particulate matter emitted from feedlots.

References

A Comparison of Wind and Water Erosion Rates among Grassland, Shrubland, and Forest Ecosystems

D. D. Breshears¹, J. J. Whicker¹, J.E. Pinder², and M. P. Johansen²
¹Los Alamos National Laboratory  ²Colorado State University

Abstract

Wind erosion is major land surface process that operates in most dryland ecosystems of the world. Although wind erosion has been quantified in several different agricultural lands and ecosystems, methodologies have generally not been sufficiently consistent to allow comparisons among different semiarid ecosystems, such as grasslands, shublands, and forests. Further, the magnitude of wind erosion relative to water erosion in a given ecosystem may indicate its relative importance as a land surface process, but studies quantifying the wind erosion relative to that of water erosion within and across ecosystems is largely lacking. Our objectives were to (1) estimate wind erosion rates in a semiarid grassland and a semiarid forest, and to compare them with rates reported for a semiarid shrubland, (2) estimate and compare projected wind erosion rates with water erosion rates in these three ecosystems, (3) discuss possible trends and hypotheses about the relative roles of wind and water erosion in semiarid grasslands, shrublands, and woodlands. Based on data from passive Bagnold sample collectors, we found that wind erosion rates at a semiarid grassland site (Aurora, Colorado, USA) and a semiarid forest (Los Alamos, New Mexico, USA) were less than an order of magnitude of those reported for a semiarid shrubland (Carlsbad, New Mexico, USA). We calibrated these wind erosion rates by comparing estimates from a Bagnold sampler to those from a another passive sampler (BSNE) with a known calibration to obtain estimates of wind erosion fluxes for the three sites: grassland, shrubland, and forest. Using meteorological data (wind and precipitation) from each site, we estimated decade-scale total amounts of wind erosion and water erosion at all three sites. Our results indicate water erosion is dominate at the grassland and forest sites, whereas wind erosion is dominant at the shrubland site. On the basis of our results, we present hypotheses about general trends in the roles of wind and water erosion across semiarid grassland, shrubland, and forest ecosystems.
Simulations to optimise sampling of aeolian sediment transport in space and time for mapping

A. Chappell, School of Environment & Life Sciences, University of Salford, Manchester, M5 4WT, UK. (E-mail: a.chappell@salford.ac.uk)

G. McTainsh, Faculty of Environmental Sciences, Griffith University, Brisbane, Queensland 4111, Australia

Craig Strong, Faculty of Environmental Sciences, Griffith University, Brisbane, Queensland 4111, Australia

John Leys, Gunnedah Research Center for Natural Resources, Department of Land and Water Conservation, Gunnedah, NSW PO Box 462, Australia

Introduction

An understanding of wind erosion requires information on the spatial and temporal variation of the magnitude of material transported by aeolian activity. Unfortunately, the sampling frequency in space and time is often inadequate to estimate the accuracy and precision of that magnitude. The problem stems from a lack of knowledge of the spatial and temporal scale of variation in aeolian transport that is often compounded by the shortage of resources (aeolian sediment traps). An innovative approach to this problem was provided by Sterk and Stein (1997). It developed work by Stein (1998) which combined variation of a property in space and time with the approach of Voltz and Webster (1990) for mapping resource limited properties. The result is a geostatistical approach that combines the spatial variation of several aeolian transport events to provide a more reliable aeolian transport model than that of observations for a single event. Consequently Sterk and Stein’s (1997) geostatistical method is attractive for mapping aeolian transport with few resources. However, it assumed that a sampling strategy static over time was suitable, that all events should be combined together regardless of differences in magnitude and scale of transport and that the model of spatial variation was constant over time. This paper aims to test these assumptions and to develop the approach further by maximising the use of the aeolian sediment traps. The objectives were to (1) model the spatial variation of aeolian transport for each event using the variogram; (2) examine the effect of alternative combinations of wind erosion events and sampling strategies (static and mobile over time) on aeolian sediment transport mapping.

Methods

The study area is the Lake Constance claypan (or playa) on the high floodplain of the Diamantina River in Diamantina National Park (DNP), western Queensland.
The claypan is approximately 5 km x 5 km and is bordered by red sand dunes on the northeast and southwest. During floods large quantities of fine grained alluvium are deposited on the floodplain that is subsequently remobilised by aeolian activity (McTainsh, 1989).

Passive sediment samplers were used in this study, based upon the design of Fryrear (1986) but modified by the addition of a rain hood to avoid sediment loss by rainfall impact, and tested by Shao et al. (1993). An alternative method to a regular grid with a spacing limited by the number of traps placed over the playa was required because it could easily miss the spatial scale of variation in the sediment transport. However, as there was little information on spatial variation of the erodibility of the playa surface with which to stratify the area, an unstratified but nested (according to anticipated scales of transport) sampling approach was used. This sampling network resulted in a total of 160 locations at which samplers could be located. These locations were logged by global positioning system (GPS). Since only 40 wind vane samplers were available at any one time, they were distributed across the study area by allocating a single sampler to a node and its related nest. The actual location of a sampler was decided by random selection of either the node or any one of the three distances along a pre-determined bearing (the nest). After a wind erosion event the sediment collected in each sampler was washed out with de-ionised water and, following oven drying, was weighed and stored. The samplers were then moved to new randomly selected locations. The samplers were located over the 25 km² playa resulting in an average density of ca. 625 000 m² per sampler. A total of eight events were sampled.

Data Analysis

It is well-documented (cf. Webster and Oliver, 1992) that the variogram requires many samples to provide a reliable model of spatial variation for a property. However, Chappell et al. (in press) have shown recently that when the scale of aeolian transport is sufficiently long and the spatial variation is sufficiently well sampled (typically using a nested sampling strategy) the variogram is reliable (Figure 1).

![Figure 1](image_url)
The model fitted to a variogram for aeolian sediment transport of each event is of the same form over time (Chappell et al., in press) regardless of the different magnitude of wind erosion events and scale of aeolian transport. This validates the previously untested assumption that the pooled within-event variogram may be calculated. However, the evidence suggests that some wind erosion events produce spatial scales of aeolian sediment transport that are more similar for some events than for others. It questions the implicit assumption that all sampled events should be combined together.

Pooled within-event variograms were calculated following Voltz and Webster (1990) and Sterk and Stein (1997) for several combinations of events by using all events, by separating events according to geomorphic (aeolian transport) information, and by an arbitrary sequential division (Figure 2). The results show that considerable variation in the model of spatial variation may be obtained depending on the combination of events that are selected. Consequently, there was a need to systematically investigate this and the effect of alternative sampling strategies on the pooled within-event calculation in order to identify the most appropriate sampling strategy for aeolian sediment transport using minimal resources.

Many wind erosion events would be required to provide a range of alternative sampling strategies. To avoid a lengthy and expensive monitoring campaign synthetic data were generated. Using the existing models of spatial variation for each wind erosion event alternative realisations of aeolian sediment transport were simulated. Stochastic simulation, specifically simulated annealing, was used here to estimate aeolian transport across the study area. These data were sampled using random, systematic and nested strategies that were either static over time or mobile over time.

Figure 2. Examples of several combinations of pooled within-event variograms for aeolian sediment transport sampled during some of the wind erosion events at Diamantina Lakes, western Queensland, Australia.

This resulted in six possible sampling strategies. All of these sampling strategies were used with three temporal combinations. Simulated sampling of this type were used to produce variograms of the aeolian sediment transport (Figure 3).
Figure 3. Pooled within-event variograms of aeolian sediment transport sampled using separate strategies during wind erosion events at Diamantina Lakes, western Queensland, Australia.

In order to compare the performance of the combinations of events and sampling strategies, 40 locations were identified at which estimates would be made. Ordinary kriging was used for each pooled within-event variogram to estimate the aeolian sediment transport at those test locations. The simulated estimates were compared with the estimates at the same test locations from ordinary kriging using the variograms of individual events. Statistical comparisons (mean absolute error and residual mean squared error) were used to identify the best performance.

Conclusions

The pooled within-event variogram is an efficient geostatistical methodology for mapping aeolian sediment transport with relatively few resources. Chappell et al. (in press) have validated one of the assumptions for that methodology. However, results here suggest that the notion that all sampled wind erosion events should be combined indiscriminately would reduce the potential efficiency of this methodology. Furthermore, these results suggest that considerable improvements over a static sampling framework can be made using a nested strategy in conjunction with this methodology. The results presented here are likely to be important for large area sampling of aeolian sediment transport that has been very difficult because of resource constraints.

References


The incidence of wind erosion as related to soil properties and geomorphic history in south-western Australia

R.J. Harper, Univ. West. Aust., Nedlands W.A. 6907 (E-mail: richardh@calm.wa.gov.au)

R.J. Gilkes, Univ. West. Aust, Nedlands W.A. 6907 (E-mail: bob.gilkes@uwa.edu.au)

M. Hill, Bureau of Rural Sciences, ACT, Australia, 2601 (E-mail: Michael.Hill@brs.gov.au)

D.J. Carter, Agric. West. Aust., Albany W.A. 6330 (E-mail: dcarter@agric.wa.gov.au)

Introduction

Wind erosion is a major form of land degradation in the dryland farming systems of south-western Australia, with severe erosion events recurring every few years (Select Committee into Land Conservation 1990).

The relationships between soil properties and wind erosion have been poorly developed in this region. Moreover, there is little soil survey information at scales suitable for farm management to allow the extension of wind erosion models at the farm scale. If, however, relationships can be deduced between wind erosion and soil properties or geomorphological features, those attributes, and hence wind erosion hazard, can be identified and mapped by landholders or consultants. Appropriate management systems can then be devised for soils according to their susceptibility to erosion, rather than by applying uniform management procedures across all soils.

The most extreme episodes of wind erosion since land development occurred near Jerramungup, Western Australia, in 1980 and 1981 with the extent estimated by a Landsat MSS remote sensing analysis (Carter and Houghton 1984). In that study, soil and geomorphological information was unavailable, this preventing an understanding of the relationships between these attributes and the incidence of wind erosion. A detailed soil and geomorphological survey has since been undertaken and estimates made of the rates of wind erosion using the $^{137}$Cs technique (Harper and Gilkes 1994). In this paper we assess the relationships between the incidence of wind erosion, soil properties and geomorphic history.

Methods

This study was undertaken in an area of 5 200 ha adjacent to Lake Cairlocup, 400 km south–east of Perth, Western Australia. The annual rainfall of 350 mm, is mostly received in winter. Farms were developed from mallee (Eucalyptus spp.) woodland in the period 1964-74, and farming involves annual rotations of cereal or legume crops with pasture.

A general-purpose soil survey, using the free survey technique, was undertaken at a scale of 1:12 500. Approximately one site was examined every three hectares, with the sampling intensity varying over the study area. Six geomorphic
surfaces, based upon geomorphic process and parent materials, were identified, with these comprised of 21 major soils (Fig. 1). These surfaces can be considered in terms of the following three groups:

1. **Deeply weathered granitic ridges and slopes:** Two geomorphic surfaces were defined in deeply weathered granitic terrain, based on the degree of landscape dissection. Unit UDL occurs where landscape dissection has been minimal. Sandy surfaced, texture contrast soils occur within these areas (Petroferric yellow Sodosols, Eutrophic yellow Kandosols, Ferric hypernatric grey Sodosols (Isbell 1996); Plinthoxeralfs, Natrixeralfs (Soil Survey Staff 1994)). Unit PDL occurs where the pallid zone, a deeper horizon of the former lateritic profile has been exposed. Sandy surfaced texture contrast soils also occur within these areas (Calcic yellow Sodosols, Natrixeralfs), often containing calcrete nodules or concretions.

![Fig. 1. Aerial photograph with overlay showing the major geomorphic units.](image_url)

2. **Valley floor:** The valley floors are poorly drained and comprise sequences of Quaternary sediments dominated by Lake Cairlocup, a 200 ha hypersaline playa. This is bounded by a series of source bordering lunettes (clay or para dunes) that extend 5 km to the southeast (Harper 1994). The lunettes and associated swales have been separated into two geomorphic surfaces — those comprised of loamy soils (Hypernatric grey sodosol, Regolithic calcic Calcarosols; Xerochrepts), which occur immediately adjacent (0-2 km) to Lake Cairlocup (Unit LSC), and those with sandy texture contrast soil profiles (Mesonatric brown Sodosols, Natrixeralfs), some distance (2-5 km) from Lake Cairlocup (Unit LSD). A third small geomorphic surface (Unit FVF) was defined for those areas of the valley floor, without playas or lunettes. This contained texture contrast soils with carbonate nodules at depth (Calcic yellow Sodosols, Natrixeralfs)

3. **Sandy aeolian deposits:** Unit SDS occurs as a discontinuous, NW-SE oriented, strip ~10 km long and 2 km wide, directly southeast of the ephemeral Cairlocup Creek. Quartzose sands, generally >100 cm deep (Mesonatic grey Sodosols, Typic Quartzipsamments) overlie a range of substrates and landscape positions. This feature represents former quartzose aeolian deposits, derived from the drainage line during former arid periods.

Samples were taken from each of the 21 major soils. Each of the 219 samples comprised 20–30 cores taken from a depth of 0-10 cm. For 133 soils with a sandy E horizon paired samples were taken from 0-10 and 10-20 cm.
A Landsat MSS analysis was used to interpret wind erosion patterns in 1980 and 1981 (Carter and Houghton 1984). This classification was superimposed onto the field mapping polygons, using ArcView, with sampling sites classified as eroded or non-eroded on the basis of this classification. Two categories were derived – those areas eroded in either year, to give a maximum extent of erosion and those areas eroded in both years, this category indicating areas most prone to erosion.

Results and Discussion

Severe wind erosion affected 1385 ha or 27% of the arable soils in the study area in 1980 or 1981, or 474 ha (9%) in both years. As a first order approximation, this resulted in a loss of soil carbon of around 4,900 t C from a total of 90,000 t in the top 20 cm. Although this carbon has been lost from the farms, it is not certain if it can be considered a net carbon emission as it may have been deposited elsewhere.

The incidence of erosion was estimated from the proportion of the different mapping units eroded across the study area. Despite strong evidence in this landscape of relict aeolian activity, the incidence of contemporary wind erosion was only partly related to past geomorphological processes (Fig 2a). Quartzose dune sands (Surface SDS) were particularly susceptible to erosion (47% eroded in either year, 18% in both years), whereas texture contrast soils formed on clayey, wind-formed lunettes (Surface LSD) were variably eroded (16%, 5%). Similarly, soils formed on deeply weathered regolith (Surface UDL) were also variably eroded (34%, 11%). Soils formed on stripped laterite (Surface PDL) and on loamy lunettes and swales close to playas (Surface LSC) were not affected by erosion. There were marked differences in the amounts of erosion within each geomorphic surface (Fig. 2b). Within Surface UDL the proportion of erosion ranged between 31 and 55% for different soils in either year (6-16% in both), in Surface LSD 0-41% (0-18%) and Surface SDS 25-75% (9-32%), with the most erodible soils being old dune systems. Overall, the incidence of erosion tended to be greatest on sandy surfaced soils and those with sand horizons deeper than 60 cm.

The risk of wind erosion was better predicted from surface soil properties, such as clay and silt contents and sand horizon depth. The contemporary particle size properties of the surface horizon of the soils may as much reflect the results of erosion, as the intrinsic properties of the soil. Therefore, samples from 10-20 cm
depth were used as a diagnostic horizon as this is below the depth of cultivation in this area (mean depth 9.2 cm).

The incidence of erosion was compared for classes based on 1% increments of clay content derived from the 10-20 cm deep sample, for two surface horizon depth classes (shallow <60 cm and deep >60 cm) (Fig. 3a). Two major trends are apparent. The first is the systematic decline in the proportion of samples that were wind eroded with increasing clay contents and the marked difference in erodibility between soils with shallow and deep sandy horizons. The proportion of sites eroded thus decreased from 45% for the 1-2% clay class to 16% for the 4-5% clay class. No site with >5% clay was eroded. There is thus a large range of wind erosion within soils that are considered as “sands”, with small differences in clay content having marked effects on erodibility. A similar trend is apparent with declining erosion with declining contents of silt, in this case erosion is confined to soils with <3% silt.

Wind erodibility is explained in terms of the strength of the soils, which is in turn related to clay content, and amount of plant cover (related to fertility and water storage). The soils that are most erodible have been pre-sorted by wind or water. Assessment criteria commonly used in field surveys, such as field texture and consistence, can indicate the likely risk of wind erosion, however these are less predictive that methods based on soil analysis.

Fig. 3 Proportion of sandy texture contrast soils eroded by wind for different classes based on physical analysis of the 10-20 cm layer. Soils with shallow (<60 cm) (■) and deep (>60 cm) (▲) sandy surface horizons. (a) Clay (%), (b) silt content (%). n = 134.

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Intensive Study Site for Monitoring, Measuring and Modeling Agricultural Field Dust Emissions

Jim Kjelgaard, Biological Systems Engineering, Washington State University, Pullman, WA 99164-6120, jkkelgaard@wsu.edu

Keith Saxton, USDA-ARS, Biological Systems Engineering, Washington State University, Pullman, WA 99164-6120, ksaxton@wsu.edu

Brian Lamb, Laboratory for Atmospheric Research, Washington State University, Pullman, WA 99164-2910, blamb@wsu.edu

Candis Claiborne, Laboratory for Atmospheric Research, Washington State University, Pullman, WA 99164-2910, claiborne@wsu.edu

Abstract

The Washington State University (WSU) USDA-ARS Conservation Unit in cooperation with the WSU Civil Engineering-Laboratory for Atmospheric Research has instrumented a large field in the Columbia Plateau (central Washington) to monitor and measure field dust emissions. The region contains soil types that predominantly consist of fine silt loams; over 90% (by weight) of the soil consist of particles that are smaller than 100 μm. During high wind events (HWE’s), the soil structure and dryland farming practices leave tilled and fallow fields (i.e. no crops planted during the growing season) extremely susceptible to wind erosion and large emissions of PM10 and smaller particles. As part of an air quality effort to quantify dust emissions, a 9-hectare section of a 300+ hectare field was instrumented and will be kept in continuous fallow over the next several years. Utility electrical power on-site will allow the operation of air quality instrumentation not previously used for agricultural dust emissions; this includes tapered element oscillating microbalance (TEOM) units, modified nephelometers (AQ-10’s), high-volume air samplers, sonic anemometers, a modified relaxed eddy accumulation flux system, intensive meteorological measurements, and threshold velocity monitors. Primary objectives for the research are to improve current model parameters for dust emission models, quantify field turbulence structure and record continuous dust concentrations at several heights through short-time step interval measurements.
Wind Erosion in Voivodina

Lj. Letic, Faculty of Forestry, 11000 Belgrade, Kneza Viseslava 1, Yugoslavia

R. Savic, Faculty of Agriculture, 21000 Novi Sad, Yugoslavia; (rassa@polj.ns.ac.yu)

Introduction

Both natural and anthropogenic conditions on the territory of Voivodina (2.15 mil. ha, northern part of Serbia, Yugoslavia) favour the occurrence of wind erosion. The continental climate of the Pannonian Plain with frequent strong winds, attaining the rates of even 40 m/s; annual precipitation sometimes even below 400 mm; large temperature amplitudes; markedly plain relief; more than 70% of the area being plow fields which are at a time without any vegetation cover and which under conditions of intensified agricultural production may be very erodible; insufficient (only about 6.5 %) and inappropriately located forest areas, are only some of the factors clearly indicating that the danger of wind erosion in Voivodina is potentially very high, and if the forecast climatic changes are to become true, the situation may be even worsened.

Destructive effects of wind erosion in Voivodina are most visible in agriculture. Because of wind erosion, the soil - one of fundamental natural resources, is degraded. The wind carries the finest humus particles, and also the nutritive and protective matter, and the just sown crop grains, pulls out and breaks young plants, denudes the roots of perennial plants, causes excessive evaporation and soil drying, and the blows of wind-borne particles damage green parts and fruit of the grown crops. Besides, wind sediments are filling in the drainage canals and water reservoirs. The loss of nutrients and moisture from the soil, repeated sowing, lower and non-uniform yields can also bring into question the profitability of the agricultural production on the areas endangered by wind erosion, and especially at the present when the intensive agricultural production is very expensive. On the other hand, the attenuation of wind erosion processes can directly lead to lowering of the production costs, and enabling, even under unfavourable climatic conditions, high yields to be more probable, and thus the production more effective.

Material and Methods

The process of wind erosion is essentially a very complex problem which requires that all its stages (initiation of movement, transportation, and deposition) should be encompassed by the study and quantification, i.e. all the relevant factors have to be included. Because of the multitude and stochastic character of essential parameters of the process, as well as because of high costs of such experiments, only fragmentary investigations have been carried out, so that is not possible to define some universal relationships. Still, on the basis of such investigations, a number of models, empirical formulae and indicators (coefficients, indexes, etc.) have been established for a quantitative description of wind erosion processes, i.e. for assessing their intensity. However, a direct, non-critical application of such models, defined for particular locations, irrespective of all their complexity and effective mimicking of natural
processes, may under the given conditions represent a certain risk, especially if the application is not followed by appropriate field investigations.

First concrete assessments and direct measurements of the intensity of wind erosion in Voivodina were carried out on the two sandy areas, which are potentially most endangered in this part of the Pannonian Plain. In the beginning of seventies, on a location in the Deliblato Sands ("European Sahara") based on measurements using a special rotating catcher of wind-borne particles (Jevtic, 1975). Afterwards, in 1980, in the Subotica-Horgos Sands a special centre was founded to monitor wind erosion on the soils of lighter composition (Letic, 1989). The investigations encompassed instrumental measurements of the intensity of wind erosion by monitoring relevant climatic parameters and the state of the soil – the object of the wind action. These investigations have been carried out simultaneously at two measurement stations, one under the conditions of intensive agricultural production with no wind protection, and the other in the forest protection belt.

Because of the justified assumption that the process of wind erosion is involved and very significant, not only on sandy soils but also well-structured soils of the type of chernozem and meadow black soil, starting from 1995 the investigations and measurements of wind erosion intensity have been carried out on a location near Novi Sad (Savic, 1999).

In this paper to point out the research of wind erosion on the soils of light mechanical composition on the Subotica-Horgos Sands. The Subotica-Horgos Sands are situated in the North part of Voivodina plain between the Danube and Tisza rivers, average length 50 km and diameter 5-10 km, area cca 24,000 ha. It is well-known orchard-grape vine region, with more than 33% under vineyards and orchards, cca 20% forests and woodlands, and over 34% under grassland.

Experimental station record: quantity of aeolian deposition, wind frequency and velocity, air and soil temperatures, air and soil humidity etc. Field data are analysed in order to qualify and quantify the deflation processes, as well as to define the conditions of climate and residual soil in which they occur.

Results and Discussion

A comparative method of stationary observation by wind-gage stations has been applied on specially selected erosion plots, of which one, used for agriculture, has not been protected ("A"), while the other ("B") has been protected with forest plantings. The obtained results suggest the existence of significant erosion processes outside the protective forest belt. On the basis of the measurements carried out during a number of years, an empirical relation has been derived for calculation of the wind erosion intensity on this and similar locations.

In the 1980-1999 period frequent variations in the wind erosion intensity and of some qualitative characteristics of the wind sediments were registered. The annual amounts of wind-borne sediments initiated in the Subotica-Horgos Sands are in the range from 0.63 to 35.87 kg/m on the unprotected area, and from 0.10 to 0.59 kg/m on protected areas (Figure 1).

In addition to the quantification of deflation processes in the researched area, the analyses of quality have been made, i.e. physical and chemical properties of the aeolian deposition were defined and compared to the same characteristics of the residual, surrounding soil. This time to point out to the following nutrients which are
removed from the soil complex of the Subotica-Horgos Sands: humus, total nitrogen, readily available phosphorus and potassium. (Letic et al., 2001).

Humus content in the aeolian deposition oscillated during the research period and ranged between 3.94-7.02 %, which is cca 5.7-18.1 (on the average 10.5) times higher content of organic matter then in the residual soil from which the particles were detached.

Total nitrogen content in the aeolian deposition ranges within the limits of 0.28-1.73 % which is 2.8-19.0 (on the average 11.1) times higher content than in the residual soil.

Readily available phosphorus content in the aeolian deposition ranges within the limits of 10.0-41.1 mg/100 g of soil, and that is cca 3.7-20.3 (on the average 9.3) times higher content of phosphorus than in the residual soil.

Readily available potassium content in the aeolian deposition oscillated within the limits of 15.7-40.0 mg/100 g of soil, and that is cca 5.8-14.3 (on the average 10.1) times higher content of potassium than in the residual soil.

The analysis of the relation of chemical characteristic of the aeolian deposition ("Ad") and the residual soil ("Rs") points to the very significant indication of soil fertilization loss affected by deflation processes. It is denoted by the "deflation coefficient" (\(\eta=Ad/Rs\)).

Also, by analysing the oscillations of the contents of humus and biogenic elements during the research period, it has been observed that the maximum and minimum of their concentration do not correspond to the maximal and minimal quantities of the aeolian deposition.

![Figure 1](image-url)  

**Figure 1.** Annual wind erosion intensities in the Subotica-Horgos Sands: "A" – Plowed field with no wind protection; "B" – area in the protective forest belt.

**Conclusions**

In natural and anthropogenic conditions on the territory of Voivodina deflation processes represent important factor of soil destruction and have also a negative effect on the other elements of the environment: water and air. Comparative
researches on the protected and unprotected erosion fields and presented results pointing out the significant degree of the vegetative cover protective effect.

The processes of accelerated wind erosion are most frequently a consequence of anthropogenic factors, inappropriate use of the soil, vegetation destruction, etc. Modern measures of wind erosion control must be complex, all-inclusive, continuous, and systematic. At that, one should constantly bear in mind the fact that there is no absolute protection from wind erosion, that is, there is no possibility of complete elimination of wind erosion processes, one can only endeavour to reduce them to a rationally acceptable level. It would be highly desirable to establish a network of measuring stations to monitor the wind erosion in Voivodina under the different natural conditions (microclimate, soil, etc.), as well as under different crops. In this way, among other things, it would be possible to check the correctness of the applied empirical methods and achieve a more reliable estimation of the erodibility of particular types of soils, protecting effect of the crop covering in particular stages of crop development, effects of different modes of soil cultivation, humidity state of the soil, and like. It should be especially pointed out the importance of establishing the amount, characteristics, and composition of the wind erosion sediments, as of the crucial factor of degradation of all the elements of the environment.

Chemical analyses of sediments indicate its increased load of nutritive matter compared to the residual soil from which the sediment was originated: humus up to 18 times, nitrogen up to 19 times, phosphorus up to 20 times, and potassium up to 14 times, and even more.

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Testing of Regional Wind Erosion Models For Environmental Auditing

J. F. Leys, Centre for Natural Resources, Dept Land Water Conservation, Gunnedah, NSW 2380, Australia (Email: jleys@dlwc.nsw.gov.au)

G. H. McTainsh, Australian School of Environmental Studies, Griffith University, Brisbane Queensland, 4111, Australia (Email: g.mctainsh@mailbox.gu.edu.au)

Y. Shao, Dept of Physics and Materials Science, City University of Hong Kong, SAR, PRC (Email: apyshao@cityu.edu.hk)

K. Tews, Australian School of Environmental Studies, Griffith University, Brisbane Queensland, 4111, Australia (Email: k.tewes@mailbox.gu.edu.au)

Introduction

In 1999, a community-government partnership in natural resource management was established in New South Wales, Australia. Catchment Management Boards (CMB) have been established with members drawn from representatives of the community, industry and government. The objective of the 18 CMB is to enhance the capacity of total catchment management to substantially improve the quality and sustainability of our state's natural resources and environment. Each of the CMB has developed a draft Blueprint to ensure the health of the landscape is improved by meeting key targets. Wind erosion monitoring is an integral part of the targets in the drier regions of the state because it monitors the performance of several of the key catchment targets including soil ground cover, soil health, and identification of areas that require regeneration and rehabilitation.

Monitoring of wind erosion over south-east Australia was undertaken by two independent methods. The first method was an integrated wind erosion modelling system as outlined by Shao (2000); the second was a dust visibility index based on Bureau of Meteorology (BOM) visibility observations.

This paper presents a methodology for validating regional scale models like the Integrated Wind Erosion Modelling System (IWEMS) against observational data. An evaluation of the process is given discussing its application to environmental auditing.

Materials and Methods

The area under study is south-east Australia. One year (1999) was selected, as it was an active wind erosion year in the study.

Integrated wind erosion modelling system

Full details of IWEMS can be found in Shao (2000). In brief, the system contains three components 1) an atmospheric-prediction model, together with land-surface model, 2) a wind erosion model, and 3) a geographic information database. The
computation of friction velocity ($u^*$) is via the Monin-Obukhov similarity theory, which depends on near surface wind speed, surface roughness and atmospheric boundary layer stability. The calculation of threshold friction velocity ($u^*_{t}$) requires information of soil type, vegetation and soil wetness where $u^*_{t} = u^*_{t\text{ (ideal)}} \times \text{modification factors}$. The modification factors are correction functions for surface roughness, soil wetness and other factors. Soil wetness is determined using a land surface parameterisation scheme (ALSIS, Atmosphere and Land Surface Interaction Scheme). The vegetation type is based on data from the Australian Resource Data and the leaf area index is derived from satellite remote sensing data of NDVI. The IWEMS system is first run over the Australian continent with a 50 km resolution. The resolution for the NSW region is 10 km. The increased resolution is achieved through a nesting procedure for both the atmospheric model and the land surface wind erosion model.

**Dust Visibility Index**

Meteorological records of dust event occurrence and intensity were obtained from the Bureau of Meteorology. Previous work has used dust event frequency (McTainsh and Pitblado, 1987) and a dust storm index (McTainsh et al., 2001, McTainsh and Tews, 1999) to measure spatial and temporal aspects of wind erosion. One of the main weaknesses of these data is the inconsistencies in meteorological observer records of the various dust event codes. The DVI overcomes this problem by using the visibility reduction at the time of the event (which is closely related to dust concentration) as the measure of wind erosion intensity, rather than the event codes. An example is given in Table 1.

Table 1. Example of weather observation records

<table>
<thead>
<tr>
<th>Location</th>
<th>Station</th>
<th>Long</th>
<th>Lat</th>
<th>Date</th>
<th>Time</th>
<th>Pres</th>
<th>Past</th>
<th>WindS (km/h)</th>
<th>WindDir</th>
<th>Vis (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>RABBIT FLAT</td>
<td>15666</td>
<td>130.0</td>
<td>48</td>
<td>20.182</td>
<td>-</td>
<td>2000090</td>
<td>1</td>
<td>7</td>
<td>33.5</td>
<td>220</td>
</tr>
<tr>
<td>BIRDSVILLE</td>
<td>38002</td>
<td>139.3</td>
<td>86</td>
<td>25.900</td>
<td>-</td>
<td>2000090</td>
<td>1</td>
<td>7</td>
<td>42.5</td>
<td>360</td>
</tr>
<tr>
<td>ALICE SPRINGS</td>
<td>15590</td>
<td>133.8</td>
<td>78</td>
<td>23.796</td>
<td>-</td>
<td>2000090</td>
<td>1</td>
<td>6</td>
<td>31.3</td>
<td>220</td>
</tr>
</tbody>
</table>

DVI was calculated using Equation 1.

\[
DVI = 0.60014 - 0.13032 \log_{10}(\text{visibility km})
\]

(1)

Equation 1 is derived from the power relationship between visibility and dust concentration measured at distance from source with a high volume air sampler (Tews 1996). The visibility / dust concentration equation was normalised to account for visibilities from 0.0456km (DVI of 1) to 100 km (DVI of 0) in line with Equation 1. DVI readings were summed for each month to indicate regions of dust activity.
Results and Discussion

The output from both methods for one month is shown in Figure 1. There is basic agreement between the methods as seen in Figure 1. The spatial resolution of each method and an understanding of what each method is measuring can account for differences in the spatial location of the dust activity. Differences in what each approach measures adds to the difficulty of a direct comparison, i.e., IWEMS predicts saltation activity with relatively high resolution (10 km pixel) and DVI measures the dust in the air at some point, or down wind of, the erosion source area at a lower spatial resolution; represented by the dots in the Figure 1b.

The results indicate that IWEMS is working well in the western (drier – rainfall decreases away from the eastern coast) areas but does not pick up the activity in the central part of the state.

Figure 1. Wind erosion distribution over SE Australia for February 1999 with predictions from (a) IWEMS and (b) DVI observations

What to measure and at what sampling interval?

The accuracy of any method will largely depend on these questions provided the scheme is sound. In the case of IWEMS, even with today’s computer/information systems it is still a complex issue to handle all the data. Therefore, those parameters that do not change rapidly over time (vegetation structure, overall roughness length, soil particle-size, salt concentration and crust strength) do not have to be updated very often. Other factors like surface roughness length, fraction of erodible area and frontal area index, need to be updated more often. Finally, wind speed and soil moisture need to be updated on short time steps. Achieving all the above becomes increasingly difficult as the update period decreases. Using remote sensing for groundcover has been undertaken using monthly NDVI. NDVI relies on changes in the greenness index, which is not always linked to cover in semiarid areas and may result in a lower cover level being assigned. Despite this, Shao and Leslie (1997) have successfully used this approach. The wind erosion model used (WEAM, Shao et al. 1996) is very sensitive to soil moisture. This parameter should be estimated at the highest resolution as should wind speed.

In the case of DVI, what to measure has been limited by the availability of data in the BOM database. The biggest limitations for the database are spatial distribution of
the stations and temporal resolution of recordings at stations (1 to 12 observations a day). As spatial distribution increases, temporal resolution decreases.

**Gaps in knowledge?**

At the regional scale, there are two competing factors, scale and process. While a good understanding of process might be available, implementing it with the required data at the correct scale is the challenge. For example, a description of the surface erodible fraction is a basic parameter for input to wind erosion models. The specification of this parameter for a 5km resolution region is not currently possible using remote sensing methods. Therefore, its substitution with particle-size data that does not change with time highlights this dilemma. Future work that estimates those factors that change rapidly in time are a priority for regional scale soil models.

**Conclusions**

Two methods have been used to assess the extent and magnitude of wind erosion in south-east Australia. There is general agreement between the modelled and the observed patterns. Differences in what each approach measures adds to the difficulty of a direct comparison, i.e., IWEMS predicts saltation activity and DVI dust transport down with of the source area. Despite this, each approach offers a tool for environmental auditing. Spatially, regions of wind erosion activity can be identified, and temporally, days (down to hours) can be identified when the activity occurred. The IWEMS also helps to explain why the erosion event occurred, e.g., high wind or low groundcover. The DVR offers a method for validating the outputs of IWEMS.

The challenge for regional wind erosion modelling is match the scale and the temporal frequency of data collection to the models requirements.

**References**


Field studies of wind erosion intensity and its variability under different vegetated sandy grasslands, Inner Mongolia

Feng-Rui Li, Hua Zhang, Tong-Hui Zhang, Yu-Lin Li, Cold and Arid Regions Environmental and Engineering Research Institute, Chinese Academy of Sciences, 260 Donggang West Road, Lanzhou 730000, PR China

Abstract

Field measurements were made with sand samplers during the two wind storms to investigate aeolian intensity and its variability for the mobile, semi-mobile, semi-fixed and fixed sand lands with different vegetation conditions.

The results showed a considerable difference in the intensity of wind erosion among the four sand lands for both measurements. On the May 15 measurement, the total windblown sand transport rate within the 20 cm height above the surface was 83.1 g cm\(^{-2}\) h\(^{-1}\) in the mobile dune, being 2.1-fold, 9.2-fold and 33.9-fold higher than the semi-mobile, semi-fixed and fixed sand lands. On the May 17 measurement, the total windblown sand transport rate was 105.7 g cm\(^{-2}\) h\(^{-1}\) in the mobile dune, being 5.6-fold, 14.1-fold and 75.6-fold higher compared to the semi-mobile, semi-fixed and fixed sand lands. This difference was primarily attributed to contrasting surface properties of the four sand lands. Of the measured surface properties, surface roughness length and vegetation cover were much more closely related to the total windblown sand transport rate (accounting for 67% and 59% of the variation in this variable, respectively) than any other surface properties such as canopy height and soil hardness. This suggests that surface roughness length and vegetation cover are the most important factors affecting the wind erosion intensity. No linear or curvilinear relationship was found between surface soil water content and the total sand transport rate or sand flux rates at individual heights, probably due to small variability in surface soil water content among the four sand lands during the measurements.

Surface roughness length of the sand land is mainly governed by the cover of vegetation accounting for 78% of the variation in this variable. Soil hardness was the next dominant component of surface roughness length, accounting for almost 50% of the variation in surface roughness length. A multivariate predictive equation, developed by regressing the surface roughness length on vegetation cover and soil hardness is applicable to the exploration of the combined effect of vegetation cover and soil hardness on surface roughness, because the predicted values matched those from field measurements reasonably well, with an \( R^2 \) value of 0.88.
Wind Erosion in Marginal Mediterranean Dryland Areas, Khanasser Valley- a Case Study

Z. Massri, ICARDA, P.O.Box 5460, Aleppo, Syria (E-mail: Z.Masri@cgiar.org)

M. Zöbisch, AIT, P.O.Box 4, K. Luang, Pathumthani, Thai.12120, (E-mail: zoebisch@ait.ac.th)

A. Bruggeman, ICARDA, P.O.Box 5460, Aleppo, Syria (E-mail: A.Bruggeman@cgiar.org)

P. Hayek, ICARDA, P.O.Box 5460, Aleppo, Syria (E-mail: P.Hayek@cgiar.org)

M. Kardous, IRA, Al-Jarf- 4119, Médenine, Tunisia. (E-mail: mouldi.kardous@ira.rnrt.tn)

Introduction

Wind erosion in the marginal drylands of the Mediterranean region is a serious problem, through its direct influence on particularly vulnerable land. Drylands are extremely susceptible to wind erosion because soils tend to be dry, poorly structured and sparsely covered by vegetation. Evidence of wind erosion is widespread throughout the dry areas of West Asia and North Africa. A major limitation in these areas is the lack of adequate and reliable rainfall to support a sustainable, protective land cover against the erosive forces of the wind. The replacement of natural, drought-tolerant species with crop cultivation and over-grazing of rangelands by small ruminants, due to growing population pressure and food demand, further exacerbate the problem.

The Khanasser Valley, where this study was conducted, is located at the fringe of the Syrian steppe, 70 km southeast of Aleppo. The Valley lies between the hill ranges of Jebel Al Hass in the west and Jebel Shbeith in the east. The soils in the valley floor are fine and moderately textured dark-brown to brown Calcisols, Gypsisols, Leptisols and Cambisols (Louis Berger International, 1982). The soils of Jebel Al-Hass and Jebel Shbeith plateaus are Inceptisols. Annual rainfall, which occurs during October and May, is normally 200-250 mm. Most households practice a combination of crop production and livestock rearing. Rainfed farming, with barley as the dominant crop, occupies the major part of the arable land. Fields are grazed by sheep after harvesting, leaving a bare disturbed soil surface, with particles loosened by the trampling of the sheep.

The purpose of this study was to quantify fluxes of wind-blown soil particles in the Valley and to assess the factors that affect the wind erosion process.

Material and Methods

During the dry summer months in 1998-2001, fluxes of soil blown particles were measured with BSNE samplers at two locations in the Khanasser Valley. One site was permanently located close to Um-Mial village on the plateau of Jebel Shbeith. This site is sparsely covered with natural vegetation. The other site was located in a rainfed barley field, after harvesting and grazing, each year at a different site in the valley.
bottom. Automated climate stations, which recorded hourly wind parameters at 2-m height, were located on the eastern plateau at Um-Mial and in the southern part of the valley bottom in Qurbatieh.

Each field was set up with 17 BSNE sampling clusters, constructed following the design of Fryrear (1986). The clusters were arranged in a radial setup covering the eight main compass directions. Radial distances from the center location are 50 and 100 m. A cluster consists of five samplers, each attached to a pivoting wind vane, mounted at 0.05, 0.1, 0.2, 0.5, and 1.0-m height, measured from the ground to the center of the inlet. The width and height of the sampler inlet areas are 20x10, 20x20, 30x20, 20x50, and 20x50 mm, for the respective heights. The trapped airborne fractions were collected on a weekly basis.

Analogous to Fryrear et al. (1998), a wind factor was computed for hourly wind speeds, assuming a threshold velocity of 5 m/s. The percentage organic matter (OM) and water-stable aggregates retained on a 0.5-mm sieve (WSA) were measured by standard methods (Ryan et al., 2001) for all five sites. The soil erodible fraction (EF), the soil crust factor (SCF), soil roughness (RR), and soil cover (SC) were computed according to procedures presented by Fryrear et al. (1998). The last two factors were visually estimated in the field. A second soil erodibility index (R) was calculated as described by Shiyatyy et al. (1972).

**Results and Discussion**

For each site, the total mass of trapped material was computed for the complete sampling period. The material trapped at the different heights was computed as an average of the 17 samplers. Fryrear and Saleh (1993) use a two-step model with a power function to describe the vertical distribution of suspended materials and an exponential equation to describe the distribution caused by saltation and creep processes. Power functions provided a good fit for the relation between the trapped material and the sampler height for all eight trials. Considering that the bottom of the inlet of the first sampler is located 4 cm above the surface, it is likely that the trapped fractions mainly represent the suspended fraction.

The absolute difference between the total mass for the 5 to 100 cm height integrated by a power function and the mass computed by a stepwise integration averaged 2.5%. For all four seasons, the airborne mass at the Um-Mial the site, located in the degraded natural vegetation of the plateau, was substantially lower than the airborne mass of the harvested and grazed barley fields in the valley floor (Table 1).
Table 1. The soil mass flux at the study sites during four summers (1998-2001).

<table>
<thead>
<tr>
<th>Site</th>
<th>Year</th>
<th>Sampling period</th>
<th>Soil mass flux at 5 heights</th>
<th>Total mass flux g/cm² width/d</th>
</tr>
</thead>
<tbody>
<tr>
<td>Um-Mial</td>
<td>1998</td>
<td>84 days</td>
<td>0.42 0.21 0.11 0.05 0.03</td>
<td>0.084</td>
</tr>
<tr>
<td>Qurbatieh</td>
<td>1998</td>
<td>84 days</td>
<td>0.58 0.36 0.26 0.19 0.16</td>
<td>0.248</td>
</tr>
<tr>
<td>Um-Mial</td>
<td>1999</td>
<td>82 days</td>
<td>0.18 0.07 0.04 0.03 0.04</td>
<td>0.051</td>
</tr>
<tr>
<td>Mgherat</td>
<td>1999</td>
<td>90 days</td>
<td>1.76 0.80 0.49 0.32 0.23</td>
<td>0.424</td>
</tr>
<tr>
<td>Um-Mial</td>
<td>2000</td>
<td>90 days</td>
<td>0.49 0.27 0.22 0.15 0.13</td>
<td>0.183</td>
</tr>
<tr>
<td>Serdah</td>
<td>2000</td>
<td>88 days</td>
<td>0.91 0.44 0.36 0.22 0.21</td>
<td>0.237</td>
</tr>
<tr>
<td>Um-Mial</td>
<td>2001</td>
<td>100 days</td>
<td>0.15 0.07 0.04 0.02 0.02</td>
<td>0.031</td>
</tr>
<tr>
<td>Rashadieh</td>
<td>2001</td>
<td>98 days</td>
<td>0.98 0.61 0.34 0.18 0.15</td>
<td>0.247</td>
</tr>
</tbody>
</table>

Average wind speeds for the four observed sampling periods varied between 3.2 and 3.9 m/s in Qurbatieh and 3.7 and 4.8 m/s in Um-Mial. The weekly sampling does not allow the establishment of direct relationships between eroded material and wind parameters, but evidence of the effect of the wind was clear for extreme events. For instance, in Um-Mial 30% of the total soil loss for the 2000 season was trapped during one weekly event. The average wind speed, average daily maximum wind speed, and the wind factor for this week were 31%, 21%, and 76% higher than the average values of these parameters for the complete season.

The airborne soil mass was negatively correlated with the organic matter content (-0.87), water stable aggregates (-0.88), soil cover (-0.76), and soil crust factor (-0.81). Correlations between the erodibility indices and the soil loss were small, 0.30 for EF and 0.31 for R. It is evident that the improvement of organic matter, soil structure, and soil cover properties can substantially reduce wind erosion.

Airborne fraction quantities decreased with height above the soil surface, while the particle-size distribution became more skewed towards the finer fractions (Fig. 1). Organic matter, available phosphorus, total nitrogen, and exchangeable potassium quantities varied considerably among sites and among heights. The enrichment ratios of organic matter and total nutrients of the trapped material versus the parent soil increased with height above the soil surface. Ratios varied between 0.8 and 3.
Figure 1. Relation of the particle-size distribution of the trapped airborne materials (g/cm²) and organic matter enrichment ratio (O.M. ratio) with height for Um-Mial site, 1999.

Conclusions

Results of four seasons of wind erosion research in Khanasser Valley, Syria, indicated that wind erosion is a serious problem in sparsely covered dryland environments. The physically removed wind-blown mass consisted of the lighter soil constituents such as organic matter, clay, and silt. The eroded material had generally higher nutrient contents than the parent soil. Nutrient loss is one of the major wind erosion hazards. Losses of the most fertile part of the soil is reducing the productivity of already poor soils and is threatening the sustainability of the natural resources in the dry areas.

Erosion mass fluxes were substantially higher in the harvested and grazed barley fields of the valley floor than in the degraded natural vegetation on the plateau. The loosely structured soils in the valley floor were more susceptible to wind erosion. Even the sparse natural vegetation and soil cover on the plateau reduced the airborne materials. The relations between eroded soil mass and the wind erosion indices indicated that improved soil and land management can reduce wind erosion. Future research work will focus on the introduction and evaluation of natural resource management practices for reducing the effect of wind erosion in fragile dryland ecosystems.

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McTainsh, G.H., Australian School of Environmental Studies, Griffith University, Brisbane, Queensland, 4111, Australia.(E-mail: g.mctainsh@mailbox.gu.edu.au)
Love, B.M., Australian School of Environmental Studies, Griffith University, Brisbane, Queensland, 4111, Australia.
Leys, J.F., Centre for Natural Resources, Department of Land and Water Conservation, Gunnedah, NSW, 2380, Australia. (E-mail: jleys@dlwc.nsw.gov.au)
Strong, C., Australian School of Environmental Studies, Griffith University, Brisbane, Queensland, 4111, Australia.

Introduction

Long term records of field-measured wind erosion rates are rare worldwide (Offer and Goossens, 2001). This is a deficiency in our research knowledge, particularly given the highly episodic nature of wind erosion processes. While the more frequently available short records of erosion rates provide valuable information, they may not provide a representative picture of the environmental factors influencing wind erosion at a location. Data are presented here from a long term field measurement project in the Channel Country of western Queensland, which started in 1994 and is ongoing. The early results from this project (1994-1997) were presented at ICAR 4 (McTainsh et al., 1999). In the present paper we present a 7 year record (1994-2000) of wind erosion rates (dust flux) and examine temporal changes in the erodibility of the same 3 land types in response to: wind conditions, rainfall, vegetation, soil surface conditions and flooding.

Materials and Methods

The field site is the remote Diamantina National Park in the Channel Country of Western Queensland. The Channel Country is an extensive floodplain channel system of the inland-draining rivers of the Lake Eyre Basin. Diamantina National Park lies within an active wind erosion region (McTainsh, 1997) and contains a variety of land types within a relatively small area, which are widespread in the Australian arid zone. The three major land types selected for this study are: claypans, dunes and downs. The claypans occupy the high floodplain of the Diamantina River, which is flooded when stage heights exceed 4.5m, which occurs every second year on average (1917-2001). The dunes are source-bordering linear dunes located on the western sides of the main rivers and the downs are the expansive very gently undulating interfluve areas between the main rivers.

The passive sediment samplers used here were designed by Fryrear (1985), and modified with a rain hood. Samples were collected at three heights (0.07m, 0.5m and 2m). Shao et al (1993) found that wind vane samplers are 90%/5% efficient for highly aggregated sandy loam soils in SE Australia, but as most of the aeolian fines in the Channel Country soils are particulate, these samplers are unlikely to be as efficient there.
Streamwise sediment fluxes were calculated from sediment concentration/sampling height curves. The relationship between sediment concentration and height has frequently been shown to be a power function for suspended sediments (Gillette and Goodwin, 1974). Five sampler arrays have been used in past studies (Leys and McTainsh, 1994) but for practical reasons in this study, the number of sampling heights were kept to three. A comparison of the dust flux curves from two arrays revealed <2% difference.

The sediment concentration and height relationship was extrapolated downwards to 0.01m, which is not as low as Nickling (1978) (0.001m), but low enough to provide a good estimate of saltation flux without any statistically-derived distortion of the loads. Dust flux (>0.5m) is the measure of wind erosion used here as it provides an estimate of the sediment lost from the immediate area. Erodibility is measured at a broad scale using the same Land type Erodibility Index (LEI) as (McTainsh et al., 1999). The LEI measures dust flux per unit of wind energy, and is calculated by dividing dust flux (F) by the mean wind speed above threshold (6m/sec) (U>t) for each measurement period.

Results and Discussion

Averaging the dust fluxes for each land type for 7 years, it is apparent that the claypans are clearly eroding at the highest rate, followed by the dunes, then the downs (Table 1). This trend is the same as for the 1994-1997, however the differences between the land types have increased.

<table>
<thead>
<tr>
<th>Land Type</th>
<th>Mean dust flux (gm⁻¹s⁻¹)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Claypans</td>
<td>0.0047</td>
</tr>
<tr>
<td>Dunes</td>
<td>0.0015</td>
</tr>
<tr>
<td>Downs</td>
<td>0.0004</td>
</tr>
</tbody>
</table>

Table 1: Mean dust flux (1994-2000) on 3 land types in the Channel Country.

Figure 1 shows yearly trends in dust flux on each land type in relation to rainfall. While annual rainfall shows a clear upward trend from 1994, the dust flux responses on each land type differ considerably. The dust fluxes on the downs are sensitive to rainfall; showing a clear negative relationship, whereas the claypans show a complex relationship with rainfall. The dunes show an initial decrease in dust flux, in common with the downs, then appear to track the changes in claypan dust flux as rainfall increases. The correlation between dune and claypan fluxes may be an artefact arising from the proximity of the dune site to the claypan, which resulted in some claypan dust being collected by the dune wind vane sampler.
These results indicate that the generally accepted negative relationship between wind erosion rates and rainfall from broadscale studies (Goudie, 1983, McTainsh and Leys, 1993) describes the situation on the downs and dunes land types (assuming that the artefact is real), because erosion on these land types is vegetation-limited. The negative relationship cannot, however be applied to claypans, which are rarely vegetated. Evidently, other factors are having a strong influence upon wind erosion on claypans.

The temporal trend of dust flux on the claypan in Figure 1 can be resolved into three periods: Period 1 (1994-1996) - decreasing dust fluxes, Period 2 (1997-1998) - increasing dust fluxes, and Period 3 (1999-2000) - decreasing dust fluxes.

   From 1994 to 1996 claypan dust fluxes decreased as annual rainfall increased. This trend appears to reflect increasing soil moisture levels.

   Dust fluxes increased sharply in this period. McTainsh et al (1999) attributed the 1997 increase to the supply of fresh erodible sediment as a result of the flooding of the claypan in early 1997. These high dust fluxes also occurred at low wind speeds, indicating that the erodibility of the claypan soils was unusually high. McTainsh et al (1999) quantified this erodibility change using the land type erodibility index (LEI). The further large increase in dust flux in 1998 cannot be so simply explained by prior flooding in 1998, and there was no significant decrease in annual rainfall. The increase may be due to more erosive conditions this year, but wind speed data are not yet available to support this assertion, or to perform an LEI analysis.

In 1999 the dust fluxes on the claypans dropped dramatically, even though annual rainfall remained high, and there was a large flood (the second largest on record) in January, 1999 which provided another fresh load of erodible sediment. Significantly, this large flood and follow-up rains brought vegetation (for the first time in at least 10 years), which may have been sufficient to significantly reduce LEI. The absence of vegetation on the claypans up to this time had been due to hypersalinity of the claypan soils, however the combination of a large flood and higher than average rainfall reduced the salinity levels in the topsoils (Mostert, 1999) sufficiently to allow initial vegetation growth.

In 2000 dust fluxes dropped to a long term low. Three significant floods occurred between January and March, 2000 and rainfall increased again. This combination of events resulted in a widespread coverage of vegetation on claypan, in response to reduced salinity levels in the claypan topsoils. The erodibility of the claypan therefore became vegetation-limited in period 3, in a similar way to the dunes and downs.

This positive shift in wind erosion conditions on the claypans may be a turning point in the history of the claypans in the area, and may represent the beginning of rehabilitation phase. The reason for this optimism is that the claypan vegetation has the best chance (in the past 200 years), of enduring as grazing pressure has been significantly reduced by the removal of cattle from the region in 1997, soon after the National Park was established.

Conclusions

Long term dust fluxes from claypans on the Diamantina River floodplain are more than three times higher than from dunes and downs in the same region. Wind erosion on the dunes and downs is negatively related to rainfall through vegetation cover effects, but on the claypans wind erosion is influenced by a more complex set of factors. Flooding of these high floodplains can both increase and decrease wind erosion rates. Deposition of fine sandy alluvium, which is highly erodible to wind, can significantly increase dust fluxes (as in 1997), but when major floods and high rainfalls converge (as in 1999 and 2000) dust fluxes are dramatically reduced. This convergence of events reduced the normally high salinity levels in claypan topsoils and stimulated vegetation growth, which in turn provided protection from erosive winds. 1999 may turn out to be a pivotal year in the history of claypan erosion in Diamantina National Park, as for the first time in 200 years this landscape is free from the grazing pressure of cattle. This may be the beginning of a natural claypan rehabilitation phase? A continuation of the present long term field measurement project will answer that question.

References


Offer and Goossens, 2001 Ten years of aeolian dust dynamics in a desert region (Negev desert, Israel): analysis of airborne dust concentration, dust accumulation and high magnitude dust events. *Journal of Arid Environments*, 47, 211-249.

Effect of Surface Conditions on PM$_{10}$ Dust Emissions, Owens Lake, CA

W.G. Nickling, Wind Erosion Laboratory, Department of Geography, University of Guelph, Guelph, ON, N1G 2W1, (e-mail: nickling@uoguelph.ca)

L.J. Brown, Wind Erosion Laboratory, Department of Geography, University of Guelph, Guelph, ON, N1G 2W1, (e-mail: laura@uoguelph.ca)

The diversion of water from the Owens River by the City of Los Angeles resulted in the drying up of Owens Lake by 1926. The present dry lakebed is subject to frequent dust storms from October to May and represents one of the largest single sources of airborne particulate matter in the Western hemisphere (Gill, 1996).

To evaluate the temporal and spatial variability of PM$_{10}$ dust emission wind tunnel tests were undertaken at 30 sites, primarily located in the Central Area of the lakebed with a few sites in the Southern portion. Tests at each site consisted of 4 components: a) soil sample collection, b) determination of threshold shear velocity, c) measurement of PM$_{10}$ dust emissions without sand feed, followed by d) measurement of PM$_{10}$ dust emissions with sand feed.

Vertical dust emissions (F) for the lake sites without-feed ranged from approximately $3.29 \times 10^{-1}$ to $5.12 \times 10^{4}$ $\mu g/m^2$s and from $7.79$ to $5.46 \times 10^{4}$ $\mu g/m^2$s for the with-feed runs. In general, there were somewhat higher F values for the with-feed runs reflecting the importance of sand bombardment as a principal mechanism for dust entrainment. Although much higher F values were expected for the with-feed runs the less than expected increase is likely attributable to the very hard surface crusts and lack of particulates available for entrainment at the surface for many of the sites examined during the wind tunnel tests.

Recent research indicates that the ratio F/q for a given surface is a direct function of the binding energies that hold particles at the surface and of the kinetic energy of impacting saltation particles as determined by particle mass. The F/q ratio ranged from $6.33 \times 10^{-8}$ m$^{-1}$ to $7.88 \times 10^{-4}$ m$^{-1}$ with a median value of $1.21 \times 10^{-5}$ m$^{-1}$ for the without-feed lake site tests. A very similar range was found for the with-feed cases ($6.61 \times 10^{-8}$ m$^{-1}$ to $2.18 \times 10^{-4}$ m$^{-1}$), although the median value was somewhat lower ($5.52 \times 10^{-6}$ m$^{-1}$). In general the F/q ratios determined from the wind tunnel testing are smaller for similar $u^*$ values than the single ratio derived by Gillette et al. (1997) from tower and sand trap measurements but are similar to those found during the 2000 wind tunnel tests and are also consistent with recent work by Alfaro et al. (1997, 1998).

The wind tunnel test results indicate the importance of changing surface conditions on dust emissions as a result of precipitation events and high evaporation rates associated with rapid drying of the surface and the formation of very hard-cemented crusts. This dramatic change in surface conditions as the season progressed resulted in a noticeable decrease in the quantity of entrainable sediment and significantly lower and somewhat irregular emissions and associated dust concentrations during the wind tunnel tests irrespective of soil type.
Wind erosion in a small catchment of grazing area in Northern Burkina Faso: influence of surface features

J.L. Rajot, IRD UR 049 LISA, UMR-CNRS 7583 Université de Paris 12 - 61, avenue du Général de Gaulle, 94010 Créteil, France (E-mail: rajot@lisa.univ-paris12.fr)

O. Ribolzi, IRD UR 049 - 956, avenue Agostino Neto BP182, Ouagadougou 01, Burkina Faso (E-mail:ribolzi@ird.bf)

J.P. Thiebaux, IRD UR 049 - 956, avenue Agostino Neto BP182, Ouagadougou 01, Burkina Faso (E-mail:thiebaux@ird.bf)

Introduction

Wind erosion studies in the Sahel mainly relate to cultivated field (e.g. Bielders et al. 2001), but there is a lack of information about natural areas where grazing is major land use. In a context of increasing population these areas are reported to degrade due to drought enhanced by overgrazing. Such a degradation seems obvious from observations: Typically the non cultivated areas show a patchwork of bare crusted soil, in some cases covered with gravel originating from overlaying laterite, and sandy soils where perennial and annual vegetation grow. These bare or graveled areas are reported as degraded. Nevertheless Casenave and Valentin (1989) suggested that a dynamic may exist leading to extension or regression of bare crusted/graveled areas according to drought and rainier years respectively. The purpose of this paper is to assess the aeolian dynamics of such typical Sahelian surface features by direct measurements of wind erosion flux.

Methods

Study area. The study area is located in the north of Burkina Faso (UTM30, WGS84, 809847 m East, 155093 m North), near Dori, 250 km North East of Ouagadougou (Fig. 1). The climate is of the Sahelian type, with a long dry season and a short rainy season from June to September (Mean rainfall 512 mm). Wind erosion measurements are performed on a small catchment because this study is a part of a larger program aiming at studying for the same surface wind and water erosions. The surface features of the catchment was mapped from aerial photographs and direct observations in the field according to the classification of Casenave and Valentin (1989, 1992). These observations allowed to select three sub-areas homogeneous in term of surface feature combination.

Wind erosion measurements. Measurements were performed during the 2001 rainy season between June 1 and October15. Wind blown sediment fluxes were obtained by using 50 masts equipped with 3 BSNE sand catchers (Fryrear, 1986) located at heights 0.05, 0.15 and 0.3 m. The masts were placed 1) approximately every 20-m on the boundaries of the sub-areas and 2) along a transect in the western side of the
catchment where the surface feature variability is higher (Fig. 1). Wind blown sediments caught in BSNE were collected after each erosion event. The horizontal fluxes were calculated at each mast by integrating the sediment flux density profile between 0 and 0.4 m height.

The mass budgets within the selected areas were calculated by subtracting outgoing from incoming wind blown sediments. Wind speed and direction were measured at 2 m height using an automatic weather station. An acoustic saltation sensor (Saltiphone) recorded the period during which the fluxes were significant. Knowing that, it was possible to estimate the mean direction of wind during each storm event, and to determine for each event the upwind and downwind limits of sub-areas. The incoming and outgoing mass fluxes along these boundaries were then calculated by linear interpolation of sediment mass fluxes measured at each mast. The transect was oriented along the East / West direction (Fig. 1) which was assumed to correspond to the more intense wind erosion events. On the transect, the BSNE masts are setup at each major surface feature change. When erosive wind direction corresponds to the transect direction (95° ± 15°), it is possible to compute a budget by subtracting downwind from upwind fluxes and dividing the result by the distance between the two considered measurement location.

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**Results and Discussion**

**Surface features.** According to Casenave and Valentin (1992) classification four types of surface features can be pointed out on the catchment (Fig. 1): i) areas of sandy soil developed on aeolian sand deposits (less than 0.7 m thick) where annual vegetation, shrubs and trees grow and where the crust type is mainly drying (DRY), ii) large areas of bare clayey soil with typical erosion crust (ERO), iii) stretches of unvegetated sandy soil with erosion crust which always develop between the two former types (ERO/S), and iv) small sand fan without vegetation deposited by water (RUN) (Ribolzi et al., 2000). The first two surface features clearly dominate and are not randomly distributed on the catchment (Fig. 1): in the upstream part (#1), only DRY are represented, on the center of the catchment (#2) ERO is patched within dominant DRY (about 30 and 80 % respectively) whereas in the downstream part (#3) the reverse is observed i.e. small patches of DRY crust within ERO (about 20 and 80 % respectively).

**Figure 1.** Situation of the study area, map of soil surface features, location of BSNE clusters, and limits of the 3 sub-areas for which wind budget was calculated.
Wind blown sediment budget. Budgets were computed from a set of 21 flux measurements out of 34 for which the standard deviation of wind direction is smaller than 20°. These 21 wind erosion events represented more than 80% of the total flux recorded during the season and are assumed to be representative of the overall wind erosion (Rajot et al. 2002). Figure 2 shows that the budgets vary considerably according to the time and space.

Eighty five percent of the total sediment mass flux measured over the study period occur before the 15 July. Such a result is well known in the Sahel and can be related to higher wind intensity and to lower soil coverage by litter and vegetation at the onset of the rainy season (Rajot 2001).

Figure 2. Wind blown sediment mass budget (Mg ha⁻¹) cumulated over the study period for the total area and the 3 sub-areas selected because of their surface feature distribution.

Budget is almost systematically positive for the upstream sub-area (#1) whereas it is systematically negative for the downstream one (#3), amounting to about + 30 Mg ha⁻¹ and -20 Mg ha⁻¹ over the measurement period, respectively. Both erosion and deposition occurred in the center sub-area (#2), but budget remained negative over the measurement period (~7 Mg ha⁻¹). These different behaviors of the sub-areas led to an almost balanced budget (~ 2 Mg ha⁻¹) at the catchment scale.

High wind blown sediment deposition was also reported by Bielders et al. (2001) in fallow land in Niger which presented the same surface feature as sub-area 1 (dry crust with annual and perennial vegetation). The deposition was ascribed to the high surface roughness of these areas. In Niger, sources of wind blown sediments were the pearl millet fields (Bielders et al., 2001). In this study, net wind erosion occurred on complex natural areas where all the different surface features encountered in the catchment are represented (Fig. 1).

The transect measurements performed across sub-areas 2 and 3 (Fig. 1) allow a better description of the processes occurring in relation with these surface features. Only 5 events met the required wind direction to be computed from the transect set up, but 2 of them are the more intense of the season. General trends appeared and can be summarized from the budget computed from the sum of these 5 events (Fig. 3). First of all, the transect revealed the high spatial variability of wind erosion at the meter scale. There is not a systematic behavior of the 2 main surface features in regards to budget:
erosion may occur on DRY surface and deposition may occur on ERO surface. Nevertheless, the larger deposition occurred on DRY surface whereas the more intense erosion occurred on ERO/S surfaces or area where such a surface is present (between 70 and 85 m), as well as on the RUN surface.

ERO/S surfaces are closely associated to DRY ones and develop on the same sandy soil. If one considers these 2 surfaces together (between 0 and 9 m and between 25 to 37 m) sediment budget is negative i.e. the small patches of sandy soil are currently submitted to net erosion as suggested by Casenave and Valentin (1989) during drought conditions.

The fact that net deposition may occur on ERO surfaces whereas sand deposits are not observed on it suggests that these sediments are mobilized by water erosion which often follows wind erosion in the Sahel (Rajot et al. 2002). Similarly the high susceptibility of RUN surfaces to wind erosion shows that water erosion produces sediment which are easily mobilized by wind erosion.

![Figure 3. Cumulated wind blown sediment budget for the 5 events with easterly winds (parallel to the transect orientation) versus distance from the west border of catchment. The various types of surface feature (see text for description) are indicated by different shades of gray on horizontal axis.](image)

**Conclusions**

Wind erosion of grazing areas in the Sahel is an highly variable phenomenon at least from the scale of the meter to that of the hundred of meters. Not only the type of surface feature plays a role in this variability, but also the size of these surfaces. In the study area, the small patches of DRY surfaces associated with ERO/S surfaces are currently submitted to intense wind erosion. Net erosion also occurs in DRY areas with patches of ERO surfaces, whereas only the largest DRY surfaces without ERO crust show a net deposition. These results suggest that the current situation of the grazing area rather corresponds to a general degradation of the environment which can be related to the current drought in the Sahel. Nevertheless, it appears that both wind and water erosions interact and must be taken into account to assess accurately the erosion in this zone.
References


The Yellow Lake Experiment

J. E. Stout, USDA-ARS, Lubbock, Texas 79415 (E-mail: jstout@lbk.ars.usda.gov)

Introduction

The term playa is often used to describe the flat-floored bottom of a closed basin that becomes at times a shallow lake (Blackwelder, 1931; Shaw & Thomas, 1989; Blank et al., 1999). On the Southern High Plains, there are two distinctly different types of playas – the small circular playas and the larger and more irregular saline playas (Reeves & Parry, 1969). Though both playa types appear to share many common features, the saline playas are not equivalent in either form or origin to the circular playas (Sabine & Holliday, 1995).

Circular playas lie on the Blackwater Draw formation well above the water table of the underlying Ogallala formation (Wood et al., 1992). Runoff that collects in the circular playas slowly drains through the soil to the underlying Ogallala aquifer flushing away salts in the process (Osterkamp & Wood, 1987). As a freshwater playa dries out; plants take root and limit aeolian deflation (Haukos & Smith, 1997).

Saline playas, on the other hand, are in close contact with the saturated zone of the Ogallala aquifer, which limits drainage (Holliday et al., 1996). A salt encrusted clay surface develops as surface runoff and groundwater discharge evaporates. High salinity levels prevent plant growth leaving the bare playa surface highly susceptible to aeolian deflation. During periods of aeolian activity, clay aggregates are dislodged and transported across the playa surface. Fine fractions become suspended and form dust plumes that may extend great distances. Coarse grains deposit near the partially vegetated playa margins where they form clay dunes, also referred to as fringing dunes or lunettes (Bowler, 1968; Reeves and Parry, 1969; Bowler, 1973). During periods of heavy rain, some of the dune sediments wash back onto the playa surface where they eventually dry out and become source material for future aeolian transport. This process sets up a continuous cycle of aeolian deflation and transport of playa sediments to the fringing dunes and fluvial erosion and transport of dune sediments back to the playa surface.

An attempt was made to study aspects of this dynamic system by measuring and monitoring some of the active physical processes that shape the saline playa environment.

Physical Setting

Located at the southern end of the Great Plains, the Southern High Plains, also known as the Llano Estacado, is an immense elevated plateau of approximately 78,000 km² (Reeves & Reeves, 1996). There are 21 large (>5 km²) topographically closed basins scattered across the Southern High Plains that contain an estimated 40 saline playas (Reeves, 1966; Sabine and Holliday, 1995; Wood and Sanford, 1995; Holliday, 1997). Yellow House Basin is the third largest closed basin on the Southern High Plains with a depression area of 74 km² (Reeves, 1966). It contains two saline playas — Yellow Lake and Illusion Lake (Fig. 1). This study focuses on Yellow Lake.

The playa surface at Yellow Lake is 4.4 km long by 0.9 km wide with a surface area of approximately 3.5 km². Partially vegetated clay dunes that extend to as high as 35 to 40 m above the playa surface line the east side of the playa. Although the water table fluctuates seasonally, it remains near the surface even when the playa is dry. At times, surface waters form a broad shallow
pool that occupies a limited fraction of the playa surface. The location of the shallow pool has been observed to slowly move about the flat lakebed as it is pushed by the wind. The portion of the playa surface not wetted by the shallow pool remains susceptible to aeolian deflation.

Figure 1. Map showing the location of Yellow Lake, the Southern High Plains (dark gray), and the Great Plains (light gray).

**Experiment**

Two sampling systems were installed on the playa surface at Yellow Lake. An acoustic water level sensor was placed near the center of the playa to record surface water depth. Another sampling system was placed near the foot of the fringing dunes on the east side of the playa to record climatic variables such as wind speed, wind direction, temperature, relative humidity, solar radiation, and precipitation. Saltation activity was measured with a piezoelectric saltation sensor mounted next to the meteorological tower.

During periods of active saltation, the piezoelectric transducer produces a signal that is used simply as an on-or-off indicator of saltation activity. Each pulse signal generated by each saltating grain that impacts the sensor is detected and if one or more impacts are detected during a given second then that second is recorded as one “saltation second” or one second of saltation activity. At the end of each hour the total number of saltation seconds are summed and output to final storage.

For each day of the experiment, the total number of saltation seconds were summed and then divided by 86,400 s to form a dimensionless value of “daily saltation activity.” Daily saltation activity expresses the fraction of time within a given day that active saltation was detected.

**Results and Discussion**

Daily saltation activity is plotted as a time series in Fig. 2. A seasonal reference frame is provided across the top of the graph. In addition, a seasonal summary of saltation activity, measured climatic factors, and water depth is presented in Table 1. The results provide a record of aeolian activity at Yellow Lake from December 1998 to June 2002.
Figure 2. Daily saltation activity measured at Yellow Lake, Texas.

Table 1. Seasonal summary for Yellow Lake, Texas.

<table>
<thead>
<tr>
<th>Season</th>
<th>Total Rainfall (mm)</th>
<th>Average Relative Humidity (%)</th>
<th>Average Water Depth (cm)</th>
<th>Maximum Water Depth (cm)</th>
<th>Average Wind Speed (m/s)</th>
<th>Total Saltation Seconds (s)</th>
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<td>3</td>
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<td>Spring 2002</td>
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<td>53</td>
<td>4</td>
<td>19</td>
<td>4.3</td>
<td>7,600</td>
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Saltation activity tends to peak during winter when dry conditions and moderately strong winds combine to produce ideal conditions for aeolian activity. This is especially true for winter 1999-2000 with a total of 306,671 saltation seconds. This value is nearly an order of magnitude
lager than any other period and this intense saltation activity is clearly visible in Fig. 2. Saltation activity is relatively low during the summer due to the unfavorable conditions of light winds, significant rainfall and high humidity. Moist conditions tend to increase threshold and this combined with light winds leads to highly intermittent saltation activity. Surprisingly, the playa surface is fairly inactive during spring when winds are the strongest. The primary reason is most likely due to significant spring rains that moistens the playa surface and essentially shuts down saltation activity despite strong winds.

Acknowledgments. I would like to thank James R. Golden for helping to design, construct, and maintain both sampling systems.

References


Particle-size characteristics of wind eroded sediments from disturbed biotic crusts in south east Australia

Craig Strong, Australian School of Environmental Studies, Griffith University, Brisbane, Queensland, 4111, Australia. (E-mail:craig.strong@mailbox.gu.edu.au)

John Leys, Department of Land and Water Conservation, Gunnedah Research Centre, Gunnedah, New South Wales, 2380, Australia. (E-mail: jleys@dlwc.nsw.gov.au)

Grant McTainsh, Australian School of Environmental Studies, Griffith University, Brisbane, Queensland, 4111, Australia. (E-mail:g.mctainsh@mailbox.gu.edu.au)

Introduction

The degrading effect of overgrazing on rangelands is well documented. Biotic crusts often stabilize the fragile soils of these grazed areas. Depending on soil characteristics and disturbance regimes, soil crust components include: mosses, lichens, algae and cyanobacteria (Belnap and Gillette, 1998). Biological crusts are susceptible to disturbance, especially in soils such as sands (Eldridge and Greene, 1994; Webb and Wilshire, 1983; Belnap and Gillette, 1997). Disturbance from compressional or shear forces acting on the vegetative components may result in the removal of biological propagules or burial. Such disturbance of propagules will decrease recolonization rates, and lower nitrogen and carbon levels, and result in the death of the organism. Whilst the impact of grazing pressure, in particular trampling, on these protective crusts is becoming better understood, the processes by which protection is offered is less well known. We examine how disturbance of these biotic crusts by grazing animals increases the erodibility of soil surfaces to wind erosion.

Materials and Methods

Experimentation was carried out on two soil types (sand and loam) in the Mallee Region of south-western New South Wales, Australia. Both soils are in areas with low grazing pressure and are classified as Tenosols (sand) and Hypercalcic Calcisols (loam) (Isbell, 1996). Two levels of crust disturbance were induced by simulated stock trampling with a constructed sheep foot roller (Leys and Eldridge, 1998). The disturbance levels created three surfaces of different erodibility: no disturbance, moderate and severe. Sediment fluxes were then measured using a portable field wind tunnel. A Coulter Multisizer measured particle-size characteristics of the eroded sediments.

Wind Tunnel

The New South Wales Department of Land and Water Conservation portable field wind tunnel used in this study is described by Raupach and Leys (1990). In brief, the tunnel is portable, open bottomed with a cross section of 0.9m x 1.2m and a working
section length of 7.5m. The tunnel has flow conditioning and a fully developed boundary layer. Wind velocity in the tunnel was set at 6m/s at a height of 0.15m and velocity maintained for one minute. The wind velocity was then increased in increments of 1.2m/s each minute until maximum velocity of 13m/s was reached. The eroded sediment was collected at each minute interval in a modified Bagnold vertically integrating trap of width 0.005m and height of 0.5m, described by Shao et al. (1993). Sediment flux (q) was determined from the weight of soil collected by the relationship \( q = \frac{m}{YT} \) and corrected for saltation overshoot (Leys et al. 1996).

**Particle-sizing technique**

Particle-size analysis (PSA) was performed using a Coulter Multisizer, on: surface soils and the eroded sediments from each disturbance surface. The Multisizer is an electrical sensing zone instrument (Lines, 1992) that counts and sizes particles and aggregates suspended in an electrically conductive liquid. The Multisizer provides high resolution results (256 size classes), with very good reproducibility and is one of few instruments capable of analyzing the very small quantities of material (McTainsh et al., 1997). To reduce errors arising from sample contamination with glass fibers (from the glass fiber filter papers), the exhaust discharge samples were analyzed using the method of Kiefert et al. (1992).

**Results and Discussion**

Sediment fluxes increased with disturbance intensity on both soils (Figure 1).

![Figure 1: Sediment flux of increasing disturbance on two soils: sand and loam.](image)

The inherent properties of each soil influence the quantity and quality of sediments removed. Particle-size distributions of the wind-eroded sediments from both soils provide evidence of wind-induced entrainment of crusted surfaces. For the undisturbed sandy soils, the biotic crust resisted saltation, offering protection to the underlying soil. However, unprotected areas between the crust cover released material with the saltating grains. This ‘puff’ of material has a particle-size similar to the parent soil. In comparison, trampling results in a coarse mode appearing with its size dependent on the level of disturbance. Moderate disturbance resulted in a mode of 420µm, while severe disturbance resulted in a 300µm mode. The severely disturbed
surface produced a winnowed fine fraction, becoming finer (one 1/4φ class) with increased level of disturbance.

The higher clay content of the loamy soil increased the overall stability of the soil enabling colonisation of more complex morphological groups of crust taxa such as squamulose and foliose lichens. These in turn increase surface roughness and soil aggregation, enhancing soil protection from wind erosion. The large surface area of mosses and squamulose lichens acts as an efficient trap, capturing silt and larger quartz grains. This lag material was removed with the undisturbed treatment. Unlike the sandy soil, the intercrust region of the loam soil was resistant to saltation due to its greater clay content. Trampling of the loam soil resulted in two sediment populations within the eroded sediments, a winnowed population sourced from the freshly exposed soil and a coarser population containing organic matter and biotic crust pieces.

Conclusions

Biological crusts offer protection to soil against wind erosion, but stock trampling reduces this protection. Sediment fluxes on two soil types increased exponentially with crust disturbance and the particle-size characteristics of the eroded sediments are dependent upon soil type and degree of stock trampling.

Reference List


Soil Roughness Degradation and Crop Residue Decomposition: Measurement and Simulation

S. J. van Donk, USDA-ARS, Manhattan, KS, 66506, (E-mail: sdonk@weru.ksu.edu)
E. L. Skidmore, USDA-ARS, Manhattan, KS, 66506 (E-mail: skidmore@weru.ksu.edu)

Introduction

The United States Department of Agriculture (USDA) Agricultural Research Service (ARS) Wind Erosion Research Unit (WERU) in Manhattan, Kansas, USA is developing a process-based Wind Erosion Prediction System (WEPS), that is able to simulate wind erosion and dust emission for different management scenarios (Hagen, 1991; Wagner, 2001). WEPS consists of a number of submodels, including models for soil roughness degradation (Hagen et al., 1995) and crop residue decomposition (Steiner et al., 1995). Accurate prediction of wind erosion depends greatly on reliable simulations by these submodels. The objective of this study was to compare roughness degradation and residue decomposition, measured on a farmer’s field near Burlington, Colorado, USA, with those simulated by WEPS.

Methods

Field measurements

Measurements, used in the present study, were taken in the context of a larger wind erosion field experiment, designed for testing the WEPS erosion submodel (van Donk and Skidmore, 2001). A field was selected 17 km south of Burlington, Colorado, USA (39.13 N, 102.30 W, elevation = 1292 m). A corn crop was grown on the field in the summer of 1998 and a sunflower crop in the summer of 1999. Wheat was planted on the 1720 m by 810 m field with a 305 mm row spacing on 29 August 2000. When we started field measurements in December 2000, the crop residue on the field was mainly corn. Measurements were taken at 15 locations on the NW corner (600 m by 415 m) of the field, on three dates: 19 December 2000, 8 March 2001, and 12 April 2001.

Ridge height for each field location was calculated as the average of the depths of four adjacent furrows, measured using a straight edge and a measuring tape. Soil random roughness was measured using a pinmeter (Wagner and Yu, 1991), positioned in parallel with a ridge. Pinmeter photographs were taken using a digital camera and analyzed using SigmaScan Pro (SPSS Inc., Chicago, IL) software. Random roughness was calculated as the standard deviation of pin positions (Allmaras, 1966), which was corrected for trends, i.e. downward or upward trends of pin positions from one side of the pinmeter to other. Such trends increase the standard deviation without contributing to soil roughness.

Above ground flat corn residue was collected within a rectangular frame of 305 mm by 584 mm. Dirt was removed from the residue using various hand tools. The residue was air dried and weighed in the laboratory. Flat residue cover was measured using a 15.2 m (50 ft) long measuring tape, counting the foot marks that covered pieces of residue. No standing residue was present on the field.
A weather station was centrally located on the NW corner of the field. Measurements included air temperature at 2.0 m (CS500, Campbell Scientific, Logan, UT1) and precipitation using a tipping bucket rain gauge (6010, Qualimetrics). The sensors were calibrated before being deployed in the field. Data were measured and recorded with a data logger (CR10X) and a solid state multiplexer (25AMT) from Campbell Scientific. Sensors were sampled every 10 s and data were recorded for 15 minute periods.

**Simulations**

The amount of precipitation is critical, especially for the simulation of the degradation of ridge height and random roughness. During the snowy period from 19 December 2000 through 8 March 2001, precipitation measured on our field differed greatly from that measured at nearby stations (Table 1). For the period from 9 March to 12 April 2001, when precipitation came mostly in the form of rain, the three stations agreed well with each other. The windswept Great Plains is a very difficult area to accurately measure the water content of snow. Using ASOS type rain gauges (the type used at Burlington airport), the office of the Colorado State Climatologist conducted a study, that showed substantial undermeasurement of snow, down to as little as 10% of actual precipitation. Errors are not linear, and are not easily corrected (Nolan Doesken, Assistant State Climatologist; personal communication).

Thus, it is very likely that Burlington airport underestimated precipitation during the winter. On our field we measured even less (Table 1). Therefore, simulations were conducted using precipitation from Burlington 4S, the station that reported the most precipitation during the winter (Table 1, Figure 1). Because of the uncertainty in precipitation, additional simulations were conducted using precipitation ‘scenario 2’

**Table 1. Precipitation (mm) measured at three nearby stations. Burlington airport and Burlington 4S are both 10 km north of the field. Precipitation from Burlington 4S and the fictitious ‘scenario 2’ were used in the simulations.**

<table>
<thead>
<tr>
<th></th>
<th>19 Dec. to 8 Mar.</th>
<th>9 Mar. to 12 Apr.</th>
</tr>
</thead>
<tbody>
<tr>
<td>Field</td>
<td>8</td>
<td>21</td>
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<tr>
<td>Burlington airport</td>
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<tr>
<td>Burlington 4S</td>
<td>43</td>
<td>24</td>
</tr>
<tr>
<td>Scenario 2</td>
<td>63</td>
<td>24</td>
</tr>
</tbody>
</table>

1 Mention of brand names is for information purposes only and does not imply endorsement by USDA-ARS.
(Table 1), that was constructed by tripling the winter precipitation of Burlington airport (3*21 = 63 mm). This scenario seems reasonable, considering that the ASOS type rain gauge used at Burlington airport underestimates snow up to 10 times.

The model for ridge height degradation is based on research by Lyles and Tatarko (1987). The model for random roughness degradation is based on work described by Zobeck and Onstad (1987) and Potter (1990). In WEPS, roughness simulation has to start immediately following roughness creation. Thus, we started simulations on the day of wheat planting (29 August 2000). Simulated ridge height and random roughness were forced to match the mean of the measured values on the first day of measurement (19 December 2000, Figure 2). No wind erosion occurred between 19 December and 12 April, so all roughness degradation was due to precipitation during this period.

Total biomass (dead crop residue plus live wheat plants) cover was estimated between 0 and 30% throughout the period of simulation. Within this range, simulated ridge height and random roughness changed little, so a more accurate estimate of biomass cover was not critical for this study. A constant biomass cover of 15% was used for all roughness simulations.

Research underlying the WEPS residue decomposition model has been reported by Schomberg et al. (1994, 1996) and by Schomberg and Steiner (1997). Decomposition greatly depends on temperature and moisture, as well as on the type of crop. For flat residue, WEPS considers both precipitation and soil water content. Since we didn’t have soil water content data, we only used precipitation (from Burlington 4S) for the simulation, which would underestimate decomposition. We, therefore, also simulated with moisture being at its optimum for decomposition, which would overestimate decomposition (Figure 3).

Results and discussion

Using precipitation from Burlington 4S, ridge height seemed overestimated (Figure 2), but the difference between simulation and measurement was not
significant ($P = 0.05$). Random roughness seemed slightly overestimated, but this was not significant either. Furthermore, when simulating using precipitation scenario 2 (Table 1), measured and simulated ridge height matched almost exactly and random roughness was underestimated, but not significantly (Figure 2). WEPS treats rain and snow the same. In reality, it may be expected that rain reduces roughness more than snow, due to its higher impact energy. Refinement of the model in this respect may be warranted.

Simulated residue decomposition was little, no matter what we assumed for moisture (Figure 3). Temperature was the most limiting factor. Decomposition picked up with warming in the spring (Figures 1 and 3). On 9 March measured residue biomass seemed greater than on 19 December, but the difference is not significant ($P = 0.05$). Simulated and measured corn residue biomass did not differ significantly from each other either.

NRCS personnel have measured 7 - 39% loss of residue biomass during the period October - March in the Northern USA (Montana, Wyoming, North Dakota). Measurements included stems, leaves, chaff, etc. (Gary Tibke, personal communication). WEPS predicted, and measurements showed, very little decomposition at Burlington, Colorado (Figure 3), where temperatures were at least as warm as those during the NRCS measurements. At least two reasons may explain this discrepancy: 1) At harvest, WEPS disregards everything but stems. Thus, subsequent decomposition only includes stems, which decompose slower than leaves and chaff. If the NRCS had measured only the loss of stem mass, losses likely would have been much smaller. 2) Some of the decrease in residue mass, measured by the NRCS, may be due to removal by wind rather than decomposition.

**Conclusions**

On a farmer’s field near Burlington, Colorado, USA, the mean ridge height of 42 mm on 19 December 2000 was reduced to 34 mm (36 mm simulated using WEPS) on 12 April 2001. The mean random roughness of 5.8 mm on 19 December was reduced to 5.2 mm (5.3 mm simulated) on 9 March. The simulation of roughness degradation is driven by precipitation, which is very difficult to measure in windy climates, especially when it comes in the form of snow. The mean corn residue biomass of 1204 kg ha$^{-1}$ on 19 December was only reduced slightly to 1174 kg ha$^{-1}$ (1144 - 1186 kg ha$^{-1}$ simulated) on 12 April. None of the differences between measured data and simulations were significant ($P = 0.05$), enhancing confidence in the ability of WEPS to simulate roughness degradation and residue decomposition.
References


Measurement and Data Analysis Methods for Field-Scale Wind Erosion Studies

Ted M. Zobeck, Wind Erosion and Water Conservation Research Unit, USDA, Agricultural Research Service, 3810 4th Street Lubbock, TX 79415 (tzobeck@Lbk.ars.usda.gov)

John E. Stout, Wind Erosion and Water Conservation Research Unit, USDA, Agricultural Research Service, 3810 4th Street Lubbock, TX 79415 (jstout@Lbk.ars.usda.gov)

R. Scott Van Pelt, Wind Erosion and Water Conservation Research Unit, USDA, Agricultural Research Service, 302 W I-20, Big Spring, TX 79720 (Svanpelt@Lbk.ars.usda.gov)

Roger Funk, Centre for Agricultural Landscape and Land Use Research (ZALF), 15374 Muncheberg, Germany (rfunk@zalf.de)

Jean Louis Rajot, IRD-LISA-Université Paris 12, 61, Avenue du Général de Gaulle 94 010 Créteil Cedex, Paris, France (rajot@lisa.univ-paris12.fr)

Geert Sterk, Wageningen University, Dept. of Environmental Sciences, Erosion and Soil & Water Conservation group, Nieuwe Kanaal 11, 6709 PA Wageningen, The Netherlands (Geert.Sterk@users.tct.wag-ur.nl)

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, field characteristics, and methods of analyses used in aeolian field studies depend upon the specific objectives of the study.

A variety of erosion models are now available to estimate windblown sediment transport. Most models have similar input variables that include the need for data related to wind and other climatic variables, soil surface and near surface characteristics and vegetative properties. However, models may vary considerably in output time period or the type of final results reported. Some models produce annual estimates while others report by erosion event or some other time interval. Some models focus on saltation flux while others also include the flux of suspended sediment. Validation and further development of present and future models will require a systematic method of data collection for the input and the output variables.

The kind of information collected in wind erosion studies depends on the purpose of the study. For instance, a modeling study requires usually much more data than an agronomic type study in which conservation measures are tested. This paper will outline important factors to consider in conducting field-scale wind erosion studies and describe most commonly used field data collection and analysis methods for use in model validation. This list is not intended to be exhaustive but will include factors used by most models.

Field Characteristics

Factors such as field shape, length, boundary conditions, and surface uniformity should all be considered when selecting or evaluating fields for field-scale wind erosion studies. The field shape and size used in field-scale wind erosion
research is generally a matter of preference and in many studies may not have practical significance; more care must be given to field orientation with respect to the prevailing wind direction. Field geometry is most important in studies to determine the potential wind erosion or ‘worst case scenario.’ Circular fields have the advantage that data collection is possible regardless of the wind direction and a range of field lengths can be measured with a minimum of samplers (Fryrear et al., 1991). If tillage ridges are to be tested in circular fields, tilling in a circular pattern allows testing the effects of ridges perpendicular to the wind direction from all wind directions. Rectangular or square fields are often preferred 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.

A field length of at least 300 m may be needed in many situations to approach maximum saltation transport capacity in bare agricultural fields. 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 250 m in a bare, fine sandy loam field with small tillage ridges (Stout and Zobeck, 1996). Smaller fields may be adequate in sandier soils. However, in a study of a sandy soil in Niger (greater than 90% sand), the maximum of saltation was never reached at 80 m (Bielders, et al., 2002).

Maintaining a clear non-erodible boundary and non-erodible upwind area is necessary to accurately determine sampling fetch distance. The edge of the eroding field should be clearly identified and stabilized to accurately determine fetch. In many instances, it is impractical or not possible to accurately determine the non-erodible boundary. For example, in many areas such as the small-scale fields in Sahelian Africa, regular, non-erodible boundaries are usually difficult to define. In these cases, the fetch may not be known so it is important to measure the input of sediment from upwind sources with samplers placed immediately upwind of the study field. If the incoming sediment is not accounted for in the analysis, the estimated soil losses may be dramatic, while in reality this is not so.

A bare, smooth, flat, dry field with unstructured, loose but uniform soil would present the most erodible condition. 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. Any changes in any of the factors listed above could cause differences in saltation flux due to differences in entrainment, transport or deposition of sediment (Stout and Zobeck, 1996). Changes in these factors within the field may limit the field length available for study. When any of these factors change, they should be carefully documented. Since, most agricultural fields are characterized by spatial variability in the many factors that determine the wind erodibility of the soil, significant spatial variation in saltation flux is usually observed.

Wind Erosion Measurements

The types of samplers used for sampling windblown (aeolian) sediment will vary depending upon the type of sediment to be measured. Windblown 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).
Estimates of creep may be most economically made by burying a bottle with the opening flush with the soil surface as suggested by Bagnold (1941). Use of a container with a round opening will facilitate calculations since the width of the opening will be the same from every direction. A rather complicated but reliable creep/saltation sampler that orients into the wind has also been used successfully (Stout and Zobeck, 1996).

Most saltation samplers actually collect samples of suspended dust as well as saltating particles. The height of sampling for saltation-size sediment is not an absolute value but rather a transition zone gradually yielding to a height dominated by suspension flow. It is difficult to place a limit on the boundary between the zone dominated by saltation flow and that dominated by suspension flow. Studies of sediment particle size distributions of a fine sandy loam placed a transition zone for the saltation and suspension flow regimes between heights of 0.1 and 0.2 m (Stout and Zobeck, 1996). This study showed that at a height of 70 cm, over 88 percent of the sample collected by Big Spring Number Eight (BSNE) samplers had a diameter of less than 90 microns. Samplers rarely need to exceed one meter in height for studies of saltation flux.

The BSNE and Modified Wilson and Cooke (MWAC) samplers (Fig. 1) appear to be the most popular for field studies of saltation. Both types of samplers are easily mounted on poles to allow sampling at multiple heights. However, the BSNE sampler opening is much larger (1050 mm² or 200 mm² for those designed for the near surface measurements) than the MWAC sampler (50 mm²) and will produce larger samples.

Sampling suspended dust may be performed with active samplers that provide a suction using a pump of some type. Nickling and Gillies (1993) describe a directionally dependent suspended sediment sampler with a 1.3 cm sampling orifice that orients into the wind. Suction is provided by a high volume pump. Four samplers are mounted to a height of ten meters. The suction on each tube is adjusted to match the ambient wind speed to provide isokinetic sampling. A recent improvement in this method was made by attaching the sediment sampling head to a DustTrak aerosol monitor described below (Bill Nickling, personal communication, Fig 2).

The DustTrak aerosol monitor is one of several commercially produced instruments now available to 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. The GRIMM Environmental Dust Monitor (GRIMM Technologies, INC) uses light scattering to measure particle size.
The Tapered Element Oscillating MicroBalance (TEOM®) continuously measures mass of a filter during air filtration by the means of a microbalance based on the change in the oscillation frequency of the support of the filter. It is equipped with a classical low volume PM10 sampling inlet. The sampling efficiency of these devices may be low in high winds unless equipped with an approximately isokinetic sampling orifice such as that shown in figure 2. Other commercially available suspended dust samplers are also available.

Other Topics

The oral paper will also discuss a variety of other measurement and data analysis techniques used in field-scale wind erosion studies. Topics will include sampler number and location, meteorological station instrumentation, elementary wind data analysis techniques, mass transport and dust emission data analysis, and soil surface characteristics including surface roughness, and aggregate and crust properties.

Figure 2. Suspended dust sampler.

Disclaimer

Mention of trade names or commercial products in this article is solely for the purpose of providing specific information and does not imply recommendation or endorsement by the US or foreign governments.

References


Numerical simulation of sand and snow drift at porous fences

S. Alhajraf, KISR, PO BOX 24885, Safat 13109, Kuwait. (shajraf@safat.kisr.edu.kw)

Introduction

Computational Fluid Dynamics (CFD) has become a key tool for a broad range of today's research studies and industrial applications. They are as varied as the dispersion of passive pollutant particles in the atmosphere to the complex flow system involving chemical reactions. The numerical prediction of the natural phenomena of wind blown particles such as sand or snow and particle deposition around objects has attracted attentions of researchers from a wide range of engineering and scientific backgrounds.

Wind blown particles occur as the result of a natural phenomenon known as an Aeolian process. The Aeolian process is divided into three stages, erosion, transportation and deposition, commencing when the wind entrains a particle from rest and carries it for some distance. When the aerodynamic forces applied to a particle become less than the gravitational force, the particle finally settles back to the ground (Bagnold 1941). A large number of field and wind tunnel studies have been conducted to study the physical behavior of wind blown particles as they approach obstacles of different shapes. Since field and wind tunnel experiments are costly, numerical modeling has been applied to many engineering and environmental applications (Ishii, 1975).

This paper presents a practical application of the numerical model based on two-phase flow theory. Two numerical experiments were conducted using this model for snow and sand drift around porous fence. The results show a good agreement with both field and wind tunnel observations.

Numerical Model

The key consideration in the numerical model is the representation of the solid particles as a second continuum flow phase superimposed upon the primary phase, the air, as described by the conventional Navier-Stokes equations (Ishii, 1975). A flow regime containing two or more flow phases of different physical properties may be treated as a multi-phase flow system, which can be solved numerically based on the theory for a multi-phase flow system (Manninen, 1996). However, considering the air as the continuous phase (carrier phase) and the particles as the discrete phase, the simplest two-phase flow model, known as the Homogenous two-phase flow model, was employed in this paper. The conservation equations can be expressed as:

$$\frac{\partial (\rho \phi)}{\partial t} + \frac{\partial}{\partial x_j} (\rho u_j \phi) = \frac{\partial}{\partial x_j} \left[ \Gamma_{\phi} \frac{\partial \phi}{\partial x_j} \right] + S_{\phi}
$$

(1)

Where $\phi$ represents the solution variable to be solved, $u$, $v$, $w$, $k$: Kinetic energy, $\varepsilon$: rate of dissipation of kinetic energy and $\alpha_p$: particle volume fraction. $x_j$ is the space components $x,y,z$. 


The flow governing equations may be formed by substituting the variable $\phi$, the diffusion coefficient $\Gamma_\phi$ and the source term $S_\phi$ with the appropriate values (Alhajraf, 2000).

Particles transported by wind usually occur in one of three modes, surface creep, saltation or suspension (Bagnold, 1941). These processes depend upon the physical properties of the particle such as size, density and on the strength of the wind velocity component parallel to the particle bed. In this paper, models of the particle transport by saltation and suspension were considered as two separate source terms ($S_p = S_{sus} + S_{sal}$) added to the particle transport equation of $\alpha_p$. The suspension source term is formulated as:

$$S_{sus} = -\beta_{sus} \frac{\partial}{\partial X_j} \left[ \alpha_p u_{Drift} \right]$$

where the diffusion velocity is $u_{Drift} = (1 - \alpha_p) u_{Rel}$ and $\beta_{sus}$ is valid between 0.05 and 0.1 in flow regimes involving sand drift, (Alhajraf, 2001).

The saltation particle zone usually has a layer thickness of a few centimetres (Bagnold, 1941). In this layer the suspension source term is modified to take into account the saltated particles. For a quartz particle with 0.25mm diameters and 2650 $kg/m^3$ density the threshold velocity was found to be about 0.22 m/s.

In this paper, the saltation source term was written in terms of particle volume fraction, the relative velocity and the dimensionless friction to threshold ratio as follows:

$$S_{sal} = \beta_{sal} \frac{\partial}{\partial X_j} \left[ \alpha_p (1 - \alpha_p) u_{Rel} U_{\phi} \right]$$

$\beta_{sal}$ is a constant that varies between 0.15 and 0.6, (Alhajraf, 2001).

The saltation source term is applied only in the control volumes adjacent to the solid boundaries. Thus, there are three possibilities in this case for the saltation source term in each control volume:

1. If the friction velocity is greater than threshold value then the source term will have a negative sign and therefore an erosion process will occur.
2. If the friction velocity is equal to threshold value then the source term will be zero and neither erosion nor deposition will occur.
3. If the friction velocity is less than threshold value then the source term will have a positive sign and therefore a deposition process will occur.

Results

The numerical computations were employed to simulate particle deposition at 50% porous fence. First run were performed for Wyoming snow fence used to protect Wyoming highway. Steady deposition profile from the field observation of Tabler, 1986 and wind tunnel measurement of Iversen, 1981 were compared with the numerical prediction.

A steady state solution was achieved with the particle deposition profile reaching an equilibrium state. Fig. 1, shows the different stages of the deposition process for different time steps. It shows that the majority of the deposition area is concentrated behind the fence with the crest of the dune found at a distance from the fence varying between 5 and 8 times the fence height $H$. The maximum height of the dune was found to be similar to the observations, about 20%H over the height of the fence.
The crest of the dune begins in the earlier stages at a distance downstream from the fence equal to about 6 to 8 times the fence height $H$. However, as more particles are deposited, the dune increases in size whilst the height of the crest point increases and moves backwards towards the fence. At the equilibrium state, the height of the crest point reaches $1.2H$ from the ground and $5H$ downstream from the position of the fence.

The full-scale, wind tunnel and numerical model deposition profiles are shown in Fig. 2a. The portion of the snow drift immediately downstream the fence in the wind tunnel experiment is just under the full-scale profile. Iversen (1981) explained this discrepancy by saying that the full-scale friction velocity is at times just above the particle threshold value while in the wind tunnel experiment it was set to some significant value above the threshold value. This is probably the reason for the lack of deposition in front of the fence in the wind tunnel experiment. It is also possible that the wind tunnel experiment had not reached the state of equilibrium.

The deposition profile predicted by the numerical model is shown to be comparable to those measured. The predicted profile shows fairly good agreement with the full-scale rather than the wind tunnel upstream and immediately down stream of the fence.

The area of deposition in front of the wall was predicted as a result of the drop in the friction velocity to values below the particle threshold in the weak zone just upstream of the fence. This behavior of the numerical model strengthens Iversen's explanation of the discrepancy between the full-scale and wind tunnel measurements. In a further investigation of this behavior, the same simulation was repeated under identical conditions with the exception of the inlet friction velocity, which was increased to a value 20% above the threshold value instead of 2% used in the previous exercise, Fig. 2b.

Fig. 3a. shows the drift around short fence of 0.762-m depth placed in the middle of the flow domain Fig. 3b shows the deposition volume where the fence spanned the whole domain width. The simulation of the domain in which the fence spans the whole width produced a dune, which is symmetrical along the length of the domain. The prediction in the case of the centrally positioned fence produced a dome shape dune due to the effect of the change in the flow field at the fence edges.

For sand drift at porous fence, the model has been employed to simulate the drift at the double row fence system used in the Kuwaiti desert to protect installations. Fig. 4, shows a photograph of the 2-Km front fence in the equilibrium state. An intermediate and equilibrium state profiles produced from the model are compared to the field measurements in Fig. 5. It is clearly shown that the front fence profile is in good agreement since they both reached the equilibrium shape. The rear fence profile over predicted the measured one and that is because the fence in the field did not reach its final capacity. This is demonstrating the potential of the model to predict the shape of the deposition profile even before it happens. Finally, Fig. 6 shows the different stages of the deposition procedure and clearly shows that the front fence reached it maximum capacity before the rear one. This behavior of the model is in a good agreement with what happens in the nature where the front fence reached the steady shape while the rear one still can trap more sand particles.

**Conclusion**

A numerical model based on the homogenous two-phase theory was introduced to simulate sand and snow drift at porous fences. The model shows a good qualitative and quantitative agreement with both field and wind tunnel observations.
In conclusion, a practical CFD tool has been developed and validated, incorporating novel physical and numerical models. The tool can be utilised by scientists and engineers to further understand the real world problem of drifting sand and snow in urban and industrial areas.

References


Figure 1: Numerical prediction against full-scale and wind tunnel measurements at 50% porous fence a. inlet friction velocity is 2% above the threshold value. b. inlet friction velocity is 20% above the threshold value.

Figure 2: 3-D deposition dune behind 50% porous fence. Isosurface of volume fraction at 0.75.
**Figure 3**: Drift formed at the 1st fence line, 2m height, facing the wind direction at KISR site. Photograph by author, December 1999.

**Figure 4**: Comparison between KISR station double row fence system and the numerical model prediction.

<table>
<thead>
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<th>T/Te</th>
<th>Deposition Curve</th>
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<td><img src="image" alt="Deposition Curve 30%" /></td>
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</tr>
<tr>
<td>60%</td>
<td><img src="image" alt="Deposition Curve 60%" /></td>
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</table>

**Figure 5**: Deposition stages at 50% porous fence. Volume fraction equal 0.75
Forecasting Dust Storms using CARMA-Dust Model and MM5 Weather Data

**Figure 6**: Deposition stages at multi row fence system using the numerical model.
Forecasting Dust Storms using CARMA-Dust Model and MM5 Weather Data

B. H. Barnum and N. S. Winstead, Johns Hopkins University, Applied Physics Laboratory, Laurel, MD 20723 (ben.Barnum@jhuapl.edu)

P. Colarco, University of Colorado, Boulder, CO (colarco@colorado.edu)

O. B. Toon, University of Colorado, Boulder, CO (toon@colorado.edu)

P. Ginoux, Georgia Tech, Atlanta, GA (Pginoux@nasa.gsfc.gov)

J. Wesely, United States Air Force Weather Agency, Offut AFB, NE, (Jeremy.Wesely@afwa.af.mil)

Lt. Amy Hakola, United States Air Force Weather Agency, Offut AFB, NE, (Amy.Hakola@afwa.af.mil)

Introduction

Dust storms throughout Saharan Africa, Middle East and Asia are estimated to place more than 200 to 5000 million tons of mineral dust into the earth’s atmosphere each year (Tegen and Fung 1994). Dust storms directly affect visibility and impact daily commercial and military operations in dust prone regions. The United States Air Force Weather Agency (AFWA) has supported the development of a dust forecast model with a 72 hour forecast capability. The dust model called CARMA (Community Aerosol Research Model from Ames) was developed by Professor Owen Toon and Dr. Pete Colarco at the University of Colorado, Boulder. The CARMA model has been modified by Johns Hopkins Applied Physics Laboratory to use daily Mesoscale Model 5th generation (MM5) weather forecasts run by the United States Air Force Weather Agency.

The latest version of the CARMA MM5 dust model can make 72 hour forecasts of surface and airborne dust concentrations in 3 different mesoscale theaters covering Saharan Africa and the Middle East, Southwest Asia and China. A new global dust source database developed by Dr. Paul Ginoux is used in the CARMA model. The dust source model is based on topographical features associated with dust sources and has been further supplemented with TOMS and AVHRR satellite data.

The forecast ability of the dust model was evaluated over a 3 month period for two of the AFWA MM5 forecast theaters; African Sahara and Middle East/Southwest Asia. The Middle East has been grouped with Southwest Asia for this evaluation. The model forecasts were compared with DMSP satellite imagery and ground observations. Each theater was broken into sub-regions for detailed evaluation of the short (6-12 hour), mid (30-36 hour) and long term (54-60 hour) forecast ability of the model. Results of the study show the dust model has good skill in forecasting dust conditions for short and medium range forecast periods.
Forecasting Dust with CARMA

The CARMA model was written as a fully scalable aerosol model to study a variety of atmospheric processes, such as cloud formation, smoke and dust aerosols (Toon et al. 1988). The version of CARMA used for dust aerosol forecasting incorporates a global dust source database developed by Paul Ginoux at NASA Goddard Flight Center. The database uses dust sources associated with topographic depressions. The database was also developed using satellite data from the Total Ozone Mapping Spectrometer (TOMS) dust Aerosol Index (AI) (Ginoux et al. 2001). The CARMA model uses 10 particle size bins covering from 0.5 μm to 10.0 μm, which will have airborne residence times greater than several hours (Tegen and Fung, 1994).

The CARMA dust model uses 22 vertical sigma pressure levels and a 90 km horizontal latitude, longitude grid spacing. The model is run with lower resolution than the MM5 weather model, which uses 41 sigma levels and 45 km horizontal resolution. This grid scheme was chosen to have approximately the same spacing as the 1° x 1° (111 km) Ginoux global database model to save run time for daily forecasting.

The dust forecast is first initialized by running the model using a 2 day (48 hour) “spin-up”. The spin-up uses the first 24 hours of MM5 forecasts generated for each of the spin-up days. The data from the spin-up portion of the model is then used as the initial dust concentration condition at the beginning of the 72 hour forecast. The dust forecast then uses the MM5 72 hour forecast data for winds, pressure, temperature, rain, etc., for the concentration predictions. The dust model outputs a set of dust concentration maps for each 3 hour time period during the 72 hour forecast.

The dust source model first calculates the surface threshold wind velocity at each grid location for each particle bin size. Where there is measurable accumulated precipitation in a 24 hour period, the threshold wind velocity is set so that no dust flux is generated at the location. The surface dust flux is then calculated for each particle size bin and the MM5 forecast 10 meter wind speed using:

\[
F_{ijr} = C*S_{ijr} * (w10m_{ij} – u_{(ijr)}) *w10m_{ij}^3
\]

Where \( C \) is a model constant, \( F_{ijr} \) is the surface dust flux in gm/m²s, \( S_{ijr} \) is the Ginoux source strength for the particle class size, \( w10m \) is the wind speed at 10 meters, and \( u_{(ijr)} \) is the threshold wind velocity for each grid location and particle bin size (Ginoux et al. 2001, Chin et al. 2001).

The dust model forecasts are displayed as a set of color images showing total dust concentration at user selected altitudes, vertical profiles and total dust loading. The images are made for each 3 hour interval in the 72 hour forecast, an example is shown in figures 1 and 2. An example of the African and Middle Eastern mesoscale theater is shown in Figure 1.
Figure 1 Example of CARMA model output showing color map of dust concentration (log scale) over Saharan Africa and Middle East for the dust storm during January 7, 2002.

Model Forecast Results

The dust model was installed and run daily at AFWA beginning February 2002. A qualitative evaluation of the model’s dust storm forecast capability was made during February through April 2002. The evaluation covered the African Sahara and Middle East/Southwest Asian theaters. The study compared dust storms located using Defense Meteorological Satellite Program imagery (DMSP) data, and ground reported observations when available. Wherever model surface forecast concentrations exceeded 1800 µgm/m³, dust storm or “dusty” conditions were considered to be present at the location. The model was scored using meteorological “skill scores” over short (6-12 hr), medium (30-36hr) and long (54-60hr) range forecasts. The skill scores used were Probability of Detection (POD), False Alarm Rate (FAR) and Critical Success Index (CSI). Each theater was divided into to sub regions for the study. The African Sahara was divided into 7 sub regions and the Middle East/Southwest Asian theater into 11 sub regions.

The average POD and FAR percentages for the two theaters are given in Table 1. The lowest CSI scores occurred in the Yemen and Oman sub regions where the POD’s were only 19, with a FAR of 0. This region of the Empty Quarter is a great sand desert, but is a relatively weak dust source in the Ginoux database. This desert region produces surface level sandstorms. Sandstorms typically have a lower TOMS AI, which is more sensitive to higher altitude dust concentrations.

<table>
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<tr>
<th>MM5 theater:</th>
<th>POD (FAR)</th>
<th>Short (6-12 hr)</th>
<th>Medium (30-36 hr)</th>
<th>Long (54-60 hr)</th>
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<td></td>
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<td>67 (15)</td>
<td>59 (18)</td>
</tr>
<tr>
<td>Middle East/Southwest Asia</td>
<td></td>
<td>61 (10)</td>
<td>62 (9)</td>
<td>52 (7)</td>
</tr>
</tbody>
</table>
Table 1: CARMA model average Probability of Detection and False alarm rate for short, medium and long range MM5 forecasts during model evaluation (Jan.-April 2002).

Results of the AFWA study show the dust model has good skill in forecasting dust conditions over short (12 hour) and medium (36 hour) forecast periods. In the Africa Sahara theater, the average POD for a 30-36 hour forecast was 67 percent with a FAR of only 15 percent. Long range forecasts of 54-60 hours had POD’s of 59 percent with FAR’s increasing to 18 percent, indicating decreasing forecast accuracy of the weather model by 60 hours.

Figure 2: Example of DMSP satellite imagery used to locate dust storms.

Conclusion

The CARMA dust model has been successfully adapted to use MM5 weather forecast data for the operational prediction of dust storms. The global dust source database developed by Ginoux et al. has been especially accurate for forecasting in Saharan Africa. The next phase of the dust project will integrate the Continental United States and East Asia and China as operational dust forecast theaters. More studies are underway to evaluate model predicted dust concentrations with data from the Puerto Rican Dust Experiment (2000) and aerosol measurements at White Sands New Mexico.

References


Dust Generation Modeling - Source Terms and Microscale Transport from Military Installations

Ronald Cionco, U.S. Army Research Laboratory, AMSRL-CI-EB, White Sands Missile Range, New Mexico 88002-5501 (E-mail: rcionco@arl.army.mil)

Donald Hooock, U.S. Army Research Laboratory, AMSRL-CI-EE, White Sands Missile Range, New Mexico 88002-5501 (E-mail: dhoock@arl.army.mil)

Introduction

Dust is a major particulate emitted from military installations, especially those that perform extensive training with tracked vehicles and high explosive artillery ranges. Concerns are frequently raised when dust exiting a military reservation consistently exceeds particulate sampling standards, when local natural vistas appear to be degraded, and especially when visible dust plumes significantly restrict local visibility for period of time. This poster presents some examples from the U.S. Army Research Laboratory measurement and modeling efforts over the past twenty years to characterize and generalize airborne military dust. This has included measurement and modeling of production source terms and cloud optical properties. Optical mass extinction coefficients tie reduced meteorological visibility to airborne dust mass concentration, composition and particle size distribution by mass. We show examples of dust transport modeling using the High Resolution Wind (HRW) Model that suggest significant reductions in near-surface visibility to less than 1 km are possible at distances of more than 10 km downwind. HRW is particularly applicable to modeling the microscale effects of complex terrain and canopies on boundary layer transport and, combined here with the Risoe RIMPUFF dispersion model, can identify regions of local concentration maxima and visibility minima due to terrain effects. Finally, dust generation from local events can often be compared in context with the historical frequency and duration of natural blowing dust (visibility reduced to less than 11 km) and dust storms (visibility less than 1 km) events. We show examples from the published global boundary later dust occurrence climatology by B. Hinds and G. Hoidale (Atmospheric Sciences Laboratory, 1975) using observation data spanning 1948 to 1973.
Wind Erosion Risk Assessment of Alberta Soils

G. M. Coen, Agriculture and Agri-Food Canada, Lethbridge, AB T1J 4B1
(coeng@em.agr.ca).

J. Tatarko, USDA-ARS-WERU, KSU, Manhattan, KS 66506.

T. C. Martin, Alberta Agriculture, Food and Rural Development, Edmonton, AB.

K. Cannon, Alberta Agriculture, Food and Rural Development, Edmonton, AB.

T. W. Goddard, Alberta Agriculture, Food and Rural Development, Edmonton, AB.

N. J. Sweetland, Agriculture and Agri-Food Canada, Lethbridge, AB.

Introduction

Alberta has significant agricultural acreage that is at risk to wind erosion. The recent availability of AGRASID (Agricultural Region of Alberta Soil Inventory Database), a seamless, standardized digital soil map at a scale of 1:100,000 (CAESA Soil Inventory Working Group, 1999) and hourly, geographically referenced spatial weather data (Shen et al., 2001) made this work possible. The WEPS (Wind Erosion Prediction System) is a process-based, continuous daily time step model (Wind Erosion Research Unit, 2001) which has the ability to respond to environmental and management variations to predict erosion events. With the use of the most recent data and understanding of wind erosion processes it should be possible to provide the most useful predictions. The objective of this study is to evaluate the use of available databases and the WEPS model to assess the susceptibility of Alberta agricultural soils to wind erosion risk and the degree of exposure Alberta soils experience under current management.

Materials and Methods

Weather files were prepared by interpolating daily weather station data to Soil Landscapes of Canada Polygons (Shen et al., 2001; Soil Inventory Staff, 1988). A survey of field management practices (Dey, 2000) supplemented by interviews with Alberta Agriculture, Food and Rural Development regional specialists was used to assign crop rotations and percentages to Ecodistrict polygons (Ecological Stratification Working Group, 1995). Soils data characterizing AGRASID polygons was prepared from the soil names and soil layer files supplemented by relationships derived from the Alberta pedon database. An unreleased version of the AGRASID file ag30smu (personal communication, May 2002, J.A. Brierley, Agriculture and Agri-Food Canada, 7000 - 113 Street, Edmonton, AB T6H 5T6) provides an estimate of the percent of each polygon that is occupied by a soil series. All files were formatted to meet the requirements of WEPS. The soils database is the most detailed (1:100,000), so each AGRASID polygon was assigned the same weather and management as the Soil Landscapes of Canada or
Ecodistrict polygon where the AGRASID polygon centroid was located. Each unique soil-crop rotation-weather combination was run using a batch procedure from the WEPS command line. The total erosion attributed to each soil-management combination was apportioned to the AGRASID polygon to estimate a mean loss per acre in the polygon. These values are then used to rank the erosion susceptibility of each polygon.

**Results and Discussion**

The WEPS 1.0 beta 8.0 release can only estimate erosion losses on a relatively homogeneous area, rectangular in shape and for a single soil type and land use. For this study a quarter section (64 ha, 160 ac) was chosen as the type situation for a WEPS run. In order to derive an estimate of erosion risk for an entire AGRASID polygon it was necessary to run WEPS for each soil type that is usually cultivated as well as the associated management files (common crop rotations) and sum their separate contributions. For the batch runs a 30(±)-year simulation was used. Total erosion was estimated \((\text{kg m}^2)\) for each combination of soil and management. For each combination, the total erosion values were then manipulated to represent the total cultivated portion of the polygon (Table 1). The average loss per ha (acre) was calculated per polygon and the result mapped (Fig. 1). The average soil loss can then be grouped into erosion risk classes.

Figure 1. Example of an erosion susceptibility map of a Township in Alberta, Canada.
WEPS is an example of a site model that provides fairly specific information given uniform environmental and management scenarios. The methodology described here provides a procedure to extrapolate site results to soil landscapes. The resulting spatial representation is appropriate to display at a map scale of 1:100,000. The format of the Soil Landscapes of Canada database (which is similar to many larger scale provincial soils databases, such as AGRASID) is fairly easily modified to match the requirements of WEPS. Some data required by WEPS is not part of these databases and must be derived from various other databases and relationships. The weather database was prepared from the Environment Canada weather records (Shen et al.) 2001 to meet the requirements of WEPS and WEPP; a significant effort but now available for continued application.

The methodology will allow temporal comparison of crop rotations used in the future with those used at present or with past management procedures thereby providing an opportunity to evaluate environmental sustainability. It will also allow a more spatially precise evaluation of the inherent wind erosion susceptibility of Alberta soils than previously published (Coote and Pettapiece, 1989; Padbury and Stushnoff, 2000).

References

http://www.agric.gov.ab.ca/agdex/000/agrasid.html


Table 1. Example of a calculation of predicted soil loss for each AGRASID polygon based on the sum of losses from each soil-crop rotation combination.

<table>
<thead>
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<th>AGRASID</th>
<th>Polygon Area</th>
<th>Soil*</th>
<th>Crop Rotation</th>
<th>Crop</th>
<th>Erosion Loss#</th>
<th>In Cult Area*</th>
<th>Mean Loss++</th>
</tr>
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<tr>
<td></td>
<td>No. ha ac</td>
<td>Cult Symbol %</td>
<td>Symbol**</td>
<td>%</td>
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</table>

Mean Erosion of Polygon 8.4 37

| 5808    | 964 | 2382 | 85 KSR | 10 | pwcb | 50 | 1.1 | 5 | 553 | 610 |
| 5808    | 964 | 2382 | 85 KSR | 10 | wfcb | 50 | 13.2 | 59 | 6810 | 7508 |
| 5808    | 964 | 2382 | 85 LET | 60 | pwcb | 50 | 0.7 | 3 | 2005 | 2211 |
| 5808    | 964 | 2382 | 85 LET | 60 | wfcb | 50 | 14.9 | 66 | 45838 | 50538 |
| 5808    | 964 | 2382 | 85 OAS | 10 | pwcb | 50 | 0.6 | 3 | 288 | 318 |
| 5808    | 964 | 2382 | 85 OAS | 10 | wfcb | 50 | 12.2 | 54 | 6268 | 6911 |
| 5808    | 964 | 2382 | 85 ZERzdb | 10 | pwcb | 50 | 0.5 | 2 | na | na |
| 5808    | 964 | 2382 | 85 ZERzdb | 10 | wfcb | 50 | 10.3 | 46 | na | na |
| 5808    | 964 | 2382 | 85 ZGW | 10 | pwcb | 50 | 74.3 | 331 | na | na |
| 5808    | 964 | 2382 | 85 ZGW | 10 | wfcb | 50 | 255.3 | 1139 | na | na |

Mean Erosion of Polygon 7.5 34

| 5815    | 2084 | 5149 | 83 LET | 40 | pwcb | 50 | 0.7 | 3 | 2255 | 2486 |
| 5815    | 2084 | 5149 | 83 LET | 40 | wfcb | 50 | 14.9 | 66 | 51552 | 56838 |
| 5815    | 2084 | 5149 | 83 RDM | 20 | pwcb | 50 | 2.8 | 13 | 4938 | 5444 |
| 5815    | 2084 | 5149 | 83 RDM | 20 | wfcb | 50 | 28.6 | 128 | 49647 | 54738 |
| 5815    | 2084 | 5149 | 83 WNY | 40 | pwcb | 50 | 2.3 | 10 | 7931 | 8744 |
| 5815    | 2084 | 5149 | 83 WNY | 40 | wfcb | 50 | 20.7 | 92 | 71691 | 79042 |

Mean Erosion of Polygon 10.8 48

| 5818    | 388 | 959 | 86 LET | 50 | pwcb | 50 | 0.7 | 3 | 541 | 596 |
| 5818    | 388 | 959 | 86 LET | 50 | wfcb | 50 | 14.9 | 66 | 12363 | 13631 |
| 5818    | 388 | 959 | 86 WNY | 50 | pwcb | 50 | 2.3 | 10 | 1902 | 2097 |
| 5818    | 388 | 959 | 86 WNY | 50 | wfcb | 50 | 20.7 | 92 | 17193 | 18956 |

Mean Erosion of Polygon 9.7 43

| 5842    | 807 | 1994 | 88 KSR | 10 | pwcb | 50 | 1.1 | 5 | 425 | 469 |
| 5842    | 807 | 1994 | 88 KSR | 10 | wfcb | 50 | 13.2 | 59 | 5236 | 5773 |
| 5842    | 807 | 1994 | 88 LET | 10 | pwcb | 50 | 0.7 | 3 | 257 | 283 |
| 5842    | 807 | 1994 | 88 LET | 10 | wfcb | 50 | 14.9 | 66 | 5874 | 6476 |
| 5842    | 807 | 1994 | 88 RDM | 35 | pwcb | 50 | 2.8 | 13 | 3938 | 4342 |
| 5842    | 807 | 1994 | 88 RDM | 35 | wfcb | 50 | 28.6 | 128 | 39599 | 43660 |
| 5842    | 807 | 1994 | 88 WNY | 35 | pwcb | 50 | 2.3 | 10 | 3163 | 3487 |
| 5842    | 807 | 1994 | 88 WNY | 35 | wfcb | 50 | 21.9 | 9 | 2853 | 3145 |
| 5842    | 807 | 1994 | 88 ZERzdb | 10 | pwcb | 50 | 0.5 | 2 | na | na |
| 5842    | 807 | 1994 | 88 ZERzdb | 10 | wfcb | 50 | 10.3 | 46 | na | na |

Mean Erosion of Polygon 8.6 38

| 5872    | 493 | 1218 | 100 LET | 20 | pwcb | 50 | 0.7 | 3 | 320 | 353 |
| 5872    | 493 | 1218 | 100 LET | 20 | wfcb | 50 | 14.9 | 66 | 7317 | 8067 |
| 5872    | 493 | 1218 | 100 WNY | 80 | pwcb | 50 | 2.3 | 10 | 4503 | 4965 |
| 5872    | 493 | 1218 | 100 WNY | 80 | wfcb | 50 | 20.7 | 92 | 40702 | 44875 |

Mean Erosion of Polygon 10.7 48

* Soil symbols beginning with 'Z' were considered to belong to the uncultivated portion of the polygon.
** Crop rotation symbol:  pwcb = peas/wheat/canola/barley,  wfcb = wheat/fallow/canola/barley.
# Rate of erosion on the portion of the polygon where the given soil/crop rotation occurs.
* Total estimated soil loss associated with the soil/crop rotation combination in the selected AGRASID polygon.
** The mean erosion rate for the cultivated portion of the AGRASID polygon.
Stochastic multivariate analysis of hydrometeorological variables for wind erosion models

Paolo D’Odorico, Department of Environmental Sciences, University of Virginia, Charlottesville, VA 22904, paolo@virginia.edu.

Thomas M. Over, Department of Geology/Geography, Eastern Illinois University, Charleston, IL 61920-3099, tmover@eiu.com

JaeChan Yoo, Department of Civil Engineering, Texas A&M University, College Station, TX 77843-3136, texasyjc@neo.tamu.edu

Introduction

The stochastic modeling of wind speed has often concentrated either on multivariate models of daily wind speed, temperature, and solar radiation (e.g., Parlange and Katz, 2000), or on the simulation of hourly wind with no account for the mutual dependence on other hydro-climatic variables (e.g., Skidmore and Tatarko, 1990). The study of aeolian erosion requires a simultaneous modeling of high-resolution (e.g., hourly) wind speed, as well as of other weather variables (such as solar radiation, precipitation, air and dew-point temperature) needed in the analysis of surface soil moisture and of soil susceptibility to wind erosion. A multivariate stochastic process is here suggested as a possible approach to the multivariate modeling of these variables at the hourly time scale. The model is fitted to the data from two locations across the High Plains of the U.S., a region particularly affected by wind-blown dusts. The performances of the model are finally tested against the data records.

Methods

The present study uses Richardson’s approach (Richardson, 1981; Parlange and Katz, 2000) to develop a multivariate model of wind components, air and dew-point temperatures at the hourly time scale. This model accounts for the diurnal variability, the autocorrelation, and the lagged cross-correlation of these variables, as well as for the inhomogeneity between rainy and dry days. The annual cycle is divided into 12 one-month-long segments; each of them is assumed to be unaffected by non-stationarities due to the seasonal cycle and is separately modeled. For each variable, the ensemble mean and standard deviation of the daily cycle are calculated month by month. Non-stationarity due to the daily cycle is removed from the time series by subtracting the mean daily cycle from the time series and dividing by the time-dependent standard deviation

\[ X_i' = \frac{X_i(t) - \langle X_i \rangle_i}{\sigma_i(t)} \quad (i=1,\ldots,4) \]  

(1)
where the subscript $i$ indicates the hydroclimatic variable under question ($X_1=W_x; X_2=W_y; X_3=T_d; X_4=T$), while $\lambda_i(t)$ and $\sigma_i(t)$ represent the daily cycle of the mean and standard deviation, respectively. The standardized variables, $X_i$, are then normalized using a Box-Cox transformation (e.g. Hipel and McLeod, 1994; p. 122)

$$X_i^*(t) = \begin{cases} X \lambda)^{\lambda_i} & \lambda_i \neq 0 \\ \ln(X_i) & \lambda_i = 0 \end{cases}$$

where values of the exponent $\lambda_i$ are selected that are able to give variables $X_i^*$ with distributions closest to normal. We show that the values of these standardized and normalized hydroclimatic variables are affected by precipitation occurrences, due to the impact that rainstorms have on air temperature, atmospheric humidity, and winds. This dependence has been assessed by separately estimating the means, $\mu_i$, and standard deviations, $\sigma_i$, of $X_i^*$ for non-rainy or dry ($j=0$), and rainy or wet ($j=1$) days. These means and variances are found to be significantly different (95% confidence limit) in the t-test and F-test, respectively, supporting the hypothesis that these hydroclimatic variables need to be modeled conditionally upon the occurrence of precipitation. This inhomogeneity has been removed by standardizing the variables $X_i^*$ on the condition that the day is dry or wet

$$X_i^*(t) = \frac{X_i^*(t) - \mu_{i,j}}{\sigma_{i,j}}, \quad (i,j=1,...,4)$$

with $j$ being 1 or 0 depending on the (hourly) occurrence or non-occurrence of precipitation, respectively.

Rainfall occurrence has been modeled at the hourly time scale as a first-order two-state Markov chain (e.g., Gabriel and Neumann, 1962), where these states represent the occurrence ($j=1$) and non-occurrence ($j=0$) of rain. The only parameters of this rainfall occurrence process is represented by the transition probabilities $P_{jk}$ ($j,k=0,1$), indicating the probability that at time $t$ the system will be in state $k$, with $j$ being the state at time $t-1$. A likelihood ratio test for the order of the Markov chain (e.g. Wilks, 1995; p. 301) shows that the values of both the AIC (Akaike’s Information Criterion) and BIC (Bayesian Information Criterion) statistics are very close in the cases of first- and second-order Markov chain. A first-order process is here chosen to model rainfall occurrences, discarding for simplicity the idea of using different orders for the different months of the year. The hourly precipitation depth in the wet hours is modeled as a random variable with gamma distribution. No account is taken for the dependence of rainfall likelihood on the time of the day, since it was not found to be significant in the data.

The hydroclimatic time series in question are modeled using a multivariate second-order autoregressive process AR(2) (e.g. Hipel and McLeod, 1994) fitted to the standardized, normalized, and homogenized variables, $X_i^*$ (equation (3)). The choice of a multivariate model is motivated by the existence of a significant ($p<0.05$) cross-correlation between most of the variables in question. The choice of a second-order autoregressive process (AR(2)) is suggested by the inability of the AR(1) model to provide a good representation of the autocorrelation functions. Previous work on the (univariate) stochastic modeling of hourly wind speed has resulted in the use of AR(2) processes (Daniel and Chen, 1991; Brown et al., 1984; Nfaoui et al., 1996). All such studies remark the complexity existing in the modeling of the autocorrelation structure of these hourly variables, due to the presence of some residuals of the daily cycle, as well as
of seasonal nonstationarity observable in these one-month-long segments of the annual cycle. This study differs from these previous analyses in the use of a multivariate approach in the modeling of hourly wind speed components, air and dew point temperatures conditionally upon the occurrence of precipitation. We believe that the mutual dependence between these variables significantly affects the estimates of soil loss by wind erosion.

A problem is found with the modeling of the dew-point temperature, due to the physical constraint $T_d < T$. Even though this condition is partly accounted for by the mean, standard deviation and cross-correlation values of $T$ and $T_d$, there are still chances to generate pairs of values of $T$ and $T_d$ with no physical meaning (i.e. with $T < T_d$). This is resolved by adjusting the values of dew-point temperature, imposing an upper bound on $T_d$. We show that this truncation does not significantly affect the statistics of $T_d$.

Solar radiation is the other variable needed in the modeling of the energy and moisture fluxes at the soil-atmosphere interface. The values of solar radiation at the ground surface depend on the sky clearness as well as on the extraterrestrial solar radiation. The latter is a deterministic function of latitude and time of the year, while the former depends on atmospheric conditions (i.e. clouds, water vapor, and aerosols). Therefore a number of stochastic weather generators (e.g., Hansen, 1999; Hensen and Mavromatis, 2001) model the sky clearness instead of surface solar radiation. Sky clearness, which is defined as the ratio between surface and extraterrestrial solar radiation, captures the random component of solar radiation. The lack of a dense network of direct measurements of solar radiation suggested (e.g., Lindsey and Farnsworth, 1997) using indirect measurements of sky clearness; the sky cover observations, $S_c$, are often adopted as proxies for sky clearness. $S_c$ is a function of water vapor, clouds and aerosols. Therefore, sky cover is modeled as a random variable conditionally on rainfall occurrence, and on the quartiles of relative humidity (which is a function of dew point temperature).

Results

The model described in the previous sections has been used to simulate wind speed components, dew point temperatures, and precipitation. The results of these simulations have been compared with the data to assess to what extent the model is able to fit the observations. The two case studies analyzed in this work refer to the cities of Dodge City (KS) and Lubbock (TX) in the years 1961-1990. Table I reports month by month both the observed (Dodge City) and the simulated values of mean and standard deviation, showing a good agreement between data (NREL, 1992) and model results.
Moreover, as expected, the model is able to provide a good representation of the mutual dependence existing between these variables at lags of 0, 1, and 2 hours. The model’s ability to reproduce the autocorrelation structure of these time series has been evaluated through the analysis of the residual autocorrelation function. In most of the months the hypothesis that the residuals are uncorrelated is true for wind components and dew point temperature, while it is not verified for the air temperature, due to the existence of some residues of the daily cycle as well as to possible non-stationarity existing in the one-month-long segments of the annual cycle. Nevertheless, even though more refined filtering techniques are being considered to remove such a periodicity, this model gives an overall realistic representation of the climatological variables needed in the modeling of wind erosion.

**Acknowledgements**

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References


Commonality in Process Based Erosion Models, Obstacles and Opportunities

Fred A. Fox Jr., Wind Erosion Research Unit, 1515 College Avenue, Manhattan, KS 66502. E-mail: fredfox@weru.ksu.edu

The United States Department of Agriculture, Agricultural Research Service, has independently developed two daily time step, process based erosion models, one to address the erosion of soil by wind (WEPS – Wind Erosion Prediction System) and one to address the erosion of soil by water (WEPP – Water Erosion Prediction Project). The two models need to simulate many of the same processes in order to accurately predict the erosion potential. Development can be enhanced if common process simulation code is used in both models.

System States used in process based erosion modeling

The defining element in any process based erosion model is the erosion process. Erosion amounts are determined by the interaction of the erosion process and it's driving forces with the state of the erodable surface. Estimating erosion amounts for varied crop management systems requires an accurate description of the time evolution of the state of the erodable surface in response to both management practices and climate driving forces.

Shown below is a comparison of the system states used to describe an erodable surface in WEPS and in WEPP. They are divided into two groups, unique state variables and common state variables to highlight obstacles and opportunities respectively for commonality in erosion modeling.

<table>
<thead>
<tr>
<th>WEPS</th>
<th>WEPP</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Driving Force</strong></td>
<td><strong>Erodable Surface Characteristics</strong></td>
</tr>
<tr>
<td>- Air density</td>
<td>- residue mass and cover fraction</td>
</tr>
<tr>
<td>- Wind direction</td>
<td>- live crop biomass and cover fraction</td>
</tr>
<tr>
<td>- anemometer height</td>
<td></td>
</tr>
<tr>
<td>- aerodynamic roughness at anemometer site</td>
<td></td>
</tr>
<tr>
<td>- Wind speed</td>
<td>- random roughness</td>
</tr>
<tr>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>For each barrier specified</strong></td>
<td></td>
</tr>
<tr>
<td>- Barrier location, height, porosity, and width</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>For each subregion specified</strong></td>
<td></td>
</tr>
<tr>
<td>- Subregion coordinates (x1,y1;x2,y2) and all following state variables</td>
<td>For each overland flow element specified</td>
</tr>
<tr>
<td>- Biomass height, stem area index, and leaf area index</td>
<td>- overland flow element slope angle</td>
</tr>
<tr>
<td>- Flat biomass cover</td>
<td>- overland flow element slope</td>
</tr>
<tr>
<td>- Allmaras random roughness</td>
<td>- overland flow element length</td>
</tr>
<tr>
<td></td>
<td>and all the following state variables</td>
</tr>
</tbody>
</table>

Not Used

Not Used
Processes modeled to predict evolution of system state in time

As was noted in a previous paper (Fox et al, 2000), many of the same processes are modeled in both WEPS and WEPP. Each process model was designed and tested to support the evolution of specific states unique to the erosion process being modeled. Decisions were made to combine sub-processes differently to capture significant state interactions while keeping computer time requirements at reasonable levels. Given that the models have been under development for the past 12 years during which time computer speed has been constantly increasing, the core design and selection of process models was very likely done with a strong consideration to meet computer time requirements by using a "reasonable approximation", not to capture process effects on the system state for the widest range of possible conditions.

Interestingly, a review of another ARS developed model, RZQM (Ma, 2001) reveals that in order to model chemical transport in soil, many of the same processes are modeled. The model processes are described as: Management - tillage, addition of manures, chemicals, or irrigation water; Potential Evapotranspiration; Sub-hourly processes - infiltration and runoff, soil water distribution, chemical transport, pesticide washoff, heat movement, actual evaporation and transpiration, plant nitrogen uptake, reconsolidation of tilled soil, and snowpack dynamics; Pesticide degradation on plant and residue surfaces and within soil layers; Organic matter / nitrogen cycle; Soil inorganic chemical equilibrium; Plant growth - Photosynthesis, nitrogen uptake, carbon and nitrogen partitioning, root growth, respiration, and mortality as influenced by temperature, soil water availability, and plant nutrient status. RZQM modelers used the same infiltration method as WEPP, the same soil water redistribution theory as WEPS and a much more complex evapotranspiration method than either WEPP or WEPS.

Spatial definitions and conflicts in process modeling

In WEPS and WEPP, processes are defined to account for changes of state and fluxes in either one or two space dimensions. One dimensional process models track changes in state and fluxes along a line, even though the results may be applied over an
area or volume. The table below summarizes the processes modeled and spatial dimensions modeled.

<table>
<thead>
<tr>
<th>WEPS</th>
<th>WEPP</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Climate generation - daily precipitation, temperature, solar radiation</strong></td>
<td>Weather generation - daily precipitation depth, duration and intensity, temperature, solar radiation, and wind</td>
</tr>
<tr>
<td>Point estimate assumed valid over wide area</td>
<td>Point estimate assumed valid over wide area</td>
</tr>
<tr>
<td><strong>Wind generation - hourly wind for 16 cardinal directions</strong></td>
<td><strong>Hydrology – daily soil water balance of rainfall, snowfall, irrigation, plant water use, drainage</strong></td>
</tr>
<tr>
<td>Point estimate assumed valid over wide area</td>
<td>One dimensional model, vertical from atmosphere into soil</td>
</tr>
<tr>
<td><strong>Hydrology – daily soil water balance of rainfall, snowfall, irrigation, plant water use, drainage</strong></td>
<td>Winter processes - snow accumulation and melting, soil freezing and thawing (hourly time step)</td>
</tr>
<tr>
<td>One dimensional model, vertical from atmosphere into soil</td>
<td>One dimensional model, vertical from atmosphere into soil</td>
</tr>
<tr>
<td><strong>Winter processes - snow accumulation and melting, soil freezing and thawing (hourly time step)</strong></td>
<td><strong>Irrigation - schedules irrigation based on soil state or fixed user provided schedule</strong></td>
</tr>
<tr>
<td><strong>Irrigation - schedules irrigation based on soil state or fixed user provided schedule</strong></td>
<td>Two and a half dimensional model, updates soil water balance in one dimension only</td>
</tr>
<tr>
<td><strong>Infiltration - Green Ampt equation, precipitation, snowmelt or irrigation event based</strong></td>
<td>One dimensional, vertical from soil surface into soil</td>
</tr>
<tr>
<td><strong>Soil surface water content – hourly modeling of evaporative flux</strong></td>
<td><strong>Overland flow hydraulics – sheet and rill flow</strong></td>
</tr>
<tr>
<td>One dimensional model, vertical from atmosphere into soil</td>
<td>One dimensional down slope</td>
</tr>
<tr>
<td><strong>Water balance - daily soil water balance of infiltration, irrigation, plant water use, percolation</strong></td>
<td><strong>Subsurface hydrology - percolation, lateral flow, resurfacing and tile drainage</strong></td>
</tr>
<tr>
<td><strong>Subsurface hydrology - percolation, lateral flow, resurfacing and tile drainage</strong></td>
<td>Two dimensional, Vertical into soil and horizontal down slope</td>
</tr>
<tr>
<td><strong>Management – soil disturbance and biomass manipulation</strong></td>
<td><strong>Soil - disturbance by tillage and natural processes</strong></td>
</tr>
<tr>
<td>One dimensional, vertical from crop into soil</td>
<td>One dimensional, vertical from soil surface into soil</td>
</tr>
<tr>
<td><strong>Soil - re-consolidation, re-aggregation of disturbed soil due to rainfall, drying, freeze/thaw, and freeze/dry events</strong></td>
<td><strong>Crop - date, water and temperature effects with additions for stem, leaf and reproductive mass partitioning</strong></td>
</tr>
<tr>
<td>One dimensional, vertical from soil surface into soil</td>
<td>One dimensional, vertical from crop into soil</td>
</tr>
<tr>
<td><strong>Crop - date, water and temperature effects with additions for stem, leaf and reproductive mass partitioning</strong></td>
<td><strong>Plant growth - date, water and temperature effects, separate field crop and rangeland modules</strong></td>
</tr>
<tr>
<td>One dimensional, vertical from crop into soil</td>
<td>One dimensional, vertical from crop into soil</td>
</tr>
<tr>
<td><strong>Residue decomposition - surface and subsurface integrating water and temperature effects</strong></td>
<td><strong>Residue decomposition and management - surface and subsurface integrating water and temperature effects</strong></td>
</tr>
<tr>
<td>One dimensional, vertical from surface residue into soil</td>
<td>One dimensional, vertical from surface residue into soil</td>
</tr>
</tbody>
</table>
Technical impediments to common code

Conceptually, process modeling can be divided into the data inputs required to model the process, the algorithms to implement that process and the states modified by the process. WEPS and WEPP are coded primarily in FORTRAN 77, where the same concepts are embodied in subroutine (or function) calls (conceptually algorithms) and a combination of arguments and common blocks (conceptually the data inputs and states modified).

Model logical structure

Based on the number of common processes represented, it is hoped that the logical structure of the two models would be very similar. This is indeed the case. At the heart of the simulation, the state of a single simulation area is updated using a daily and sub-daily loop. A comparison of time scales and process ordering used in the two models follows:

<table>
<thead>
<tr>
<th>WEPS</th>
<th>WEPP</th>
</tr>
</thead>
<tbody>
<tr>
<td>do all subregions</td>
<td>do overland flow elements</td>
</tr>
<tr>
<td>hydrology - subdaily calculations</td>
<td>decomp - tillage and weather process residue effects</td>
</tr>
<tr>
<td>management - tillage, planting, harvesting</td>
<td>soil - tillage and weather process soil effects</td>
</tr>
<tr>
<td>soil - weather processes</td>
<td>aspect - solar energy balance</td>
</tr>
<tr>
<td>crop - plant growth</td>
<td>winter - energy balance in winter - subdaily calculations</td>
</tr>
<tr>
<td>decomposition</td>
<td>irrig - irrigation flows</td>
</tr>
<tr>
<td>end do</td>
<td>irs - infiltration runoff simulation</td>
</tr>
<tr>
<td>erosion</td>
<td>watbal - soil water balance</td>
</tr>
<tr>
<td></td>
<td>newtil - plant growth</td>
</tr>
<tr>
<td></td>
<td>route - sediment routing down hillslope</td>
</tr>
<tr>
<td></td>
<td>sloss - resultant sediment loss on hillslope</td>
</tr>
<tr>
<td></td>
<td>watershed - channel and impoundment routing</td>
</tr>
</tbody>
</table>

Model code structure

The degree of impediment to using common code for processes common to both models is directly proportional to the magnitude of common block usage. Implementing a common module requires defining the module with the appropriate input and output definitions, passed to the module as parameters, and then implementing WEPS or WEPP wrapper code to include the appropriate common blocks and pass the values to the common module. Common blocks are heavily used in present model implementations.

Opportunities
The comparisons above reveal several patterns in the implementation of erosion model code. Opportunities for employing common code are clearly shown in areas where the space, time dimensionality is the same and where the processes are similarly modularized. Climate generation of all but wind are external modules and are presently common. The plant growth process is the most likely candidate for common code being modularized identically. Management processes and their effect on crop, soil and residue are the next logical candidates, with a different modularization of the sub-elements, but full encapsulation of the overall process effects. A common module would need to include additional elements to describe the states needed for both erosion modules. The most beneficial and most difficult process to make common is hydrology. Spatial definitions are in conflict and the time scales vary for different elements. With continued development of process based models, analyses similar to this should be done to minimize the reinvention and re-implementation of modeling code and promote cooperative development efforts.

References


Comparison of wind erosion measurements in Germany with simulated soil losses by WEPS

R. Funk, ZALF, 15374 Müncheberg, Germany (rfunk@zalf.de)

E. L. Skidmore, USDA-ARS, Manhattan, Kansas 66506 (skidmore@weru.ksu.edu)

L. J. Hagen, USDA-ARS, Manhattan, Kansas 66506 (hagen@weru.ksu.edu)

Introduction

Wind erosion is a serious problem in the northeastern parts of Germany. These regions are characterized by sandy soils, low precipitation and the transition to continental dry climatic conditions. The highest climatic erosivity in March and April coincide with the lowest resistance of the soils against wind erosion, caused by seedbed preparations and drilling. The problematic situation is increased by bare or only sparse covered soil surfaces as well by shelterbelts without leaves. Therefore wind erosion occurs especially in spring on fields of sugar beets, corn and other summer crops (Frielinghaus & Schmidt 1993).

Measurements of wind erosion have been carried out since 1991 beginning with a German project to develop a wind erosion model (Kuntze et al. 1989, Kruse 1994). In the first 3 years as much erosion data as possible should be collected in combination with all relevant meteorological parameters. For that aim an erosion plot of 2.25 ha was installed and equipped with sediment traps and a meteorological station to measure the wind erosion processes in a high spatial and temporal resolution (Funk 1995). These data are the basis for the first comparison between measured and simulated soil losses by wind erosion in Germany with the Wind Erosion Prediction System (Hagen et al. 1995).

Methods

Soil transport was measured with two sampler designs. Four automatic SUSTRA (SUspension Sediment TRAp, Fig. 1) were placed in the centre of the field with inlet heights in 5, 15, 25 and 45 cm. Weight of trapped sediment, wind speed and direction were stored as 10-minutes average in a data logger. Additionally 16 MWAC (Modified Wilson and Cooke Catcher, Fig. 2) were used, which were arranged in a grid of 25 m to measure the spatial distribution of the horizontal fluxes.
Calculations for single events were made with the WEPS erosion submodel (Hagen 1997). Unfortunately not all needed input parameter of field surface conditions were measured and therefore some had to be estimated. The estimation is based on following assumptions:

1. for the first erosion event after a tillage operation the roughness parameters (random roughness and ridge height) were varied to set the model used threshold wind speed equal to the measured;
2. the first erosion event after a tillage operation always got the highest roughness data and a non-crusted surface;
3. following erosion events always got decreasing roughness values (or at least the same), increasing parts of crust fraction and decreasing parts of loose erodible material depending on rainfall and erosion between the single events.

The surface crust cover fraction was calculated in dependency on cumulated rainfall with an equation of Zobeck and Popham (1992, in Hagen et al. 1995).

This gradually change of the inputs was continued until the next tillage operation. These assumptions seem to be reasonable to describe several erosion events in succession. The accuracy of the simulation depends much more on the relations between the events than on a good fit to one single event. Ridge width and ridge spacing were kept constant to 100 mm and 150 mm respectively, because the influence of these parameters was not so important.

**Results**

All in all 21 erosion events were selected and compared with simulated soil losses by WEPS. The results show a very good agreement between the measured and simulated soil losses with $R^2 = 0.98$ for the SUSTRA and $R^2 = 0.93$ for the MWAC.

Table 1: Measured and simulated soil losses for selected erosion events

<table>
<thead>
<tr>
<th>storm date</th>
<th>SUSTR A (kg/m)</th>
<th>BOSTRA (kg/m)</th>
<th>WEPS (kg/m)</th>
<th>Soil loss (kg/m²)</th>
<th>Random roughness</th>
<th>Ridge height (mm)</th>
<th>Crust fraction of LEM</th>
<th>fast tillage operation</th>
<th>cumulated precip. (mm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>14/04/92</td>
<td>33.8</td>
<td>44.7</td>
<td>27.9</td>
<td>0.37</td>
<td>4.0</td>
<td>20</td>
<td>0</td>
<td>1</td>
<td>08/04/92 4.8</td>
</tr>
<tr>
<td>21/04/92</td>
<td>936.4</td>
<td>714.9</td>
<td>860</td>
<td>10.46</td>
<td>3.0</td>
<td>0</td>
<td>0</td>
<td>1</td>
<td>7.4</td>
</tr>
<tr>
<td>04/05/92</td>
<td>7.3</td>
<td>8.8</td>
<td>6.8</td>
<td>0.15</td>
<td>3.0</td>
<td>0</td>
<td>0.42</td>
<td>0.4</td>
<td>25.3</td>
</tr>
<tr>
<td>12/05/92</td>
<td>7.5</td>
<td>23.9</td>
<td>29</td>
<td>0.37</td>
<td>4.0</td>
<td>10</td>
<td>0</td>
<td>1</td>
<td>05/05/92 11.6</td>
</tr>
<tr>
<td>15/05/92</td>
<td>1.85</td>
<td>7.3</td>
<td>8.7</td>
<td>0.11</td>
<td>3.0</td>
<td>0</td>
<td>0.42</td>
<td>1</td>
<td>27.3</td>
</tr>
<tr>
<td>18/05/92</td>
<td>5.4</td>
<td>43.3</td>
<td>50.5</td>
<td>0.59</td>
<td>3.0</td>
<td>0</td>
<td>0.42</td>
<td>1</td>
<td>27.5</td>
</tr>
<tr>
<td>27/05/92</td>
<td>31.5</td>
<td>52.6</td>
<td>42</td>
<td>0.53</td>
<td>3.0</td>
<td>0</td>
<td>0.42</td>
<td>1</td>
<td>27.5</td>
</tr>
<tr>
<td>05/06/92</td>
<td>101.2</td>
<td>190.2</td>
<td>151</td>
<td>1.8</td>
<td>3.0</td>
<td>0</td>
<td>0.42</td>
<td>1</td>
<td>03/06/92 0.1</td>
</tr>
<tr>
<td>10/06/92</td>
<td>27.9</td>
<td>52.5</td>
<td>45</td>
<td>0.5</td>
<td>3.0</td>
<td>0</td>
<td>0.37</td>
<td>1</td>
<td>5.6</td>
</tr>
<tr>
<td>29/07/92</td>
<td>274.1</td>
<td>254.5</td>
<td>225</td>
<td>2.4</td>
<td>2.0</td>
<td>0</td>
<td>0.42</td>
<td>1</td>
<td>11/06/92 27.3</td>
</tr>
<tr>
<td>08/04/93</td>
<td>7.77</td>
<td>6.5</td>
<td>7.8</td>
<td>0.11</td>
<td>4.0</td>
<td>25</td>
<td>0.37</td>
<td>1</td>
<td>29/03/93 20.3</td>
</tr>
<tr>
<td>20/04/93</td>
<td>50.3</td>
<td>126.2</td>
<td>101</td>
<td>1.29</td>
<td>4.0</td>
<td>10</td>
<td>0.6</td>
<td>1</td>
<td>4.5</td>
</tr>
<tr>
<td>23/04/93</td>
<td>22.6</td>
<td>18.2</td>
<td>24.4</td>
<td>0.3</td>
<td>3.0</td>
<td>5</td>
<td>0.37</td>
<td>1</td>
<td>4.5</td>
</tr>
<tr>
<td>26/04/93</td>
<td>39.9</td>
<td>29.6</td>
<td>57.3</td>
<td>0.64</td>
<td>3.0</td>
<td>5</td>
<td>0.4</td>
<td>0.6</td>
<td>4.5</td>
</tr>
<tr>
<td>30/04/93</td>
<td>36.4</td>
<td>45.6</td>
<td>30.9</td>
<td>0.4</td>
<td>2.5</td>
<td>5</td>
<td>0.4</td>
<td>0.6</td>
<td>4.5</td>
</tr>
<tr>
<td>10/05/93</td>
<td>46.6</td>
<td>48.6</td>
<td>44.6</td>
<td>0.64</td>
<td>2.0</td>
<td>0</td>
<td>0.4</td>
<td>0.2</td>
<td>5.3</td>
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<tr>
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<td>22.4</td>
<td>32.5</td>
<td>0.35</td>
<td>2.0</td>
<td>0</td>
<td>0.4</td>
<td>0.2</td>
<td>5.3</td>
</tr>
<tr>
<td>02/06/93</td>
<td>15.5</td>
<td>22.9</td>
<td>28.4</td>
<td>0.33</td>
<td>3.0</td>
<td>15</td>
<td>0.4</td>
<td>1</td>
<td>26/05/93 18.6</td>
</tr>
<tr>
<td>16/06/93</td>
<td>11.7</td>
<td>25.8</td>
<td>43.6</td>
<td>0.68</td>
<td>3.0</td>
<td>10</td>
<td>0.6</td>
<td>1</td>
<td>60.8</td>
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<tr>
<td>08/07/93</td>
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<td>449.1</td>
<td>292</td>
<td>3.53</td>
<td>4.0</td>
<td>0</td>
<td>0.43</td>
<td>0.2</td>
<td>22/06/93 0</td>
</tr>
<tr>
<td>27/07/93</td>
<td>31.3</td>
<td>37.6</td>
<td>56.2</td>
<td>0.78</td>
<td>3.0</td>
<td>0</td>
<td>0.43</td>
<td>0.2</td>
<td>28.4</td>
</tr>
</tbody>
</table>

Conclusions

The first comparison between measured and simulated soil losses by WEPS in Germany shows satisfying results. This includes the total soil loss for an event, the spatial variations on the field and the temporal changes in transport capacity. The estimation of all missing parameters was handled very carefully with respect to all available information, to reduce the uncertainty and to minimize subjectivity.

References


http://www.weru.ksu.edu/symposium/proceed/hagen.pdf


Modeling of Wind-blown Dust Emissions at Owens (dry) Lake, CA

Dale Gillette, Air Resources Laboratory, NOAA, Research Triangle Park, NC 27711 (gillette.dale@epa.gov)

Duane Ono, Great Basin Unified Air Pollution Control Dist., 157 Short Street, Bishop, CA 93514

Kenneth Richmond, MFG, Inc., 19203 36th Avenue West, Suite 101, Lynwood, WA 98036 (krichmond@mfgsea.com)

Abstract

A method for estimating $F_a$, the vertical flux of wind erosion resuspension particles smaller than 10 micrometers (PM10), was developed for Owens Lake, a large dust-source area in California. Owing to the size of the potentially dust-emitting dry lake bed (about 130 km$^2$) and the large effort and expense to rigorously measure vertical mass fluxes using micrometeorological methods, an alternate method was developed to use cheaper and more easily accomplished measurements. The method used the model of Shao that expresses the vertical particle flux of suspended particles produced by impact of saltating particles

$$F_a = K' q$$

where $\Psi$ is binding energy, $K'$ is a constant, $q$ is the integrated horizontal flux of airborne particle mass from the ground to 1 meter height and $m_d$ is the mass per particle. We assumed that the ratio $F_a/q$ [equivalent to $K' m_d/\Psi$] could be taken as a constant for scales of a few hundred square meters of the lake-bed and could be evaluated by using field-scale experimental data. Our method required detailed large scale measurements of $q$, high quality measurements of PM10 at several locations near the shoreline of Owens Lake and use of a transport model for emitted dust from the lake surface. We estimated $F_a/q$ values for the most active areas of dust emissions at Owens Lake as follows:

1. Sand fluxes ($q$) were measured in a grid of 130 sand flux samplers. The grid has a separation distances of 1 km.
2. Concentration of PM10 were measured at several locations along the shoreline of Owens Lake.
3. Concentrations were modeled using CALPUFF. Initially, a “first guess” value for $F_a/q$ was used with the 130 $q$ values to give the $F_a$ (vertical fluxes of PM10) for each 1 km$^2$ area of Owens Lake. These $F_a$ values were used in the model and the concentration field is calculated.
4. The ratios of the calculated concentrations at the locations of the TEOM instruments to the actual concentrations are found. Using the mean of these ratios the “first guess” $F_a/q$ value is adjusted so that the predicted and measured concentrations agree–one $F_a/q$ value is found for each hour of a dust storm.
(5) Owens Lake was divided into three areas having similar surface characteristics. Using appropriate data selection, one-hour $F_a/q$ values that apply to the three distinct source areas of the lake were calculated.
Long-term simulation of dust distribution with the GOCART model: Correlation with the North Atlantic Oscillation

Paul Ginoux, GEST-UMBC, NASA GSFC, Code 916, MD-20771 (ginoux@rondo.gsfc.nasa.gov)
Joe Prospero, RMSAS, U. of Miami, FL-33149 (jprospero@rsmas.miami.edu)
Omar Torres, JCET-UMBC, NASA GSFC, Code 916, MD-20771 (torres@tparty.gsfc.nasa.gov)

Introduction

Long term measurements of surface concentration at Barbados and Miami show a strong year-to-year variability that is apparently linked to various meteorological factors including climate conditions in North Africa (Prospero, 1999). Optical thickness retrieved from Total Ozone Mapping Spectrometer (TOMS) satellite since 1979 presents also a significant inter-annual variation over the North Atlantic (Torres et al., 2002). If one wants to understand the observed inter-annual variability and eventually to predict dust distribution, transport models are useful to determine which processes (e.g. emission, transport, or removal) are driving such variability. The GeorgiaTech/Goddard Ozone Chemistry Aerosol Radiation Transport (GOCART) model is used to simulate dust distribution from 1981 to 1997. The model results are compared with observed surface concentration and TOMS aerosol index, and are correlated with the North Atlantic Oscillation Index, defined by Hurrell (1995).

GOCART Model Description

The GOCART model is a multi-components aerosol transport model which simulates the global distribution of dust, sulfate, carbonaceous and sea-salt aerosols. The detailed description of the model components have been given elsewhere (Chin et al., 2000; Ginoux et al., 2001; Chin et al., 2002), here we give a summary of the dust component. The model has a horizontal resolution of 2° latitude by 2.5° longitude and 20-30 vertical sigma layers, and uses the assimilated meteorological fields generated from the Goddard Earth Observing System Data Assimilation System (GEOS DAS). The particle size distribution is discretized into four size bins: 0.1-1, 1-3, 3-6 m. The continuity equations of mass concentration for each size bins are treated independently. The equations include emission, transport and removal processes. A new approach has been used in GOCART to identify the dust sources. Using Total Ozone Mapping Spectrometer (TOMS) aerosol index, Prospero (2002) have identified and characterized the geomorphological characterization of the major dust sources. Based on this analysis and the previous work by Herman et al. (1997), Ginoux et al. (2001) have defined a global dust-source function. Figure 1 shows the comparison between the dust source function and climatological TOMS Aerosol Index. The function is constructed as the probability of sediments accumulated in topographic depressions with bare surface. Dust emission depends on the cubic power of the surface winds speed, and the threshold velocity of
wind erosion. Transport processes include advection by winds, convection by clouds, and diffusion by turbulent mixing. Removal processes include gravitational settling, surface deposition, and wet deposition (scavenging in convective updrafts and rainout/washout in large scale precipitation).

Figure 1. Comparison between GOCART dust sources and the TOMS Aerosol Index.

**GOCART Simulation and Comparison with the Observations**

The dust distribution has been simulated from January 1981 to 1998. Figure 2 shows the comparison of the monthly dust concentration observed at Barbados (13.17°N, 59.43°W) and simulates with GOCART. The model captures the main seasonal and year-to-year variability.

Figure 2. Comparison between observed (dash line) and simulated (bold line) monthly mean surface concentration at Barbados from January 1981 to December 1997.

Torres et al. (2002) have derived optical properties from the backscattered near-ultraviolet from Total Ozone Mapping Spectrometer (TOMS) on board the Nimbus 7 (1979-1992) and the Earth Probe (mid-1996 to present). The advantages of the near-UV technique are the high sensitivity to absorbing aerosol which can be separated from non-absorbing aerosol (e.g. sulfate), and the aerosol retrieval over both land and ocean. In
case of dust particle, the optical thickness is a function of the observed radiances, the viewing angles, the altitude of the dust plume and the single scattering, assuming the refractive index. Instead of calculating the dust optical thickness from GOCART and comparing with Torres et al. (2002) data, we calculate an equivalent TOMS AI. The advantage is that there is only one set of assumptions and they are on the model side. The refractive index of dust particles is assumed to be 1.58, for the real part, and 0.0064, for the imaginary part. A detailed description of the TOMS aerosol index (TOMS AI) is given by Herman et al. (1997). Figure 3 shows the correlation coefficient between TOMS AI and the Index calculated with GOCART results. Over all the dusty regions, the correlation is very strong (higher than 0.8). The slope of the correlation (not shown) is around one which suggests that GOCART model reproduces correctly the amplitude of the variability.

Figure 3. Correlation coefficient between monthly mean TOMS AI and the equivalent index calculated with GOCART, for 1981 to 1992.

Correlation with the North Atlantic Oscillation

The NAO index is defined from the difference between normalized sea-level atmospheric pressures between Lisbon, Portugal, and Stykkisholmur, Iceland (Hurrell, 1995). Winters with high NAO indices are characterized by a deepening of the Icelandic low associated with a stronger Azores anticyclone. This yields to higher surface pressure and drier conditions over Northern Africa. During low NAO conditions, there is an increase of precipitation over the Mediterranean and North Africa. The dust emission, loading, deposition have been correlated with the NAO Index. There are significant correlations in winter, but none in summer. The main reason is that the pressure gradient between Iceland and Portugal is much weaker in summer.
Figure 3 shows the correlation coefficient between dust emission from GOCART and the NAO Index. Over the Bodele depression (Lake Chad region) the correlation is higher than 0.8. The Bodele depression is the most active dust source in winter. Our results suggest that the NAO is a driving force for modulating dust emission in winter. Dust loading is significantly correlated with the NAO Index over the North Atlantic but not over Africa.

Conclusions

Dust size distribution is simulated with the GOCART model which is driven by assimilated meteorology. One major feature of GOCART dust emission is the use of an original dust source inventory based on topography and vegetation. The dust distribution has been simulated from 1981 and compare with surface concentration at Barbados and the TOMS AI. The GOCART results are well correlated with the observations. The simulated inter-annual variability has been correlated with the NAO Index. We find significant correlation with dust emission, loading and deposition, and the NAO Index in winter. Our results suggest that the inter-annual variability of dust over the Atlantic is a combination of variable dust emission and transport, both forced by the NAO.

References


Prediction Success with Integrated, Process-Based Wind-Erosion Model

J. M. Gregory, College of Engineering, Texas Tech University, Lubbock, TX 79409 (james.gregory@coe.ttu.edu)

M. M. Darwish, Engineering Technology Department, Texas Tech University, Lubbock, TX 79409 (mukadess.darwish@coe.ttu.edu)

Introduction

The value of any prediction tool is measured by how well the results from the tool match accurate measurements from the process to be modeled. The value is also measured by how easy data can be obtained to run the model, how easy the model is to use, and how wide a range of conditions over which the model can be used. It is desirable to have both an accurate and robust model.

The TEAM (Texas Tech Erosion Analysis Model) is an integration of many mathematical models each derived or developed to describe specific physical processes that affect the wind erosion process. To provide stability and efficiency in calculations, any differential mathematical form has been integrated to describe the process in real instead of differential space. TEAM has two major functions: (1) a maximum-transport component and (2) a length factor, which varies between zero and one. Input variables to each of these functions, such as threshold friction velocity, are often a function of other processes and variables. The combination of these variables defines a system of equations that use inputs of wind speed, relative humidity, soil particle size distribution, clay content, residue and vegetative cover, soil aggregate cover, field length, and wind break height and porosity to model wind erosion. Soil ridges perpendicular to wind direction are treated as cover if the ridge is solid or if the surface of the ridge is covered with stable aggregates. Ridges parallel to wind direction are considered non-effective as a treatment to reduce wind erosion.

Objective

The objective of this paper is to overview some of the unique features of TEAM and discuss the success of TEAM as a prediction tool to model wind erosion. Some limitations of TEAM will also be discussed.

Overview and Discussion

TEAM is a dynamic model that predicts real time rate of soil movement by wind. It does have two time-limiting processes. TEAM includes a gust factor to account for variable wind conditions. The gust factor is based on a 2- to 5-second minimum time period to avalanche particle movement with the saltation process. This is equal to about 10 particle bounces. TEAM also uses an average wind profile to calculate friction.
velocity. Greeley and Iversen (1985) have indicated that 30 minutes is sufficient to determine an average profile. Our experience is that 10 minutes is usually sufficient to determine a reasonable average profile for given wind conditions. The gust factor we use is the ratio of the gust factor for 2 to 5 seconds, which is approximately constant in this time range, over that for the time period used to determine the average wind profile. The gust factor for a 1-hour time period is 1.0. That for a 2- to 5-second period is about 1.5. Thus, if hourly wind speeds are used as wind input, we use a gust factor of 1.5. This factor goes down as we reduce the period over which we average the wind speed. A practical lower time limit for TEAM is about 10 minutes, although it has been used to generate soil detachment and dust generation with the pressure wave and wind velocity resulting from explosions.

There is considerable confusion in the literature about the effects of soil moisture on threshold friction velocity. TEAM calculates threshold friction velocity based on relative humidity of the air. The equation for threshold friction velocity was developed from data from Darwish (1991) and Belly (1964). These researchers were careful to make measurements with the soil system in equilibrium with the air above the soil. Some researchers have attempted to measure soil moisture for surface conditions by collecting surface soil 0.5 to 1 cm deep. This technique grossly miss calculates the soil moisture at the surface when the relative humidity of the air is not in equilibrium with the soil, especially for sandy soils with very low unsaturated hydraulic conductivity. A simple calculation shows the error of this process. We have noticed other researcher misuse our threshold friction equation by using measured soil moisture near the surface instead of using the equations to convert air relative humidity to surface soil moisture then making the calculation. While this process of calculating threshold friction velocity is generally an advantage for the TEAM model, it caused one limitation in use. TEAM will under estimate surface soil moisture immediately after a rain. TEAM moisture equations are not accurate until stage I drying (surface appears wet) has pasted. For sands, the time for stage I drying is often less than one-half day. We correct for this problem with long-term simulations by multiplying by a factor that expresses the probability of the surface being wet, which is based on soil type and the probability of getting rain (Solorzano-Campos, 1990).

TEAM is an energy-based energy-driven wind erosion model. The maximum transport component of TEAM uses the average sized non-clay particle to determine threshold friction velocity. Theoretically, we should use the mode of the particle size distribution; however, the average, D_{50}, is a close approximation of the mode and is less confusing to the user. Thus, the model is more robust by using the mean particle size with little loss in accuracy. TEAM also uses the D_{75} or average of the upper one-half of the particle size distribution less than 1 mm in diameter as an indicator of potential to return wind energy to the soil surface through the saltation process. This 1-mm upper limit is very close to the 0.8-mm diameter used by USDA researchers (Chepil and Woodruff, 1963). Theoretically, wind velocities exist during some dust storms that could move particle sizes larger than 1 mm. Generally, however, soils do not contain particles in the 1 to 8 mm range (Pettijohn et al., 1972). Winds high enough to move particles larger than 8 mm, such as roofing gravel, sometime exist but are too short in duration to contribute to significant soil movement. Particles less than 0.08 mm in diameter return to the surface too slowly to transfer significant detachment energy. Thus, the difference
between \(D_{75}\) and 0.08 is used to consider the influence of particle size distribution on the maximum transport rate. The ideas are similar to work by Bagnold (1941).

When wind speeds are high compared to threshold wind velocity, the maximum transport rate is a cubic function of wind speed and varies with the soil particle size distribution as observed by Bagnold (1941). The maximum transport rate also varies with relative humidity because the energy needed to detach loose soil particles varies with the threshold friction velocity. This component of TEAM has a theoretical basis and was verified with wind tunnel measurements (Wilson, 1994). Thus, the maximum transport component of TEAM is both dynamic and robust including effects of wind, water, and soil.

The length effect for TEAM was first derived by Gregory (1984) and later re-derived as reported by Wilson (1994) to provide a more stable and theoretically based function for the avalanching effect. The original derivation considered length of field, soil erodibility, and cover conditions. The newer derivation also considers wind speed, soil moisture through relative humidity, and particle size distribution. With the newer derivation, the length factor is dynamic through out a windstorm. For relatively dry pure sand, the length before maximum transport is achieved varies between 2 and 4 m depending on wind speed. For soil conditions, especially with about 30 percent clay content, the length until maximum transport is achieved can be several hundred meters. The length factor in TEAM is robust enough to describe the full S-shape of soil detachment and transport with length of erosion area.

The current version of TEAM is easy to calibrate to new field and weather conditions. It requires wind speed at a known height, relative humidity, clay content, soil particle size distribution to get \(D_{50}\) and \(D_{75}\), field length, aggregate diameter and height of cover, residue cover and plant canopy cover and height. Once soil and cover conditions are known and entered only wind and relative humidity are needed to simulate an erosion event. The soil erodibility that is the primary controlling variable in the length function is calibrated from clay percentage for solid soil conditions typical after a wet period or a tilled condition after a wet period. Both an advantage and limitation occurs with this variable. If several cycles of light rainfall occur that do not saturate the soil, the weathering associated with the wetting and drying cycles will greatly reduce the soil strength. TEAM has a weathered or "drought condition" calibration, which increases soil erodibility by four times the normal amount (Singh, 1992). This alternative erodibility selection increases the robustness of TEAM but leaves the user with a judgment as to which setting to use. The current TEAM does not have a set of equations to automatically calculate this observed weathering effect. More research is needed to develop weathering equations that are both accurate and robust for field use.

**Results**

TEAM generally predicts measured results with an \(R^2\) of 0.8 or better and generally has an \(\alpha\) level of 0.01 or lower indicating it is not a random fitting of numbers. Wind-tunnel data generally does not have gust effects unless the experiment was designed to produce gust. Thus, the gust factor for wind-tunnel data is 1.0. Otherwise, calibration for wind-tunnel data is the same as field data, provided the wind profile in the wind tunnel is in equilibrium with the surface roughness conditions of the experiment.
We have attempted to show the prediction success of TEAM for a variety of sites and conditions in Table 1. We are especially pleased with the success in predicting the results for the Big Spring, Texas data. We calibrated for this site using the average clay percentage, soil particle size distribution, and the reported description of surface and previous rainfall history to determine whether to use normal or weathered condition. We entered the average of 10-minute wind velocity and relative humidity from date and time of cleaning samplers to date and time for new cleaning. TEAM predicted the correct downwind total movement and predicted the correct movement with length of field.

<table>
<thead>
<tr>
<th>Type of Prediction</th>
<th>( R^2 )</th>
<th>( \alpha )</th>
<th>Data Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>Transport rate as a function of field length and soil type assuming a constant value for wind speed and humidity</td>
<td>0.93</td>
<td>0.001</td>
<td>Chepil (1957)</td>
</tr>
<tr>
<td>Transport rate as a function of friction velocity and relative humidity in a wind tunnel for 0.55-mm diameter sand</td>
<td>0.97</td>
<td>0.001</td>
<td>Wilson (1994)</td>
</tr>
<tr>
<td>Total amount of soil moved as a function of length and variations in wind velocity and relative humidity and soil conditions for the 5 largest or most complete small movement individual storms</td>
<td>0.82</td>
<td>0.001</td>
<td>Big Spring, TX field data collected by Bill Fryrear</td>
</tr>
<tr>
<td>Total amount of soil moved as a function of length and variations in wind velocity and relative humidity and soil condition for the largest storm</td>
<td>0.89</td>
<td>0.001</td>
<td>Big Spring, TX field data collected by Bill Fryrear</td>
</tr>
</tbody>
</table>

Conclusions

TEAM is a comprehensive wind erosion model that accurately predicts particle movement ranging from sands to soils with zero to 30 percent clay content. It accurately responds to both wind and relative humidity weather data and accurately predicts changes in particle movement as a function of length. While the focus of this paper has been on predictions for bare soil conditions, the cover factor was verified with several data sets by Gregory (1984). It is, thus, close to a complete wind erosion model. It currently has one weakness--it lacks the ability to adjust soil erodibility on a continuous bases as a function of rainfall and wetting and drying cycles. It currently appears to have the correct calibration for both non-weathered and fully weathered conditions but does not have a function to adjust erodibility between these two conditions.

TEAM has several factors often not included in other wind erosion models. These factors include relative humidity, wind gust factor, particle size distribution, and a dynamic length factor. It also self adjusts displacement height and aerodynamic roughness as a function of surface roughness conditions. Even with all of these considerations, TEAM is relatively easy to use and quickly calculates soil movement. A
more complicated programming of TEAM even allows up to 10 sequential length segments that can have different soil and/or cover conditions. In this form, TEAM can consider wind breaks and also predict deposition. TEAM is, thus, a very robust model.

Summary

TEAM is a process-based prediction tool that simulates the movement of sand or soil by wind. Many of the functions were originally derived from a differential expression describing the process. All of these differential forms, however, were integrated to obtain final function in real time and space. This integrated form increases both stability and speed in calculations.

As shown in the Results section, TEAM is a successful predictor of either sand or soil movement caused by wind. Part of the success of TEAM is due to the inclusion of unique features such as a gust factor, relative humidity, particle size distribution, and a dynamic length factor. TEAM is, thus, a very robust yet relatively easy to use single-event wind erosion prediction tool.

References


Validation of WEPS Erosion Predictions for Single Wind Events

L.J. Hagen, USDA, ARS, Wind Erosion Research Unit, Kansas State Univ., Manhattan, Kansas
(hagen@weru.ksu.edu)

Introduction

The Global Change and Terrestrial Ecosystems Soil Erosion Network (GCTE-SEN) has conducted a model validation exercise for water erosion models (www.nmw.ac.uk/GCTEFocus3/networks/erosion.htm). Similar to water erosion models, wind erosion models also are widely used to design control practices and to estimate both on-site and off-site erosion impacts. But most wind erosion models have not had extensive validation. Hence, a GCTE-SEN model validation project has been initiated for wind erosion models. Data on selected storm events collected during the last decade by ARS scientists and various cooperators (Fryrear et al., 1991) were distributed to participating scientists for model validation tests (Zobeck, et al., 2001). In this study, we compared observed soil loss with simulated soil loss predictions for individual storms using the erosion submodel of the Wind Erosion Prediction System (WEPS), as part of the GCTE-SEN model validation exercise.

The WEPS model is a process-based, daily time-step model that simulates weather, field conditions, and wind erosion on crop lands (Hagen et al., 1995; Wagner, 1996). The WEPS has a modular structure that includes a daily weather simulator along with an hourly wind speed simulator. There are five additional submodels in WEPS, and these simulate crop growth, residue decomposition, hydrology, soil status, and management operations. When wind speed exceeds the threshold for erosion, the erosion submodel simulates erosion on a subhourly basis.

During erosion, the horizontal saltation/creep soil discharge has limited transport capacity, while the horizontal suspension soil discharge has nearly unlimited transport capacity from individual fields. Hence, the erosion submodel simulates these as separate components of the total erosion for each wind direction (Hagen, Wagner, and Skidmore, 1999). Based on conservation of mass, the saltation/creep discharge is simulated with two sources (entrainment of loose, mobile soil and entrainment of soil abraded from clods and crust) and three sinks (breakage of saltation/creep to suspension-size, trapping of saltation/creep, and interception by plant stalks). Similarly, the suspension component is simulated with three sources (entrainment of loose soil, entrainment of material abraded from clods and crust, and breakage from saltation/creep to suspension-size). Simulating the saltation/creep and suspension components separately, greatly facilitates estimating off-site erosion impacts (Wagner and Hagen, 2001).

Methods

The experimental sites were 2.5-ha., tilled circular areas located in larger fields that did not erode. Soil sediment samplers (Fryrear, 1986) were arranged in vertical clusters to sample the horizontal soil discharge of eroding soil between the surface and
one m height. Thirteen clusters were installed within each circular field. Six clusters were located at 60-degree intervals on each of two concentric circles with radii of 55 and 87 m. The remaining cluster was located at the center of the circle along with a meteorological tower, associated weather transducers, and data logger.

The horizontal soil flux (kg m$^{-2}$) from sediment samplers in each cluster was integrated to a height of two m to estimate the horizontal soil discharge (kg m$^{-1}$) at each cluster. The wind direction and distance to upwind field boundaries for each cluster were also calculated for each storm event. In our analyses, we fitted an empirical equation to the point-discharge cluster data to estimate the total soil discharge at 180 m downwind and divided the result by 180 to estimate the observed soil loss per unit area. The empirical equation providing the best least-squares error fit to most of the point-discharge data was

$$q = a + bX^c$$

(1)

where $q$ is the downwind horizontal discharge for a storm (kg m$^{-1}$), $X$ is downwind distance from a nonerodible boundary, and $a$, $b$, $c$, are empirical coefficients.

Wind statistics provided for each daily storm included maximum speed, average speed, and a wind factor (Fryrear, Saleh, and Bilbro, 1998). These statistics were used to calculate three parameters (scale, shape, and zero intercept) for a Weibull cumulative distribution of daily wind speed. Using the Weibull distribution, a synthetic distribution of subhourly wind speeds was generated that was symmetric about the daily maximum wind speed. These data were used to drive the erosion submodel simulations.

This study included data from 46 storms in seven locations in six states (Table 1.)

Table 1. Test sites and surface soil characteristics.

<table>
<thead>
<tr>
<th>Location</th>
<th>Soil Texture</th>
<th>Sand</th>
<th>Silt</th>
<th>Clay</th>
<th>Organic Matter</th>
<th>Calcium Carbonate</th>
<th>Number of storms</th>
</tr>
</thead>
<tbody>
<tr>
<td>Eads, CO</td>
<td>Clay loam</td>
<td>29.3</td>
<td>38.6</td>
<td>32.1</td>
<td>1.6</td>
<td>1.0</td>
<td>2</td>
</tr>
<tr>
<td>Elkhart, KS</td>
<td>Fine, sandy loam</td>
<td>68.1</td>
<td>21.5</td>
<td>10.4</td>
<td>0.7</td>
<td>0.0</td>
<td>1</td>
</tr>
<tr>
<td>Kennett, MO</td>
<td>Sand</td>
<td>90.0</td>
<td>7.1</td>
<td>2.9</td>
<td>0.7</td>
<td>0.2</td>
<td>8</td>
</tr>
<tr>
<td>Sidney, NB</td>
<td>Loam</td>
<td>39.8</td>
<td>42.9</td>
<td>17.4</td>
<td>2.3</td>
<td>0.0</td>
<td>4</td>
</tr>
<tr>
<td>Big Spring, TX</td>
<td>Loamy sand</td>
<td>83.6</td>
<td>8.4</td>
<td>8.0</td>
<td>0.3</td>
<td>0.0</td>
<td>24</td>
</tr>
<tr>
<td>Mabton, WA</td>
<td>Loamy sand</td>
<td>82.3</td>
<td>12.8</td>
<td>4.9</td>
<td>0.8</td>
<td>0.0</td>
<td>5</td>
</tr>
<tr>
<td>Prosser, WA</td>
<td>Silt loam</td>
<td>44.2</td>
<td>50.2</td>
<td>5.7</td>
<td>1.1</td>
<td>0.0</td>
<td>2</td>
</tr>
</tbody>
</table>
Results and Discussion

The average storm loss from the cluster measurements extrapolated to 180 m downwind was 0.82 kg m\(^{-2}\), while the average predicted soil loss was 0.64 kg m\(^{-2}\). The maximum differences between observed and predicted loss occurred during large erosion events where the predicted values were frequently less than those observed (Fig. 1). Validation of another model reported a similar response with this data set (Zobeck et al., 2001). Reasons for the differences include the scatter in the cluster data along the wind direction which suggested the initial field surfaces were not always uniform as assumed in the model. There were also uncertainties about some of the input field surface conditions when they were not measured close to the storm dates.

Linear regression of the storm data showed reasonable agreement between predicted and observed ($R^2 = 0.71$) with an intercept greater than zero. However, nonlinear regression using Eq. 1 showed that for storm losses less than 2 kg m\(^{-2}\) the predictions were close to the 1:1 line, and the intercept was slightly less than zero.

Figure 1. Measured versus predicted soil loss for 46 wind erosion storms.
References


Spatially explicit regional wind erosion and dust emission modeling: Incorporating large- and small-scale variability

G. S. Okin, Department of Environmental Sciences, University of Virginia, PO Box 400123, Charlottesville, VA 22904-4123, (okin@geog.ucsb.edu)

D. A. Gillette, Air Resources Laboratory (MD-81), Applied Modeling Research Branch, Research Triangle Park, NC 27711, (gillette.dale@epamail.epa.gov)

Introduction

Despite the importance of desert dust in global and regional scales, it is often unclear in detail where it is produced and what role humans play in mediating its production. The dust observed over North Africa, for example, certainly originates in the Sahara and Sahel regions, but current technologies do not allow unique identification of the loci in these landscapes of the greatest dust emission. Thus, controversy remains about the extent to which land use contributes to the atmospheric mineral dust. Recent work by Prospero et al. (In Press) and Ginoux et al. (2001) suggests that the overwhelming majority of desert dust comes from closed basins in arid areas related to now-dry or ephemeral lakes. They argue further that humans do not significantly perturb the dust cycle, a conclusion supported by Guelle et al. (2000). This point of view contrasts sharply with that of Tegen and Fung (1995) who suggest that land use may in fact dramatically affect the amount of dust emitted in desert regions.

We have created a spatially explicit wind erosion and dust flux model (SWEMO) that allows estimation of wind erosion and dust flux across a landscape by incorporating spatial distributions of important parameters. This approach provides a powerful basis for trying to understand how vegetation and soil interact in the landscape to create the dust sources. This approach is therefore applicable in trying to understand the most important or persistent dust sources in an area. The goal of SWEMO is to integrate soil and vegetation parameters from field studies or remote sensing in a robust model of dust sources. By explicitly incorporating random variations in derived parameter (e.g. lateral cover, threshold shear velocity on vegetated surfaces) and mass flux estimation in a Monte Carlo framework, SWEMO can accommodate the inherently nonlinear nature of wind erosion and dust flux. The inclusion of random variation in SWEMO highlights the importance of small but intense deflation surfaces on landscape-scale wind erosion and dust flux estimates.

Model Description

Wind erosion depends on several parameters that vary as a function of soil and vegetation cover. By using maps of dominant vegetation type and soil texture, SWEMO is able to impose spatial variability by allowing the main parameters that determine wind erosion and dust flux (see Table 1) to vary according to the specific soil and vegetation
found at any location. However, even if categorical maps of soil texture and vegetation type, with polygons labeled, for example, “sandy loam” and “creosote”, are 100% accurate, they do not represent the full variability of the landscape: among other things, the size and spacing of plants varies even among areas with the same polygon labels. Therefore, a stochastic modeling approach is implemented in SWEMO, allowing parameters to vary within a specific range. As a result, SWEMO is able to model, statistically, small areas not well represented by local averages. A small hole in vegetation, such as a natural disturbance, a road, a dry river, or a dry lake, may account for the majority of dust emitted in an area, but be insignificant on the scale at which most maps are produced.

SWEMO uses maps of soil texture and vegetation, in addition to knowledge of vegetation cover and size parameters, to derive maps of threshold shear velocity for a vegetates surface ($u_{ts}$) and $z_o$. For each cell in the model, a histogram of shear velocity is derived from a histogram of wind speed at one height using the value of $z_o$ at that cell. A mass flux equation (Shao and Raupach, 1993) is then evaluated for each cell to derive an estimate of total horizontal flux, $Q_{tot}$. A soil-texture based value of the ratio of vertical flux to horizontal flux is use to calculate vertical flux, $F_a$. The processing stream for SWEMO is depicted in Figure 1.

### Table 1. Relations between wind erosion model parameters and vegetation/soil parameters

<table>
<thead>
<tr>
<th>Model Parameter</th>
<th>Vegetation/Soil Parameter</th>
</tr>
</thead>
<tbody>
<tr>
<td>Threshold shear velocity of soil</td>
<td>$u_{ts}$</td>
</tr>
<tr>
<td>Displacement height</td>
<td>$D$</td>
</tr>
<tr>
<td>Roughness height</td>
<td>$z_o$</td>
</tr>
<tr>
<td>Basal/Frontal area ratio</td>
<td>$\sigma$</td>
</tr>
<tr>
<td>Drag Coefficient ratio</td>
<td>$\beta$</td>
</tr>
<tr>
<td>Lateral Cover</td>
<td>$\lambda$</td>
</tr>
<tr>
<td>Fractional Cover</td>
<td>$C$</td>
</tr>
<tr>
<td>Number Density</td>
<td>$N$</td>
</tr>
<tr>
<td></td>
<td>Soil grain size, crusting, disturbance</td>
</tr>
<tr>
<td></td>
<td>Plant height &amp; density</td>
</tr>
<tr>
<td></td>
<td>Plant height &amp; density</td>
</tr>
<tr>
<td></td>
<td>Plant height &amp; radius</td>
</tr>
<tr>
<td></td>
<td>Approx. constant (~100)</td>
</tr>
<tr>
<td></td>
<td>Plant height, radius, &amp; number density</td>
</tr>
<tr>
<td></td>
<td>Found directly from images</td>
</tr>
<tr>
<td></td>
<td>Fractional cover &amp; plant radius</td>
</tr>
</tbody>
</table>

SWEMO is depicted in Figure 1.

### Methods

In the present study, a site with all requisite data (soil maps, vegetation maps, wind data, and a wealth of ongoing ecological research) was chosen to test SWEMO. The Jornada Basin in south-central New Mexico is a part of the National Science Foundation’s Long-Term Ecological Research (LTER) network, and as such provides a wealth of required and ancillary data. The Jornada del Muerto basin lies approximately 30 km northeast of Las Cruces, NM, in the Chihuahuan Desert ecosystem.
Portions of the Jornada Basin have been mapped by the US Soil Conservation Service (1980). Each polygon in this soil map is labeled with a soil texture of the dominant soil type in that polygon, which allows estimation of $u_{*ts}$ and $F_a/Q_{Tot}$. The particle-limitation coefficient is assumed to be 1.0 for the entire study area. Values of $u_{*ts}$ used in this study were derived from mean values provided by Gillette (1988). Values of $F_a/Q_{Tot}$ were estimated using data from Gillette et al. (1997).

A vegetation map of portions of the Jornada Basin was made available to this study through the Jornada LTER project (digital data produced by R. Gibbens, R. McNeely, and B. Nolen). This map contained information on spatial distribution of the dominant plant communities in the basin: grassland, mesquite, creosote, tarbush, snakeweed, other shrubs, and no vegetation. Fractional cover for the grassland and snakeweed cover types and plant diameter and height for all vegetation cover types were derived from ongoing vegetation monitoring. Fractional cover for the creosote and

**Figure 1.** Processing stream for the spatially-explicit wind erosion and dust flux model (SWEMO)
tarbush cover types was derived from ongoing vegetation monitoring data as part of the Small Mammal Exclosure experiment at Jornada. Fractional cover for the creosote vegetation cover type was taken from Okin and Gillette (2001). Plants were assumed to be cylindrical in shape.

Wind monitoring by one of the present authors (D. Gillette) has been ongoing at several sites in the Jornada for many years. Data from one windy season (March 28, 2000 – July 10, 2000) was used in this study.

Results and Conclusion

By combining soil and vegetation maps with reasonable values for erosion-related soil and plant parameters, SWEMO has the potential of modeling wind erosion and dust flux on regional and landscape scales using relatively simple relations. The explicit incorporation of sub-grid variability through Monte Carlo simulation in the model allows more accurate estimation of derived-parameter and mass flux.

Because of the highly nonlinear nature of wind erosion and dust emission, they are highly sensitive to heterogeneity in the landscape. In particular, as heterogeneity increases, mass flux also increases for all surfaces. While increasing variability increases the incidence of both high and low erodibility surfaces, the high erodibility surfaces account for the bulk of the mass flux. Therefore, the modeling scheme used in SWEMO allows implicit modeling of erosion “hot spots” and more realistic estimation than modeling using mean surface parameters.

Acknowledgments

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References


**Time-averaged numerical modelling of airflow over an idealised transverse dune**

Parsons, D.R. Department of Geography, University of Sheffield, Western Bank, Sheffield, S10 2TN, UK (d.parsons@sheffield.ac.uk)

Wiggs, G.F.S. Department of Geography, University of Sheffield, Western Bank, Sheffield, S10 2TN, UK (g.wiggs@sheffield.ac.uk)

Walker, I.J. Department of Geography, University of Victoria, Victoria, British Columbia, V8W3P5, Canada (ijwalker@uvic.ca)

Garvey, B.G. Department of Geography, University of Sheffield, Western Bank, Sheffield, S10 2TN, UK (ggp99bgg@sheffield.ac.uk)

Ferguson, R.I. Department of Geography, University of Sheffield, Western Bank, Sheffield, S10 2TN, UK (r.ferguson@sheffield.ac.uk)

**Introduction**

There has been much recent interest in the complex interactions evident between sand dune morphology, windflow and sediment transport (McKenna-Neumann et al, 1997, 2000; Wiggs, 2001; Walker and Nickling, 2002). Research concerning secondary flow regimes governing dune/flow interactions and the existence of a dynamic equilibrium between dune morphology and windflow patterns has dominated recent dune literature (see reviews by Nickling and McKenna-Neumann, 1999; and Wiggs, 2001). The research focus on the dynamics of dune windward slopes (e.g. Lancaster et al, 1996; Frank and Kocurek, 1996a; Wiggs et al, 1996; McKenna-Neumann et al, 2000) has been complemented by similar field and wind tunnel studies investigating flow patterns and sediment dynamics in the turbulent lee-side eddies downwind of transverse dune crests (Frank and Kocurek, 1996b; Walker and Nickling, 2002). The acceleration of windspeed and surface shear stress to a maximum on the stoss slopes of sand dunes followed by a lee-side re-circulating eddy and region of flow recovery are now well documented. Progress is still hampered, however, by the small number of field sites investigated and by the limited dune geometries that have been experimented upon in wind tunnel studies. Additionally, due to design limitations, regions of highly turbulent or reverse flows have not been quantified in wind tunnel experiments, resulting in deficient understanding of the flow structure over dune forms.

Whilst physical experimentation has provided us with a substantial insight into a number of flow-form interactions (flow acceleration, crestal separation, re-circulation, re-attachment etc.), questions remain as to the sensitivity, structure, and dynamic function of these interactions with changing dune geometry. An adaptable and rapid method by which our understanding of flow patterns could be further improved involves the mathematical modelling of the flow field over different dune geometries. Previous attempts to model the turbulent boundary layer over isolated dunes (e.g. Howard et al,
The models used to calculate the flow structure over dune forms often have severe limitations. For example, Stam (1997) applied an analytical flow model based on a boundary-layer model (e.g. Jackson and Hunt, 1975), which is unable to solve the reverse flow in lee-side eddies often present over dune forms. This limits the calculation of flow structures to low angle dunes where lee-side eddies are not present. With the recent proliferation of field and wind tunnel data concerning dune processes it is now appropriate to apply new refinements in numerical calculations of flow fields around bedforms to questions of dune flow dynamics.

Numerical flow models (Computational Fluid Dynamics, CFD) have been widely applied in engineering disciplines for many years. In the last few years, there has been a proliferation of the use of CFD in the fields of geomorphology and hydrology (see Bates and Lane, 2000). These models enable an improved simulation of important processes providing prediction fields that allow considerable insight into the spatial distribution of these processes. CFD modelling has offered a new methodology that is complementary to traditional field and laboratory approaches. Indeed, the models can provide details of the flow field that are often difficult to measure and offer controlled conditions in which certain aspects of the experimental set up can be varied rapidly.

This paper represents the first stage in applying a numerical code to flow field patterns around dunes and validates a new numerical model against wind tunnel derived experimental data (from Walker and Nickling, in press). It provides details of model background set-up and application. Once validation is complete, the model may be used to examine the variation in the flow field around both isolated and closely-spaced dunes of differing geometries. This latter procedure will be the subject of an accompanying paper.

**Results: the numerical model**

In order to ascertain the capabilities of the model it was initially used to predict the measured flow velocities for the wind tunnel experiment of Walker and Nickling (in press). The model validation was based on 415 predicted points within the model domain, which coincided with the locations of measurements taken in the wind tunnel.
For the downstream velocity component the agreement is generally very good with a high correlation coefficient and the 1:1 line in close agreement for the higher velocity values (figure 1a). However, there is a significant zone of disagreement for the lower velocities, where the predicted velocities are negative whilst the measured are positive. These points are from the lee separation zone, where, due to design limitations, the probe was unable to resolve the highly turbulent and negative velocities apparent, which were correctly predicted by the model (figure 2). Although slightly weaker, the agreement between measured and predicted values of vertical velocity (figure 1b) is still high with most points clustered close to the origin. Nevertheless, there is considerable disagreement in three distinct areas: (1) a vertical line of points where the model predicts very high positive values in comparison with the slightly negative measured values; (2) a
group of points in the bottom left of the graph where the measured values are higher than
the predicted; (3) a small group of points that extends towards the top right between the
line of best fit and the 1:1 line. The first group of points is explained by the high positive
vertical velocities predicted by the model flowing up the lee of the dune in the separation
zone that are not correctly measured by the probe. The second area is explained by the
separation re-attachment, where the model predicts lower velocities overall, perhaps due
to energy dissipation through high turbulence intensities that would again compromise
the measurement probes performance. Finally, the third area of disagreement is due to
the model slightly over predicting the stoss side vertical velocity components.

Although there are notable differences between the measured and modelled results,
they are primarily due to the limitations with the measuring instrument rather that the
numerical model. In regions were the instrument is known to perform well the match is
very good. We therefore have confidence in the performance of the numerical model to
use it for both the investigation of the flow structures over the dune, particularly in the
lee, and for future experiments where different aspects of dune form can be investigated.

Results: the flow structure

Results indicate that the model predicts the flow patterns over the dune very
accurately. Figure 2 illustrates the predicted stream wise and vertical velocity over the
dune.

![Figure 2: Streamwise and vertical velocity contour maps](image-url)
The model produces regions of flow stagnation at the toe, acceleration up the stoss slope and a convincing region of flow separation in the lee of the dune. These results show broad similarity to a range of studies into flow dynamics over dunes, and permit a more in-depth analysis of the flow structures.

Conclusion

The numerical model presented in this study successfully predicts the flow structure over an idealized transverse dune and validates extremely well to experimental wind tunnel measurements. Moreover, the model seems to correctly simulate the flow in regions of high turbulence and flow reversal, where experimental limitations are unable to provide flow information. Thus, the model provides a complete picture of the flow structure, which is spatially much richer than results produced by current wind tunnel experiments and field studies. The ease of use and flexibility of the modelling allows testing of a variety of isolated and complex dune morphologies, which are subjects of forthcoming papers.

References


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Self-limiting Blowing Dust Mechanism

R.E. Peterson, Texas Tech University, Lubbock, Texas 79409 (E-mail: richard.peterson@ttu.edu)

Introduction

Historically the highest frequency of blowing dust days for the United States has occurred on the High Plains of Texas, centered on Lubbock (Changery, 1983). The synoptic climatology (i.e., the nature and distribution of the weather patterns) for blowing dust events across this region has been detailed by Wignier (1984) and Bernier (1995); see also Wigner and Peterson (1987). Data was available spanning about four decades for four stations: Amarillo NWS, Lubbock NWS, Reese AFB and Midland NWS. Unfortunately, with the closure of Reese AFB and the implementation of the automatic observational system (ASOS) by the National Weather Service, the continuity of the blowing dust data set has been broken.

For the four reporting stations there are significant differences in the dust source regions and also variations in the importance of possible high-wind generating mechanisms (e.g., strong cold fronts, deep cyclones, thunderstorm outflows etc.). The greatest dust frequency has been recorded in the center of the region at Lubbock NWS and Reese AFB (about 20 km from Lubbock in a more rural location); for these sites about 50% of the blowing dust hours occur triggered by the daytime mixing down of higher momentum air.

It is typical that the dry, cloudless air allows for substantial nighttime radiational cooling and the establishment of a strong but shallow surface-based temperature inversion. Due to the relatively low latitude (~33N), strong solar heating after dawn leads to a dry convectively unstable layer which evolves upward from the surface; this most often eventually eliminates the inversion. Thereafter, convective mixing proceeds rapidly to greater heights, with compensating downward motion transporting the winds with dust-raising potential to the ground. The greatest effect occurs during the winter. During this season, cloudiness is a minimum, the winds overhead are most vigorous and the ground cover is at a minimum.

There are at least two factors which can limit or curtail the heating responsible for the mixing-down process: cloudiness (either advected from elsewhere or generated by the mixing in situ) and/or dust raised by the winds. This paper considers a model for the self-limitation of the blowing dust event due to the increasing aerosol load in the atmosphere.

Clear-air Model

For the daytime convective modification of the lower atmosphere, Carson (1968) presented an analytic model which can be adapted to include the effect of the blockage of solar heating due to airborne dust. The formulation uses as input the initial vertical profile of the atmosphere, T(z) or θ(z), and the time variation of the surface heating, H(0, t). One of the key products of interest is the depth of the mixed layer as a function of time, h(t).
The one-dimensional (height), time-dependent heat equation is integrated over the depth of the mixed layer, incorporating the heating contributions both from the earth’s surface, as well as the reservoir of potentially warm air aloft.

The heat equation is expressed in terms of the individual change of potential temperature, $\theta$; changes are brought about by the vertical flux of sensible heating. In the decomposition of the total derivative of $\theta$ into the local time change and the vertical advection, allowance is made for large-scale vertical motion, $w(z)$:

$$\frac{\partial H}{\partial z} = -\rho c_p \left[ \frac{\partial \theta}{\partial t} + w \frac{\partial \theta}{\partial z} \right].$$

The simplest initial situation has a surface-based temperature inversion with a constant lapse rate of $\theta$ surmounted by a deep stable layer with a different constant lapse rate. Once heating begins, a surface-based layer with constant $\theta$ evolves; a very shallow superadiabatic layer near the ground is ignored.

With the assumption that the stable air layer above the mixed layer is dominated by a linear lapse rate, $\gamma$, then the vertical motion is constrained to vary linearly with height; the constant of proportionality, $\beta$ (subsidence parameter), is thus equivalent to the horizontal convergence (or divergence) of the upper layer.

A major element is the parameterization of the heat brought into the mixed layer from the layer above, $H(h, t)$; it is taken as proportional to the surface sensible heat flux:

$$H(h, t) = -A H(0, t)$$

where $A$ is taken to lie between zero and one. This allows the formulation of the governing differential equation for the mixed layer depth:

$$\frac{dh}{dt} + 2\beta h^2 = 2[H(0, t) - 2H(h, t)]/\rho c_p \gamma(t).$$

The resulting solution for the depth of the mixed layer is a function of the rate of subsidence aloft, and the time integral of the heating function and the lapse rate of the atmosphere aloft:

$$h^2(t) = h^2(0) \exp(-2\beta t) + 2\exp(-2\beta t) \int \exp(2\beta \tau) \left\{ \left[H(0, \tau) - 2H(h, \tau)\right]/\rho c_p \gamma(\tau) \right\} d\tau.$$

The integral limits are from 0 to $t$; $h(0) = 0$ if the initial inversion extends to the surface.

If the heating rate is taken as a simple sine function of the time since sunrise

$$H(0, t) = H \sin \Omega t,$$

where $H$ and $\Omega$ depend on the latitude and time of year,

then an analytic solution results:

$$h^2(t) = \left\{2(1 + 2A)H/\gamma(0)(\beta^2 + \Omega^2\rho c_p)\right\} \exp(-2\beta t) \left[\exp(\beta t)(\beta \sin \Omega t - \Omega \cos \Omega t) + \Omega\right].$$
Typical solutions for most parameter selections yield an almost linear increase with time in the depth of the mixed layer for half a dozen hours or so, with a leveling off as the heat input diminishes. For sufficiently strong large-scale subsidence though, the mixed layer depth may begin to decrease even while heating is substantial.

**Dusty Model**

The vertical mixing induced by the daytime heating not only redistributes the potential temperature but also other properties: humidity, momentum and of course dust. The effect is to produce constant values of each parameter over the mixed depth. The temperature lapse rate becomes dry-adiabatic (i.e., constant potential temperature with height). For humidity mixing usually results in a large decrease in the surface mixing ratio and the relative humidity. For momentum there is usually an increase in surface wind speed, accompanied by a shift of direction becoming increasingly like that of the winds aloft. As the surface wind increases, once a critical speed is attained, dust will be raised and be lifted.

As the dust volume increases, sunlight – and the surface heating - will be reduced. For simplicity the decrease may be taken as proportional to the amount of dust in the air, $D$:

$$H'(z, t) = H(z, t) - cD.$$

Likewise the amount of dust may be taken as proportional to the square of the wind speed, $v$:

$$D \propto v^2.$$

If the initial wind profile increases linearly with height

$$v(z, 0) \propto z,$$

then integrating over the depth of the mixed layer, 0 to $h(t)$, yields a time variation of the mean layer wind (which develops at the surface) proportional to the time variation of the square of the mixed depth, $h^2(t)$.

Reducing the assumed heating function by a term dependent on the square of the mixed depth in the governing differential equation allows the dust reduction factor to be combined with the subsidence term. The net effect is analogous to increasing the subsidence from aloft. This in turn limits and perhaps even reverses the growth of the mixed depth. The increased dust load then is a self-limiting development.

**Discussion**

For blowing dust events triggered by the mixing-down mechanism, the aerosol may be expected to fill the mixed layer with time. Whether dust is raised will depend of course on the availability of a dust source and therefore the direction of the wind. (In the recent years sources have become even more restricted due to the maintenance of ground cover throughout the year – the conservation reserve program.)
In the near future it may be possible to monitor the growth of the mixed layer within blowing dust events using observational platforms at the West Texas Mesonet site by serial launches of rawinsondes and/or by means of the lower atmosphere sounder. (Actual aerosol measurements may be made from the 200 m meteorological tower. This data will provide a test of the model predictions of mixed layer growth, in clear conditions as well as with blowing dust.

Conclusions

A simple model has been presented for the time evolution of the surface-based mixed layer assuming that wind-generated dust reduces the surface heating. It is hypothesized that the dust loading will lead to a limitation of the mixed-layer growth in a manner analogous to the role of large-scale subsidence.

References


Representation of Land-Surface Processes in Models of Wind Erosion

Michael R Raupach, CSIRO Land and Water, Canberra, ACT 2602, Australia (Email: Michael.Raupach@csiro.au)

Hua Lu, CSIRO Land and Water, Canberra, ACT 2602, Australia (Email: Hua.Lu@csiro.au)

Introduction

Quantitative assessment of wind erosion through mathematical modeling provides a useful tool to understand the spatial distribution and temporal dynamics of wind erosion and its impacts on climate and environmental changes. As much of the crucial action takes place at or just above land surfaces, a key issue is the correct representation of the land surface processes involved in particle uplift and deposition, and their dependencies on soil, vegetation and atmospheric variables.

This paper (of which the present abstract is a brief, incomplete summary) has two primary aims: to review the progress in key areas of process-based wind erosion and aeolian transport modelling during last decade, with an emphasis on surface processes, and to identify uncertainties and new research directions. Areas covered include (1) process-based saltation and dust emission models; (2) effects of vegetation on threshold friction velocity and dust uplift; (3) effects of vegetation on particle deposition; and (4) integrated wind erosion and dust transport modelling at large scales.

Saltation and Dust Emission Models

Sand grains saltating over a surface of loose fine particles excavate ovoid-shaped craters, by "saltation bombardment". Both field and wind tunnel experiments show that dust emission is mainly caused by this process. The resulting dust emission can be quantified by considering the relative values of sand-grain impact stresses on the surface and the soil surface strength. Lu (2000) showed that for typical values of impacting particle velocity and angle, the maximum surface pressure is about two orders of magnitude larger than the strength of the eroding soil surface. Under this condition, plastic deformation is the dominant mechanism for soil displacement. Lu and Shao (1999) derived an analytic dust emission model based on this idea, which yields the following prediction for the dust emission rate (for all dust particle sizes) caused by saltation of sand grains of diameter \( d \):

\[
F(d) = \frac{C_p g f \rho_b^{0.24 + C_p u_s^2}}{2p} Q(d)
\]  

(1)

where \( Q(d) \) is horizontal sand flux for particles with diameter \( d \); \( F(d) \) is the resulting vertical dust flux; \( \rho_p \) and \( \rho_b \) are the saltating-particle and soil bulk densities; \( u_s \) is the wind friction velocity; \( g \) is gravitational acceleration; \( f \) is the fraction of fine (dust) particles contained in the eroding soil; \( p \) is the plastic flow pressure of soil during impaction; \( C_p \) is the released fraction of dust contained in the volume removed by saltation bombardment; and \( C_p \) is a dimensionless coefficient. Integration of Equation (1) over a given sand particle size distribution gives the total dust emission rate. Lu and Shao (1999) showed that this model
compares well with field measurements from sandy to sandy loam soils, but poorly for soils with surface structure, such as clay soils.

There are strengths and weaknesses of process based dust emission models, such as Equation (1). Strengths include (a) insight into the dependencies of dust emission processes on soil properties, wind speed, and the intensity of saltation; and (b) representation of supply limited dust emission through the parameters $p$, $p_b$, $C_a$ and $f$, which relate to soil properties. Weaknesses include (a) issues in the application at large spatial and temporal scales, because of large-scale variability in microphysical parameters, and (more fundamentally) variability in the dominant basic processes; (b) difficulties in measuring or calculating parameters (although all parameters have physical meaning). For example, Lu (2000) showed by sensitivity analysis that the most sensitive parameter is $p$, which relates to the state of surface crusting but is very hard to measure. Its value changes with soil moisture and by freezing and thawing processes.

A model for dust emission by saltation bombardment depends on the existence of a saltation model. For transport-limited saltation over a loose sand surface, there is general agreement that the horizontal sand flux $Q$ is proportional to $u_{*}^{m_{*}}[1 - f(u_{*}/u_{t})]$ where $u_{*}$ is the threshold friction velocity for the eroding surface and $m = 3$. Different authors propose slightly different functional forms for the term $f(u_{*}/u_{t})$ which accounts for the threshold of sand drifting (Greeley and Iversen 1985), subject to the requirements that $f = 1$ for $u_{*}/u_{t} < 1$ (no drifting below threshold) and $f \rightarrow 0$ as $u_{*}/u_{t} \rightarrow \infty$ (so that $Q$ is proportional to $u_{*}^{3}$ at high wind speeds).

The situation for supply-limited saltation is not as clear, and depends on the mechanism by which the saltation supply is restricted. Possibilities include sheltering by vegetation (treated in the next section), moisture, and surface crusting.

**Threshold Friction Velocity and its Dependence on Vegetation Cover**

We first consider the inherent or bare-surface threshold velocity for particle uplift by wind, in the absence of vegetation. This is well known to depend on the balance of three forces: gravity, drag and interparticle cohesion (Greeley and Iversen 1985). Recent work (Lu and Raupach 2002) shows that consistent agreement with data for particle uplift in both air and water flows can be obtained from a simple expression of the form

$$\rho_s u_t^2 = A \left[ (\rho_p - \rho_s) g d + B/d \right]$$

where $\rho_p$ and $\rho_s$ are the particle and air densities, $d$ is particle diameter and $A$ and $B$ are empirical coefficients. The first term in this expression accounts for the gravity-drag interaction (dominating the threshold condition for large particles) and the second for the drag-cohesion interaction (dominant for small particles). As shown in Figure 1, this expression is successful in predicting laboratory observations of $u_t$, both in air (with $A = 0.0123$, $B = 3x10^{-8}$ N m$^{-1}$) and in water (with $A = 0.05$, $B = 7.6x10^{-5}$ N m$^{-1}$).
Vegetation has major effects on threshold velocity. An initial estimate of these can be obtained from consideration of drag partition, the ratio of the stress on the ground surface to the total stress including both ground and vegetation. Raupach et al. (1993) used drag partition theory to produce a formula for the effect of vegetation on wind erosion threshold:

\[ \frac{u_{\text{GR}}^2}{u_{\text{GR}}^2} = \frac{1}{(1-\sigma\lambda)(1+m\beta\lambda)} \]  

(3)

where \( u_{\text{GR}} \) and \( u_{\text{GR}} \) are the threshold friction velocities for bare-soil and vegetated (roughened) surfaces, respectively, \( \sigma \) is the basal-to-frontal area ratio, \( m \) is a parameter accounting for non-uniformity in the surface stress, and \( \beta = C_R/C_S \), where \( C_R \) is the drag coefficient for isolated roughness elements and \( C_S \) is that for the soil surface. Recent work has generally confirmed the validity of this model, while suggesting revised interpretations of some coefficients (especially \( m \)) and better values for the drag coefficients \( C_R \) for standing vegetation elements.

Effects of Vegetation on Particle Deposition

Besides particle uplift, the other major surface process affecting particle transport by wind is deposition. In general, the particle deposition flux \( D \) to a surface can be represented as \( D = W_{d}C \), where \( C \) is the particle concentration at a reference level above the surface and \( W_{d} \) is the deposition velocity. This is a transfer coefficient which can be expressed as a parallel sum of conductances over three pathways, gravitational settling (at terminal velocity \( W_i \)), impaction (with conductance \( G_{\text{imp}} \)), and Brownian diffusion (with conductance \( G_{\text{brown}} \)). Thus, \( W_{d} = W_i + G_{\text{imp}} + G_{\text{brown}} \). These terms depend quite differently on particle diameter, with deposition being dominated by sediment for large particles, impaction for particles in the range 1 – 50 \( \mu \)m, and Brownian diffusion for very small particles. The impaction and Brownian-diffusion conductances are strong functions of surface roughness, in ways that can be described well by a simple, single-layer model (Raupach et al. 2001); see Figure 2.
Integrated Wind Erosion and Dust Transport Modelling

The dust emitted during wind erosion affects the global climate changes and energy balance. Predicting these effects requires quantitative estimations of the spatial and temporal variations of the source location, rate, transport pathways and deposition area of the dust. One way to model the systems behavior of dust transport is through an integrated system coupling an atmospheric model, a dust emission model, a dust transport and a dust deposition model and linking to a GIS data base. Such integrated approaches have been applied to simulate dust storms at regional to continental scale (Shao and Leslie 1997; Lu and Shao 2001). The parameterization of large-scale models of this kind is not simple, but a path forward is offered by “model-data fusion” approaches now being implemented in earth system science.

References


Numerical Simulation of Drag Partition over Rough Surfaces

Yaping Shao
Dept. of Physics and Materials Science, City University of Hong Kong, SAR

An Li
School of Mathematics, The University of New South Wales, Sydney

Introduction

The momentum flux, $\tau$, generated by a turbulent flow over a rough surface is attributed mainly to a pressure drag on roughness elements, $\tau_r$, and a skin drag on the underlying surface, $\tau_s$, i.e.,

$$\tau = \tau_r + \tau_s$$

The prediction of $\tau$ and the partition of $\tau$ into $\tau_r$ and $\tau_s$ are important to the studies of aeolian processes, because $\tau_s$ is responsible for the movement of soil particles. The problem of drag partition has been under investigations for over 60 years. A review of the early studies on the subject has been given by Wooding et al. (1973). Marshall (1971) carried out a wind tunnel experiment to examine the effect of roughness element on drag partition. Marshall placed cylinders and hemispheres in the wind tunnel and measured the total drag and the pressure drags on individual elements. Based on Marshall's dataset, Wooding et al. (1973), Arya (1975) and Raupach (1992) developed theories for drag partition.

Despite the success of the theories in describing wind tunnel data, three important questions remain unanswered. The first one is that the wind tunnel experiments are limited to a small sample of simple geometries and arrangements of roughness elements. It is not clear whether the proposed theories are universal and how they should be modified if the quantities, such as roughness-element size and arrangement, differ from the wind tunnel experiment. The second question is that while the functional forms of the Arya (1975) model and the Raupach (1992) model have captured the basic features of drag partition, the key hypotheses which underpin these theories have not been fully tested and hence the physics involved in drag partition is not well understood. The third question is that natural surfaces usually consist of roughness elements of various sizes that are complicated distributed in space. It is not certain whether simple models can be applied to natural surfaces. Even if they can be, it is not clear how the model parameters can be estimated. The field measurements of Wyatt and Nickling (1997) indicate that drag partition over sparsely vegetated surfaces can be different from that over surfaces with simple roughness elements.

It is difficult to examine the above problems through wind tunnel experiments. Computational fluid dynamic models offer an alternative. With the rapid development of computational techniques, numerical models for turbulent flows can be applied
effectively to simulating flows over rough surfaces with roughness elements of any shape, size and arrangements. Hence, numerical experiments provide a great deal of information that is not available from wind tunnel and/or field studies. In this study, we present a numerical simulation of drag partition over rough surfaces. A large-eddy flow model with a k-e closure is run with very high resolution to simulate fully developed shear flows over regular and irregular arrays of cylinders of various sizes mounted on a smooth surface. The skin drag on the ground surface and the pressure drag on the roughness elements are computed. These numerical simulations allow an examination of the interactions of turbulent wakes arising from different roughness elements. They offer an independent validation of the existing theories and a possible foundation for developing new theories of drag partition over natural surfaces. In this study, we first describe the methodology used for the computational simulation of drag partition and then compare the numerical results with the measurements of Marshall (1971) and the theories of Wooding et al. (1973), Arya (1975) and Raupach (1992). We then apply the numerical simulation technique to examine how drag partition depends on the geometry and arrangements of roughness elements.

The Large-eddy Flow Model

A large-eddy simulation model with a k-e closure is used to generate turbulent flows over roughness elements. The large-eddy type of flow simulation is important for the study because the spatial and temporal resolutions must be high in order to represent the rapidly varying turbulent flows around roughness elements. The domain configuration for the simulation is illustrated in Figure 1. It consists of three 1m high and 1.3m wide sections. The first and last sections are 1m long. The length of the second section is adjustable according to the number and size of the roughness elements placed on a plate in this section. The plate has the same length as the second section, but is narrower. The average vertical resolution in all three sections is 10 mm. Grids in the vertical direction are non-uniform with a logarithmically decreasing resolution with height. The horizontal resolution in the first and last sections in the x direction is uniformly 10 mm. Grids in the lateral direction in all sections are the same but non-uniform. Grids on the test plate (x-y plane) are uniform with resolution of 1x1 to 3x3 mm², while grids on the remaining parts of the second section are uniform with resolution 1x5 to 3x5 mm². Thus, grids in the x-y plane of the first section and last section are uniform with resolution 10x1 to 10x3 mm². The maximum computational domain contains 500x100x100 nodes.
Drag Partition

We first simulate the wind tunnel experiment of Marshall (1971) and compare the numerical results with observed data and the estimates based on existing theories. To mimic Marshall's experimental configuration, identical cylinders of h=25.4mm are placed regularly on the surface, and the number of cylinders is increased one by one to allow λ to increase from 0 to 0.1. The cylinder height-diameter ratio h/d is allowed to vary between 0.5 to 2, and the cylinder spacing-height ratio D/s/h between 2 and 30. The flow velocity at z=46 mm is maintained at 20.3ms⁻¹ identical to the free stream flow speed used in Marshall's experiment. Figure~2 shows an example of the velocity distributions.

Figure 1: Domain configuration for the drag partition simulation.

Figure 2. A cross-section of velocity distribution around two cylinders of d=h=25.4 mm at z=10 mm for uH=20.3 ms⁻¹.

Figure~3 compares the simulated (τ/τ0)¹/² and τ/τi with Marshall's (1971) data and the predictions of Raupach (1992), Arya (1975) and Wooding et al. (1973). The observed
values of \((\tau_r/\tau_t)^{1/2}\) are in good agreement with the simulated data, except for the two values at around \(\lambda=0.02\), which are slightly larger. The simulated values of \((\tau_r/\tau_t)^{1/2}\) are also in good agreement with the predictions of Raupach (1992) and Arya (1975) for \(\lambda\) between 0 and 0.2. The simulated \(\tau_r/\tau_t\) is also in good agreement with Marshall's data and the predictions of Raupach (1992) and Arya (1975).

Figure 3: Simulated \((\tau_r/\tau_t)^{1/2}\) and \((\tau_s/\tau_t)\) for \(u_H=20.3\) ms\(^{-1}\) are compared with Marshall's (1971) measurements and the estimates using the theories of Raupach (1992), Arya (1975) and Wooding et al. (1973). The roughness density \(\lambda\) increases with the numbers of cylinders of \(d=h=25.4\) mm.

To study how drag and drag partition depend on the random distribution of roughness elements, we carried out experiments using five roughness elements. Three cases are considered. In Case 1, the five elements differ from each other both in height and diameter; in Case 2, they have the same diameter (25.4 mm) but differ in height; in Case 3, they have the same height (25.4 mm) but differ in diameter. There are five fixed locations on the ground surface and for each test the roughness elements are randomly placed at these locations. Figure 4 shows an example of the flow field.
Figure 4: Instantaneous velocity and pressure fields on the x-y plane at z=10 mm for \( u_H = 20.3 \text{ ms}^{-1} \).

The numerical simulations suggest that the arrangement of roughness elements for given \( \lambda \) does affect the pressure drag on roughness elements, the skin drag on exposed ground surface and the drag partition. However, if \( \lambda \) is large, the impact appears to be relatively small. Therefore, applying the theories of Raupach (1992), Arya (1975) and Wooding et al. (1973) to irregular arrays of non-uniform cylinders to predict drag and drag partition produce only a small error. Because of the limitations in the number of numerical simulations, we have not been able to establish a general relationship between the scatter and the variance of the roughness element size. Therefore the above conclusion should be considered as preliminary.
Numerical Simulation of Northeast Asian Dust Storms Using an Integrated Wind Erosion Modelling System

Yaping Shao, Dept of Physics and Materials Science, The City University of Hong Kong SAR, P. R. China

Introduction

The emission of dust from the deserts and the adjacent areas in Northeast Asia contributes greatly to the global mineral aerosol balance. To predict continental scale dust storm activities, it is necessary to develop an integrated wind erosion modeling system coupling various dynamic models and a geographic information database. Westphal et al. [1988], Joussaume [1990] and Gillette and Hanson [1989] provided early examples of such an approach. Shao and Leslie [1997] and Lu and Shao [2001] developed a fully integrated wind erosion modeling system that simulates all stages of wind erosion, from particle entrainment and transport to deposition. In this study, this integrated modeling system is further developed by incorporating the new wind erosion scheme of Shao [2001].

Northeast Asian dust storms were active between March and May 2002 and a severe event occurred on 19 and 20 March 2002. We carried out intensive numerical experiments using the integrated wind erosion modeling system and were able to successfully predict all major dust storm events during the period between March and May. The simulation area is (30E, 5N) to (180E, 65N) with a spatial resolution of 50km. The area of data analysis is (72E, 5N) to (148E, 53N). The atmospheric data required for model initialization and boundary conditions are derived from the T213-GCM of the China Meteorological Administration.

Results

The 20 March 2002 northeast Asian dust storm is associated with the development of an intense cyclone located in the vicinity of (115E, 47N), accompanied by very strong NW and WNW winds behind the cold front, reaching 16ms⁻¹. As a consequence, wide spread dust storms occurred in Northeast China. The predicted dust concentration is shown in Figure 1 for 08hr 20 March 2002. The prediction shows wide spread dust storms occurring in South Mongolia and Northwest China. During the next hours, the system moved further eastward, affecting a much larger area. These predictions are in excellent agreement with the surface observations, demonstrating the capacity of the integrated modeling system. It can also be shown that the spatial and temporal evolutions of entire dust storm episode are well predicted. The results presented in this study are genuine predictions, because only the atmospheric model is forced using pre-specified boundary conditions. The system has the capacity of predicting many other physical variables for the quantification of dust cycle, including dust emission, transport and deposition, apart from dust concentration and load.
Figure 1 Simulated near surface dust concentration in ug/m3 for the 20 March 2002 dust storm event.

References


Using WEPS with Measured Data

John Tatarko, USDA-ARS-WERU, 1007 Throckmorton Hall, Kansas State University, Manhattan, Kansas 66506 (jt@weru.ksu.edu)

Larry Wagner, USDA-ARS-WERU, 1007 Throckmorton Hall, Kansas State University, Manhattan, Kansas 66506 (wagner@weru.ksu.edu)

Introduction

The Wind Erosion Prediction System (WEPS) is designed to simulate soil loss by wind from cultivated fields by simulating weather and field conditions (Wagner, 1997). However, in some situations, WEPS may be run using measured or simulated data from other models. This is typically done to validate various components or submodels of WEPS, particularly the erosion portion of the model. For example, a user may have measured soil loss data and limited weather and soil data. They can then input the measured weather and soil data to compare the model soil loss with the measured loss. Some users may also wish to use WEPS in a “predictive mode” where measured, “real time” field data is used in conjunction with weather predictions to estimate future soil loss from fields. This paper will explore the use of WEPS with measured or other simulated data.

WEPS is a process-based, continuous, daily time-step model that simulates weather, field conditions, and erosion. It has the capability of simulating spatial and temporal variability of a field’s soil, crop, and residue conditions and soil loss/deposition within a field. The saltation/creep, suspension, and PM10 components of eroding material are also reported separately and by direction. The WEPS model is modular in design with submodels that simulate weather, soil conditions, crop growth, residue decomposition, management operations, and soil loss by wind. It is designed to be used by the USDA-NRCS under a wide range of conditions throughout the U.S. However, with proper inputs, WEPS is easily adapted to other parts of the world.

Procedures

In typical applications, input files are created within the user interface that makes these files available to the science portion of the model to calculate field conditions and erosion. WEPS requires several types of information to complete a simulation run. However, these files can be modified with measured data and input into WEPS under certain constraints. All of these files except the management file, may be easily altered using a standard text editor or the WEPS user interface to reflect measured data. All input files must be formatted to meet the requirements for WEPS (USDA-ARS-WERU, 2001). A description of these input files and considerations for their creation with measured data are listed below. See the WEPS User Manual (USDA-ARS-WERU, 2001) for detailed description of each input file.

- **Run** file (default is ‘weps.run’) - This file contains general information for a simulation run including the dates of the simulation, the field and barrier
dimensions, the field location, and the path and names of the other input files (described below). The ‘run file’ parameters can be modified to match the parameters for the field simulated. The list of the other input files should specify the path and name of measured data to be used. This file also contains comments (indicated by a ‘#’ in column one) which describes each line of input data to aid in checking and modifying input data.

- **Weather files** - WEPS runs are made for multiple years in full year increments beginning on January 1. If only a partial year of weather data is available (typical), the user has two options. One is to substitute measured data within the simulated weather file and observe the output for the period with measured data. The other option is to use the stand alone Erosion model (described below) for single day simulations. Two weather files are required for the full WEPS model.

  Windgen file (default is ‘*.win’) - This file contains both the wind speed (m s\(^{-1}\)) on a subdaily time step and one wind direction (degrees clockwise from North) for each day of the simulation. The subdaily wind speeds are by default the average hourly speeds (i.e., 24, 1 hourly averages) but can be of other time steps of equal length (e.g., 96, 15 minute averages or 8, 3 hour averages).

  Cligen file (default is ‘*.cli’) - The Cligen weather generator was developed for use with the Water Erosion Prediction Project (WEPP) (Flanagan, et.al., 2001) and is used by WEPS to simulate other weather parameters. The input file created by Cligen includes precipitation amount (mm), duration (hr), time to peak (fraction of duration), and peak intensity (mm hr\(^{-1}\)) as well as maximum and minimum air temperature (°C), solar radiation (ly d\(^{-1}\)), and dew point temperature (°C). This file also contains historical monthly averages for maximum and minimum temperature (°C) which are required by WEPS.

- **Soil file** (default is ‘*.ifc’) - This file contains the initial soil conditions at the start of a simulation run. The soil and management submodels then simulate the changes in these conditions as affected by weather, management, and erosion for each simulation day. Even intrinsic parameters such as particle size distribution will change with tillage as layers are mixed. If simulated soil parameters vary significantly from measured values, it is recommended that the user use the stand alone Erosion model (described below). The soil input file includes the taxonomic order, number and thickness (mm) of soil layers, detailed particle size distribution (fraction), wet and dry bulk density (Mg m\(^{-3}\)), aggregate stability (ln(J m\(^{-2}\))), density (Mg m\(^{-3}\)), and size distribution (fraction), soil crust properties (varies), random and oriented (ridge) roughness (mm), soil water characterization parameters (varies), dry albedo (fraction), organic matter (fraction), pH, calcium carbonate (fraction), and cation exchange capacity (meq 100g\(^{-1}\)). This file also contains comments (indicated by a ‘#’ in column one) which describes each line of input data to aid in checking and modifying input data.

- **Management file** (default is ‘*.man’) - This file contains parameters for the
manipulation of soil and biomass properties as a result of various management operations performed on the field on a given date. These operations include planting, harvesting, cultivation, defoliation, fertilization, and irrigation. The management file should only be altered using the Management Crop Rotation Editor for WEPS (MCREW) to guarantee that parameters are correct. MCREW is accessed through the WEPS user interface.

The Erosion submodel can also be operated as a stand alone model to simulate erosion for a single storm (i.e., daily). A stand alone erosion model user interface has been created and is available for easy input of parameters, run the model, and view outputs. Input parameters that must be provided for the day include the field and barrier dimensions as well as biomass, soil, hydrology, and weather parameters. Wind speed can be entered either as Weibull distribution parameters or listed as average wind speeds for each time period throughout the day. Specific definitions of these parameters are documented within the user interface.

For assistance using measured data with WEPS or to obtain the WEPS model or the standalone Erosion model, please go to {http://www.weru.ksu.edu/weps} or contact one of the authors at the e-mail addresses listed above.

References


Emission of Soil Dust Aerosol: Anthropogenic Contribution and Future Changes

Ina Tegen, MPI for Biogeochemistry, Jena, Germany (ITEGEN@BGC-JENA.MPG.DE)

S.P. Harrison, MPI for Biogeochemistry, Jena, Germany (SHARRIS@BGC-JENA.MPG.DE)

K.E. Kohfeld, MPI for Biogeochemistry, Jena, Germany (KEK@BGC-JENA.MPG.DE)

S. Engelstaedter, MPI for Biogeochemistry, Jena, Germany (SENGEL@BGC-JENA.MPG.DE)

M.Werner, MPI for Biogeochemistry, Jena, Germany (MWERNER@BGC-JENA.MPG.DE)

Introduction

Soil dust aerosol is an important factor in the climate system, impacting on the radiation balance, atmospheric chemistry and biogeochemical cycles. Anthropogenic disturbance of soil surfaces has been estimated to contribute as much as 50% to the modern global dust load in the atmosphere (Intergovernmental Panel on Climate Change 2001), but this estimate is highly uncertain. We need to quantify this contribution more accurately in order to understand the historical dust record and estimate future changes in dust emissions.

Method

We compare a global compilation of dust storm frequencies, based on data from more than 2000 stations, to vegetation cover and the distribution of cultivated and rangeland areas. The global emission of natural and anthropogenic dust is computed with a dust source model, which explicitly includes the contribution of preferential source areas and vegetation phenology.

Results and Discussion

We find only slightly higher dust storm frequencies in rangeland and cropland areas than in undisturbed regions (Figure 1). The increase is small, and, according to our model anthropogenically-disturbed soils contribute only 5 to 15% to the modern global dust emission. Using these results, we estimate the role of future changes in anthropogenic dust loading in offsetting natural changes in dust emissions under a global warming scenario. First results indicate that the change in natural emissions under a global warming scenario for 2050 is small in the global mean (ca. 10% decrease in dust emissions), but regional effects (e.g. in east Asia) may be considerable.
Figure 1: Comparison of global dust storm frequencies in areas of natural vegetation, croplands and rangelands. The relationship between dust storm frequencies and vegetation cover (left) shows that maximum dust storm frequencies in cultivated areas do not exceed those in natural areas. The bars (right panel) indicate the 25th to 75th percentiles of the data.

References

Validation of the Wind Erosion Equation (WEQ) for Discrete Periods

R. S. Van Pelt, USDA-ARS, Big Spring, Texas 79720 (svanpelt@lbk.ars.usda.gov)

T. M. Zobeck, USDA-ARS, Lubbock, Texas 79415 (tzobeck@lbk.ars.usda.gov)

Introduction

In the United States, the USDA Natural Resources Conservation Service (NRCS) is the primary agency charged with the task of reducing wind erosion associated with production agriculture. The NRCS has used the Wind Erosion Equation (WEQ) (Woodruff and Siddoway, 1965) to assess the effects of field management on the potential for wind erosion for the last two and a half decades. The WEQ uses inputs of soil erodibility, ridge roughness, locally calibrated climatic factors, field length, and vegetative cover to predict the potential annual wind erosion for a given field and set of management variables. By using this model, the NRCS is able to direct producers toward crop management systems that effectively reduce erosion.

Wind erosion modeling efforts by the USDA Agricultural Research Service (ARS) over the last decade have necessitated the collection of several large bodies of wind erosion and weather data from many diverse locations in the United States. This effort has been facilitated by the development of technology and equipment that have enabled the measurement of wind erosion losses on storm event basis (Fryrear, 1986; Stout and Zobeck, 1996). The availability of field measurements has improved the description of erosion losses across a field (Stout, 1990) and also permits the validation of wind erosion models. We tested WEQ against much of the aforementioned body of data in order to determine the accuracy of its predictions.

The Global Change and Terrestrial Ecosystems Soil Erosion Network (GCTE-SEN) has recently conducted a model validation exercise for water erosion models (See September - October 1996, J. Soil & Water Conserv.). Several models also are available to estimate wind erosion losses, but few studies have compared the output from these models with field-measured data. In this study, as part of a GCTE-SEN project, we evaluate how well predictions of erosion made with a commonly used model compare with data collected from eroding fields. We also investigate appropriate factors that may be changed to calibrate WEQ for local climate and soil conditions.

Methods

Seven sites from six states across the United States were chosen to validate WEQ. The site locations, years of comparisons, soils classification, soil erodibility index (I), and climate factor (C’) are presented in Table 1. The sites were described, instrumented, and the erosion data collected by USDA-ARS and USDA-NRCS personnel. All the sites were a 100 m radius circular field (~ 2.5 ha) outfitted with a weather station and 13 erosion sampling stations. Weather data collected included rainfall, wind speed, and
wind direction at one minute intervals. Soil surface condition data including ridge height, random roughness, crusting, percent erodible fraction in the absence of crusting, and standing and flat plant residues were collected several times a season.

Soil saltation and suspension loads at each of the 13 field locations were estimated by taking the weight of sediment collected in individual samplers at those locations and calculating the transport load (Fryrear et al., 1998). Creep load was estimated for each of the 13 locations in a similar manner based upon transported soil weights collected at 4 locations in the field. Field

Table 1. Test site locations, soils and climate factors.

<table>
<thead>
<tr>
<th>Location</th>
<th>Years of Comparison</th>
<th>Soil Classification</th>
<th>I</th>
<th>C</th>
</tr>
</thead>
<tbody>
<tr>
<td>Big Spring, TX</td>
<td>1989, 90, 93, 94, 95, 96, 97</td>
<td>Amarillo fine sandy loam</td>
<td>86</td>
<td>60</td>
</tr>
<tr>
<td>Kennett, MO</td>
<td>1993, 94</td>
<td>Farrenburg fine sandy loam</td>
<td>86</td>
<td>10</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Malden loamy fine sand</td>
<td>134</td>
<td></td>
</tr>
<tr>
<td>Eads, CO</td>
<td>1991</td>
<td>Wiley silt loam</td>
<td>56</td>
<td>90</td>
</tr>
<tr>
<td>Elkhart, KS</td>
<td>1992</td>
<td>Dalhart fine sandy loam</td>
<td>86</td>
<td>70</td>
</tr>
<tr>
<td>Sidney, NE</td>
<td>1990</td>
<td>Alliance silt loam</td>
<td>56</td>
<td>50</td>
</tr>
<tr>
<td>Prosser, WA</td>
<td>1992</td>
<td>Shano silt loam</td>
<td>56</td>
<td>55</td>
</tr>
<tr>
<td>Mabton, WA</td>
<td>1991</td>
<td>Quincy loamy fine sand</td>
<td>134</td>
<td>50</td>
</tr>
</tbody>
</table>

soil loss for each event was calculated using soil transport estimates from selected locations across the field. Since all these field erosion observations are calculated estimates based upon actual measured observations, we will refer to the erosion data as observed estimates.

A spreadsheet version of WEQ developed by Mike Sporcic and Leigh Nelson of the USDA-NRCS in Spokane WA was used to evaluate the accuracy of WEQ. This spreadsheet version requires user input of field identification and width, tillage direction, field orientation, field length to width ratio. Input values for C’ value (climatic factor), soil I (erodibility), and soil wind erodibility group were taken from the appropriate sections of Part 502 National Agronomy Handbook (USDA-SCS, 1978). Internal menus allow selection of local average wind data tables for the specified location, crop, and field management. The crop and field management is entered by user specified date, allowing the spreadsheet to calculate ridge height and spacing, standing biomass, standing and flat residue, and random roughness for the period extending from that entry to the next. The model predicts the potential erosion for each of these management periods and sums them to obtain average annual wind erosion.

We entered previous crop and tillage management information as was presented in the field notes and records. In order to obtain a better fit of the menu choices and internal calculations with field observations, we occasionally adjusted tillage operations to create residue and roughness effects similar to those observed and sparsely planted crops during periods where photographs indicated that weeds were developing in the fields. Management dates were chosen to coincide with the dates of field sampler installation and removal so that the WEQ predictions of erosion could be summed to coincide with the period of actual field data collection.
Results and Discussion

A summary of WEQ simulation results and comparisons with observed estimates is presented in Table 2. Averaged across all sites and years of comparison, WEQ predicted only 53.3% of the observed estimated erosion. WEQ only predicted 37.5% of the observed estimated erosion at Big Spring for all 7 years of comparisons. If we remove the comparisons for 1996, a year with much lower than average winds, WEQ only predicted 22.6% of the observed estimated erosion. Similar results were noted for most of the other sites with the exception of relatively good agreement between WEQ predictions and observed estimated erosion for Eads, CO and Elkhart, KS both of which are located near the area where WEQ was developed. Highly variable results were noted between the two years of comparison at Kennett, MO. Dissimilar surface conditions between the two years of comparison are reflected in the differences in WEQ predicted erosion, but the differences in observed estimated erosion between the two years was much greater than in the predicted erosion. On-site wind data indicated that 1993 was much windier than 1994 resulting in observed estimated erosion in 1993 of more than 20 times that noted in 1994.

WEQ uses statistical wind distribution data for input wind parameters and therefore should not be expected to match each year’s observed estimated erosion since the magnitude, duration, and direction of erosive winds vary from year to year. Woodruff and Siddoway (1965) state that wind speeds are normally distributed. Analysis of 5 minute average and daily average wind speed for Big Spring, TX indicates that this is not the case. Winds of erosive velocity typically occur only a few hours out of the day and a few days per month. Long periods of sub-erosive velocity winds separate the extreme events. Additionally, detailed saltation activity data for Big Spring, TX would indicate that daily average wind speed is a very poor predictor of wind erosion activity. It was not uncommon to note three orders of magnitude higher saltation activity on a day with a lower daily average wind speed than another day within the same week.

Attempts to create simulation conditions that would allow WEQ to more closely predict the erosion for 1989 and 1990 at Big Spring, TX, two average years, yielded interesting results. We at first assumed that the wind velocity parameter in the C’ factor was perhaps underestimated for this location. C’ values were varied from 60 to 150, 2.5 times the given value for the location and the maximum value indicated for any location, and yet WEQ only predicted 57.3% and 60.2% of the observed estimated erosion for 1989 and 1990 respectively. The C’ factor had to be increased to 222, a value nearly 4 times that published for the site, in order for the predicted erosion to be 95.6% and 104.4% of the observed estimated erosion for 1989 and 1990, respectively. Woodruff and Armbrust (1968) recommend use of a monthly C’ that improves the accuracy of WEQ simulations. While monthly values for the erosive season at Big Spring are larger than the annual value, they are still too low to allow good agreement between WEQ predictions and observed estimates in the two years investigated. Further, there is no provision for varying the C’ value within years of simulation in this version of WEQ so management periods would have to be separated by months, individual annual simulations run with C’ values for each month, and the appropriate soil loss figures would have to be summed across 12 simulations for a single year’s prediction. This
problem could, however be solved by modification of the spreadsheet and a table of monthly C’ values by location.

Table 2. Summary of WEQ predicted erosion, observed estimated erosion, and comparison.

<table>
<thead>
<tr>
<th>Location</th>
<th>Year</th>
<th>Comparison Period</th>
<th>WEQ Predicted (T/ac)</th>
<th>Observed Estimates (T/ac)</th>
<th>WEQ/Observed (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Big Spring, TX</td>
<td>1989</td>
<td>01/12/89 - 05/03/89</td>
<td>20.02</td>
<td>96.11</td>
<td>20.8</td>
</tr>
<tr>
<td>Big Spring, TX</td>
<td>1990</td>
<td>01/05/90 - 05/04/90</td>
<td>20.79</td>
<td>93.52</td>
<td>22.2</td>
</tr>
<tr>
<td>Big Spring, TX</td>
<td>1993</td>
<td>03/16/93 - 06/01/93</td>
<td>14.16</td>
<td>128.42</td>
<td>11.0</td>
</tr>
<tr>
<td>Big Spring, TX</td>
<td>1994</td>
<td>01/06/94 - 05/18/94</td>
<td>20.81</td>
<td>76.57</td>
<td>27.2</td>
</tr>
<tr>
<td>Big Spring, TX</td>
<td>1995</td>
<td>01/11/95 - 05/15/95</td>
<td>21.75</td>
<td>117.31</td>
<td>18.5</td>
</tr>
<tr>
<td>Big Spring, TX</td>
<td>1996</td>
<td>01/12/96 - 05/16/96</td>
<td>22.63</td>
<td>17.80</td>
<td>127.1</td>
</tr>
<tr>
<td>Big Spring, TX</td>
<td>1997</td>
<td>01/23/97 - 05/23/97</td>
<td>21.67</td>
<td>60.82</td>
<td>35.6</td>
</tr>
<tr>
<td>Eads, CO</td>
<td>1991</td>
<td>10/30/90 - 05/07/91</td>
<td>10.54</td>
<td>10.84</td>
<td>97.2</td>
</tr>
<tr>
<td>Elkhart, KS</td>
<td>1992</td>
<td>01/01/92 - 10/15/92</td>
<td>87.95</td>
<td>69.16</td>
<td>127.2</td>
</tr>
<tr>
<td>Kennett, MO</td>
<td>1993</td>
<td>12/02/92 - 06/17/93</td>
<td>6.83</td>
<td>61.26</td>
<td>11.1</td>
</tr>
<tr>
<td>Kennett, MO</td>
<td>1994</td>
<td>11/18/93 - 05/05/94</td>
<td>3.98</td>
<td>2.86</td>
<td>139.4</td>
</tr>
<tr>
<td>Prosser, WA</td>
<td>1992</td>
<td>06/10/92 - 06/15/93</td>
<td>0.74</td>
<td>1.43</td>
<td>51.8</td>
</tr>
<tr>
<td>Sidney, NE</td>
<td>1990</td>
<td>10/24/89 - 04/24/90</td>
<td>0.8</td>
<td>1.96</td>
<td>40.7</td>
</tr>
</tbody>
</table>

While holding the value for C’ at 150, the soil erodibility index, I, was increased to 134 to provide a prediction of 98.4% and 105% of the observed estimate for 1989 and 1990 respectively. If we returned the C’ value to 60 and increased the soil erodibility index, I, to the maximum value of 310, WEQ predicted 101.4% and 108.1% of the observed estimates for the respective years. Although reasonably good fits can be obtained by varying the I value alone, it would be difficult to predict erosion for a soil more erodible than a fine sandy loam and many agricultural soils in this area are loamy fine sands and fine sands. It should be pointed out, however that the soil erodibility
index, I, values for WEQ are based upon the percentage of soil aggregates in the upper inch of soil larger than 0.84 mm. While this value may be appropriate for freshly tilled soils, rainfall often results in the disintegration of aggregates, crusting of the soil, and creation of a surface mantle of sandy abrader material. This sandy abrader is usually the first material to move during a wind event and thus the apparent texture of the surface soil would approximate the soil for which a soil erodibility index, I, of 310 would be appropriate. This adjustment of soil surface conditions would also be easier to implement in areas where detailed information on actual wind erosion rates were not available to allow C’ adjustment for local model calibration.

**Conclusions**

WEQ tended to underestimate the observed estimated erosion in most cases but the performance did improve with local calibration for the Big Spring, Texas location. Increasing the annual input value of C’ to nearly four times the published value did allow close agreement between predicted and observed erosion as did increasing the value of I to the upper limit and combinations of increased C’ and I input values. The use of increased values of I could be explained by the texture of the surface mantle of fine sand resulting from the rain induced disintegration of soil aggregates and points to the importance of soil surface conditions in controlling wind erosion and the necessity of careful characterization of these soil surface conditions when running predictive wind erosion models.

**References**


Validation of the Wind Erosion Stochastic Simulator (WESS) and the Revised Wind Erosion Equation (RWEQ) for Single Events

R. Scott Van Pelt, USDA-ARS, Big Spring, Texas (E-mail: svanpelt@lbk.ars.usda.gov)
Ted M. Zobeck, USDA-ARS, Lubbock, Texas (E-mail: tzobeck@lbk.ars.usda.gov)
Ken N. Potter, USDA-ARS, Temple, Texas (E-mail: potter@brc.tamus.edu)
John E. Stout, USDA-ARS, Lubbock, Texas (E-mail: jstout@lbk.ars.usda.gov)
Thomas W. Popham, USDA-ARS, Stillwater, Oklahoma (E-mail:tpopham@pswcrl.ars.usda.gov)

Introduction

The Environmental Policy Integrated Climate model (EPIC) has been used to evaluate policy effects on soil erosion. Recently, WESS, the wind erosion submodel of EPIC, was successfully evaluated against observed estimates of erosion from a clay loam soil for several wind events in Alberta, Canada (Potter, et al., 1998). This study was conducted for a single season and surface roughness and vegetative cover did not vary sufficiently during the period of investigation to assess the effectiveness of the model under soil surface conditions other than those noted.

USDA Agricultural Research Service (ARS) scientists and engineers have recently released RWEQ (http://www.csrl.ars.usda.gov/wewc/rweq/readme.htm). RWEQ makes annual or period estimates of wind erosion based on a single event wind erosion model that includes factors for wind and rainfall, soil roughness, the erodible fraction of soil, crusting, and surface residues (Fryrear et al., 1998a, 1998b). A previous test of 11 wind erosion events found the correlation between observed and estimated maximum transport (Qmax) and field soil loss (SL) to be 0.82 and 0.97, respectively (Fryrear et al., 1998a).

Wind erosion modeling efforts conducted by the ARS over the last decade have necessitated the collection of several large bodies of wind erosion and weather data from many diverse locations in the United States. This effort has been facilitated by the development of technology and equipment that have enabled the measurement of wind erosion losses on storm event basis (Fryrear, 1986; Stout and Zobeck, 1996). The availability of field measurements has improved the description of erosion losses across a field (Stout, 1990) and also permits the validation of wind erosion models. We tested RWEQ and WESS against much of the aforementioned body of data in order to determine the accuracy of their predictions.

Methods

Seven sites from six states across the United States were chosen to validate RWEQ. Individual storm event data at Big Spring, Texas (24 storms during 7 years of data collection) were used to validate WESS. The sites were described, instrumented, and the erosion data collected by USDA-ARS and USDA-NRCS personnel. All the sites were a 100 m radius circular field (~ 2.5 ha) outfitted with a weather station and 13 erosion sampling stations (Fryrear et al., 1998b). Weather data collected included 2 m wind speed and direction at one minute intervals. For the purpose of validating WESS and RWEQ, these wind data were averaged over 10 minute intervals. Soil surface condition data including ridge height, random roughness, percent erodible fraction, and standing and flat plant residues were collected several times during the data collection season.
Soil saltation and suspension loads at each of the 13 field locations were estimated by taking the weight of sediment collected in individual samplers at those locations and calculating the transport load (Fryrear and Saleh, 1993). Creep load was estimated for each of the 13 locations in a similar manner based upon transported soil weights collected at 4 locations in the field. Field soil loss for each event was calculated using soil transport estimates from selected locations across the field. Since all these field erosion observations are calculated estimates based upon actual measured observations, we will refer to the erosion data as observed estimates.

WESS simulations were run for the dates of 24 storm events in the seven years of erosion observations at Big Spring, TX. The period between field sampler servicing was rounded to the nearest multiple whole day and the 10 minute average wind data for that event was input along with the soil surface conditions reported for the site on that date. Soil roughness was calculated according to Potter and Zobeck (1990) when pin roughness data were available and were estimated by comparing chain roughness data (Saleh, 1993) with surface photographs when only chain roughness data were available. Initial soil moisture conditions were inferred from rainfall and temperature data for each event. The simulations for each event were run at distances from protected surface (DPS) that coincided most closely with the DPS of field samplers as these distances varied with the mean wind direction of each event.

RWEQ simulations were run for 41 storm events at six locations across the United States. RWEQ uses a number of input factors to estimate Qmax and SL. The wind factor (WF), erodible fraction (EF), surface crust factor (SCF), roughness (K’), and crop on the ground factor (COG) were determined for each erosion event. Details of the procedures to determine these values are described by Fryrear et al. (1998a, 1998b). The WF for each estimated storm was computed from one minute wind speed measurements averaged for the entire 24 hour day (Fryrear et al., 1998a). EF was determined for each location by dry sieving surface soil samples (Chepil, 1962). SCF was defined as a function of percent clay and organic matter (OM) content. K’ was determined using measured roughness parameters and the amount of flat and standing residue were used to estimate COG (Fryrear et al., 1998b). Since the EF, K’ and COG factors were often measured only two or three times per season, estimates of these factors were often necessary. Degradation of aggregates and roughness were estimated using cumulative rainfall and linear interpolation between measurement dates was used to estimate COG.

Results and Discussion

WESS was used to predict the erosion resulting from 24 of the individual wind events at Big Spring, TX during the seven years that wind erosion data was collected at this location. Erosion estimates using WESS under-predicted 9 events, accurately predicted 8 events, and over-predicted 7 events. In general, the events that WESS under-predicted were large magnitude storms with observed erosion estimates >1.0 kg m⁻² and the events that WESS over-predicted were small storms with observed estimates < 0.2 kg m⁻². WESS gave the most accurate predictions for events that had observed estimated erosion from 0.2 kg m⁻² to 1.0 kg m⁻². Since the large magnitude storms, some of which had observed estimates of greater than 5.0 kg m⁻² at distances greater than 60 m from protected surface, have a much larger effect on annual erosion than the small storms, it is evident that WESS would tend to under-predict erosion on an annual basis.

Plots of the comparisons between WESS predicted erosion and observed estimated erosion for 4 events at Big Spring, TX are presented in Figure 1. The plot for the 4/22/89 event shows paired estimates with low variability and the relatively high accuracy of prediction for this event. The plot
Figure 1. Plots of WESS predicted (solid lines) vs. observed estimated erosion (diamonds) by distance from protected surface (DPS) for four selected storm events at Big Spring, TX.

For the 4/25/94 event shows observed estimates with low variability and the typical under-prediction for large events. Both of these plots show the inability of WESS to accurately predict erosion rates at DPS of less than 60 m. Although this source of error would become increasingly insignificant as field size increases, it does point out a problem either with the use of transport load to estimate erosion rates or with the form of the equation used by this and other models to predict the erosion rate. This problem is worthy of further investigation.

Plots for the 1/29/90 and 2/12/90 events demonstrate another problem with prediction of wind erosion. These two events occurred 14 days apart, there was no rain or other change in surface conditions between the events, the wind speed and duration were nearly identical, WESS predicted similar erosion curves, and yet there is a great difference in the observed estimated erosion. There was a 150° shift in wind direction between the storms, but this was a circular field in a broad open area. In spite of the sophistication of our data collection and predictive models, there are probably sources of variability in any field that we may not ever be able to quantify and predict.

In general, RWEQ underestimated Qmax and SL. Close inspection of the data by location revealed that the estimates were not consistently higher or lower for most of the locations. Observed estimates for Qmax and SL were higher than RWEQ predictions in about 58% of the cases investigated, while observed estimates for S were higher than RWEQ predictions in only 22% of the events analyzed.

Simple linear regressions of observed estimates vs. RWEQ predictions of Qmax and SL revealed significant (P<0.05) correlations for Qmax and SL with correlation coefficients of 0.70 and 0.62, respectively. Figure 2 illustrates the relationship of the observed estimates and RWEQ predictions of Qmax for all storm events investigated. The relation of observed estimates and RWEQ predictions of SL is presented in Figure 3.

The results of the investigation with RWEQ are very encouraging. Erosion was measured at locations that varied considerably with respect to soil, climate, and wind patterns and yet the RWEQ predictions were correlated within an order of magnitude of the observed estimates.
References


Wind Erosion Modelling in a Sahelian Environment

Saskia M. Visser\textsuperscript{a} Geert Sterk\textsuperscript{b} Judith J.C. Snevangers\textsuperscript{b}

\textsuperscript{a}Erosion and Soil & Water Conservation Group, Department of Environmental Sciences, Wageningen University, Nieuwe kanaal 11, 6709 PA Wageningen, The Netherlands. email: Saskia.Visser@users.tct.wag-ur.nl

\textsuperscript{b}Netherlands Institute of Applied Geoscience TNO, Delft, The Netherlands.

Introduction
On a field scale, observations of wind-blown sediment transport often show a considerable spatial variation caused by differences in soil characteristics, surface roughness, topography, vegetation and soil crusting. In practice it is almost impossible to measure all factors that explain the variation in observed sediment transport. If we add the fact that fields of farmers in a sahelian environment are more heterogeneous in soil management, living and dead vegetation cover and application of fertilizer than the average fields of "western" farmers, we can expect considerable variation in erosion and deposition of sediment due to wind--blown mass transport in a sahelian field.

In the sahel soil conservation tools are limited, it is important to warily manage them. A good simulation model, which correctly predicts erosion and deposition at the field scale, will be a useful tool in successful application of soil conservation measures.

In this article we discuss three modelling approaches and their performance in the Sahel. Field data of wind erosion measurements in northern Burkina Faso are used to run a geo-statistical (spatial modelling), an empirical (RWEQ) and a deterministic (WEPS) model, in order to determine which model predicts best the spatial variability in wind blown mass transport in the Sahel. Results of the different modelling approaches will be discussed.

Materials and methods
During the early rain season of 2001, field data of 12 wind erosion events were collected at 3 sites with different geomorphologic settings; a degraded site, a valley and a dune, in northern Burkina Faso. Each site of 80 x 80 m. is instrumented with 17 Modified Wilson and Cook catchers. The catchers are regularly distributed so that in each wind direction a line of 5 catchers is formed. In one line each of the catchers are 15 m. apart. At the dune and the degraded zone weathers stations were installed. Since field in the valley floor was situated less then 500 m from the degraded zone, weather data of the degraded zone was also used for the valley floor. Wind speed and wind direction were measured every one-minute at a height of 2 m. At the degraded site a wind profile is continuously measured at 0.5, 1, 2 and 3 m. Mass transport was determined by sampling over 5 heights and intergrating over the height.

The Revised Soil Erosion Equation (RWEQ)
The Revised Soil Erosion Equation (RWEQ) uses a number of input factors to estimate transport capacity, critical field length and field soil loss. The wind factor (WF), erodible fraction (EF), surface crust factor (SCF), roughness (K') and crop on the ground factor (COG) were determined for each erosion event. For further details about the procedure to determine these values we refer to Fryrear et al (1998a and 1998b). The WF for each storm was computed from one-minute wind measurements, EF was calculated as a function of percent sand, silt, clay organic matter and calcium carbonate content (Fryrear et al., 1998). SCF was defined as a function of percent clay and organic matter content (Fryrear et al. 1998). K’ was determined using measured roughness parameters and COG was estimated using the amount of flat and standing residue (Fryrear et al., 1998). In the dry period the input parameters were determined once a month and in the wet period these parameters were determined once a week.
In RWEQ equations 1 and 2 are used to estimate maximum sediment transport \((Q_{\text{max}})\) and critical field length \((S)\).

\[
Q_{\text{max}} = 109.8 \times ( WF \times EF \times SCF \times K^* \times COG )
\]

\[
S = 150.7 \times ( WF \times EF \times SCF \times K^* \times COG )^{0.371}
\]

Finally we used equation 3 to determine mass transport at a certain distance downwind.

\[
Q(x) = Q_{\text{max}} \times [1 - e^{- \frac{x}{S}}]
\]

Maximum sediment transport \((Q_{\text{max}})\) and critical field length \((S)\) can be calculated from the field measured data by performing least squares non-linear regression using equation 3.

**The Wind Erosion Prediction Project (WEPS)**

The WEPS erosion sub-model decides if erosion can occur based on the current soil surface roughness (oriented and random), flat and standing biomass, aggregate size distribution, crust and rock cover, loose erodible material on the crust and the soil surface wetness (Hagen, 1995). The aerodynamic roughness \((z_0)\) is calculated using information about ridge and random roughness and vegetation characteristics. The friction velocity \((u^*)\) using eq (4).

\[
u^* = \frac{0.4 \times U}{\ln\left(\frac{z}{z_0}\right)}
\]

In which \(U\) is the wind speed at height \((z)\), 0.4 is the von Kármán constant and \(z_0\) is the aerodynamic roughness. Since the friction velocity is the driving force behind the model, it is very important to correctly predict this parameter.

At the degraded site we measured wind speed at 0.5, 1, 2 and 3 \(m\). The universal velocity distribution eq. 5 was used to determine the friction velocity and the aerodynamic roughness for each storm at the degraded site.

\[
U = \frac{u^* \times 0.4 \ln\left(\frac{z}{z_0}\right)}{z_0}
\]

The observed friction velocity and aerodynamic roughness are compared to the values predicted by WEPS.

**The spatial modelling approach**

Due to the fact that it is almost impossible to measure the distribution of all factors that account for the spatial variation in sediment transport, Sterk and Stein, 1997 suggest an alternative modelling approach for sediment transport modelling. By making use of geostatistics they produced storm-based maps of sediment transport.

In geostatistics the variogram is used to model the spatial variability. A common rule is that at least 30-50 measurements are needed to obtain representative variograms. As we have only 17 measurements per storm, we decided to combine the measurements of all storms in one overall variogram per site. We could do this by first standardising the sediment transport values (standardised = [value-mean]/variance), as a strong relation exists between mean and variance of the measurements of each storm. The variogram was modelled using a spherical model with a nugget effect. This model was tested using cross-validation. During cross-validation the variogram model is used to re-predict actual observations from neighbouring observations. With a perfect variogram the mean cross-validation Z-score should be 0, with a standard deviation of 1. To obtain spatial maps per storm that honor the high variability at close distances in the field, we used the overall variogram model in conditional stochastic simulation. The geostatistical variography and model were carried out with the geostatistical software package GSTAT (www.gstat.org).
Results and Discussion

RWEQ

Fig. 1 is an example plot of soil losses across a field with a critical field length (s) of 50 m. The RWEQ model assumes a non-eroding boundary around a field, based on this assumptions highest soil losses will be predicted in the zone near the non-eroding boundary.

![Graph](image)

Fig 1: Relationship between mass transport, soil loss and average soil loss from RWEQ using s=50 m and Qmax=1.0kg/m. (Fryrear et al, 1998)

In the Sahel the fields are not surrounded by a non-eroding boundary. The edges of the fields are often indicated by a single tree, a path or a large rock. None of our research sites had a non-eroding boundary, so we do not know at which point in the predicted plot the measurement were made. Therefore it is virtually impossible to compare observed soil transport with soil transport predicted by the model as it is. Preliminary results show that it is possible to transpose the plot over a distance a using equation 5.

\[
Q(x) = Q_{\text{max}} [1 - e^{-\frac{x-a}{s}}]
\]

(5)

With this formula the plot will cross the Y-axis at point (0, Q) (the first field measurement) and cross the x-axis at point (-a, 0). By performing least squares non-linear regression analysis using equation 5, a good estimation for Qmax and S can be obtained. Only now the observed data can be compared with the predicted data.

WEPS

The degraded site is a completely bare area with an extremely smooth surface (RR=1.21). Since no vegetation is present and the area is not cultivated (no oriented roughness) the aerodynamic roughness is only determined by the random roughness.

The meteostation was situated in such a way that winds coming from the direction North-West till East had a free run without obstacles. Table 1 shows average wind speed of the storm and the observed and predicted \(u_0\) and \(z_0\).

<table>
<thead>
<tr>
<th>Date</th>
<th>(U) (m/s)</th>
<th>WEPS (Z_0) (mm)</th>
<th>WEPS (u_0) (m/s)</th>
<th>Obs (Z_0) (mm)</th>
<th>Obs (u_0) (m/s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>22-5-01</td>
<td>8.64</td>
<td>0.363</td>
<td>0.395</td>
<td>4.40</td>
<td>0.823</td>
</tr>
<tr>
<td>3-6-01</td>
<td>7.17</td>
<td>0.363</td>
<td>0.327</td>
<td>6.25</td>
<td>0.903</td>
</tr>
<tr>
<td>9-6-01</td>
<td>8.80</td>
<td>0.363</td>
<td>0.402</td>
<td>2.17</td>
<td>0.880</td>
</tr>
<tr>
<td>19-6-01</td>
<td>8.52</td>
<td>0.363</td>
<td>0.388</td>
<td>6.72</td>
<td>0.783</td>
</tr>
<tr>
<td>22-6-01</td>
<td>9.21</td>
<td>0.363</td>
<td>0.420</td>
<td>6.90</td>
<td>0.750</td>
</tr>
<tr>
<td>29-6-01</td>
<td>7.49</td>
<td>0.363</td>
<td>0.341</td>
<td>6.25</td>
<td>0.861</td>
</tr>
<tr>
<td>11-7-01</td>
<td>8.40</td>
<td>0.363</td>
<td>0.383</td>
<td>4.07</td>
<td>0.854</td>
</tr>
<tr>
<td>13-7-01</td>
<td>8.19</td>
<td>0.363</td>
<td>0.374</td>
<td>5.19</td>
<td>0.837</td>
</tr>
</tbody>
</table>
In general, WEPS predicts the aerodynamic roughness a factor 10-20 too low and the predicted friction velocity is half of the measured friction velocity. Apparently the aerodynamic roughness at smooth bare surfaces is not only determined by the random roughness. Due to the too low prediction of the aerodynamic roughness the prediction of the friction velocity is low. Therefore further investigation on the behaviour of wind profiles on these kind of smooth surfaces is necessary. WEPS can not simply be applied for smooth soil surfaces in the Sahel. We suggest to add a formula that better predicts the aerodynamic roughness for smooth surfaces in the Sahel.

The spatial modelling approach

The model parameters of the overall variogram model and the cross-validation can be found in table 2. The valley variogram showed only noise, so no spherical model could be fitted. Comparing the parameters of the dune and degraded site, it is clear that the nugget to sill ratio is very high, indicating a large contribution of short distance variability for both sites.

<table>
<thead>
<tr>
<th>Site</th>
<th>Co</th>
<th>A</th>
<th>Total Mean Z</th>
<th>Total St-dev Z</th>
</tr>
</thead>
<tbody>
<tr>
<td>Degraded site</td>
<td>0.40</td>
<td>63.81</td>
<td>0.07</td>
<td>0.91</td>
</tr>
<tr>
<td>Dune</td>
<td>0.40</td>
<td>24.46</td>
<td>0.13</td>
<td>0.84</td>
</tr>
<tr>
<td>Valley</td>
<td>0.62</td>
<td>0.16</td>
<td>0.75</td>
<td></td>
</tr>
</tbody>
</table>

*Co=nugget constant, C=sill parameter, a=range parameter*

Further, the range value for the degraded site is much larger than the range value for the dune site, indicating that the variability pattern is over larger distances for the degraded site than for the dune site. This can easily be explained by lack of residue cover, the more even crustation pattern and the lack of vegetation at the degraded site.

The cross validation of the variogram of the degraded site was ok, but for the valley and dune site the cross validation is less well. The only way to obtain better variograms, is to carry out more measurements per storm. Unfortunately, sampling at a density sufficient for variogram modelling cannot be done in wind erosion studies, because of the high cost of equipment and labour and because such a high density of catchers will probably alternate the wind field and no realistic pattern of erosion and deposition will be measured.

Therefore, despite the weak variograms, we performed conditional stochastic simulation with these variograms to obtain mass transport maps at the three sites for all 12 storms. The produced maps emphasise the strong variations in mass transport across short distances. Based on this, is can be stated that, in Sahelian regions, it will be more useful to distinguish erosion and deposition areas in the field than to describe soil loss per unit area, as is traditionally done. Windblown sediment transport is highly controlled by vegetation cover and soil crustation.

These factors are not evenly distributed over the sites of the dune and the valley, whereas the degraded site is more or less homogeneous for these factors. We expect that if we combine the map of the conditional stochastic simulation with maps of the spatial distribution of soil crustation and vegetation cover, we will have enough information about the possible spatial variation of erosion and deposition for better application of soil conservation measures.

References


Overview and Current Status of WEPS 1.0

L. E. Wagner, USDA-ARS, Manhattan, KS 66506 (E-mail: wagner@weru.ksu.edu)

Introduction

The Wind Erosion Prediction System (WEPS) is a process-based, daily time-step, computer model that predicts soil erosion by wind via simulation of the physical processes controlling wind erosion. It is intended to replace the predominately empirical Wind Erosion Equation (WEQ) (Woodruff and Siddoway, 1965) as a prediction tool for planning soil conservation systems, conducting environmental planning, or assessing off-site impacts caused by wind erosion. WEPS incorporates improved technology for computing soil loss by wind from agricultural fields. It also provides additional capabilities not present in WEQ such as separate calculation and reporting of creep/saltation size particles, suspension loss, and PM-10 emission estimates from the field (Wagner, 1996).

WEPS 1.0 is the first implementation of WEPS to be released and includes a graphical user interface to simplify use of the model. WEPS 1.0 implements a subset of all the features and capabilities originally envisioned for WEPS. However, it still includes the daily time-step organizational structure, input databases, process-based erosion prediction technology, as well as most other key elements advertised for WEPS in the past. Since the United States Department of Agriculture, Natural Resource Conservation Service (USDA-NRCS) is expected to be one of the primary users of this release, the mix of features in WEPS 1.0 was determined based upon their current primary needs. Future releases of WEPS are projected to include additional functionality and other capabilities as customer demand dictates. It is expected that users trained in the use of WEPS 1.0 will find their knowledge, databases, management/crop rotation files, etc. for WEPS 1.0 to be directly transferable to future versions of WEPS.

Background

Soil erosion by wind is a surface phenomenon. It is initiated when the wind speed exceeds the threshold for a given soil and its surface condition. After initiation, the duration and severity of an erosion event depend on the wind speed and the evolution of that surface condition. Therefore, WEPS requires wind speed and direction to simulate the erosion process. These and other weather variables are used in WEPS to drive temporal changes in hydrology, soil erodibility, crop growth, and residue decomposition. WEPS tracks changes in the current soil state and surface condition within the model due to these weather driven processes. Stochastic climate generators, WINDGEN (Skidmore and Tatarko, 1990) and CLIGEN (Nicks et al., 1987), are used to provide WEPS with the necessary daily weather data and subdaily wind information. The simulation region in WEPS is a field or, at most, a few adjacent fields. Areas of the simulation region that have differing soil, management, or cropping conditions are treated as separate subregions. WEPS can output soil
loss/deposition over user-selected time intervals from user-specified accounting regions within the simulation region. This allows the WEPS user to obtain output over various spatial scales within the simulation region. WEPS also provides users with individual soil-loss components of creep/saltation, suspension, and PM10 size fractions. These components are particularly useful as an aid in estimating off-site impacts of wind erosion.

WEPS 1.0

WEPS 1.0 is the first public release of WEPS. To meet NRCS’s immediate needs for improved capabilities of wind erosion prediction; to allow other users earlier access to WEPS technology; and to facilitate getting that first product “out the door”, it was necessary to limit the scope of what WEPS 1.0 would do. Thus, the following design decisions were made:

1) WEPS 1.0 would be released with a graphical user interface to assist users in making WEPS simulation runs. For NRCS users to apply WEPS in the manner expected, they would need to be able to select and modify inputs quickly and easily. Likewise, users want the output presented in a convenient form that is easy to interpret.

2) WEPS 1.0 would use the full WEPS core science model. No separate, “scaled down” version of the WEPS simulation code would be developed. As the WEPS science code evolves, WEPS 1.0 would continue to use the “latest” version of that code. This approach ensures that WEPS 1.0 will benefit from bug fixes and simulation enhancements in the WEPS science code.

3) WEPS 1.0 would handle a single homogeneous subregion, i.e., the simulation region would be represented by only one soil type and management/crop rotation sequence. This allows the WEPS 1.0 developers to focus on the core physical simulation processes in WEPS and not the code to handle multiple subregions.

4) WEPS 1.0 would have only one defined accounting region, the entire simulation field. Because only one subregion is used, the need for additional accounting regions would be limited. This approach also reduces the number of potential output options with which a user is confronted.

5) WEPS 1.0 would provide a limited selection of time scales for output reports: a) half-month period averages per rotation year; b) monthly averages per rotation year; c) rotation year averages; d) long-term yearly averages; and e) yearly results for a few selected data, such as crop harvest yields.

6) WEPS 1.0 would use the same soil, plant growth, management operation, and climate databases to be used in future WEPS releases. Planning for this level of compatibility with future versions of WEPS should make it easier for WEPS users to upgrade to later, more functional, WEPS releases.

7) WEPS 1.0 would limit wind barrier placement to field boundaries only. This would simplify the graphical user interface code for placing and displaying wind barriers. Also, because WEPS 1.0 would deal with only a single subregion, placing barriers within the simulation region would be of limited benefit.
User Interface

WEPS 1.0 requires the following user input selections to perform a simulation run: 1) site location, which determines the appropriate weather station data records to be used by the climate generators; 2) rectangular site dimensions and orientation with respect to North; 3) dominate soil type; 4) management/crop rotation practices applied to the field; and 5) any wind barriers along the field boundaries (optional). The objective of the graphical user interface is to provide an easy way for a user to make those input selections via choice lists from databases and pre-built templates. The interface also gives the ability to modify individual input parameter values for any given simulation run. In addition, the user interface provides a convenient method to select and view desired WEPS output information. The most current beta versions of WEPS are available for download from: http://www.weru.ksu.edu/weps

Current WEPS 1.0 Status

To facilitate feedback to the WEPS development team, training sessions were conducted with NRCS state and regional level specialists across the country. It became obvious from those workshops, NRCS would require a well tested science model and a robust user interface designed to simplify use of the model. Also, the need for fully populated and quality reviewed national level soil, plant, and management operation databases would be required for NRCS field office use of WEPS 1.0. The identified deficiencies in the science model, interface, and databases were documented and are currently being addressed.

In addition, to help ensure that the final release of WEPS 1.0 comes to fruition, WERU has developed a formal Agreement to be signed by ARS and NRCS. The Agreement stipulates what features will be included in WEPS 1.0 and what capabilities will not by the scheduled delivery date to NRCS of June 30, 2003. The specific tasks addressing all documented deficiencies and missing features required by NRCS for WEPS 1.0 have been enumerated and a timeline schedule outlining the work to be completed are included in the Agreement. Currently, work is progressing based upon the tasks outlined in the Agreement. A copy of the most recent version of the Agreement is accessible at: http://www.weru.ksu.edu/weps/Agreement/Agreement.pdf

WEPS 2.0

In the future, WEPS will better represent spatially the effects of varying terrain, changing soil types, wind barriers, strip-cropping systems, ridge till, contour-farming practices, and emergency tillage on wind erosion. In addition, WEPS will be expected to more accurately estimate the potential offsite effects of wind erosion on air and water quality, roadside visibility and possible health consequences. To do so, the following capabilities will need to become available in subsequent versions of WEPS:

1) The ability to simulate more complex sites, which consist of two or more dominant soil types and possibly have different management practices applied to specific areas within the field site. This will be handled in WEPS through the use of...
of multiple subregions or areas representing a single soil type and set of management practices.

2) The ability to specify elevation variations to describe the simulation region terrain more accurately and to better describe deviations in ridge and row directions that are due to the topography and/or shape of the field (contour farming practices and tilling parallel with field boundaries).

3) The ability to handle more advanced wind barriers (those with seasonal changes in height and silhouette area) and erosion traps (regions that are considered to only “collect” moving soil and not be potential transmission sources, e.g. ponds, grass waterways, road ditches) and have them placed anywhere within the simulation region.

Summary

WEPS is a process-based, daily time-step, computer model that predicts soil erosion via simulation of the physical processes controlling wind erosion. WEPS 1.0 is the first implementation of WEPS for use by the USDA-NRCS and is intended primarily for soil conservation and environmental planning. It includes a graphical user interface to allow the user to easily select climate stations, specify field-site dimensions, pick a predominant soil type, and describe any border field wind barriers and management practices applied to an agricultural field.

References


Simulated Global Atmospheric Dust Distribution: Sensitivity to Regional Topography, Geomorphology, and Hydrology

Charles S. Zender, Earth System Science Dept., UC Irvine, Irvine, CA 92697 (zender@uci.edu)

David J. Newman, Earth System Science Dept., UC Irvine, Irvine, CA 92697 (newman@uci.edu)

Introduction

Explaining the observed spatial heterogeneity of global dust emissions is important for understanding the relative role of natural and anthropogenic processes in the present day dust emissions, and thus for predicting future trends in dust production. We identify three related geomorphologic and hydrologic constraints which may contribute to this heterogeneity. Using the global Dust Entrainment and Deposition (DEAD) model (Zender et al., 2002), we investigate the implications of each process relative to a control simulation with no imposed spatial heterogeneity. We show that Aeolian erosion appears linked to regional surface geomorphology and runoff on large spatial scales.

Satellite observations show that persistent maxima in the observed mineral aerosol distribution are associated with topographic basins where loose alluvial sediments may accumulate (Prospero et al., 2002). These sediments may be responsible for greater dust emission efficiencies (i.e., erodibility) than otherwise comparable locations which lack hydrologically disturbed/renewed sediments. Dust models which attempt to account for sediment-rich source regions succeed in reproducing significant spatial features of the dust distribution (Ginoux et al., 2001; Zender et al., 2002). However, the extent to which parameterizations of source efficiency are required to explain the observed dust record, and the physical basis of such parameterizations, have not yet been adequately investigated.

Dust emissions are directly related to wind speed, atmospheric stability, surface roughness, vegetative cover, gross soil texture, and soil moisture. In addition, emissions show large variations attributable to other soil characteristics such as parent soil (saltator) texture, fine particle aggregation, soil modulus of rupture, and degree of disturbance. The second group of properties are related to current and past hydrologic activity because precipitation and surface runoff in and upstream of dust sources are linked to local soil abundance, size, chemical properties, and disturbance history.

We quantify the dust source efficiency $S$ as the ratio of actual vertical dust mass flux $F_d$ to the mass flux $F_{d,0}$ mobilized from an idealized surface in the absence of regional geographic influences. Thus $S$ is intended to represent the influence of regional topography, geomorphology, and hydrology on dust emissions. If regional geomorphologic and hydrologic processes are unimportant in explaining present day dust emissions, then global simulations using $S = 1$ should perform no worse than simulations which account for regional heterogeneity.

Methods

Four one-year global dust simulations are performed, a control and three experiments. In the control simulation, called “Unity”, regional geographic properties are neglected so that $S = 1$ globally. The first experiment, called “Topo”, tests the influence of topographic basins using the
so-called basin factor of Ginoux et al. (2001). Ginoux et al. parameterized $S$ as the fifth power of the ratio of the local height above a regional minimum height to the total elevation range of the surrounding $10^\circ \times 10^\circ$ region. The second experiment, called “Geo” defines $S$ proportional to geomorphologic basin area. The basin area is the upstream area from which surface runoff may reach a given location. We computed the basin area from a Digital Elevation Map using standard hydrologic techniques. The third experiment, called “Hydro”, defines $S$ proportional the surface hydrologic flow accumulation through each point. In this case, surface runoff was obtained from a 20 year present-day simulation of the Community Land Model.

We evaluate the importance and efficacy of each process in the DEAD model (Zender et al., 2002) forced by interpolated 6-hourly NCEP winds for the years 1994–1996. The four three-year simulations were tuned a posteriori to produce identical annual mean atmospheric mass burdens. Thus differences in the statistics of the dust distribution are due to the spatial distribution of sources, and to regional differences in the efficiency of sink processes, especially wet deposition.

**Results**

Figure 1 shows the annual mean dust burden simulated using four different source emission factors. The experiments show many features which are controlled by regional geomorphologic influences. All three experiments show enhanced emissions from the Bodele depression that are seen in observations (Prospero et al., 2002) but missing from the control. This regional behavior is consistent with the hypothesis that hydrologic basins are richer in dust source materials, and thus more efficient dust sources than nearby regions with comparable meteorology but less erodible
source material. All experiments improve overestimated emissions near the Indian Ocean, but the Hydrologic experiment overestimates emissions in the Himalayan region.

The differences between the topographic basin experiment and the control simulation show the important impact that topography has on emissions. Assigning enhanced erodibility to topographic minima (Ginoux et al., 2001) realistically increases emissions in the central Sahara around the Bodele depression, in the Tarim Basin in eastern China, in the Aral Sea region, and in the Lake Eyre region in central Australia. Emissions are reduced in the eastern Sahara near Sudan, in Namibia, in South America near Patagonia, and in western Australia.

The differences between topographic basins and geomorphologic basins show the impact of realistically defined basins. The geomorphologic basins enhance emissions from the Namibian regions and reduce emissions from the Algerian regions of the western Sahara. These changes significantly improve model agreement with station observations of dust concentration and seasonal cycle obtained from the Rosenstiel School of Marine and Atmospheric Sciences aerosol network (not shown). Of the four erodibility assumptions tested, the geomorphologic basins produced the closest overall agreement with observations.

The differences between geomorphologic basins and hydrologic basins show the regions in which present day hydrological transport and disturbances would be expected to have the greatest impact. Regions with increased emissions include east China, the Indus River valley in Pakistan, and the Mesopotamian region in Iraq. Each of these regions has a high concentration of sediments from the upstream regions it drains. The $S$ in active river valleys are much larger than $S$ in desert basins, so it is somewhat surprising that the simulated dust distribution is not completely dominated by Earth’s major coastal estuaries in this case. This is because other constraints in the dust model (Zender et al., 2002), primarily vegetation and soil moisture, control dust emissions in these wet regions.

Regions where present day hydrology decreases emissions include the Aral Sea region, the eastern Sahara, and Patagonia. These regions are known dust sources, and whether the present day hydrologic source/disturbance hypothesis improves model performance there should be examined using in situ or satellite observations. We note that basing erodibility on present day hydrology does not test the alternate hypothesis that present day erodibility is controlled by sediments accumulated in past climates when runoff differed. Indeed, our future work will include testing the hypothesis that present day emissions from Africa are from deposits accumulated during the moister Saharan climate of the mid-Holocene.

Table 1 compares the annual mean dust budget of the control to the three experiments. The Geo and Hydro experiments have $\sim 20\%$ shorter turnover times than Topo because their emission regions are closer to regions where wet deposition is significant.

**Conclusions**

Global simulations which account for regional geomorphological and hydrological influences show significant spatial distinctions from simulations which do not impose spatial heterogeneity in dust source efficiency. In many cases the signature of regional geomorphologic and hydrologic influences improves the dust simulations relative to observations. However, the model only parameterizes natural processes relevant to sediment accumulation and disturbance. Dust observations, on the other hand, record anthropogenic contributions (e.g., disturbance) to source efficiency $S$. The present work is a first step toward partitioning observed source enhancement between natural
Table 1: Climatological Budget Statistics

<table>
<thead>
<tr>
<th>Quantity</th>
<th>Unity</th>
<th>Topo</th>
<th>Geo</th>
<th>Hydro</th>
</tr>
</thead>
<tbody>
<tr>
<td>Burden(^a)</td>
<td>22.7</td>
<td>22.7</td>
<td>22.7</td>
<td>22.7</td>
</tr>
<tr>
<td>Emission(^b)</td>
<td>2066</td>
<td>1718</td>
<td>2070</td>
<td>2654</td>
</tr>
<tr>
<td>Lifetime(^c)</td>
<td>4.0</td>
<td>4.8</td>
<td>4.2</td>
<td>4.2</td>
</tr>
<tr>
<td>(\tau) Global(^d)</td>
<td>0.036</td>
<td>0.037</td>
<td>0.037</td>
<td>0.037</td>
</tr>
</tbody>
</table>

\(^a\) Prescribed atmospheric burden in Tg
\(^b\) Emissions of \(D < 10 \ \mu m\) particles in Tg yr\(^{-1}\)
\(^c\) Turnover time in days
\(^d\) Optical depth at 0.63 \(\mu m\)

and anthropogenic factors.

**References**


Partially closed aeolian/alluvial sedimentary systems on Mojave Desert alluvial fans

C.T. Baldwin, Dept. Geog. & Geol./TRIES, Sam Houston State Univ., Huntsville TX 77341 (E-mail: baldwin@shsu.edu)

R. Rush, TRIES, SHSU, Huntsville TX 77341 (E-mail: ENV_RNR@shsu.edu)

J. Brint, TRIES, SHSU, Huntsville TX 77341 (E-mail: stdjwb13@shsu.edu)

The Marine Corps training base at Twenty Nine Palms, CA, situated in the southern Mojave Desert includes a series of high relief, north-west/south-east trending rock ridges and intervening plains and valleys. The rock ridges partially block the prevailing north -west to south-east aeolian sand transport paths such that alluvial fans on west facing slopes are mantled with sand, including transverse and swept dunes. In some places sand climbs sufficiently high to leak through the ridges and to drape east facing slopes and to continue on to the next ridge.

The sand covered fan surfaces are cut by deep, sometimes meandering alluvial drainage channels that incise the mid fan surfaces and terminate lower down on the lower and distal fan as delta-like units or splays. These channels give rise to a system in which water is transported across the upper and mid-fan surfaces in quasi-stable channels that lack the more extensive dispersal characteristics of normal alluvial fan channels. At the upper end of these channels they connect to rock-bounded or coarse debris-bounded chute-like channels of the upper alluvial fan. In other cases alluvial fan channels have coalesced to flow along the boundary at the top of the aeolian mantled surface of the fan (i.e. obliquely across the upper fan) so beheading alluvial drainage across the mid and lower fan. In both cases the flash flood dominated alluvial systems return stored increments of aeolian sediments to the distal fan and beyond where they are subsequently reworked back up the fan surface by the wind. As a result a partially closed loop of sand transport is developed on these fan systems.

Troops training in the area are reasonably familiar with the hazards attached to flash flood events in the highly confined rock channels and canyons of the higher parts of the rock ridges. Unfortunately, the presence of a constantly rejuvenated supply of sand to the surfaces of the extensively used fan systems makes them particularly hazardous to military personnel because they too now exhibit zones of highly confined flow. Further, most of the fixed plant and facilities of the base are situated in a zone of closed sand circulation requiring costly stabilization and removal of sand washed down by flooding events.
Modelling to reconstruct recent wind erosion history of fields in eastern England.

A. Chappell, University of Salford, Greater Manchester, U.K. (a.chappell@salford.ac.uk)

A. D. Thomas, University of Salford, Greater Manchester, U.K. (a.d.thomas@salford.ac.uk)

Introduction

Wind erosion seriously damages crops, jeopardises sustainability and creates dust that is a significant air pollutant in many parts of northwestern Europe. In East Anglia, UK a recent survey showed that farmers expect moderate damage to crops from wind erosion once every three or four years and severe damage once in ten years. Moreover, off-site costs are probably many times those of the on-site costs, based on experience in the USA (Piper 1989). Despite considerable anecdotal evidence there is a dearth of quantitative evidence to support wind erosion and deposition rates (soil flux). This is probably because aeolian activity is dependent on the spatial and temporal variation of the erodibility (Zobeck, 1991) and erosivity conditions. This complexity renders all but the longest and spatially intensive monitoring campaigns unrepresentative of the wind erosion pattern. The selectivity of aeolian processes removes the smaller size fractions which can be considerably enriched in soil nutrients (Zobeck and Fryrear, 1986). However, the affect of this may not be apparent because net soil nutrient loss is small as a consequence of more intensive use of fertilisers and slurries sprayed onto and injected into the soil. Thus, in the U.K., the affect of soil loss by wind on agricultural production is not well documented or understood.

The study utilises an accumulation of soil in a boundary adjacent to a field to validate a model of wind erosion. The study field is located near Thetford in East Anglia, U.K. It typically undergoes a three-year crop rotation of sugar beet, winter wheat and carrots. The easterly end of the field has a large (ca. 1 m) accumulation of soil which began forming in the mid 1950s shortly after a fence was constructed forming a boundary to the easterly end of the field. It is postulated that historical wind erosion events produced the accumulation of aeolian material and recorded its intensity and the type of material removed. The aim here is to construct a model to predict the concentration of nutrients in eroded material under different conditions (plough depths, erosion rates, enrichment ratio and soil nutrient content). Variations in the nutrient content of the dated wind blown accumulation can then be used to examine the impact of historical wind erosion on the field. Further, as historical land use information is available, the impact of different crop types on erosion can be determined.
Methods

Samples were obtained from the soil accumulation in order to verify the output from the model. A pit was dug in the accumulation and a monolith was removed and sampled every 2 cm along its length. All samples were measured for pH to indicate the redox soil status and mineral magnetic susceptibility which was intended to measure the enhanced levels of secondary ferrimagnetic minerals (e.g. haematite) in the upper soil horizons (Mullins, 1977). Fewer samples were measured for soil chemical properties including total nitrogen (N), available phosphorous (Olsen’s P) and exchangeable potassium (K) as indicators of the soil nutrient status.

The Model

For each year since 1954 the model predicts the concentration of P in the accumulation of wind blown material (Equation 1, 2). Phosphorus was chosen initially because it is relatively immobile in the sediment and is a key plant nutrient. Various outputs from the model were plotted against the actual P concentration from the dated profile obtained from the wind blown accumulation. The model accounts for the mixing of subsurface material deficient in P into surface soil after ploughing and the lowering of the soil surface through erosion. The model used a nutrient enrichment factor for the eroded material of two times (based on empirical data from the study site) and assumes there is a small annual accumulation of P due to the excess of artificial inputs over and above plant uptake.

The P concentration in the wind blown material was predicted using the following:

\[ P_{surf}^t = \left[ \frac{D}{P} \times P_{sub}^t \right] + \left( P \times \frac{P_{surf}^{t-1} - D}{P_{surf}^t} \right) \times \left( P_{surf}^{t-1} + F \right) \]  
Equation 1

Where:
- \( P_{surf}^t \) is the predicted P concentration at the soil surface (integrated through the ploughed depth)
- \( P_{surf}^{t-1} \) is the previous year’s P concentration at the soil surface
- D is the depth of soil loss
- P is the plough depth
- \( P_{sub}^t \) is the subsurface (below plough depth) P concentration
- F is the change in soil surface P concentration after fertiliser additions and plant uptake at the end of the year

And the predicted concentration in the eroded material is:

\[ \text{Predicted P concentration in eroded material} = P_{surf}^t \times ER \]  
Equation 2

Where ER is the enrichment ratio of the eroded material to the soil surface concentration.
Discussion

Initial findings suggest that inter annual concentrations of wind blown nutrient content can be predicted using a simple model accounting for mixing in the plough layer and enrichment. It is clear that there have been significant historical losses of P as a result of wind erosion. In this sandy, nutrient deficient soil, fertiliser inputs are increasingly necessary to maintain soil fertility.

References


The causes and processes of sandy desertification of typical region in north China

Duan Zhenghu*** Xiao Honglang * Wang Gang**
* Shapotou Desert Experimental Research Station, Cold and Arid Regions Environmental and Engineering Research Institute, Chinese Academy of Sciences; 260 Donggang West Road, Lanzhou, 730000, P.R. of China; and ** State Key Laboratory of Arid Agroecology, Lanzhou University, 298 Tianshui Road, Lanzhou, 730000, P.R. of China

Abstract
Improper farming practices, overgrazing, the conversion of rangeland to croplands in marginal areas and uncontrolled expansion of urban and rural settlement at the cost of cultivable land are among the major causes of land degradation in northern China. The purpose of this study was to discuss the major causes of land degradation in the area. Six sites receiving different amounts of annual precipitation, annual evapotranspiration, and annual temperatures and with different vegetation types were selected to represent the major agricultural areas in northern China. The major soil properties that can be linked to land degradation were studied. Desertification in northern China is taking place through loss of soil fertility and productivity, overgrazing and water and wind erosion. Erosion by wind and water is considered the major cause of land degradation in the area. The soils contain little organic matter and their alkaline reaction reduce the availability of phosphorous and macronutrients and consequently lead to very low crop yields.
Chemical contaminants, globally transported dust and downstream ecosystems

V. H. Garrison, US Geological Survey, St. Petersburg, FL
W. Foreman, US Geological Survey, Denver, CO
M. Majewski, US Geological Survey, Sacramento, CA
C. Holmes, US Geological Survey, St. Petersburg, FL
E.A. Shinn, US Geological Survey, St. Petersburg, FL
D. Griffin, US Geological Survey, St. Petersburg, FL
C. Kellogg, US Geological Survey, St. Petersburg, FL
R. Smith, American International School, Bamako, Mali
M. Ranneberger, American International School, Bamako, Mali

Viable microorganisms, macro- and micronutrients, trace metals and a variety of chemical contaminants, predicted to be carried in global dust systems and deposited in the oceans and on land, may play important roles in the complex changes occurring in downstream ecosystems. We are specifically focusing on coral reef and human health. Every year, hundreds of millions of tons of African dust are transported from the Sahara and Sahel across the Atlantic to the Caribbean, and southeastern U.S. A similar global system transports dust from the Gobi and Takla Makan deserts across Korea, Japan, and the northern Pacific to the Hawaiian Islands. The Asian system periodically reaches the western U.S. and at times transits North America to the western Atlantic Ocean. Fine soil particles have been transported globally for millions of years, however the quantities of dust have increased dramatically since the early 1970s. Global climate systems, local meteorology, geomorphology of source areas, and regional land use practices affect the quantities of dust transported. The quality of the dust has also changed. Dust may transport a variety of microorganisms and chemicals that hitch hike on or within the small particles. We have recently begun a pilot project to test our hypothesis that Sahelian dust air masses transport chemical contaminants (from the burning of biomass and waste, and use of antibiotics, pharmaceuticals, and pesticides in dust source regions) thousands of kilometers to the Americas, and that those contaminants may be working synergistically to adversely affect coral reef and human health.
Aeolian Dust Captured on Isolated Surfaces along a Transect from the Mojave Desert to the Colorado Plateau, USA

Goldstein, Harland, U.S. Geological Survey, MS-980, Federal Center, Denver, CO 80225 (hgoldstein@usgs.gov)

Reheis, M., U.S. Geological Survey, MS-980, Federal Center, Denver, CO 80225 (mreheis@usgs.gov)

Reynolds, R., U.S. Geological Survey, MS-980, Federal Center, Denver, CO 80225 (rreynolds@usgs.gov)

Yount, J., U.S. Geological Survey, MS-980, Federal Center, Denver, CO 80225 (jyount@usgs.gov)

Simmons, K., U.S. Geological Survey, MS-980, Federal Center, Denver, CO 80225 (ksimmons@usgs.gov)

Lamothe, P., U.S. Geological Survey, MS-973, Federal Center, Denver, CO 80225 (plamothe@usgs.gov)

Luiszer, F., University of Colorado at Boulder, Geological Sciences, 399 UCB, Boulder, CO 80309-0399 (Fredrick.Luiszer@Colorado.edu)

This study centers on the role of dust with respect to nutrient inputs and influence on hydrologic properties of soil. As such, conditions of dust emission, dust sources, and dust flux under variable climatic and land-use histories bear on interrelations between landscape evolution and ecosystem dynamics. We are examining the causes of local to regional variations in texture and composition of arid-land soils with respect to past and modern dust inputs that may have varied due to changes in climate and dust sources.

In a previous study, aeolian dust was identified in fine-grained sediment captured in small depressions (potholes) on the central Colorado Plateau (Reynolds et al., 2001). The potholes were sampled (typically to 5-cm depth) on high, isolated surfaces to eliminate the possibility of contamination by colluvial or alluvial sedimentation. Primary evidence for the aeolian origin for dust in the pothole sediments was the presence of detrital magnetite, which yielded moderately high magnetic susceptibility (MS). Rocks in which the potholes formed were devoid of magnetite and yield negligible MS values. Physical and chemical properties of dust in these potholes suggested dust sources outside the Colorado Plateau. For example, a high proportion of the silt fraction consists of < 10 µm particles that are characteristic of far-traveled dust, and the content of magnetite and trace elements implied that some dust originated in geologic terrain characterized by felsic igneous rocks, such as the Mojave Desert of southeastern California (Reynolds et al., 2001). The dust in potholes on the Colorado Plateau was not dated. The uppermost samples were collected from biologic soil crust in the upper ~1 cm on a surface that has been stable for over at least 70 years (Belnap and Gardner,
1993); thus, the sampled dust likely represented deposition over the past several centuries to the present day.

The current study was designed to investigate further the possibility that some dust in Colorado Plateau soils originated from sources in the Mojave Desert. Following the strategy of Reynolds et al. (2000), we collected samples from potholes on high, isolated surfaces, developed mostly on Jurassic aeolian sandstone, along a transect from the central Mojave Desert northeast to the original study area on the central Colorado Plateau. Initially, physical and chemical parameters of the silt- and clay-size fraction were compared with those of associated bedrock to confirm the presence of aeolian dust. In addition, the spatial distributions of geochemical and mineralogic parameters in the pothole sediment were examined for regional trends along the southwest-to-northeast transect. If found, such trends might reveal some contribution of Mojave dust sources to fine-grained sediment on the Colorado Plateau.

Magnetic, isotopic, and chemical data indicate the presence of aeolian dust in the potholes. First, MS values for the silt and clay (fines) fraction range from \(1.40 \times 10^{-6}\) to \(1.59 \times 10^{-7}\) m\(^3\) kg\(^{-1}\), whereas MS values for the associated bedrock is typically \(< 7.27 \times 10^{-9}\) m\(^3\) kg\(^{-1}\). As in the case of the earlier study, the relatively high MS of the fines in these potholes is generated by the presence of detrital magnetite and titanomagnetite. These minerals are lacking in associated bedrock, and thus must have been deposited as atmospheric dust. Similarly, magnetic measurements of hematite content show much higher hematite contents in the fines compared to those in associated bedrock. Second, \(^{87}\text{Sr} / {^{86}\text{Sr}}\) values of the carbonate fraction (acetic-acid leach) in the pothole fines range between 0.7082 - 0.7109 and are less radiogenic than are leach fractions from the associated bedrock samples (0.7110 - 0.7144), indicating that the pothole sediment did not originate by weathering of the bedrock. Third, chemical data obtained by inductively coupled plasma spectroscopy (ICP-AES and ICP-MS) show increased elemental content in the pothole fines compared to the bedrock. For example, when all pothole sites are considered together, elemental contents of Ti and Zr in the fine fraction of the pothole sediment is as much as 7x that of the bedrock. Textural data are consistent with the presence of far-traveled dust in the pothole sediments. Silt and clay make up more than 30% of the pothole sediment, in distinct contrast to less than 15% in the bedrock. Normalized distributions within the \(<63\mu m\) size fraction reveal a high proportion of particles \(<12\mu m\) in diameter. These sizes are typical of particles that are capable of being transported over large distances (>100 km) (Yaalon and Ganor, 1973; Goudie, 1978).

Spatial trends are found in some magnetic, chemical, and textural parameters of the potholes fines. MS values (magnetite contents) decrease from southwest to northeast. This trend likely is related to physical sorting during transport of dust from magnetite-rich terrain in arid regions west of the Colorado Plateau. Similarly, Ti, Fe, and Zn generally decrease eastward in abundance. Zr, however, shows an eastward increase. Within the trends of eastward decreasing contents of magnetite and Ti, a few spikes of high MS and correspondingly high values of Ti likely reflect local dust contributions related to nearby exposures of basalt. Finally, the particle-size distribution within the silt fraction is coarser in pothole sediment west of the Colorado Plateau than in pothole sediment on the Colorado Plateau.

Data from fine-grained sediment in potholes on isolated surfaces are consistent with sources of dust in Colorado Plateau soils from the Mojave Desert. As such, processes acting in distant regions, for example the Mojave Desert, can influence ecosystem dynamics hundreds of kilometers away on the Colorado Plateau. However, many other dust sources have surely contributed. Therefore, we need to further evaluate the effects of dilution by dust from local sources on the chemical and magnetic parameters we have measured. Our results suggest that aeolian dust can strongly influence soil geochemistry over large areas of the southwestern U.S. The results also provide insight into the
causes of regional variations in texture and composition of arid-land soil with respect to modern ecosystem dynamics.

References


How Wind Erosion Processes Affect Selection and Performance of Erosion Control Systems

L. J. Hagen, USDA, ARS, Wind Erosion Research Unit, Kansas State Univ., Manhattan, Kansas (E-mail: hagen@weru.ksu.edu)

Introduction

During the past decade, the physical processes governing erosion of soil by wind have been investigated by a number of researchers (Anderson et al., 1991; Armbrust and Bilbro, 1997; Gillette et al., 1997; Hagen et al.; 1992; Hagen et al.; 1999; Marticorena et al., 1997; and Mirzamostafa et al., 1998). The results of these investigations have improved our understanding of several wind erosion phenomenon. Among these are estimates of erosion threshold wind speeds, entrainment rates of loose soil, abrasion rates of crust/clods, breakage rates of saltation/creep, interception efficiencies of plants, and trapping rates of eroding soil. But to fully utilize our improved understanding of these processes, wind erosion models must incorporate them.

However, even models that incorporate most of these processes still rely upon the model user to optimize the design of individual wind erosion control systems. Generally, the goal of optimization is to achieve acceptable erosion control at minimum cost for a given land management system. While implementing erosion controls may or may not provide short term positive economic returns for the land manager, failure to control erosion often generates large offsite costs (Huszar and Piper, 1986).

Effectively mitigating offsite impacts may include the need to consider other factors in addition to total soil loss. For example, emissions of PM-10 (i.e., particulate matter less than 10 microns in diameter) from farmlands and other disturbed areas near the rural-urban interface may prevent urban areas from meeting air quality standards. Other critical offsite impacts are caused both nutrient and sediment transport to water bodies, pesticide movement to non-target areas, and decreased visibilities that hinder transportation. For all these cases, erosion control specialists need to understand the influence of various processes on their design parameters. They also may need to use specialized designs that vary depending on the location of critical offsite targets.

The objectives of this report are twofold. First, to present a brief overview of several of the wind erosion processes, and second, to suggest how these processes might influence selection and design of wind erosion controls.

Erosion Processes and Design of Controls

Soil Roughness

Currently, millions of hectares are protected from erosive winds by the combined use of soil roughness and immobile soil aggregates. It is a fragile control system and subject to failure when the soil aggregation decreases during prolonged droughts. The macro roughness of bare soil, as measured by conventional pin meters, controls the aerodynamic roughness of the surface. But the macro roughness is not highly correlated to the surface aggregate size distribution (Wagner and Hagen, 1992). Hence, these should be treated as separate parameters in design of erosion control systems.

Oriented soil roughness such as tillage ridges control erosion primarily by trapping and
sheltering mobile particles between the ridges. However, their level of efficacy is variable depending on the wind regime speed and directional variability as well as surface conditions. They are most effective when the tops of the ridged surface are armored with immobile aggregates or residues. This raises the threshold wind speed at which erosion begins and sharply reduces the transport rate of mobile particles. Experiments show that as the clod cover on ridges is increased and mobile soil decreased, saltating particles no longer impact loose soil so the threshold velocity increases from dynamic to the static threshold (Hagen and Armbrust, 1992).

When ridges are not fully armored with immobile material and considerable soil movement is anticipated, the designer should specify creation of large ridges in order to maintain sufficient storage capacity in their sheltered region for the mobile soil particles. When erosive winds parallel to ridges are expected, random roughness and perhaps furrow dikes are useful to further enhance erosion control. Ridges are generally constructed parallel to crop strips. However, in some wind regimes, an enhancement of erosion control can be gained by separately optimizing both the angle of the crop strips and the tillage ridges.

The shearing stress on ridge tops and the turbulent diffusion above ridges is enhanced compared to a smooth surface. Thus, if tillage ridges are composed of mostly suspension-size soil aggregates, the ridges may increase rather than control erosion for this field condition.

**Downwind Field Length to Nonerodible Boundaries**

Reducing field length limits the opportunity for saltating particles to abrade and breakdown the immobile soil aggregates and crust cover over long downwind distances. Computer simulations of wind erosion on long fields with initially uniform surfaces, revealed that for some surface conditions there were intermediate field lengths that produced a significant maximum of soil loss per unit area. Many fields that are relatively stable often include small areas that may begin to erode and thereby destabilize the downwind area. In such cases, limiting field length can prevent large downwind areas from becoming unstable and eroding. Typically, strip cropping is used to limit field length, but any control device that traps the saltation may be used. The trapping capacity of downwind traps also should be evaluated in the design process.

**Standing and Flat Vegetation**

Standing vegetation controls erosion primarily by reducing wind shearing stress on the surface and, secondarily by intercepting particles of moving soil (Hagen, 1996). Obviously, the efficacy of the shear reduction on erosion is not a constant but depends on the particular wind regime. Nevertheless, residue stalks are generally 5 to 10 times more effective than flat. Hence, design of land management practices that reduce the flattening of standing vegetation are warranted in erosion prone areas.

Flat vegetation controls erosion mainly by sheltering the surface from impacting particles and, secondarily by reducing shear stress on the surface. At a fixed wind speed, one can readily estimate the reduction in emission rate of loose sand by estimating the shelter area provided by flat residue of a given diameter and assuming an average particle impact angles of 12 degrees above horizontal. Flat residue often tends to blow away, so management practices to keep it in place are useful.

Flat residue is more effective than standing in controlling water erosion. Thus, where both are significant, the designer must aim for an optimum combination of standing and flat residues to attain target values of erosion control.

**Wind Barriers**

Wind barriers may be composed of natural or manufactured materials. Their main function is
to reduce wind speeds near the surface both slightly upwind and as well as downwind from the barrier. In general, barriers are useful for controlling wind erosion, but there are a number of challenges in designing optimum barrier systems. Barriers are most effective when combined with other erosion controls, because their zone of erosion control expands with increases in threshold velocity of the soil surface. Generally, it is critical to obtain barrier porosities less 70 percent and orient the barrier normal to the erosive winds during periods when the soil is erodible. Meeting these criteria can often require natural barriers to be composed of at least two rows.

Tall barriers on the downwind side of wide fields may trap much of the saltation/creep component of the eroding soil on the field, so models reporting only net field loss may suggest there was acceptable erosion control. Hence, when downwind trapping occurs, one challenge is to present sufficient information to those designing controls to show that there may be both a loss and deposition problem occurring on the same field.

A typical design with barriers and strip crops is to place the barriers on the strip borders. But barriers typically trap soil moving near the surface, and thus, serve as a non-erodible field boundary. Hence, in some design situations placing the barrier in the center of each strip can serve to reduce field lengths and thus, improve the level of erosion control.

Finally, one case where barriers may not be useful occurs when barriers are widely spaced on a surface composed mainly of saltation-size particles such as a coarse sand (Schwartz et al., 1997). In this case, the wind entrains sand until it reaches transport capacity and transports the load of sand to the nearest barrier. This process is then repeated across the field. Without barriers, the net removal from the field surface is equal to that with barriers, but rearrangement of surface sand on the field surface may be increased by the barriers.

**Conclusions**

In general, abatement of wind erosion must be achieved by combining a number of control mechanisms in a single control system. Significant progress toward optimizing wind erosion control systems can be achieved by considering how wind erosion processes affect the erodible surface conditions in particular wind regimes.
References


Introduction

Modern dusts in the Southwestern U.S. commonly contain, or are accompanied by, larger amounts of ordinarily-rare trace elements than can be explained by the compositions of the common minerals that constitute the dusts, or by the average composition of the earth’s crust. Records of deposition of dusts and trace elements from other places and other times give a similar picture (pre-industrial dusts preserved in Antarctic ice, and central European peat bog records). The degrees of enrichment (“enrichment factors”) of trace elements in the ancient dusts from those other places, and fine-grained modern dusts from the SW U.S., are similar. In addition to knowing the degrees of enrichment, it is necessary to calculate the amounts of excess trace elements deposited per unit area over time (mass fluxes) as they accompany the dusts, to allow evaluation of the consistency of “source strengths” of trace element supply, through time and across regions. This comparison can be made between regions of the earth, and between pre-industrial and modern times. It is then possible to compare the calculated source strengths with known sources of trace elements to the atmosphere, such as volcanic emissions.

The ordinarily-rare trace elements Pb, Cd, Cu, Se, and others are among those that are present in excess amounts in modern dusts deposited from the atmosphere, and modern atmospheric load material, relative to the rocks and soils that are the sources of the bulk of the dusts (Bowen, 1979). It has been a question whether this enrichment is due to natural processes, or industrial processes, or some combination (Duce et al., 1975; Weiss et al., 1978; Heidam, 1985; Mart, 1983; ref’s therein). Analyses of modern dusts we collected in the Southwestern U.S. over several years confirm that many trace elements (Zn, Cu, Pb, Cd, As, Se, Sb, Bi) are much more abundant in at least finer-grained dusts than in the average crust of the earth.

Besides the information from the Southwestern U.S., there are two other studies that contain information on amounts of dust and their accompanying trace elements, and that present or allow
extraction of information about the flux rates: a study of dusts in Antarctic ice representing pre-
industrial atmospheric deposition (Matsumoto and Hinkley, 2001), and a study of long-term
deposition in European peat bogs (Shotyk et al., 2002). In addition, there is a new estimate of the
source strength of trace elements from worldwide volcano emissions (Hinkley et al, 1999), one of the
natural sources of trace elements to the atmospheric load.

Methods

To obtain winter-season, high-altitude dust, snow pack strata were collected each early Spring
from 1997 to 2002 in the Southwestern U.S., under clean conditions. For year-around, low-elevation
dust, dry-deposition samples were collected on greased glass plates. Snow samples were reduced in
volume in the laboratory under flowing filtered nitrogen, and both kinds of samples were digested and
analyzed for a major, minor and trace elements by ICP-MS. Masses of dusts in samples were
estimated by summing the masses, as oxides, of elements measured in bulk analyses. Calculations of
“enrichment factors” of elements were done by comparing the concentration of an element in dust in a
sample to the concentration in average crustal material. Mineral identity, grain size and shape, and an
independent check on flux rate and bulk composition were provided by microbeam (SEM) methods.

Results and discussion

SEM analyses of samples indicate that the dusts are composed of common minerals (quartz,
feldspars, micas, clays, carbonates; small amounts of pyroxenes and amphiboles; smaller amounts of
accessory minerals such as zircon and rare earth minerals; also pollen grains), and that the mode
(assemblage of actual mineral species present) is consistent with the bulk chemical results.

In dusts in the U.S. Southwest, “enrichment factors” for trace elements (the factor by which
the concentration of an element is greater than in the average crust of the earth) are commonly
between 10 and 100, although they range by element and by sample type. Enrichment factors are
especially high in finer-grained, farther transported dusts, namely dust in snow pack strata with low
concentrations of dust, and in dry deposition samples in which only small amounts of dust are
deposited on collection plates per unit time. As for the pre-industrial dust in polar ice and the
European peat bog deposits, enrichment factors commonly range between about 10 and more than
100. The polar dust is clearly fine-grained and far-transported, the peat bog dust may be a mixture of
near-source and far-transported dust.

A possible argument to dismiss the observed similarity in degree of trace element enrichment
in modern SW U.S. dust and pre-industrial Antarctic or European peat bog dusts is that polar dusts
and dusts over heavily-vegetated Europe, being fine-grained (large surface/mass ratio) and far-
transported, must have been enriched to the maximum extent, because of “exposure” time during
atmospheric transport; but that dusts within the dry US Southwest (possibly coarser, more locally
transported) could have acquired equivalent amounts of trace elements only if a supplementary
(modern; anthropogenic or unique local) source were available. However, this argument cannot be
evaluated at present because it is not clear that either of the older dusts had finer grain size than the
fine component of U.S. SW dust, and it is not clear that either kind of dust has been transported over
greater distances, because it has been shown that fine-grained dusts of uniform composition appear to
constitute a hemispheric or global background atmospheric load, which can be identified at low-
energy times at both polar and dusty continental interior sites of deposition (Hinkley et al., 1997).
The worldwide, “background” atmospheric dust is not dependent on local source regions (Hinkley et
This fine dust may be a scavenger of trace metals (possibly from natural sources) during its long residence in the atmosphere, and may account for a large portion of the universality of trace element enrichments in dust.

Industrial trace metal pollution has been documented in northern polar and even Antarctic ice (Sherrel et al., 2000; Rosman et al., 1994). However, these modern increases in elemental inputs have been stated only as concentrations in ice, not as changes in “enrichment factors” of the dusts present in the ice, partly because reliable measurements of both dust content and trace element content have not been made together in the same studies. Those studies, and other documented trace metal pollution in the world today suggest that trace element loadings of S.W. US dusts have likely increased in modern industrial times, but the extent is unknown. Local terranes have been proposed as sources of trace element enrichment for some S.W. dusts collected in specific nearby regions (Reheis et al, 2002).

**Conclusions**

The finer-grained component of atmospheric dusts in the Southwestern U.S. is significantly “enriched” in many ordinarily-rare trace elements, commonly to factors of 10-100 above the amounts that would be expected if the sources were unaltered average crustal material. This degree of enrichment in dusts in the SW U.S. is about the same as in pre-industrial dusts preserved in Antarctic ice and in European peat bogs, which have natural sources of enrichment. The similar levels of trace element enrichment of the of dusts from the two very different times and among the three locations appear to indicate that naturally-high amounts of trace elements, as seen in the pre-industrial samples, are at the very least a significant component of what is seen today in dust in the Southwest, and that the trace element loads of the modern and ancient dust are indistinguishable at present state of knowledge. The remaining task is to compare the absolute fluxes of key trace elements (mass deposited per unit area per unit time) for such different locations and the different time periods they represent.
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elements in atmospheric aerosol since 12,370 \(^{14}\)C yr BP, and their variation with Holocene
Regional Evaluation of Wind Erosion from Loess Lands: A Case Study of Pengyang County, Ningxia, China

Yubao Li, Inner Mongolia Agricultural University, China (li_yob@hotmail.com)

E. L. Skidmore, USDA-ARS, Wind Erosion Research Unit, Kansas State University, Manhattan, Kansas 66506 (skidmore@weru.ksu.edu)

Baoping Sun, Beijing Forestry University, China (sunbp@mail.bjfu.edu.cn)

Introduction

The Loess Plateau in China is notorious for severe soil erosion by water. Soil erosion on the Loess Plateau contributes 80% of the annual sediments of 1.6 billion tons into the Yellow River (Liu, 1985). Recently, a nationwide soil erosion survey recognized that wind erosion from loess lands is a serious problem. Creep and saltation sediments from croplands and grasslands collect in gullies during the wind erosion seasons, and are then carried into rivers during the rainy seasons. So, wind erosion in loess region not only degrades land resources and pollutes local and off-site atmospheric environments, but also contributes sediments into rivers.

Zobeck et al. (2000) identified four problems that may exist in the use of field models and GIS: assumption of homogeneity, methods to derive the attribute values needed, merging data, and scale differences. Most current wind erosion models simulate an isolated homogeneous field (Hagen, 1991). The influences of sand-borne flow from upwind field are hard to estimate when it is scaled up. Evaluation of wind erosion on a regional scale in China is still a challenge. The objective of the project was to determine erosion severity for each land use according to The National Standard for Soil Erosion Classification (China Ministry of Water Resources, 1994), using Pengyang County, Ningxia as an example.

Methods

Pengyang County, located in south Ningxia Autonomous Region, China, is a semiarid region with an annual precipitation of 490mm, and average annual temperature of 5.3°C. The region has
Table 1  Classification criteria for soil erosion severity as a national standard

<table>
<thead>
<tr>
<th>Grades</th>
<th>Average erosion module</th>
<th>Average erosion dep</th>
</tr>
</thead>
<tbody>
<tr>
<td>1 Tolerable (Trace)</td>
<td>&lt;200</td>
<td>&lt;0.15</td>
</tr>
<tr>
<td>2 Slight</td>
<td>200-2500</td>
<td>0.15-1.9</td>
</tr>
<tr>
<td>3 Medium</td>
<td>2500-5000</td>
<td>1.9-3.7</td>
</tr>
<tr>
<td>4 Severe</td>
<td>5000-8000</td>
<td>3.7-5.9</td>
</tr>
<tr>
<td>5 Very severe</td>
<td>8000-15000</td>
<td>5.9-11.1</td>
</tr>
<tr>
<td>6 Extremely severe</td>
<td>&gt;15000</td>
<td>&gt;11.1</td>
</tr>
</tbody>
</table>

an average of 24 days of gale (≥17m/sec.) and signs of soil blown by wind can be observed on some barren croplands in spring. Its total land area is 2524.7 km² and total soil area is 2389.2 km², in which 93.3% is covered by loess (Wang, 1990). There are four soil groups in the region, Heilu, Loessal and Alluvial soils have developed from loess materials and Gray Cinnamon Forest soil is distributed in mountainous areas. The loess in the region is several to two hundred meters deep with gully density of 1-3 km²/km². As parent material, loess includes two major mineral constituents, quartz (49-64%) and feldspar (30-43%). Lime content is as high as 8-16%. Soil grain size composition is 17.0-24.0% of 0.25-0.05mm, 56.0-61.5% of 0.05-0.01mm, 6.0-7.5% of 0.01-0.005mm, 5.5-6.0% of 0.005-0.001mm and 8.0-8.5 of <0.001mm. The regional wind erosion evaluation is based on The National Standard for Soil Erosion Classification (Table 1) issued by China Ministry of Water Resources (China Ministry of Water Resources, 1994). Grasslands and croplands were evaluated separately. The SQL language was used to execute evaluation regulations in GIS environment.

GIS Database of Land Use

A GIS database is essential to regional wind erosion evaluation. Processed scenes of Landsat Thematic Mapper(TM) images of Ningxia (July, 1996) and Ningxia Land Use Map provided by the Remote Sensing Institute of Chinese Academy of Sciences are base maps. Ningxia Relief Map (1:100,000), the Soil Type Map (1:350,000) and Vegetation Map (1:500,000) were consulted as references when completing the GISdatabase. For each land use, we added fields including geomorphological element, soil group and subgroup, annual precipitation and gale days, autumn and summer crop type, and irrigation. The database and base map is established in ARC/INFO geographical information system. Region Manager, a geographic information system specialized in water and soil conservation planning was employed to realize evaluation regulations(Shi and Sun, 1996). Field validations were conducted in July and October of 1999 and April and June of 2000.

Evaluation Indicator System

Factor systems selected as indicators to evaluate erosion intensities are different between grassland and cropland. Classification of these indicators was shown in Table 2. Erodible Fraction was added latter as an indicator. Grains less than 0.84mm in diameter were defined erodible. Erodible fraction referred to the percentage of erodible particles in the surface soil. Eighteen soil samples from 6 land types collected within 5 mm were directly sieved on the sites in October 15 of 1999, using a shovel to collect dry surface soil, then weighted in laboratory after air-dried. The sieves were manually operated. Each natural soil sample was around 5 kg.
Table 2  Indicators and their gradation for wind erosion evaluation

<table>
<thead>
<tr>
<th>Indicators</th>
<th>Grades classified</th>
</tr>
</thead>
<tbody>
<tr>
<td>Vegetation Coverage(%)</td>
<td>High: &gt;70, medium: 70-30, low: &lt;30</td>
</tr>
<tr>
<td>Erodible Fraction(%)</td>
<td>&lt;30, 30-40, &gt;40</td>
</tr>
<tr>
<td>Landforms</td>
<td>gully, valley, lower slope and small basin; high table, hilltop and middle and upper slope</td>
</tr>
<tr>
<td>Farming Management</td>
<td>irrigated, nonirrigated</td>
</tr>
<tr>
<td>Irrigation</td>
<td>autumn, summer</td>
</tr>
<tr>
<td>Crop Type</td>
<td>arbor, shrub, none</td>
</tr>
</tbody>
</table>

Erosion Severity Ranges in the Region

Plowed and harrowed cropland was defined as the most erodible in terms of the condition of vegetation, roughness and aggregation. Precipitation and the gale days in 1999 were 442.6 mm and 31 days, which were slightly drier and windier than the recent 10-year average. Three pieces of dry croplands in Pengyang County, Ningxia were selected to measure the soil loss from October 15th, 1999 to May 20th, 2000 (an erosive rain event occurred on May 26). Two hundred and fifteen metal rods (200mm long and 5mm in diameter) with millimeter marks were put vertically in the 3 pieces of lands. Each rod occupied an area of 3m×3m, evenly distributed in the lands.

Field observations by Sun and Li were conducted in April and June 2000. For the least erodible lands, erosion severity of Trace (average erosion modulus less than 200 Ton km⁻² yr⁻¹) were determined according to: 1) if wind erosion signs or tracks could be found, and 2) if sediments in gullies and furrows around the lands could be observed.

Results and Discussion

Land Types and Land Use Types

To meet the needs for regional evaluation of wind erosion intensity, the land use type classification system included 3 hierarchies. The first class separated grassland and cropland from others. The second is small landform, basically including slopes and gullies. The third covered information of vegetation coverage or cropping. As shown in Table 3, croplands occupied more than 70% of the total land area of the county, among which 72% was slope farmlands. In total, 125 polygons were validated. Interpretation error is 27.3%. The classifications of geomorphological forms and crop types presented the majority of total error.
Table 3 Areas for land types and land use types in 1999 of the Pengyang County.

<table>
<thead>
<tr>
<th>Land use type</th>
<th>Area</th>
<th>Land use type</th>
<th>Area</th>
</tr>
</thead>
<tbody>
<tr>
<td>TOTAL</td>
<td>252,466.7</td>
<td>2.1 slope</td>
<td>129,306.4</td>
</tr>
<tr>
<td>1 Grasslands</td>
<td>55,200.0</td>
<td>2.1.1 summer crop slope</td>
<td>41,378.1</td>
</tr>
<tr>
<td>1.1 gully</td>
<td>16,339.2</td>
<td>2.1.2 autumn crop slope</td>
<td>87,928.3</td>
</tr>
<tr>
<td>1.1.1 high grass gully</td>
<td>14,705.6</td>
<td>2.2 valley</td>
<td>49,293.6</td>
</tr>
<tr>
<td>1.1.2 medium grass gully</td>
<td>58,821.1</td>
<td>2.2.1 summer crop valley</td>
<td>83,799.9</td>
</tr>
<tr>
<td>1.1.3 low grass gully</td>
<td>89,866.6</td>
<td>2.2.2 autumn crop valley</td>
<td>32,876.7</td>
</tr>
<tr>
<td>1.2 hilltop</td>
<td>38,860.8</td>
<td>2.2.3 irrigated valley</td>
<td>80,370.0</td>
</tr>
<tr>
<td>1.2.1 high grass hilltop</td>
<td>34,975.1</td>
<td>3 others</td>
<td>18,666.7</td>
</tr>
<tr>
<td>1.2.2 medium grass hilltop</td>
<td>198,190.0</td>
<td>3.1 forest</td>
<td>26,666.7</td>
</tr>
<tr>
<td>1.2.3 low grass hilltop</td>
<td>155,443.3</td>
<td>3.2 water</td>
<td>26,666.7</td>
</tr>
<tr>
<td></td>
<td></td>
<td>3.3 others</td>
<td>133,333.3</td>
</tr>
</tbody>
</table>

Erosion severity evaluation

As supposed, the plowed and harrowed croplands have more erodible particles in surface layer (Table 4). The average erosion depth from the tested croplands was 3.19 mm, falling into Medium erosion severity. So wind erosion severity for all the lands in the estimated region is between Trace to Medium. The field observation determined that the slightest grade of severity, Trace included 7 land use types, accounting for 23.1% of the total soil area while the severest grade, Medium, occupied 34.8%.

Table 4 The results of wind erosion severity evaluation for Pengyang County.

<table>
<thead>
<tr>
<th>Erosion intensity</th>
<th>Land types</th>
<th>Erodible frac Management</th>
<th>Area</th>
</tr>
</thead>
<tbody>
<tr>
<td>1 Tolerable(Trace)</td>
<td>1.1.1 high grass gully</td>
<td>Grazing</td>
<td>5,882.1</td>
</tr>
<tr>
<td>Area: 58,322.4 hm²</td>
<td>1.1.2 medium grass gully</td>
<td>Grazing</td>
<td>8,986.6</td>
</tr>
<tr>
<td>Percent: 23.1%</td>
<td>1.1.3 low grass gully</td>
<td>Grazing</td>
<td>3,497.5</td>
</tr>
<tr>
<td></td>
<td>1.2.1 high grass hilltop</td>
<td>Grazing</td>
<td>19,819.0</td>
</tr>
<tr>
<td></td>
<td>1.2.2 medium grass hilltop</td>
<td>Grazing</td>
<td>2,666.7</td>
</tr>
<tr>
<td></td>
<td>3.1 forest</td>
<td>--</td>
<td>2,666.7</td>
</tr>
<tr>
<td></td>
<td>3.2 water</td>
<td>--</td>
<td>13,333.3</td>
</tr>
<tr>
<td></td>
<td>3.3 others</td>
<td>--</td>
<td>--</td>
</tr>
<tr>
<td>2 Slight</td>
<td>1.2.3 low grass hilltop</td>
<td>Grazing</td>
<td>15,544.3</td>
</tr>
<tr>
<td>Area: 106,216.0 hm²</td>
<td>2.1.1 summer crop slope</td>
<td>Plowed in early autumn</td>
<td>41,378.1</td>
</tr>
<tr>
<td>Percent: 42.1%</td>
<td>2.2.1 summer crop valley</td>
<td>Plowed in early autumn</td>
<td>8,379.9</td>
</tr>
<tr>
<td></td>
<td>2.2.2 autumn crop valley</td>
<td>Plowed in late autumn</td>
<td>32,876.7</td>
</tr>
<tr>
<td></td>
<td>2.2.3 irrigated valley</td>
<td>Plowed in late autumn</td>
<td>80,370.0</td>
</tr>
<tr>
<td>3 Medium</td>
<td>2.1.2 autumn crop slope</td>
<td>Plowed in late autumn</td>
<td>87,928.3</td>
</tr>
<tr>
<td>Percent: 34.8%</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
Conclusions

RS and GIS technologies were employed to evaluate wind erosion severity from loess soils on a regional scale. The results show that wind erosion from traditionally managed dry croplands is a problem in the studied region. A GIS database proved essential to regional evaluation of soil erosion. Field observations to determine the ranges of erosion severity from the most and least erodible lands may improve personal subjectivity from experts’ evaluation system.

References


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Transport and dispersion of blowing dust in the Mexico Basin

M.T. López, R. Villasenor, A. I. Quintanar, and V. Mora
Instituto Mexicano del Petroleo
Eje Central Lazaro Cardenas 152, 07730, Mexico D.F.

Introduction

The Mexico City Metropolitan Area (MCMA) is one of the world’s largest metropolitan areas, containing nearly 20 million inhabitants within the Mexico City Basin. Mountainous terrain, at subtropical latitude and at high elevation surrounds the region where the MCMA lies. As in many large cities, and especially the ones located in valleys with high solar radiation, Mexico City experiences air pollution problems, especially for ozone and suspended particles. A recent study [1] shows that the daily standard for PM$_{10}$ has been exceeded on more than 40% of the days in some years (although in 1999 the standard was exceeded on fewer than 10% of the days).

High concentrations of PM$_{10}$ in the MCMA can be attributed to a combination of meteorological conditions and emission patterns. A typical pattern that produces high PM$_{10}$ concentrations in the Mexico Basin cannot be described by the typical conceptual description of most mid-latitude valleys and basins [2,3]. Collins and Scott [4] stated that air quality problems in the Mexico Basin are exacerbated by strong temperature inversions that form within the elevated basin. A more recent study of the boundary layer evolution and diurnal flow circulation over the Mexico Basin and Mexican plateau [3] contradicts this assertion. With data collected in a measurement campaign in February and March 1997 [5,6] Whiteman and co-workers[3] were able to show the absence of a temperature inversion usually associated to mid-latitude basins without the presence of reversing valley wind systems. For March they also noticed that the mean morning low-level stability was only marginally greater than in the free atmosphere surrounding the Mexican plateau at the same altitude. In this sense, the Mexico Basin does not exhibit one of the chief meteorological characteristics of mid-latitude basins, namely the formation of strong nighttime temperature inversions. Rather the atmospheric wind conditions inherent to the MCMA play a major role in the air quality problem. Furthermore, the particulate problem in the Mexico Basin region during the end of winter or beginning of spring is not solely attributed to manmade sources but can be influenced by non-local sources of natural origin surrounding the periphery to the north and northeast sector of the Basin [7,8].

The Mexico Basin periodically experiences windblown dust events that cause exceedances of the national ambient air quality standard for PM$_{10}$ in the densely inhabited areas of the MCMA. Blowing dust normally involves local entrainment of dust and is associated with moderate or large winds occurring in early spring when temperatures are high and humidity is low. Climatological summary of the airborne dust environment in the region reveals that the intensity and frequency of dust events downwind of geological sources takes place in the
month of March [9]. Analysis of the IMADA database using chemical concentrations of five crustal species (Al, Si, Fe, Mg, K and Ca) shows [8] that geological material was the major contributor to PM$_{10}$. In fact, 40-55% of the PM$_{10}$ mass was of geological origin at two sites (NET and XAL). This means that deposition mass fluxes at local sites near potential soil sources are quite significant when compared to the rest of the sources in the urban area.

Particulate matter was considered in the 1989 Emission Inventory (EI) as TSP, and an updated version produced in 1994. Particle emission was attributed mainly to soil erosion encompassing approximately 95% of the total particles [10]. In the 94-EI, particle emissions from sources other than erosion were considered for the first time. Primary PM$_{10}$ emissions from the most important sources, including the industrial sector, were considered in the 1996 EI. The 1998-EI, which includes economic activity data is presented in the Federal District Government web page [11] with a soil erosion apportionment of 40%. This last inventory does not report emission estimates for TSP, but contains estimates for primary sources of PM$_{10}$ and PM$_{2.5}$. Differences in methodology and changes in activity data among the 1994, 1996 and 1998 emissions inventories have made difficult to correlate emissions figures with pollution control strategies in the MCMA.

The latest phase of the Program to Improve the Air Quality in the Valley of Mexico, also known as Pro Aire [12], is about to go into effect for the next ten years. Before this program becomes officially implemented the ability to model transport and dispersion of PM$_{10}$ is necessary for areas that may be in non-attainment status for the PM$_{10}$ standard, and to assess the PM contribution the emission inventory.

Previous air quality studies on MCMA focused on the origin of high concentrations of particles including the spatial and temporal distribution of some gaseous pollutants [13,14]. Some drawbacks to those studies can be noticed. First, the domain of study excluded the regions that are prone to soil erosion. Second, the size of the domain did not fully consider the entire mountainous range that surrounds the urban area to the east and west, which caused computational difficulties for the meteorological runs.

Air Quality Field Study

The ambient monitoring component of IMADA-AVER took place over a four-week period from February 23 through March 22, 1997 [5]. Of the 28 days monitored with particle samplers, samples taken from March 2 through March 19, 1997 (18 days) are analyzed for elements, ions, and elemental and organic carbon. This period contains three distinct episodes of pollution buildup and cleanout. The period before March 8 was relatively dry, while the subsequent episodes occurred during moist weather conditions, interspersed with fogs, clouds, and rainstorms. Meteorological data included radar wind profiles, remote acoustic sounding system (RASS) temperature sensors, and temperature and humidity profiles by airsonde and surface meteorological towers [13]. For particle measurements the following air quality monitoring stations within the MCMA were used: Xalostoc (XAL), Merced (MER), Cerro de la Estrella (CES), Pedregal (PED), Netzahualcoyotl (NET) and Tlalnepantla (TLA). Samples from the first three sites were collected every 6 hours and for the last 3 every 24 hr. From
these particle measurements the total PM\textsubscript{10} and PM\textsubscript{2.5} concentrations as well as their geological origin were determined \cite{15,16}.

**Modeling Episodes**

In this study, days 5 and 6 of March 1997 were chosen for modeling because they are a manifestation of contrasting meteorological features and distinct air quality conditions. For instance, on day 5 maximum temperature and relative humidity were recorded as 28°C and 50%, while on day 6 these two drastically changed to 18°C and 80%, resulting in a significant visibility increase. Additionally on day 5 the highest PM concentrations were observed dropping to practically negligible values the following day when compared to national ambient air quality standards. It must be noted that these days were not considered in previous studies of transport and dispersion of pollutants due to the fact that attention was focussed on the meteorology and transport of gaseous pollutants within the MCMA. Also these days were considered by other authors to be atypical in the sense that synoptic conditions led to unusually strong wind within the MCMA \cite{3}.

**Wind Erosion Model**

Emissions from rural areas are primarily concerned with agricultural activity on the Mexico Basin. Windblown dust sources on the north and northeast areas outside of the MCMA invariably intensify in the month of March and the impact of these fugitive PM emissions has never been assessed through the use of mathematical modeling. When these sources are disturbed their ability to emit windblown dust is enhanced during dry periods of high wind events. These emissions are typically associated with disturbed land, such as agricultural fields under cultivation, or uncultivated soil with minimum or no vegetation coverage at all. In this work, use was made of an existing algorithm that was developed using wind tunnels and field studies to produce a wind-erosion-prediction equation \cite{17}. The wind erosion equation is currently the most widely used method for assessing average annual soil loss by wind from agricultural fields. The Natural Resources Conservation Service (NRCS) and other national agencies throughout the US use it. Saxton and co-workers \cite{18} present a newer method to estimate wind erosion and dust emissions and concentrations on an event basis but lack of field measurements inherent to the MCMA impedes its application at this time. Details on the wind erosion equation and the used values to apply the equation for the MCMA soil sources can be found elsewhere \cite{11}.

**Modeling Approach**

The CALMET/CALPUFF modeling pair was selected as a combination of a wind field generator and a pollutant transport model. The CALMET model along with CALGRID \cite{19} were developed by the state of California for the purpose of modeling photochemical oxidant formation and transport in 1987. CALGRID was then integrated into the CALMET/CALPUFF modeling framework to create a complete modeling system for both reactive and non-reactive pollutants. Three main components integrate the selected modeling system: CALMET (a diagnostic 3-D meteorological model), CALPUFF (the transport and dispersion model), and CALPOST (a postprocessing package) \cite{21}. Each of these programs has a graphical user

interface (GUI). In addition to these components, there are several other processors that may be used to prepare geophysical (land use and terrain) data in many standard formats, meteorological data (surface, upper air, precipitation, and buoy data), and interfaces to other models such as the Penn State/NCAR Mesoscale Model (MM5).

The diagnostic meteorological model, CALMET

The first component of this system, CALMET, is a diagnostic wind field generator [20] that uses surface and upper air meteorological data to predict winds and turbulence parameters in each grid of the modeling domain for each hour of a modeling period. Meteorological surface stations used for this work are located throughout the Mexico City Valley. The ten meteorological stations that conform the surface network were deployed as follows: Merced (478.5,2147.5), Chalco (509.5, 2128.4), Tacubaya (479.0, 2145.0), Teotihuacan (515.7,2176.0), UNAM (480.0,2136.2), Tulancingo (566.0,2220.5), ENEP Acatlan (474.6,2154.5), Tlanepantla (478.5,2159.3), Pedegral (478.6,2136.7), Hangares (491.3,2147.5). The upper air soundings were released at four sites with coordinates given as Chalco (509.5,2128.4), Cuautitlan (480.1,2177.1), UNAM (480.0,2136.2), and Teotihuacan (515.7,2176.0). More specifically, the origin (southwest corner) of the computational domain in UTM coordinates was (434,2080) kilometers. It was assumed that a five-kilometer horizontal resolution was reasonably accurate for model resolution, while allowing for an acceptable execution time. The orthogonal axis extends to the north and east creating a uniform grid system of horizontal squares of surface area equal to 25 km². CALMET performed 24-hour simulations on two consecutive days for the 14 UTM zone. The maximum radius of influence over land in both the surface layer and aloft was taken as 5 kilometers with a maximum acceptable divergence in the divergence minimization procedure of 5.0x10⁻⁶. The complex topography of the MCMA considers 13 land use categories.

The Air Quality Model CALPUFF

Once the predicted tridimensional wind field and micro-meteorological variables are generated and the area source emission of soil dust inventoried, these are input into the next component of the modeling system, CALPUFF. The transport and dispersion model, CALPUFF, advects “puffs” of PM emitted from modeled sources, simulating the dispersion and transformation process at each grid cell. In our case the pollutant has been assumed to behave as a passive scalar and hence no chemical transformation takes place. The primary output files from the non-steady-state Lagrangian Gaussian puff model contain hourly concentrations at all receptor locations but only MER, CER, and XAL were used for comparison.

The CALPUFF model uses the same grid system as CALMET, consisting of 9 layers over the 28x32 horizontal grid cells. The vertical layers were specified with variable spacing at heights of 20, 80, 160, 300, 600, 1000, 1500, 2000, and 2130 meters. The vertical concentration distribution in the near field is considered to be Gaussian. Dispersion coefficients are computed from internally calculated from velocity variances using micrometeorological variables supplied by CALMET. CALPUFF models dry deposition and the emitted PM₁₀ species are modeled assuming they behave as particles. The mean and standard deviation are used to compute a deposition velocity for size-ranges, and these are then averaged to obtain a mean deposition velocity. Some miscellaneous dry deposition parameters include the
reference cuticle resistance, 30.0 s cm\(^{-1}\), and the reference ground resistance, 10.0 s cm\(^{-1}\). The area sources (tons/m\(^2\)/year) were taken as a composite of irregular polygon surfaces that match a soil eroded maps obtained with a satellite imaging technique [7]. The total surface area susceptible of erosion is approximately 1200 squared kilometers, which is comparable to the MCMA. The geophysical parameters that define the soil dust properties were taken as default values as given in reference [17] since these have not been measured.

Results

Figure 1 shows the location of agricultural non-irrigated areas as dark regions, which are usually dry from January to May and hence these locations are susceptible to windblown dust events. Figure 1 also shows the political boundaries of the MCMA where all major socioeconomic activity takes place. Significant terrain features are found to the southwest and southeast of the domain. Between the two mountain ranges there is a mountain pass where topographically confined airflow is channeled into and outside the MCMA. The surrounding mountain ranges act as barriers to air pollutants restricting the horizontal ventilation.

Figure 2 shows two panels for day 5 of March at 08:00 hours as simulated by CALMET and CALPUFF. Figure 2a shows the surface wind vectors and windblown dust concentration contours while Figure 2b presents mixed layer depths. Predictions at this hour of the day are similar in shape and intensity to the earlier hours of simulation. At 08:00 hours light-to-moderate easterly winds blow over the two major dust sources located northeast of the MCMA causing entrainment and suspension of dust. Over the eastern mountain range, cool air descends with a westerly component, leading to air streams that causes minor convergence at the center of the MCMA. Two scenarios are observed in Figure 2 in regard to the soil sources. First, winds with a northerly component are able to advect dust plumes to the southeastern part of the MCMA where both monitoring receptors and CALPUFF results show lower PM concentrations. Second, the mountains on the southeast corner of the domain impede the penetration of the plumes forcing suspended dust to circle around the high mountainous range (altitudes of up to 3400 meters asl). The mixing layer height remains relatively shallow over the domain of interest for much of the morning period, but PM concentrations tend to remain relatively high near the erosion sources. The western portion of the MCMA remains unaffected by windblown dust due to in part to westerly winds that counteract the effect of the spreading dust cloud.

The predictions at 14:00 hours represents a transitional period characterized by low concentration of particulate matter measured at the monitoring locations within the MCMA from 10:00 to 15:00 hours (not shown). A significant increase in mixed layer heights over the MCMA going from 500 m to 2000 m in less than 5 hour highlights this period. In addition, the wind field displays large wind speeds (about 7 m/s) relative to early hours with a northeasterly component throughout much of the MCMA. This is in accord with the lower PM concentrations measured (not shown) during this time over the MCMA particularly at the industrial and urban areas.
Figure 1: Map of topography, political boundaries, receptors sites and soil dust areas (dark gray) of the MCMA.

Figure 3 shows similar vector and scalar fields as Figure 2 except shown at hour 19:00. The winds have a well-defined northeasterly component all over the physical domain and they are responsible for transporting measurable amounts of geological dust to the densely populated areas of the MCMA. This constitutes the second most important scenario of March 5 during this hour in which PM concentrations over the southern part of the MCMA attained a second maximum before they slowly decayed in the late evening hours.
Figure 2b: Mixed layer depth and topography for March 5 at 08:00 hr.

Figure 3a: Surface winds and dust concentrations for March 5 at 19:00 hr.
The diurnal evolution of predicted PM concentrations of geological origin for day 6 show (not shown) a similar spatial behavior as those found during the simulation of day 5 but with significantly lower values. The key feature to note here is that from day 5 to 6 a strong northerly basin-to-valley wind was driven by a cold and humid air mass as evidenced from a large temperature drop to the north of the MCMA in the evening of day 5. This cold air mass pushed its way through the basin and helped maintained the strong northerly wind blowing thorough most of day 6, thus resulting in a cleaner airshed and a high visibility index.

Figure 4 shows the scatter diagrams of predicted windblown dust concentrations (Cp) versus the observed ones (Co) for days 5 and 6. On day 5 the predicted concentrations show a much better correlation to the geological component of measured PM concentrations when compared to day 6. The significant agreement between predictions and observations for day 5 suggest that much of the suspended PM in the airshed came from soil sources located on the northeast sector of the MCMA. The poor agreement between predictions and observations for day 6 indicates that the PM concentrations measured at the receptor sites had a more significant contribution of local geological sources than agricultural soil sources.
Conclusions

Two contrasting scenarios were studied and simulated during days 5 and 6 of March 1997 with observational data from the IMADA campaign. On day 5 high wind blown dust concentrations were measured at selected sites within the MCMA. The simulation of the spatial and temporal evolution of PM concentrations showed reasonable agreement with the observed particle measurements of geological origin for day 5. Furthermore, on day 5 an important portion of the measured concentration was of geological origin while on day 6 this situation was reversed in the sense that sources other than geological contributed to the total PM concentrations. Until now there have not been any systematic long term studies on blowing dust events in the MCMA that have been undertaken at the local scale.

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Aeolian Processes in Old and New Industrialized Areas (A Case Study of Silesian Upland– Southern Poland)

J. Pelka-Gosciniak, University of Silesia, 41-200 Sosnowiec, Bedzinska 60, Poland (E-mail: pelka@us.edu.pl)

J. Wach, University of Silesia, 41-200 Sosnowiec, Bedzinska 60, Poland (E-mail: jwach@ultra.cto.us.edu.pl)

T. Szczypek, University of Silesia, 41-200 Sosnowiec, Bedzinska 60, Poland (E-mail: kgf@ultra.cto.us.edu.pl)

Upper Silesia is the most industrialized region in Poland with predominating black coal mining, iron metallurgy as well as zinc and lead ores mining and metallurgy.

The results of historical development in Silesian Upland are old and new industrialized areas, where on a rather large scale aeolian processes could occur. These processes are still going on (Szczypek, Wach, 1991).

The area of old industry is connected with the Middle Ages, when in eastern part of Silesian Upland silver and lead ores exploitation and metallurgy was started. The result of it was large demand for wood and charcoal to heat in the contemporary primitive works. Therefore, the process of deforestation in large sandy areas followed. Those times and later in the 16th century typical dune landscape was formed under the influence of westerly winds. Then so-called Bledow and Starczynow “Deserts” were originated, character of which remained almost till present times. These sandy areas originate from the Pleistocene and they are an effect of fluvioglacial and river waters accumulation. (Szczypek, Wach, 1991; Pelka-Goscinia, 2000).

During hundred years the type of aeolian relief here gradually changed. Initially it mostly had a deflative character. Wide deflation areas dominated here and large dunes developed only at the border with humid river valley running across Bledow Desert as well as at the foothill of the Upper Jurassic cuesta, which limits the desert to the east.

The detailed observations of relief changes were possible at least on the base of arial photo analysis. They indicate that from 1950 till now and especially from 1970, when human being repeatedly interfered into this desert in a form of introduction of bushes of willow *Salix acutifolia* and *Salix arenaria*, the aeolian relief type gradually changed into accumulative and systems of changing in time and space small transverse as well as longitudinal dune forms appeared. Now deflation fields and dunes are to a large degree fixed by vegetation and “desert” landscape gradually disappears.

New industrialized areas include central and eastern parts of the Silesian Upland. They originate from the turn of the 19th century as well as from the 20th century and they are connected with the black coal mining and with zinc and lead ores metallurgy. Aeolian processes consist here in an intensive deflation of material from dumping grounds being an effect of mining and metallurgy (photo 1). This material often contains substances which are harmful for human health, e.g. heavy metals (Dulias, Pelka-Goscinia, Radosz, 2002). It is transported even to distances of some km and reaches the neighboring towns. Apart from it in the area discussed
large sandpits appeared, material from which is used in mines as stowing sand. Therefore, on bare sandy substratum to a large scale aeolian processes occur, creating many different forms. Some years ago the development of typical anthropogenic scarp dune was here observed, which – in relation to human interference – was more or less intensively translocated accordingly to predominating westerly winds and it covered the border of pine forest (photo 2). Now, in result of processes of land reclamation this form was damaged (Pelka, 1994; Szczypek, Wach, 1999).

Photo 1. Deflation on dumping ground west of Olkusz (photo by T. Szczypek)

Photo 2. Anthropogenic scarp dune covering the pine forest (photo by T. Szczypek)

In both cases: old and new industrialized areas the development of aeolian processes was initiated by human impact on the neighboring environment (Szczypek, 1995; Pelka-Gosciniaiak, 2000). The human being created the proper conditions for aeolian processes to develop in a
natural way, giving these areas the “desert” character and causing the specific, new circulation of sandy material (Maszlej, Pelka-Gosciniak, 2001).

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The PM$_{10}$ and PM$_{2.5}$ Dust Generation Potential of Soils/Sediments in the Southern Aral Sea Basin, Uzbekistan

A. Singer, Hebrew University of Jerusalem, Rehovot 76100, Israel (singer@agri.huji.ac.il)

T. Zobeck, ARS, Lubbock, Texas 79401, U.S.A. (tzobeck@lbk.ars.usda.gov)

L. Poberezsky, Vodproject, Tashkent 70000, Uzbekistan (waterprj@globalnet.uz)

E. Argaman, Hebrew University of Jerusalem, Rehovot 76100, Israel (argaman@agri.huji.ac.il)

Introduction

Extensive desiccation of the Aral Sea in Central Asia has exposed large portions of the former sea bed. Enormous dust storms emanate from the area and have disastrous ecological consequences. Hundreds of thousands of tons of dust are deposited annually in areas to the S and SW of the sea (Fig. 1). The dust constitutes a major threat to the health of the population. Since the dust contains large amounts of salts, dust deposition causes severe salinization of both waterbodies and huge tracts of agricultural lands, some of the latter intensively cultivated (Rafikov, 1999). The area most afflicted by this catastrophe is the Southern Aral Sea Basin in Uzbekistan. The exposed surfaces include wetlands in the delta close to the Amu Darya River bed, with transitions to Solonchak soils, commonly with a salt crust. Takyr and Takyr-like soils exhibit a fine-grained crust and are somewhat more removed from the river bed. Shallow, stony soils occupy the elevated terrain. The desiccated and exposed Aral Sea bed includes a variety of sediments/soils, the most prominent of which are Solonchak-like soils.

The objective of this study was to assess the contribution of the major soil/sediment surfaces in the Southern Aral Sea Basin to the dust generation potential of this region.

Methods

Eight crusts and soils/sediments from 7 sites, representative of these surfaces, were sampled and their major characteristics (particle size distribution, organic carbon content, carbonate content, salt content and composition) that are related to dust generation, determined. The PM$_{10}$ and PM$_{2.5}$ dust generation potential of the materials was postulated as a general indicator of their dust generation capability, and was determined in the laboratory using the Lubbock Dust Generation, Analysis and Sampling System (LDGASS) (Fig. 2).
Results

The highest amount of PM$_{10}$ dust (579.3 mg m$^{-3}$) was generated from the Takyr crust material (Table 1). The lowest by one Solonchak salt crust material (39.6 mg m$^{-3}$). Salt crusts from the desiccated Aral Sea bottom generated intermediate amounts of dust. The distribution curve obtained for the Takyr crust PM$_{10}$ dust is distinctly unimodal, with the maximum at 3 µm
The distribution curve obtained for the Solonchak salt crust is bimodal, with the major peak at about 1.5 µm and a secondary one at 5 µm. Apparently, high potentials for dust generation are related to high proportions of very fine aggregates (<70 µm, obtained by dry sieving), characteristic for Takyr crusts. Salt crusts seem to generate much lower amounts of PM$_{10}$ dusts. This is due to the densely interlocking matrix of the salt crystallites forming the crust (Singer et al., 2001). However, under field conditions, blowing winds charged with saltating sand grains that exert abrasive forces on the crusts, break interparticle bonds and dislodge fine particles.

Table 1

<table>
<thead>
<tr>
<th>Site</th>
<th>Sediment source</th>
<th>Dust (PM$_{10}$) conc. mg.m$^{-3}$</th>
<th>Dust (PM$_{2.5}$) conc. mg.m$^{-3}$</th>
<th>Relative PM$_{2.5}$ conc. %</th>
</tr>
</thead>
<tbody>
<tr>
<td>I</td>
<td>Salt crust</td>
<td>39.6</td>
<td>19.1</td>
<td>48.2</td>
</tr>
<tr>
<td></td>
<td>1-15 cm</td>
<td>81.6</td>
<td>16.5</td>
<td>20.2</td>
</tr>
<tr>
<td>II</td>
<td>Takyr crust</td>
<td>593.3</td>
<td>261.1</td>
<td>45.1</td>
</tr>
<tr>
<td>III</td>
<td>Soil, 0-23 cm</td>
<td>379.5</td>
<td>135.0</td>
<td>34.0</td>
</tr>
<tr>
<td>IV</td>
<td>Solonchak (wet)</td>
<td>115.3</td>
<td>25.6</td>
<td>22.2</td>
</tr>
<tr>
<td>V</td>
<td>Solonchak (dry)</td>
<td>520.5</td>
<td>167.7</td>
<td>32.2</td>
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<tr>
<td>VI</td>
<td>Desiccated sea bottom crust</td>
<td>252.3</td>
<td>85.1</td>
<td>33.7</td>
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<tr>
<td>VII</td>
<td>Desiccated sea bottom crust</td>
<td>111.6</td>
<td>25.4</td>
<td>22.8</td>
</tr>
</tbody>
</table>
Conclusions

The experimental results indicate that the Takyr and Takyr-like soils, that occupy over 1 million ha in the Southern Aral Sea Basin, constitute the surfaces with the highest potential for being the source for the severe dust storms of the area. Second to the Takyr soils, the Solonchaks and Solonchak-like soils, also with an extent of over 1 million ha, contribute highly saline dust. To these must be added a large, as yet unsurveyed, proportion of the approximately 4 million ha of exposed sea bed, that exhibit Solonchak-like characteristics.

References


Sediment Deposition in an Attic Near a Region of Dust Provenance: Implications for Historic Regional Dust Dispersion and Deposition Patterns

R. Scott Van Pelt, USDA-ARS, Big Spring, Texas 79720 (svanpelt@lbk.ars.usda.gov)

Ted M. Zobeck, USDA-ARS, Lubbock, Texas 79415 (tzobeck@lbk.ars.usda.gov)

Thomas E. Gill, Texas Tech Univ., Lubbock, Texas 79409 (tgill@TTU.EDU)

Introduction

Fugitive dust is a frequent product of wind erosion and is often the most visible evidence of wind erosion downwind from actively eroding fields. While wind erosion may occur on any soil surface, it is most prevalent in semi-arid climates. Fugitive dust negatively impacts human activity and the environment and, while it has long been recognized that dust is a human health concern, only recently has the environmental cost associated with the dust-borne transport of plant nutrients (Zobeck and Fryrear, 1986), soluble and chelatable metallic salts (Prospero, 1999), and pesticides (Larney et al., 1999) been considered. These contaminants affect human health when they are transported over population centers (Prospero, 1999) and impact the nutrient loading of waters flowing from adjacent watersheds (Wood and Sanford, 1995).

Recent advances in weather data availability, satellite imagery, geographical information systems (GIS), and computer modeling have made it possible to predict, measure, and track plumes of dust emanating from selected source regions (Saxton et al., 2000). The particle size distribution of soil-derived dust most closely approximates a two parameter Weibull function (Zobeck et al., 1999). With increasing distance and time in transport, the larger particles with higher terminal velocities settle out of the airmass leaving progressively finer and finer particles in suspension.

A large body of data exists on the physical and chemical properties of soil-derived dust. Elemental and mineralogical analyses have been used to identify the source regions of dust deposited in Arctic ice caps and other depositional surfaces. More recently, biological fingerprinting has been used to more closely match mineral aerosols with the soil from which they were winnowed. While current technology is allowing characterization of atmospheric dust, little information is available concerning recent historic trends in dust transport and the physical, chemical, and microbiological characteristics of that dust.

Dust deposited in attics has recently been used to explore the history of aerosol contaminants in a given locality. Cidziel and Hodge (2000) collected dust from several attics in southern Nevada and Utah and analyzed the dust for trace elements and pesticides. Attics provide an ideal location for archived dust because they are rarely cleaned and the dust that settles in them is protected from rain and ultraviolet radiation. Attics typically have very small vents in relation to the total cross section and very small velocities of air movement with little turbulence. In such a system we hypothesized that the particles would settle in a very short distance compared to the regional pattern in the windy and turbulent conditions often associated with dust generation. Such a system should provide a physical scale model of the deposition...
basin for a dust source area. Our study was undertaken to investigate the properties of dust sampled in attic built in 1954 and to determine whether the chemical, physical, and microbiological characteristics of dust deposited at different locations in the same attic might be representative of regional patterns.

**Methods**

Big Spring, Texas is located at 32° 15' north latitude 101° 30' west longitude near the southern terminus of the Southern High Plains, a locally important source region for dust generation. Approximately one half of the surrounding land is utilized for dryland production of summer crops including cotton and grain sorghum. During the winter and spring, the fields are fallowed and susceptible to erosion by the often gusty winds that blow predominantly from the west and southwest from January through May.

Very near the center of Big Spring, a rectangular two story building was added to an existing church structure in 1954. The long axis of the building is oriented approximately into the direction of the winds that prevail during the erosion season. The attic is ventilated with a small louvered vent at the top of the gable approximately 8 m above the ground on the upwind side and by a louvered bell tower on the downwind side. Dust samples were collected from a measured area approximately 0.4 m downwind (toward the center of the attic) from the vent and at 1.1 m intervals up to the center of the building. A total of 13 samples varying from 5.9 to 81.9 g each were collected. The samples were placed in pre-weighed soil cans, labeled, dried in an oven at 60° C for 48 hours. The samples were then passed through a 60 mesh (250 μm) screen to remove building debris and macrobiological materials and weighed to the nearest 0.001 g.

Total dust deposition, expressed as kg m⁻², was calculated for each sample and particle size analysis was performed on 0.3 g dust samples diluted in a sodium hexametaphosphate solution and sonically dispersed on a Beckman-Coulter LS230 laser/PID particle size analyzer. Percentages of organic carbon and total nitrogen were determined with an Elementar C/N analyzer. Although results are not presently available, additional chemical analyses including trace metal contents and ²³⁹+²⁴⁰Pu content are currently underway.

Total dust deposition, particle size analysis, percent N, and percent organic C data were regressed as a function of distance from the upwind vent using a simple exponential decay function of the form:

\[ y = a \left( e^{x/b} \right) \]

where a is a scaling factor and b is a shape factor. The location 1.5 m from the vent was excluded from the fit for total deposition due to evidence of surface replacement after the date of construction.

**Results**

The total dust deposition, mean particle diameter, median particle diameter, modal particle diameter, 5th and 95th percentile particle diameters (d₅ and d₉₅, respectively), percent nitrogen (N), and percent organic carbon (C) are presented for each sample in Table 1 along with the fitted a and b parameters and coefficients of determination for Eq. 1. All measures were highly correlated with distance from the vent with the exception of percent N. The appearance of sand-sized particles in the locations within 3 m of the vent that is located 8 m above the ground at a
distance of 4 km from the nearest farm field is evidence of the height of large particle
entrainment possible in the turbulent, fast moving dust storms that frequent the Big Spring area.
The total dust deposition drops very rapidly from the point of entry at the attic vent. The
exponential decay curve for total deposition was much more eccentric than for the other physical
factors as evidenced by the relatively small value for the fitted shape parameter, b. This
phenomenon may be attributed to several factors including the rapid settling of the larger
particles upon entry to the relatively still attic. Particle momentum would drop rapidly upon
entry and fast moving large particles entering the attic space would rapidly slow to the attic
velocity and settle very close to the vent. Finally, days with little wind and atmospheric dust
would result in attic airmass velocities very near zero and particles settling close to the vent.

Table 1. Measures of physical and preliminary chemical attributes of the attic dust samples

<table>
<thead>
<tr>
<th>Distance (m)</th>
<th>Deposition kg m⁻²</th>
<th>Mean d µm</th>
<th>Median d µm</th>
<th>Modal d µm</th>
<th>d₅ µm</th>
<th>d₉₅ µm</th>
<th>Percent N</th>
<th>Percent C</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.4</td>
<td>1.0009</td>
<td>48.5</td>
<td>35.4</td>
<td>55.1</td>
<td>1.1</td>
<td>133.7</td>
<td>0.14</td>
<td>3.18</td>
</tr>
<tr>
<td>1.5</td>
<td>*0.5560</td>
<td>42.0</td>
<td>28.2</td>
<td>50.2</td>
<td>1.1</td>
<td>111.0</td>
<td>0.25</td>
<td>3.51</td>
</tr>
<tr>
<td>2.6</td>
<td>0.6082</td>
<td>39.1</td>
<td>28.1</td>
<td>50.2</td>
<td>1.0</td>
<td>101.1</td>
<td>0.15</td>
<td>3.43</td>
</tr>
<tr>
<td>3.7</td>
<td>0.4632</td>
<td>35.5</td>
<td>22.1</td>
<td>41.7</td>
<td>1.0</td>
<td>96.6</td>
<td>0.19</td>
<td>4.31</td>
</tr>
<tr>
<td>4.8</td>
<td>0.3499</td>
<td>32.7</td>
<td>19.1</td>
<td>38.0</td>
<td>1.0</td>
<td>89.4</td>
<td>0.18</td>
<td>4.42</td>
</tr>
<tr>
<td>5.9</td>
<td>0.2487</td>
<td>29.7</td>
<td>17.5</td>
<td>38.0</td>
<td>0.9</td>
<td>89.4</td>
<td>0.13</td>
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<td>7.0</td>
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<td>31.5</td>
<td>17.0</td>
<td>34.6</td>
<td>0.9</td>
<td>93.9</td>
<td>0.13</td>
<td>5.72</td>
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<tr>
<td>8.1</td>
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<td>24.4</td>
<td>15.0</td>
<td>25.3</td>
<td>0.9</td>
<td>76.4</td>
<td>0.11</td>
<td>5.92</td>
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<tr>
<td>9.2</td>
<td>0.1077</td>
<td>24.1</td>
<td>13.7</td>
<td>20.7</td>
<td>0.9</td>
<td>77.7</td>
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<tr>
<td>10.3</td>
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<td>23.2</td>
<td>12.7</td>
<td>19.8</td>
<td>0.8</td>
<td>76.4</td>
<td>0.12</td>
<td>6.87</td>
</tr>
<tr>
<td>11.4</td>
<td>0.0759</td>
<td>21.9</td>
<td>11.8</td>
<td>19.5</td>
<td>0.8</td>
<td>71.9</td>
<td>0.12</td>
<td>7.48</td>
</tr>
<tr>
<td>12.5</td>
<td>0.0738</td>
<td>20.4</td>
<td>11.0</td>
<td>18.0</td>
<td>0.8</td>
<td>69.6</td>
<td>0.14</td>
<td>7.52</td>
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<tr>
<td>13.6</td>
<td>0.0764</td>
<td>20.4</td>
<td>11.3</td>
<td>18.0</td>
<td>0.8</td>
<td>68.6</td>
<td>0.12</td>
<td>7.95</td>
</tr>
</tbody>
</table>

* Identified as an outlier and excluded from curve fit due to evidence of surface
replacement after 1954
size distribution with increasing distance from the vent. This phenomenon is consistent with patterns observed in regional dispersion and deposition.

The percent organic C increased with increasing distance from the vent. This phenomenon may be explained by the physical behavior of the two forms of organic C. The structured forms of organic C often have irregular shapes, lower densities, and lower terminal velocities than mineral particles of similar size. Humic forms of organic C are usually found as surface coatings and finer particles with higher surface area to volume ratios have higher percentages of humic C.

It is unclear why there was no observed correlation of N with distance from the vent except to note that soil N in surrounding soils is very low and the addition of nitrogenous foreign materials such as the proteins in small plant detritus, small insect body parts, and animal waste products not sieved from the samples may have eclipsed any differences arriving and sorting with the mineral portion of the samples. Analysis of bulk dust samples collected in an adjacent attic built in 1930 had lower N percentages than those collected in this 1954 addition. Local farmers greatly increased the use of nitrogen fertilizers during the 1950s and the increased surface soil N levels are apparently expressed in the percentage N of the fugitive dust transported from the fields.

Conclusions

The exponential decay of total deposition and particle sizes with increased distance and time of transport is consistent with patterns noted in regional dust dispersion and transport. It would appear that this and other attics may offer scale physical models of regional dust dispersion and deposition that will allow relatively undisturbed sample collection to determine the physical and chemical nature of dust that has been transported over and deposited in nearby population centers and watersheds in recent history.
References


High-resolution stable isotope geochemistry of aeolian dust from Owens Lake, California

Russell T. Winn, Department of Geosciences, Texas Tech University, Lubbock, TX 79409-1053: (email FeO4y@aol.com)

Thomas E. Gill, Department of Civil Engineering and Department of Geosciences, Texas Tech University, Lubbock, TX 79409-2101: (email tom.gill@ttu.edu)

Haraldur R. Karlsson, Department of Geosciences, Texas Tech University, Lubbock, TX 79409-1053: (email hrkar@ttacs.ttu.edu) (presently on sabbatical leave at the Nordic Volcanological Institute, Reykjavik, Iceland: email halli@norvol.hi.is)

Introduction

Geochemical characteristics such as trace element and isotopic compositions have been used to trace the transport of mineral dust to its source. An inherent assumption in such studies is that geochemical signatures are preserved during dust formation and aeolian transport. Here we test this premise by examining the stable isotopic ratios of carbonate in dusts from Owens (dry) Lake, California, a large desiccated playa subject to wind erosion of unconsolidated surface sediments. The dust emissions from Owens Lake have been the subject of many papers presented at previous International Conferences on Aeolian Research (e.g. Gillette et al., 1996; Niemeyer et al., 1999)

Materials and Methods

In the Lake Owens Dust Experiment (LODE) of 1993, aeolian sediments were captured in BSNE samplers (Fryrear, 1986) during their generation at 6 heights above the playa surface in seven sites along a one-kilometer linear transect parallel to prevailing winds during three dust events over a two-week period (Gillette et al., 1996), as the lakebed evolved from an evaporite-covered to a clastic-covered surface (Cahill et al., 1996). For four selected sites up to five heights (10, 20, 50, 60, 100 cm) were measured for isotopic composition. Approximately one hundred dust samples, free of organic C, were digested in phosphoric acid and the CO₂ produced from carbonates at 25.2°C was analyzed for C and O stable isotopes. Isotopic compositions were determined on a VG-SIRA 12 gas-source ratio mass spectrometer equipped with a micro-inlet system. Results are reported using the conventional notation using V-PDB as the international standard and are in ‰ units. The acid fractionation factor for calcite (1.01025) was assumed in calculating the ¹⁸O value of the carbonate. Precision of analysis is generally better than 0.1‰.

Results and Discussion

The overall variation in delta values among the samples is –0.5 to +4 and –5.5 to –2.5 ‰ for ¹³C and ¹⁸O, respectively. The relatively high ¹⁸O values are consistent with hydrologically closed system conditions (evaporation) for the carbonate generation.
Both isotope systems show variations with sampling height, collection site, and time. The first major dust storm of the season (3/11 - N wind) produced carbonates with the lowest $^{18}$O and $^{13}$C values and generally the most extreme decrease in these heavy isotope values with height at the four sites studied. The dust showed progressively higher $^{18}$O and $^{13}$C downwind with a constant shift of nearly 2‰ independent of height. The second storm (3/17 - S wind) showed a decrease in $^{18}$O and $^{13}$C with increasing height above the playa surface at each locality, though less pronounced than in the first event. There is less variation between sites (1‰) than the first storm but the order of enrichment among the sites is still the same. The third storm (3/23 - S wind) produced the smallest variation with height at each site and the smallest variation between sites (0.5‰). Overall, there is a linear relationship between C and O isotopes for the carbonates in Owens Lake dust (Figure 1), and their $^{18}$O and $^{13}$C values increased (i.e. the windborne sediments became more enriched in the heavier isotopes of both elements) (Figure 2) with each successive dust emission event after the initial erosion of the efflorescent crust (Cahill et al., 1996).

Figure 1. All stable isotope data points for Owens Lake carbonate samples analyzed in this study.

![Graph](image1.png)

Figure 2. Example site data shows isotopic decrease with height, increase with sequential events.

![Graph](image2.png)
There are several possible sources of the dust carbonate. To the north of the sampling transect on Owens (dry) Lake was located a large salt pan and to the south, sand dunes. These, however, do not appear to have contributed significant carbonate dust to the collectors during the storms nor varied along the sampling transect since the isotopic enrichment did not appear to depend on wind direction.

A more likely source of the carbonate is the lakebed sediment itself. During the flooding of pluvial Lake Owens, detrital carbonate was brought in and subsequently during evaporation authigenic carbonates have formed. The change in the isotopic composition of the carbonates in Owens Lake dust over successive emission events in both horizontal and vertical directions appears to reflect a change in the dust source materials exposed on the playa from authigenic (evaporitic) type to a more clastic type consistent with the physical changes to the wind-exposed playa surface observed during the sequence of dust storms by the LODE researchers. The isotopic composition of the dust could thus be considered a mixture between these sources, reflecting the quickly-changing chemistry of the lakebed sediments exposed to the wind.

It is also possible that wicking of the groundwater is redepositing a thin new layer of evaporite minerals between each dust event (Saint-Amand et al., 1986). These new minerals might differ slightly in their isotopic composition from pre-existing playa materials due to subtle variations in conditions of deposition, accounting for the slight shift between the storms. The dust collected during the first event represents a large set of efflorescent evaporites that accumulated during the previous winter. The particles from the subsequent dust storms may contain newly formed sets of evaporites deposited on the playa surface between the individual storms. The emplacement of new material between dust events could account for the increase in $^{18}$O and $^{13}$C values.

![Figure 3. Stable isotope data, all samples: saltation zone 10-20 cm, suspension zone 50-100 cm.](image)

There appears to be a distinct isotopic difference between saltation and suspension layers in these samples (Figure 3). Fryrear and Saleh (1993) described sedimentological differences between such distinct layers immediately above the soil surface during wind erosion, with the
“transition” below which saltation dominates and above which suspension dominates generally at a height of 30 to 50 cm above the soil surface. Gillette et al. (1997) demonstrated that such an effect was well-defined for this set of Owens Lake samples; it is also well-indicated in the stable isotope analyses shown in Figure 3. The saltation-dominated particles collected at 10 and 20 cm above the surface have a more homogeneous isotopic signature. This signature ranges from approximately +2 to +4 and −4.5 to −2.5‰ for $^{13}\text{C}$ and $^{18}\text{O}$ respectively for all three storms, or a 2‰ positive shift. The 20 cm samples are slightly more homogeneous than the 10 cm samples. The suspension-dominated particles from the 50, 60, and 100 cm heights are much more heterogeneous, including more negative values for both carbon and oxygen. The ranges are −0.5 to +4 and −5.5 to −2.5‰ for $^{13}\text{C}$ and $^{18}\text{O}$ respectively for all three storms.

Conclusions and Further Plans

Our study suggests that it may be possible to trace the origins of aeolian dust to its source using stable isotopes in carbonates as long as the individual sources differ by at least several ‰ from each other. There does appear to be an intrinsic isotopic difference between the suspension and saltation components of wind-eroded material. This research indicates that wind-eroded dust becomes more varied in isotopic composition with increase in height above the soil surface, representing mixing of aerosols from a larger source area. Saltating grains near the surface represent a much smaller source area and may not give an accurate sample of the overall lakebed geochemistry and sedimentology. Thus, care must be taken when comparing the geochemistry of aeolian dusts to that of potential source sediments. We will be using X-ray diffraction and trace-element analysis data next to investigate if the saltation and suspension layers also differ significantly in chemistry and mineralogy.

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References


Sources of wind-blown dust in Northern China: towards developing an Asian Dust Databank

Jie Xuan Program in Atmospheric & Oceanic Sciences, University of Colorado at Boulder, Boulder, CO 80309. (email: xuan@colorado.edu)

Irina N. Sokolik Program in Atmospheric & Oceanic Sciences, University of Colorado at Boulder, Boulder, CO 80309. (email: sokolik@lasp.colorado.edu)

Introduction

Wind-blown mineral dust causes diverse effects on human health, oceanic ecosystem, regional environment and climate (Gao et al., 2001; Sokolik et al., 2001). The magnitude of the dust impacts depends on the amount and physical and chemical properties of atmospheric dust, which are largely controlled by the dust source. Currently, the strength of major dust sources is poorly constrained mainly due to both the complexity of dust emission mechanisms and a lack of data.

Northern China is the second world’s largest source of atmospheric dust, but less studied than the Saharan and US sources. The main objective of our ongoing work is to set up a framework for comprehensive characterization of dust sources and atmospheric dust properties in Asia by combining data and modeling. In particular, we have been compiling the Asian Dust Databank which currently comprises the geographical, pedological and climatological data for Northern China (Xuan and Sokolik, 2001). The data were used to quantify the source strength in terms of emission rates, as well as to characterize dust sources in Northern China. We identified three broad types of dust sources: Type1. Deserts in dry-agricultural areas, Type2. Gobi-deserts and deserts located on the plateaus, and Type3. Deserts and gobi-deserts located in topographical lows. Types 1, 2 and 3 sources contribute to the annual mean PM$_{10}$ emission 1%, 36% and 63%, respectively. Although the maximum of dust emission occurs in spring, each source type has a distinct seasonal cycle. We suggest that extreme aridity and strong winds are the main factors controlling the dust emission in Northern China. Here we present a brief description of the Asian Dust Databank and three types of dust sources in Northern China.

Developing the Asian Dust Databank

The Asian Dust Databank has been designed to record diverse classes of the data related to dust production, transport, and effects in the atmosphere. The Databank is subdivided into three major data categories such as (1) Climate data, (2) Soil and dust emission data, and (3) Data for Case Studies: Meteorology and dust events of Spring 2001. Currently, we are compiling a new category consisting of geochemical data to better characterize the chemical and mineralogical properties of surface soils and wind-blown dust.

The first category consists of the climate data: monthly means for the 1961-1990 time period from 647 surface weather stations and 124 aerological stations in China. The data include all routine surface weather observations and aerological sounding data such as air pressure,
geopotential height, air temperature, dew-point deficit, relative humidity, wind direction, wind speed, precipitation, visibility, cloud, and snow cover, etc.

The second category consists of soil data (soil texture, composition, vegetation cover, land use, etc.) and dust emission rates of PM$_{10}$, PM$_{30}$ and PM$_{50}$, i.e., the dust particles smaller than 10, 30 and 50 micron in diameter, respectively. Dust emission rates were calculated on the climatic (30-year mean) scale at a spatial resolution of about 2.5° x 2.5° degrees (92 grid points across Northern China) (Xuan, 1999, Xuan et al., 2000, Xuan and Sokolik, 2001). The calculations were based on modified U.S. EPA formulas. To illustrate, Figure 1 shows the distribution of PM$_{10}$ emission rates in Northern China (Xuan and Sokolik, 2001). One can point out a strong east-west graduate in emission rates.

The third category of data is dedicated to case studies. Currently, we are focusing on the Spring of 2001 in the scope of the ACE-Asia field experiment. This category includes three subcategories: (a) surface data, (b) aerological sounding data, (c) visibility data and (d) satellite imagery.

The key feature of the Asian Dust Databank is that it has been specifically designed for dust emission and transport studies. The main difference from other existing databases (such as FAO and NSCD soil databases) is that it includes not only soil properties but also meteorological and satellite data. Thus, it adds an additional, vertical dimension to the traditional GIS-type databases. This is a critical issue since dust mobilization and transport and hence the dust effects in the atmosphere are controlled by both meteorological characteristics and land surface features.

**Identification and characterization of dust sources in Northern China**

It is recognized that dust mobilization is a complex process controlled by various surface properties and local meteorological conditions. Thus, a combination of these factors must be considered to quantify the strength of a dust source. To guide the characterization of dust sources, we introduced an integrated set of the following factors: the frequency of dust storm occurrence, wind speed, aridity and precipitation, morphology and composition of surface soil, and dust emission rates (Xuan and Sokolik, 2001). Combining these factors, we performed comprehensive, comparative characterization of dust sources in Northern China, identifying their commonalities and specific features that affect dust emission. We also addressed the effects of human activities and land use change on dust emission in this region.

We identified and characterized three broad types of dust sources in Northern China. Figure 2 shows that these sources spread from east to west, reflecting a rapid decrease in the precipitation and the increase in aridity and emission rates. The relative contribution of Types 1, 2, and 3 sources to the annual mean dust emission is about 1%, 36% and 63%, respectively.

**Type 1. Deserts in dry-agricultural areas.**

Type 1 sources consist of the Hulun Buir Desert, Horqin Desert, Hunshandake Desert, Mu Us Desert, and Hobq Desert. The name “dry-agricultural” means that farming strongly depends on irrigation because of the limited rain fall (200–400 mm/yr, with 75% in the summer season). This is an arid area in which farmlands (about 6.82x10$^4$ km$^2$), dry-grasslands (58.11x10$^4$ km$^2$) and sand deserts (13.76x10$^4$ km$^2$) are alternatively distributed. There are no Gobi-deserts in this region.
Because of the cold waves from Siberia, the wind speed is rather high in the winter and spring seasons. The 30-year climatic data show that the Type 1 sources lie in the high-frequency duststorm region. Our calculations predict annual mean emission of PM$_{10}$, PM$_{30}$ and PM$_{50}$ in Type 1 sources of 0.1, 0.2 and 0.5 million tons, respectively.

Type 2. Gobi-deserts and deserts located on plain plateaus.
This type of sources mainly consists of the Central Gobi-desert, Ulan Buh Desert, Tengger Desert, Badain Jaran Desert and the gobi-deserts in the Hexi Corridor. The Central Gobi-desert is located in the west part of the Inner-Mongolia Plateau, while the other three deserts are located on the Alxa Plateau. The existence of gobi-deserts suggests that it is an extremely dry area. Precipitation is about 50–200 mm/yr and the vegetation coverage is less than 10% on average.

Our calculation shows that annual mean emissions of PM$_{10}$, PM$_{30}$ and PM$_{50}$ in Type 2 sources are 2.9, 6.9 and 13.2 million tons, respectively. The large part of the Central Gobi-desert extends into the Republic of Mongolia. Because of various similarities between Chinese and Mongolian parts, the latter can be considered as the Type 2 source. Assuming the same emission rates for Mongolian and Chinese parts of the Central Gobi-desert, we estimated that about 4.7 million tons of PM$_{10}$ dust might be emitted annually into the atmosphere from the Mongolian part, while the Chinese part emits about 2.7 million tons.

Type 3. Deserts and gobi-deserts located in topographical lows.
This source type mainly comprises the Taklimakan Desert, Gurbantunggut Desert, Kumtag Desert, Hashun Gobi-desert, Turpan-Hami Basin (gobi-desert) and Tsaidam Basin (gobi-desert and playa). All these deserts and gobi-deserts are located in the basins or on flanks of high mountains. The Taklimakan Desert, being surrounded by high mountains and located far from the oceans, receives extremely low annual precipitation (0-20 mm/yr). Almost the entire Taklimakan is devoid of plant cover. As a result, the Taklimakan Desert is a prodigious dust source, being a main supplier of wind-blown dust in East Asia. Overall for Type 3 sources, our calculations give annual mean emissions of PM$_{10}$, PM$_{30}$ and PM$_{50}$ dust of 5.4, 17.6 and 28.9 million tons, respectively. The relative contributions of these sources to the total dust PM$_{10}$, PM$_{30}$ and PM$_{50}$ in Northern China is 63%, 71% and 68%, respectively.

Summary
We have been compiling the Asian Dust Databank which currently comprises the geographical, pedological and climatological data for Northern China. These data enabled us to perform comprehensive characterization of dust sources and to estimate their strengths on the climatic (30-year mean) scale. Our main findings are the following:

i). The Taklimakan Desert is a main source of atmospheric dust in Northern China. The next important sources are the Central Gobi-desert and the deserts on the Alxa Plateau - the Ulan Buh Desert, Tengger Desert and Badain Jaran Desert. The Loess Plateau appears to be a weak dust source.

ii). We identified three broad types of dust sources in Northern China. In addition to Types 1, 2, and 3, the Hexi Corridor and Tsaidam Basin are two unique dust sources we have identified. The former has characters of the all three types and is currently included in Type 2, and the latter is currently included in Type 3, having characteristics of alpine deserts. Relative
contributions of Types 1, 2, and 3 sources to the annual mean dust emission is about 1%, 36% and 63%, respectively.

iii). Analyses of both the spatial distribution and seasonal variation of dust emission rates demonstrate that extreme aridity and strong winds are the main factors controlling dust emission in Northern China.

iv). Human activities, mainly exhaustive farming, over-grazing and improper use of limited water resources in arid and semi-arid lands, have seriously damaged the natural environment in Northern China, causing the expansion of dust sources. In particular, Type 1 sources seem to be more vulnerable due to rapid population growth in this region.

**Acknowledgments.**
This work was supported by the National Science Foundation (grant ATM-0002746)

**References**


Figure 1. Annual mean PM$_{10}$ dust emission rates (t ha$^{-1}$ yr).

Figure 2. Three types of dust sources in Northern China. The dots show the main weather stations; H: Hexi Corridor; T: Tsaidam Basin.
Changes of soil environment and their effect on crop productivity in desertification processes of sandy cropland

Zhao Halin, Zhang Tonghui*, Zhao Xueyong Cui Jianyuan

Cold and Arid Regions Environment & Engineering Research Institute, Chinese Academy of Sciences, Lanzhou, 73000, P. R. of China
*e-mail: zhangth@ns.lzb.ac.cn

Introduction

Soil is a primary base for all terrestrial plants living on it. Soil environment, such as soil nutrient, soil moisture and soil temperature, plays an important role in the composition and abundant of plant species, and have in the productivity of land (Cheng Weixin. 1985. Liang Zhenxing, 1994.). For higher and stable output, there must be a health soil environment for crop growth. There have been many literatures of deep insights on the relationship between crop yields and soil moisture and nutrients, supplying scientific principles for improving soil environment and making a progress in agriculture (Xie Xianqun, 1992. Zhao Aifen, 1999. Wang Deshui. 1995.)

Rainfed Cropland is a larger portion of the total cropland in Horqin Sandy Land. And its productivity is very low, large area of the cropland has been degraded into desertified land due to dry and very windy spring, loose soil texture vulnerable to wind erosion and mismanagement (Cheng Weixin. 1985. Zhu Zhenda, 1994. Liu Xinmin, 1993.). Scientists have paid much attention on the causes, processes and damages of desertification (Zhu Zhenda, 1994. Liu Xinmin,1993. Li Yulin, 2000. Zhu Zhenda, 1980. Zhao Halin, 1998. Zhao Halin, 1999.) However, few papers on soil environmental evolvement and its influence on land productivity have been reported in this area.

Horqin Sandy Land located in the semi-arid area of agro-pasturage in eastern part of Inner Mongolia Autonomy Region. In comparison with Maowusu Sandy Land and Hunshandake Sandy Land, Horqin has a better water and thermal condition. Therefore, it has been a base for food and animal production in Inner Mongolia Autonomy Region since 1949 (Wang Deshui. 1995. Liu Xinmin, 1996.). Now, Horqin is one of the most severe desertified areas in northern part of China because of over-cultivation of grassland into cropland, over-grazing and over-collection of fuel-wood by local farmer. The aim of this paper is to present the changes of soil environment with the development of desertification, assess the effect of desertification on land productivity, and provide some suggestions on soil improvement of desertified croplands.

Materials and Methods

Study Site

The experiment site situated in Naimanqi (120°19’—121°35’ E, 42°14’—43°32’ N, a.s.l. 340-360 m), which locate in Horqin Sandy Land, Inner Mongolia, China. It is a semi-arid and continental climate. The annual mean temperature is 6.4°C. The annual accumulated air temperature that is more than and equal to 10°C is 3151.2°C. The frostless period is 130-167 days. The annual mean precipitation is 365 mm, and the annual mean evaporation is 1972 mm. The annual mean wind speed is 3.4 m/s, and the frequency of wind with more than 5 m/s, which is recognised as threshold of wind velocity leading to wind erosion, is 524 times per year. The typical soils are sandy meadow soil and aeolian sandy soil. The characteristics of those
soils are higher sand particle contents, low soil nutrient, and bad plasticity, fragile to erode by wind erosion. The study site is on aeolian sandy soil, whose thickness is from 30 to 40 cm depth, and under which is alluvial coarse sandy soil.

**Methods**

A transect along the gradient of desertification was set up in a large cropland with gentle topography and wind eroded. According to the degree of desertification, four type of cropland can be distinguished, non-desertification (ND), slight-desertification (LD), moderate-desertification (MD) and severe-desertification (SD), each including four replication and two permanent quadrates, one for determining soil moisture, the other for investigating of crop growth. The cultivar green bean was sown on 20th Apr. and harvested on 10th Sep. Nitrogen was applied as urea (46% N) at 75 kg/ha. Items measured were height, basal diameter, LAI, above-ground biomass by harvesting at adjacent plot, soil temperatures (at 0, 5, 10, 15, 20, 30cm depth, respectively) by Digital Thermometer (Chino-ND500, Japan), the heat value of biomass by oxygen bomb calorimeter (PARR-3430) and soil moisture (at 0-5, 6-10, 11-20, 21-30, 31-40, 51-60cm depth, respectively) by gravimetric method at intervals of every 10 days after 14th May. Statistical analyses of correlation and regression were carried out with the SPSS programme (SPSS 10, 1998).

**Results**

**Changes of soil particle size due to desertification**

The composition of soil particle size at four treatments shows in Table 1. Along with development of desertification, compared with ND, the sand content in SD is increased by 140%, but the clay content in SD is decreased by 82.3%. The hygroscopic water content was decreased by 82.1% because of high sand content in SD. The more severe desertification, the higher sand content of the cropland soil is for wind blowing the clay away. (Zhao Halin, 1986. Zhou Ruilian, 1995).

<table>
<thead>
<tr>
<th>Treatments</th>
<th>0.05-1.0 mm</th>
<th>0.005-0.05 mm</th>
<th>&lt;0.005 mm</th>
<th>Hygroscopic moisture</th>
<th>wish off</th>
</tr>
</thead>
<tbody>
<tr>
<td>ND</td>
<td>32.2</td>
<td>33.0</td>
<td>30.6</td>
<td>2.35</td>
<td>3.2</td>
</tr>
<tr>
<td>LD</td>
<td>52.0</td>
<td>38.9</td>
<td>7.9</td>
<td>0.74</td>
<td>1.2</td>
</tr>
<tr>
<td>MD</td>
<td>52.8</td>
<td>42.0</td>
<td>5.1</td>
<td>0.49</td>
<td>0.7</td>
</tr>
<tr>
<td>SD</td>
<td>77.3</td>
<td>26.5</td>
<td>4.8</td>
<td>0.42</td>
<td>0.4</td>
</tr>
</tbody>
</table>

**Changes of soil nutrient**

Along with the development of desertification, the soil nutrient content of cropland decreased dramatically as shown in table 2. Comparing to ND, the OM, total N, total P, total K, available N, available P, and available K decreased by 66.2%, 72.8%, 59.4%, 9.3%, 69.0%, 27.5%, and 56.3% in SD, respectively.

Soil nutrient environment is a basic element for crop growth and reproduction. Soil nutrient content directly affect crop yield. Therefore, desertification for deteriorating the soil environment popularly recognized as the leading factor resulted in low and unstable crop yield in this region.

It also was found that soil OM, total N, available N, available K were decreased consistently, however the total P and available P presented a fluctuating trend as shown in table 2.
Table 2: Changes of soil nutrient in desertification process

<table>
<thead>
<tr>
<th>Items</th>
<th>Organic matte (%)</th>
<th>Total N £ (%©)</th>
<th>Total P £ (%©)</th>
<th>Total K© (£%©)</th>
<th>Available N© mg/kg ©</th>
<th>Available P© mg/kg ©</th>
<th>Available K© mg/kg ©</th>
</tr>
</thead>
<tbody>
<tr>
<td>ND</td>
<td>1.48</td>
<td>0.103</td>
<td>0.032</td>
<td>2.70</td>
<td>42.0</td>
<td>4.0</td>
<td>167.0</td>
</tr>
<tr>
<td>LD</td>
<td>0.61</td>
<td>0.039</td>
<td>0.017</td>
<td>2.75</td>
<td>28.0</td>
<td>5.0</td>
<td>85.0</td>
</tr>
<tr>
<td>MD</td>
<td>0.59</td>
<td>0.039</td>
<td>0.025</td>
<td>2.56</td>
<td>18.0</td>
<td>3.0</td>
<td>85.0</td>
</tr>
<tr>
<td>SD</td>
<td>0.51</td>
<td>0.028</td>
<td>0.013</td>
<td>2.45</td>
<td>13.0</td>
<td>2.9</td>
<td>73.0</td>
</tr>
</tbody>
</table>

Changes of soil moisture

Changes of soil moisture in all treatments were shown in Fig. 1. The mean value of soil moisture in whole growth period in ND is significantly higher than that in LD, MD, and SD. Along with desertification, the soil moisture of cropland was diminished. As compared to ND, the soil moisture in LD, MD, SD, had was reduced to 46.4%© 54.3% and 74.8%, respectively. The soil moisture in ND was 12.4% at drought period, 21.4% at monsoon period, and still around 9.2% during the most drought phase. It was correspondingly 2.4%© 5.8%, and 2.3% in SD. Apparently, the water demand of green bean for growth can be met in ND. Nevertheless, it was difficult for crop to growth under such lower soil moisture in SD.

Changes of soil temperature

With coarse particle increase in soil, soil thermal efficiency and thermal exchanges became poorer in desertification cropland (Li Yulin, 2000). Meanwhile, the fluctuation of soil temperature at topsoil layers was larger than that at deep ones, and the soil temperature at tillage layer increased in SD. The average soil temperature at 5-30 cm depth was low in ND, and high in SD (Fig. 2). Soil temperature at ND, LD, MD and SD was 21.5°C, 22.0°C, 22.1°C and 22.4°C, respectively. Difference of soil temperatures between the maximum and minimum was 0.9°C. Cropland with high soil temperature in summer, had a disadvantage for biomass accumulation and soil moisture maintenance (Li Yulin, 2000.).

Formation of crop biomass

The processes of crop biomass forming at ND, LD, MD, and SD were shown in Fig. 3. In ND the rapid growth period of green bean was from 4 July to 25 Aug., but in SD from 15 July to 4 Aug. The daily rate of biomass accumulation in ND increased continuously and steadily. However, in desertified area, the rate increased irregularly. The more serious desertification occurred in cropland, the larger fluctuation of the daily rate of biomass accumulation can be found in Table. 3. All those indicated that desertification had brought out the changes of soil environment and influenced significantly on the rhythm of crop growth. The developing of desertification would result in remarkable decreasing of crop biomass. The biomass in LD, MD and SD had decreased by 15.3%©31.1%, 60.5%, compared to ND, respectively.
Fig. 2 Seasonal changes of soil temperature in 5-30 cm depth
a. non-desertification; b. slight-desertification; c. moderate-desertification;
d. severe-desertification

Fig. 3 Formation processes of biomass in farmlands

Table 3 Accumulating rates of biomass in farmlands (g/days)

<table>
<thead>
<tr>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>ND</td>
<td>0.8</td>
<td>0.8</td>
<td>2.2</td>
<td>3.8</td>
<td>4.0</td>
<td>5.0</td>
<td>6.1</td>
</tr>
<tr>
<td>LD</td>
<td>1.1</td>
<td>0.2</td>
<td>0.5</td>
<td>8.7</td>
<td>6.0</td>
<td>3.6</td>
<td>0.0</td>
</tr>
<tr>
<td>MD</td>
<td>0.2</td>
<td>0.7</td>
<td>0.3</td>
<td>7.4</td>
<td>6.0</td>
<td>2.0</td>
<td>0.0</td>
</tr>
<tr>
<td>SD</td>
<td>0.1</td>
<td>0.2</td>
<td>0.0</td>
<td>1.8</td>
<td>0.4</td>
<td>4.5</td>
<td>2.8</td>
</tr>
</tbody>
</table>

Comparison of photosynthetic efficiency in different treatments

The heat values of biomasses from different treatments in unit area were shown in Fig. 4. The heat value in ND is 3.585 MJ/m², larger than that in LD. The biomass of green bean in ND was 2.36 times that in SD. The photosynthetic efficiency were 0.37% & 0.23% & 0.22% & 0.16% in ND, LD, MD and SD, respectively. These indicated that the photosynthetic efficiency and the energy that input to cropland-system decreased with cropland desertification. Thus, desertification was harmful
Relationship between factors of soil environment and cropland productivity

As above, in the processes of desertification the cropland productivity decreased remarkably because of the continuous deterioration of soil environment. The results of correlation analysis showed the crop yield had a close relationship to available N, soil moisture and total K, and their relative coefficient were 0.9423 and 0.9096, respectively. However, the soil temperature had a negative effect on crop yield, and the relative coefficient was 0.9475 (Table 4). Analysis also revealed that soil OM had a significant relationship with the amount of soil particle size at 0.005mm. The soil OM and the amount of soil particle size at 0.005mm had a considerable effect on soil moisture. The soil temperature was increased due to a high content of coarse sand and low soil moisture at desertified cropland. Soil moisture, soil OM content and total nitrogen took a great effect on the content of soil available nitrogen. Those results showed that crop yield was decreased remarkably at desertified cropland due to soil coarse sand increase, soil moisture declining, soil temperature increase, and most of clay and soil OM blown out by wind erosion in the processes of desertification.

Table 4. Coefficient of correlation in the biomass with environment factors

<table>
<thead>
<tr>
<th>Items</th>
<th>corre.coef.</th>
<th>Items</th>
<th>corre.coef.</th>
<th>Items</th>
<th>corre.coef.</th>
</tr>
</thead>
<tbody>
<tr>
<td>water content</td>
<td>0.9202*</td>
<td>Soil temp.</td>
<td>-0.9475*</td>
<td>Organic matte</td>
<td>0.7972</td>
</tr>
<tr>
<td>Available N</td>
<td>0.9423*</td>
<td>Available P</td>
<td>0.6956</td>
<td>Available k</td>
<td>0.7854</td>
</tr>
<tr>
<td>Total N</td>
<td>0.7972</td>
<td>Total P</td>
<td>0.7622</td>
<td>Total K</td>
<td>0.9096*</td>
</tr>
</tbody>
</table>

*p£¼0.05, **p£¼0.01
Discuss and Conclusion

Most part of semi-arid and agro-pasturage region in North China is undergoing desertification. Rainfed system, as the larger part of cropland in this region is uncovered at most part of a year (from the middle of Sep. to early of May) because of a short growth period in this region. Therefore, the cropland is easy to erode under a strong and frequent wind condition. In the processes of cropland desertification, the clay and soil OM had been blown away and the coarse content of soil increased in the topsoil. In this study, by comparison with ND, the soil OM and clay were decreased by 84.3%, 66.2%, but the soil particle size at 0.05-1mm increased 2.33-fold in SD. The soil environment of soil moisture, soil temperature and nutrient had deteriorated significantly because of soil OM and clay loss by erosion. At first, the soil water capacity and soil moisture reduced. The average of soil moisture in ND was 15.1% during growth period, but 3.1% in SD. During a dry time, the former was 9.2% and the later was 2.3%. Soil temperature was increased and fertility decreased because of a low water capacity and a high coarse content in soil. By comparison with ND, the average of soil temperature in SD was higher 0.9°C during growth period, the maximum margin at 2.8°C. The OM, total N, total P, total K, available N, available P, and available K in SD were decreased by 66.2%, 72.8%, 59.4%, 9.3%, 69.0%, 27.5%, and 56.3%, compared to that in ND, respectively.

The spring is more windy and frequent in semi-arid and desertification region of North China on the desertified cropland. The local farmers usually sow crop such as millet, sorghum and green bean, which are drought-enduring crops with short growth period. These croplands were originally used as grassland for grazing. After conversion of grassland to farmland, for first several years the farmer used to plant these crops

The biomass of green bean in LD was decreased by 15.3% compared with that in ND, 31.9% in MD and 60.5% in SD, respectively. Though the causes for the incline of crop yield are complex much, the deficient of water and the lack of nitrogen have taken a vital role in the formation of crop biomass in this region.

The available nitrogen and soil moisture in SD, which was 13.0mgkg⁻² and 2.3 % accounted for 31% and 25% that in ND, have not met apparently the demand of crop survival. In a dry year, or temporal drought, the crop yield in SD decreased significantly and was sometimes nothing to harvest. The results of correlation analysis on crop yield and available N and soil moisture had shown this exchange too. With the formation processes of green bean biomass, the soil environment deterioration resulted in rapid curtail of growth period, photosynthetic efficient and daily rate of biomass accumulation declining. This conclusion is the same as that of related researchers in effect of soil environment deterioration on maize yield. The decline of crop yield has a negative correlation with soil temperature, and the correlation coefficient is -0.9907. Apparently, rising soil temperature is a vital reason for declining of agricultural production because of cropland desertification.

In the semi-arid and desertification region of North China, desertified area has spread at a rate of 2000 km² per year recently, in which 23% resulted from cropland desertification. There are two reasons leading to cropland desertification. One is not immediately to set up windbreak for cropland protection from soil erosion. The cropland, therefore, is easy to be deteriorated by desertification. As a result, the land productivity was reduced absolutely.

The other is that the fallow management system adopted by local farmer does not fit in with the climate and soil condition in this region. This management depends on the natural potential capacity of land for agricultural production []. The cropland was abandoned because of severe desertification occurring or land productivity declining. The dry and bare surface of abandoned cropland is easy to
be desertified in coming spring.

Therefore, strengthening the protection and management of rainfed land not only has significance for agricultural production, but also a vital role for desertification control in this region. Based on our research achievements of many years, we think it is a set of good comprehensive measures for agriculture production and soil erosion control in this region to strengthen construction of shelterbelt, to develop irrigation cropland and to increase use of organic manure in no-irrigated cropland. Therefore, we should extent intercrop and rotation systems of grass and food crop in rainfed cropland.

Reference

Natural sources of Aeolian dust in Amman, and selected Areas of Jordan.

Hadeel Al Dwaikat, PhD student; School of Science and Environment, Coventry University, UK

Ian Foster, School of Science and Environment, Coventry University, UK.

Nasfat Hunjul, Natural Resources Authority, Amman , Jordan.

Adrian Wood, School of Science and Environment, Coventry University, UK

Joan Lees, School of Science and Environment, Coventry University, UK

Abstract

The overall aim of this research programme is to characterise potential sources of atmospheric dusts deposited in the Amman (urban), Mafraq (rural) and Azraq (desert) areas of Jordan. These dusts derive from a number of sources, some of which may be remote from the point of deposition and will reflect long range transport during dust storms. The potential dust sources will be characterised by their mineral magnetic, radionuclide and geochemical signatures which will be quantitatively compared, using an un-mixing model, with the same signatures of deposited dusts in each area.

The aim of this paper is to identify the potential natural and anthropogenic sources of dust and the impact that human activity may have in increasing the availability of dust from natural sources for transport.

Geologically it is important to identify those rocks and minerals that could weather by natural processes to provide material fine enough to be entrained by the wind. Part of this study therefore focuses on an analysis of the geological materials capable of producing such fine sediments. The major natural process increasing air pollution in the research areas is the dust storm (sandstorm) which affects the desert areas more than the highlands. However, in north-eastern Jordan over the last decade, changes in agricultural practices, especially the clearance of surface stone layers in the basalt regions for cultivation, may have increased the availability of dust in these regions for aeolian transport. Soil erosion in the hilly areas causes landslides while tree cutting and overgrazing by large numbers of livestock may also increase sediment availability.

Small localised anthropogenic sources of contaminated dust may also be released in rural and desert areas but these pollutants will be rapidly dispersed and diluted by wind, washout by rain or through deposition with large quantities of natural dusts.

Direct contributions to air pollution primarily affects urban areas where the density of building, industry and vehicles prevents pollutants from being dispersed. Urban air pollutants include particulate matter, heavy metals and acidifying gases such as $\text{SO}_4$, $\text{NO}_x$. The city of greater Amman is suffering from serious air pollution problems from nearby industries.

Other activities, especially quarrying, mining and building construction sites in and around Amman may also make additional natural geological materials available for transport. These include limestone extraction for building construction and cement production in the south of the city. Another major source of dust to the west of Amman derives from the Al Fuhays cement factory because most of the prevailing winds are from the west. The most important source of dust in the north is from the sandstone quarries of the Safout area. The eastern areas of Amman and Ruseifa have been greatly affected by emissions from major industrial plants, phosphate mining and phosphate processing. These phosphate deposits, and processed phosphates, are characterised by elevated levels of U-235.
Migration of parabolic dunes at Aberffraw on the island of Anglesey, north Wales

Bailey, S.D., School of Earth Sciences, Birkbeck University of London, Malet Street, London WC1E 7HX bsimon@blueyonder.co.uk

Bristow, C.S., School of Earth Sciences, Birkbeck University of London, Malet Street, London WC1E 7HX c.bristow@ucl.ac.uk

Aberffraw is a 1km wide and 3km long transgressive dune field that extends inland along a northeast – southwest trending valley from a southwest facing beach, Traeth Mawr. The prevailing wind is from the southwest and both the parabolic dunes and the valley within which they lie are parallel to the prevailing wind.

The dune field at Aberffraw includes two foredune ridges and three rows of active compound parabolic dunes. At the landward end is a lake, Llyn Coron, which has been formed by dunes migrating up the valley and damming the river, Afon Ffraw. Between the parabolic dunes are gently sloping interdune areas with a close cropped vegetation. The parabolic dunes at Aberffraw have been migrating inland across the interdunes areas. Rates of parabolic dune migration are derived from three sets of aerial photographs taken in 1940, 1982 and 1993. The aerial photographs have been scanned and manipulated in ArcView GIS software. Registration of the aerial photographs to an Ordnance Survey map was performed using ground control points (GCPs), common fixed features which are identifiable on both the aerial photographs and the baseline map (e.g. road intersections, road corners, buildings). Attempts to correct for the inherent distortions of aerial photography (sensor and panoramic distortion) are made during registration. Standardising the projection of the photographs to a common baseline allows meaningful spatial analysis, and the dune ridges, trailing edges, and areas of bare sand were mapped from each photograph as a series of overlays.

Rates of dune ridge migration are calculated from the spatial distance between linear trend lines applied to sections of dune ridges for 1940 and 1993. Trend lines were only fitted to sections where continuity of dune form was maintained over the given time period. Rates of parabolic dune migration range from a minimum 0 m/yr to a maximum of 2.63 m/yr, with an average migration rate of 1.1 m/yr.

At the same time, the foredune ridge has been accreting and prograded 60 m onto the beach. The rate of foreslope accretion is approximately 1.09 m/yr. The volume of sand accumulated in the foredune ridge between 1940 and 1993 is estimated at 475,000m³, a sediment accumulation rate of approximately 20,000 Tonnes/yr.
Linear dune development along flood-eroded desert margins in central Australia.

Mary C. Bourke, School of Geography, University of Oxford, Mansfield Road, Oxford. (Mary.bourke@geog.ox.ac.uk)

Introduction

During the late Pleistocene and Holocene a series of large floods eroded kilometre-wide swaths of linear dunes (15 m high) along the Simpson Desert margin. The floods emplaced coarse sand and gravel braid channels and backwater clay pans in dune swales. Since abandonment in the late Pleistocene, dunes have reformed across the surface of the paleoflood termini. This paper will describe dune formation along flood eroded desert margins in central Australia.

Methods

Preliminary observations from satellite images (TM, 30 m/pixel) were verified and mapped in the field. Two transects were surveyed across the Pleistocene (847 m long) and reformed dunes (560 m long). Samples were taken for Optically Stimulated Luminescence, SEM and textural analysis.

Results

The reformed dunes extend directly from the flood-truncated end of the Pleistocene dunes and their swales. They are also linear but are smaller and their spacing closer (see table 1). Preliminary luminescence age estimates indicate a significant lag between the timing of sediment supply and the initiation of dunes.

<table>
<thead>
<tr>
<th>Average</th>
<th>Pleistocene Dunes</th>
<th>Reformed Dunes</th>
</tr>
</thead>
<tbody>
<tr>
<td>Dune height (m)</td>
<td>15</td>
<td>7.4</td>
</tr>
<tr>
<td>Dune width (m)</td>
<td>74.95</td>
<td>89.5</td>
</tr>
<tr>
<td>Dune spacing (m)</td>
<td>315</td>
<td>138.7</td>
</tr>
<tr>
<td>Swale width (m)</td>
<td>77</td>
<td>49.65</td>
</tr>
<tr>
<td>Dune area (m²)</td>
<td>2505.2</td>
<td>1414.52</td>
</tr>
</tbody>
</table>

Table 1: Dimensions of linear dunes on flood-eroded desert margins in central Australia
Optical dating chronologies of dune reactivation in the south-eastern Arabian Peninsula

Helen Bray and Stephen Stokes, Oxford Luminescence Research Group, School of Geography and the Environment, University of Oxford, Mansfield Road, Oxford OX1 3TB, UK

Abstract

The Liwa Oasis of the United Arab Emirates contains some of the largest and most aerially extensive mega-barchanoid sand dunes on a global scale. They extend southwards into Saudi Arabia, forming the upwind margin of the Rub Al Khali sand sea. Locally the dunes reach up to 150 m in height, and are separated by interdune sabkhas (width up to 2-3 km). Past research in this area and the wider Rub Al Khali has highlighted evidence for precession-associated wet periods in the preserved record at 6-10 and 25-35 ka BP. These humid episodes have generally been indicated using radiocarbon dating techniques on lacustrine deposits, speleothems and palaeogroundwater surfaces. Evidence for dry periods in this area which would presumably be associated with dune mobilisation are not well known.

Here we present optical dating results on samples of aeolian sediment from long (c. >200m) drill cores extracted from the largest dune filed of the Liwa area which provide a detailed sequence of both the wet and dry phase deposits. These are used to consolidate understanding of Late Quaternary environmental change in the region. Optical dating of the Liwa Oasis core sediments using the single aliquot regeneration (SAR) protocol (Murray and Wintle, 2000), amongst other techniques, has outlined a number of phases of rapid dune deposition. The most recent period of reactivation began at around 3 ka BP, after a hiatus in deposition since 5 ka BP. These preliminary results suggest that, during the the initial periods of dry phases of climate cycles, large bodies of sand are transported and deposited. Following the initial phase of rapid aeolian accumulation, the system appears to remain in stasis. Further dating of the sediment cores will be presented to establish a tighter chronological control on this, and other dry phases.
Origin of Iowa’s Sand Prairies

Steven H. Emerman, Department of Biology and Environmental Science, Simpson College, Indianola, Iowa 50125 (emerman@storm.simpson.edu)

Brian R. Depew, Department of Biology and Environmental Science, Simpson College, Indianola, Iowa 50125 (depew@storm.simpson.edu)

Lisa K. Anderson, Department of Biology and Environmental Science, Simpson College, Indianola, Iowa 50125 (andersol@storm.simpson.edu)

Introduction

The sand prairies of Iowa are generally regarded as eolian sand blown from adjacent river valleys about 4000 years B.P. (Ruhe, 1969; Prior, 1991; Fleckenstein, 1992). The eolian origin is based primarily on the proximity of the sand prairies to river valleys and the timing is based solely on evidence for a Hypsithermal period ending about 4000 years B.P. (Prior, 1991). In fact, the sand prairies of Iowa have been little studied geologically aside from studies of wetlands occurring in the sand prairies (Knapp, 1983; Thompson et al., 1992). Our initial objective was to determine whether the sand prairies were actually eolian deposits. The leading alternative is that the sand prairies are remnants of sand bars that were abandoned by the widespread stream downcutting, which occurred during the Holocene (Prior, 1991).

A classic paper by Visher (1969) showed that conclusions can be drawn regarding the depositional environment of a sand deposit from the grain size distribution. Visher (1969) plotted the cumulative percentage of sand on a probability scale as a function of phi $\phi$, where $\phi = -\log_2$ (grain diameter in mm). The advantage of a probability scale is that normally distributed data fall on a straight line. Visher (1969) showed that log-probability plots of grain size distributions fit well to three connected straight lines. Visher (1969) interpreted the straight line for the coarse grains as the sand population transported by traction. The straight line for the fine grains corresponds to the sand population transported by suspension. The straight line for the intermediate grain sizes corresponds to the sand population transported by saltation. Visher (1969) examined thousands of sand samples from a wide variety of depositional environments and found that the depositional environment could be related to (1) the grain size that marks the transition between traction and saltation, (2) the grain size that marks the transition between saltation and suspension, (3) the fractions of the sand sample in the traction, saltation and suspension populations, and (4) the sorting in the traction, saltation and suspension populations. Since the publication of Visher’s (1969) paper, further studies have applied his methodology to determine the depositional environments of unconsolidated sands and sandstones (e.g., Glaister and Nelson, 1974). Middleton (1976) and Sagoe and Visher (1977) have investigated the theoretical basis for the transition grain sizes.

Methods

This study focused on the five sand prairies that are held in Iowa state preserves, which are Behrens Ponds and Woodland (Linn County), Cedar Hills Sand Prairie (Black Hawk County), Kish-Ke-Kosh Prairie (Jasper County), Marietta Sand Prairie (Marshall County) and Rock Island (Linn County). Nine sand samples were collected from each of the five sand prairie state preserves. For each sample, about 2 kg of soil was collected between depths 10 – 20 cm. Care was taken to minimize the visual and environmental impact on state land by removing the intact vegetation with its upper roots, collecting the soil from beneath the root zone, and then replacing the vegetation. In order to deflocculate the clay-sized particles, the samples were blended with Calgon (60 g/L) at a ratio of 1 mL Calgon to 4 g of field-moist soil. The clay and silt were then removed by washing the samples through
a 0.063 mm sieve. After drying the sand at 90 °C for four hours, the sand samples were passed through 18 sieve sizes using an electric sieve shaker.

**Results and Discussion**

Fig. 1 shows a typical grain size distribution for the sand prairie at Behrens Ponds and Woodland State Preserve. Table 1 summarizes preliminary results for the sand prairies in comparison to Visher’s (1969) results for eolian and fluvial sands. The sand prairies clearly do not have a fluvial origin. On the other hand, with the exception of Kish-Ke-Kosh Prairie, the parameters do not fall within the ranges predicted by Visher (1969) for eolian sands. For the sand prairies, the grain sizes that mark the traction / saltation and saltation / suspension transitions are shifted toward coarser grains (smaller ø). In other words, a sand grain so coarse that it could only roll in the formation of a typical eolian deposit could bounce in the formation of the Iowa sand prairies. Moreover, a sand grain so coarse that it could only bounce in the formation of a typical eolian deposit could be carried in suspension in the formation of the Iowa sand prairies. In addition, the fraction of sand in the suspension population is considerably higher than is found in typical eolian sand deposits.

The differences between the Iowa sand prairies and typical eolian deposits can be explained by assuming that the sand that blew from river valleys to form the sand prairies traveled through grass prior to deposition. Grass presents a very rough and heterogeneous surface to the wind and the effect is to increase the turbulence and gustiness of the wind (Thom, 1971; Jackson, 1981; Raupach, 1991; Livingstone and Warren, 1996). Therefore, the grain size for transition between saltation and suspension will be shifted toward coarser grains and the percentage of grains in the suspension population will increase. Moreover, traction will be impeded by the clumps of grass. Coarse grains will accumulate on the windward side of grass clumps until the impact of an unusually large number of descending saltating grains causes them to be forced over the grass clump. The effect will be to force grains that would normally roll due to their size into a manner of “pseudo-saltation,” thus shifting the transition from traction to saltation toward coarser grains.
Figure 1. Typical sand size distribution for Behrens Ponds and Woodland. The transition from traction to saltation occurs at grain size $\phi = 0.74$ (0.60 mm). The transition from saltation to suspension occurs at grain size $\phi = 1.73$ (0.30 mm).

Table 1. Comparison of sand prairies of Iowa with Visher’s (1969) results for eolian and fluvial sands.

<table>
<thead>
<tr>
<th></th>
<th>Transition Traction / Saltation ($\phi$)</th>
<th>Transition Saltation / Suspension ($\phi$)</th>
<th>Suspension Population (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Behrens Ponds and Woodland</td>
<td>0.69 ± 0.02</td>
<td>2.3 ± 0.2</td>
<td>9 ± 4</td>
</tr>
<tr>
<td>Cedar Hills Sand Prairie</td>
<td>1.0 ± 0.1</td>
<td>2.3 ± 0.1</td>
<td>24 ± 5</td>
</tr>
<tr>
<td>Kish-Ke-Kosh Prairie</td>
<td>1.3 ± 0.1</td>
<td>3.75 ± 0.00</td>
<td>0 ± 0</td>
</tr>
<tr>
<td>Marietta Sand Prairie</td>
<td>0.82 ± 0.04</td>
<td>2.71 ± 0.03</td>
<td>5.4 ± 0.4</td>
</tr>
<tr>
<td>Rock Island</td>
<td>0.58 ± 0.08</td>
<td>2.39 ± 0.07</td>
<td>10 ± 2</td>
</tr>
<tr>
<td>Eolian Sands (Visher, 1969)</td>
<td>1.0 – 2.0</td>
<td>3.0 – 4.0</td>
<td>1 – 3</td>
</tr>
<tr>
<td>Fluvial Sands (Visher, 1969)</td>
<td>-1.5 - -1.0</td>
<td>2.75 – 3.50</td>
<td>2 - 35</td>
</tr>
</tbody>
</table>

References


Aeolian influences on the soils and landforms in semi-arid south-western Australia

R.J. Harper, Univ. West. Aust., Nedlands W.A. 6907 (E-mail: richardh@calm.wa.gov.au)

R.J. Gilkes, Univ. West. Aust, Nedlands W.A. 6907 (E-mail: bob.gilkes@uwa.edu.au)

Introduction

Aeolian influences on the soils and geomorphology of semi-arid south-western Australia have been previously reported in several studies. Within a landscape developed on deeply weathered, granitic rocks there are several overt aeolian features, these including clayey and sandy saltation deposits (“lunettes”) adjacent to playas (Stephens and Crocker 1946), gypseous and clayey deposits (“lake parna”) downwind of playas (Bettenay 1962), the wind-induced shaping of playas (Killigrew and Gilkes 1974) and source bordering sand dune systems adjacent to ephemeral stream lines (Beard 1982). Severe, recurrent wind erosion is a contemporary hazard for sustainable land use (Harper et al. 2002).

Developing an understanding the processes that have distributed materials across landscapes is important as it provides a conceptual framework for more efficient soil survey and mineral exploration. This in turn may provide a basis for better land management. Similarly, aeolian features provide a means of interpreting the past responses of these landscape to climate change and changes in hydrology induced by agricultural development. In this paper we describe the aeolian features that occur in a study area on the Yilgarn Craton, near Cairlocup, Western Australia, with soils and landscapes typical of much of the region.

Regional setting: the south-western Australian physical environment

South-western Australia is dominated by the Yilgarn Craton, one of the former nuclei of Gondwanaland. This extends 900 km in a north-south direction, and has an east-west width of 700 km. It is comprised mainly of granites and gneisses with the widespread intrusion of dolerite dykes (Johnstone et al. 1973).

The drainage pattern of the Yilgarn Craton comprises inland areas with broad, flat floored valleys with sluggish drainage, whereas well defined drainage lines occur on the western and southern periphery. Progressive changes in valley form occur along specific drainage lines. Going from inland towards the coast these include greater relative relief, a change from flat to incised valley floors and steeper drainage gradients. Whereas the inland drainage lines can be up to 15 km wide and be in landscapes with up to 60 m relief, closer to the coast drainage lines are narrow and relief ranges up to 300 m (Mulcahy 1967).

Deep weathering has been extensive across the region, with resultant profiles often extending to depths of 50 m or more. These weathering profiles occur on diverse rocks and are often termed laterites (Gilkes et al. 1973), with a distinctive sequence of horizons. These include a sandy surface horizon underlain successively by ferricrete, weathered clays, saprolite and parent rock. The differential stripping of these profiles in relation to drainage lines provides an array of soil parent materials. In the upper catchments of the palaeo-river systems there is little relief, and the limited stripping that does occur is to local rather than regional base levels. Products of weathering, such as sand and solutes,
tend to be retained in these landscapes. It is in such an environment that the present study was based.

**Study area and methods**

Details of the Cairlocup study area, which is 400 km south-east of Perth the capital of Western Australia, and the methods employed are described in Harper et al. (2002).

**Unequivocal aeolian features**

*Playas*: Several playas occur in the valley floor, those containing water being elliptical in shape, as with others in the region. This elliptical shape has been interpreted as being due to the action of wind driven waves and currents (Killigrew and Gilkes 1974). Drilling in the valley floor indicated the occurrence of sediments to at least 30 m depth. Several playas were vegetated at the time of land development but have subsequently become salinized.

*Lunettes and lunette arrays*: Clayey aeolian saltation deposits occur either as single members or multiple arrays on the south-eastern shore of both contemporary and relict playas. These invariably have crescentic shapes, these forming in response to the elliptical shape of the playa shoreline. In some instances lunettes have shapes indicative of playas larger than those currently present. Similarly, the valley floor is dominated by a seven member, 5 km long, multiple lunette array, the lunettes ranging in height from 1-14 m (Fig. 1). Two and three member lunette arrays have been previously reported from in Texas (Reeves 1965) and South Australia (Campbell 1968).

![Figure 1](image-url)  
**Fig. 1.** NW-SW transect surveyed from the major playa, Lake Cairlocup. The distant lunettes are very subtle with amplitudes of 1- 2 m.

There are systematic changes in several soil attributes with distance along the Lake Cairlocup lunette array. Not only are the more distant lunettes relatively subtle in morphology (~2 m high), but there are also systematic differences in soil profile composition and morphology along the lunette array. Whereas profiles close to Lake Cairlocup are alkaline throughout, have loamy surface horizons and contain relatively large amounts (>25%) of diffuse carbonates, distant profiles have sandy surface horizons, contain carbonates both as nodules and soft segregations in discrete horizons, and have acidic upper horizons.

*Sand dunes and sheets*: Sand dunes and sheets occur both in the valley floor and surrounding hills within a distinct 10 km long and 2 km wide strip, downwind of the ephemeral Cairlocup Creek. These are most likely saltation deposits associated with a former source-bordering sand dune system.
Contemporary wind erosion: Severe and recurrent, contemporary wind erosion represents a hazard to sustainable land use. This is described in more detail in Harper et al. (2002).

Inferred or subtle aeolian features

A more widespread, but subtle, aeolian influence on the soils is likely.

Dust deposits: The occurrence of dust deposits can be inferred from the presence of the clayey lunettes. These are aeolian saltation deposits and their formation would have been accompanied by the evolution of aeolian suspension loads. Although a discrete silty horizon does not occur, field evidence for dust accessions includes the occurrence of calcareous and illitic materials in soils south-east of the major playa, in an apparent plume. Soils outside this plume are mostly acidic and kaolinitic.

Asymmetrical slope deposits: Deep sandy soils occur on many south-easterly slopes and may result from the interaction of topography with sand transport. Such soils occur both in the lateritised granitic terrain, and valley floor lunettes. The occurrence of these may be due to the preferential aeolian deposition of sands on the lee-side of ridges, due to the interference of topography on wind flow and hence sand transport.

Discussion

Lunettes as palaeo-hydrologic indicators

The onset of dryland salinity is a major problem in south-western Australia with 3 Mha of land considered at risk. This salinity has been caused by the replacement of deep rooted xerophytic vegetation with shallow rooted annual plants and the consequent rise of water tables.

Clayey lunettes are playa shoreline deposits and the occurrence of extensive lunette arrays can be used to interpret previous landscape responses to fluctuations in hydrology. Playas have responded to changes in regional hydrology by expanding during pluvial periods and contracting with arid phases. The formation of clayey lunettes requires drying of the playa bed, and sandy lunettes may result from sand transport by currents to the shores of water-filled playas, hence lunette composition may be related to regional hydrological, and hence climatic conditions (Bowler 1983).

The multiple lunette arrays are stranded shore-line deposits, deposited by playas migrating, or receding, to the north-west. They have not been previously described in the broad valley floors of south-western Australia. Many of the lunettes have shapes indicating formation adjacent to playas much larger than presently occur. Together, this indicates a wetter landscape in the past, with more extensive salinization of the valley floors. Many of these areas had a cover of natural vegetation at the time of land development, with this indicating, in a general sense, that salinization is reversible.

Aeolian influences on soil patterns

A strong aeolian influence is evident across the Cairlocup landscape, with features such as elliptical playa shapes, the orientation of the lunettes with respect to source playas and parabolic blowouts in clayey lunettes suggesting that the geomorphologically most effective winds have been from the north-west. Moreover, the lunette arrays are chronosequences, and their consistent orientation suggests uniformity of winds over a long period.

These features are relatively simply recognised. It is clear however, that there have been more extensive aeolian influences on this landscape, these including the extensive lunette arrays with very subtle lunettes, source bordering sand deposits that occur up to 10
km from source, topographically controlled saltation deposits and a thin veneer of aeolian dusts. The occurrence of these features have implications for soil surveys in such environments, particularly where these often rely on topographic models of soil distribution and mineral exploration where bedrock properties are often inferred from analysis of surficial materials.

References


Trace elements deposited with dusts in Southwestern U.S. -enrichments, fluxes, comparison with records from elsewhere

Todd K. Hinkley, U.S. Geol. Surv. Federal Center, ms 980, Denver CO 80225 (thinkley@usgs.gov)
Paul J. Lamothe, U.S. Geological Survey, Denver (plamothe@usgs.gov)
Gregory P. Meeker, U.S. Geological Survey Denver (gmeeker@usgsprobe.cr.usgs.gov)
Xiao Jiang, Townsend Management Group, Denver CO (xiaoj@usgs.gov)
Mark E. Miller, Grand Staircase-Escalante National Monument UT
Robert Fulton, Calif. State Univ., Desert Studies Center, Baker CA

Introduction

Modern dusts in the Southwestern U.S. commonly contain, or are accompanied by, larger amounts of ordinarily-rare trace elements than can be explained by the compositions of the common minerals that constitute the dusts, or by the average composition of the earth's crust. Records of deposition of dusts and trace elements from other places and other times give a similar picture (pre-industrial dusts preserved in Antarctic ice, and central European peat bog records). The degrees of enrichment (“enrichment factors”) of trace elements in the ancient dusts from those other places, and fine-grained modern dusts from the SW U.S., are similar. In addition to knowing the degrees of enrichment, it is necessary to calculate the amounts of excess trace elements deposited per unit area over time (mass fluxes) as they accompany the dusts, to allow evaluation of the consistency of “source strengths” of trace element supply, through time and across regions. This comparison can be made between regions of the earth, and between pre-industrial and modern times. It is then possible to compare the calculated source strengths with known sources of trace elements to the atmosphere, such as volcanic emissions.

The ordinarily-rare trace elements Pb, Cd, Cu, Se, and others are among those that are present in excess amounts in modern dusts deposited from the atmosphere, and modern atmospheric load material, relative to the rocks and soils that are the sources of the bulk of the dusts (Bowen, 1979). It has been a question whether this enrichment is due to natural processes, or industrial processes, or some combination (Duce et al., 1975; Weiss et al., 1978; Heidam, 1985; Mart, 1983; ref’s therein). Analyses of modern dusts we collected in the Southwestern U.S. over several years confirm that many trace elements (Zn, Cu, Pb, Cd, As, Se, Sb, Bi) are much more abundant in at least finer-grained dusts than in the average crust of the earth.

Besides the information from the Southwestern U.S., there are two other studies that contain information on amounts of dust and their accompanying trace elements, and that present or allow extraction of information about the flux rates: a study of dusts in Antarctic ice representing pre-industrial atmospheric deposition (Matsumoto and Hinkley, 2001), and a study of long-term deposition in European peat bogs (Shotyk et al., 2002). In addition, there is a new estimate of the source strength of trace elements from worldwide volcano emissions (Hinkley et al, 1999), one of the natural sources of trace elements to the atmospheric load.
Methods

To obtain winter-season, high-altitude dust, snow pack strata were collected each early Spring from 1997 to 2002 in the Southwestern U.S., under clean conditions. For year-around, low-elevation dust, dry-deposition samples were collected on greased glass plates. Snow samples were reduced in volume in the laboratory under flowing filtered nitrogen, and both kinds of samples were digested and analyzed for a major, minor and trace elements by ICP-MS. Masses of dusts in samples were estimated by summing the masses, as oxides, of elements measured in bulk analyses. Calculations of “enrichment factors” of elements were done by comparing the concentration of an element in dust in a sample to the concentration in average crustal material. Mineral identity, grain size and shape, and an independent check on flux rate and bulk composition were provided by microbeam (SEM) methods.

Results and Discussion

SEM analyses of samples indicate that the dusts are composed of common minerals (quartz, feldspars, micas, clays, carbonates; small amounts of pyroxenes and amphiboles; smaller amounts of accessory minerals such as zircon and rare earth minerals; also pollen grains), and that the mode (assemblage of actual mineral species present) is consistent with the bulk chemical results.

In dusts in the U.S. Southwest, “enrichment factors” for trace elements (the factor by which the concentration of an element is greater than in the average crust of the earth) are commonly between 10 and 100, although they range by element and by sample type. Enrichment factors are especially high in finer-grained, farther transported dusts, namely dust in snow pack strata with low concentrations of dust, and in dry deposition samples in which only small amounts of dust are deposited on collection plates per unit time. As for the pre-industrial dust in polar ice and the European peat bog deposits, enrichment factors commonly range between about 10 and more than 100. The polar dust is clearly fine-grained and far-transported, the peat bog dust may be a mixture of near-source and far-transported dust.

A possible argument to dismiss the observed similarity in degree of trace element enrichment in modern SW U.S. dust and pre-industrial Antarctic or European peat bog dusts is that polar dusts and dusts over heavily-vegetated Europe, being fine-grained (large surface/mass ratio) and far-transported, must have been enriched to the maximum extent, because of “exposure” time during atmospheric transport; but that dusts within the dry US Southwest (possibly coarser, more locally transported) could have acquired equivalent amounts of trace elements only if a supplementary (modern; anthropogenic or unique local) source were available. However, this argument cannot be evaluated at present because it is not clear that either of the older dusts had finer grain size than the fine component of U.S. SW dust, and it is not clear that either kind of dust has been transported over greater distances, because it has been shown that fine-grained dusts of uniform composition appear to constitute a hemispheric or global background atmospheric load, which can be identified at low-energy times at both polar and dusty continental interior sites of deposition (Hinkley et al., 1997). The worldwide, “background” atmospheric dust is not dependent on local source regions (Hinkley et al., 1997). This fine dust may be a scavenger of trace metals (possibly from natural sources) during its long residence in the atmosphere, and may account for a large portion of the universality of trace element enrichments in dust.

Industrial trace metal pollution has been documented in northern polar and even Antarctic ice (Sherrel et al., 2000; Rosman et al., 1994). However, these modern increases in elemental inputs have been stated only as concentrations in ice, not as changes in “enrichment factors” of the dusts present in the ice, partly because reliable measurements of both dust content and trace element content have not been made together in the same studies. Those studies, and other documented trace metal pollution in the world today suggest that trace element loadings of S.W. US dusts have likely increased in modern
industrial times, but the extent is unknown. Local terranes have been proposed as sources of trace element enrichment for some S.W. dusts collected in specific nearby regions (Reheis et al, 2002).

Conclusions

The finer-grained component of atmospheric dusts in the Southwestern U.S. is significantly “enriched” in many ordinarily-rare trace elements, commonly to factors of 10-100 above the amounts that would be expected if the sources were unaltered average crustal material. This degree of enrichment in dusts in the SW U.S. is about the same as in pre-industrial dusts preserved in Antarctic ice and in European peat bogs, which have natural sources of enrichment. The similar levels of trace element enrichment of the of dusts from the two very different times and among the three locations appear to indicate that naturally-high amounts of trace elements, as seen in the pre-industrial samples, are at the very least a significant component of what is seen today in dust in the Southwest, and that the trace element loads of the modern and ancient dust are indistinguishable at present state of knowledge. The remaining task is to compare the absolute fluxes of key trace elements (mass deposited per unit area per unit time) for such different locations and the different time periods they represent.

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Factors affecting sand dune mobility on the Navajo Nation, Arizona, U.S.A.

M.M. Hiza, U.S. Geological Survey, Flagstaff, Arizona 86001 (mhiza@usgs.gov)

Introduction

The interaction among natural and invasive vegetation species, dune mobility, and changes in soil moisture conditions are currently under examination on the 71,000 km² Navajo Nation. Sand-dune deposits that cover one-third of this area are being assessed for potential sand dune mobility by combining surficial mapping with data gathered on rainfall, temperature, and vegetation cover and its characteristics. Meteorological data are used to calculate the ratio of precipitation to potential evapotranspiration (P/PE) that has been shown to be a critical factor controlling dune mobility, because of its direct link to the amount of stabilizing vegetation (Lancaster and Helm, 2000; Muhs and Holliday, 1995; Muhs and Maat, 1993). Thresholds in P/PE for changes in dune mobility are based on observations by Muhs and Holiday (1995), with transitions from mostly stable to mostly active at P/PE = 0.315, and from mostly active to fully active at P/PE = 0.125. This study has classified Navajo Nation dune fields as active or stabilized, and the degree of dune mobility has been compared to local conditions of calculated P/PE. New findings in this study generally corroborate previous work, but suggest that certain vegetation species may alter these thresholds and exert an overriding influence on dune mobility.

Discussion and Results

Calculated annual values of P/PE of data, from 40 meteorological stations within and adjacent to the Navajo Nation, exhibit a wide range of values that reflect distinct local variations in climate as well as temporal variations in soil moisture. Based on these calculations, and the known climatic variability of the past 50 years, sand dunes on the Navajo Nation exist under meteorological conditions that promote the entire spectrum of dune mobility, from mostly stable, mostly active, to fully active during periods of drought. Arid regions within the Navajo Nation have consistently low P/PE, with median values for areas such as Monument Valley as low as 0.145, indicating that dunes in these areas are “normally” on the threshold of becoming fully active. (P/PE values for drought years in these regions are typically from 0.07 to 0.10, well below the threshold of 0.125 for fully active dunes.) In contrast, Betatakin, in the western Navajo uplands has a median P/PE of 0.487, indicating that this region typically has enough moisture for dunes to remain fully stable.

Work to date has shown that areas with active and largely active sand dunes generally have a P/PE < 0.3, corroborating work of Muhs and Holliday (1995). However, dunes that are covered with native vegetation and stable are also found in these areas despite low soil moisture. In addition, mostly active dunes in the areas of moderate P/PE (0.25-0.35) are closely associated with Russian Thistle (tumbleweed), an invasive annual that requires minimal moisture to germinate. Active to mostly active dunes associated with this type of vegetation have been mapped in areas that differ significantly in calculated soil moisture across the Navajo Nation. The relation of dune mobility to vegetation may be altered due to the presence of this invasive plant, which is an annual designed to detach and blow away during dry, windy periods.
An examination of temporal variation in P/PE for areas on the Navajo Nation indicate a trend of decreasing soil moisture balance starting in 1988, with P/PE ratios in the eastern Navajo Nation that are below 0.3 for the first time in 50 years. These data suggest a change to warmer and drier conditions for the southern Colorado Plateau, beginning in 1988 and continuing to the present. Interviews with elderly Navajo residents who live in areas with active dunes near Tuba City and Chinle, Arizona, indicate that dunes in these areas have become more active in recent years. Dunes at a local home site in the Tuba City area have become mobile in the last five years, covering corrals and collapsing a house. Significant changes in sand migration were observed during two drought years: 1996 and 1999. However, an earlier, more significant dry period in 1988-1990 did not result in similar changes in mobility. Possible explanations for increased dune activity in years of less significant drought are 1) recent drought years are part a continuation of drier conditions, whereas the earliest significant drought years (1988-1990) were preceded by wetter conditions, and 2) the recent addition of Russian Thistle at the Tuba City site may have contributed to the rapid change in dune mobility during dry periods. Because Russian Thistle invades areas that are disturbed by human activities, land use factors may compound changes in dune stability that occur as a result of climatic variation and alter the ability of these areas to become stabilized by native vegetation during periods of increased moisture.

References


Wind-Strength Variations Inferred from Quartz Grain-Size Trends in the Lower Cutler Beds Loessite (Pennsylvanian-Permian, Utah, U.S.A.)

N. Hoang, School of Geology and Geophysics, University of Oklahoma, 100 E. Boyd Street, Norman, OK 73019 (ngochoang@ou.edu)

M.J. Soreghan School of Geology and Geophysics, University of Oklahoma, 100 E. Boyd Street, Norman, OK 73019 (msoreg@ou.edu)

G.S. Soreghan, School of Geology and Geophysics, University of Oklahoma, 100 E. Boyd Street, Norman, OK 73019 (lsoreg@ou.edu)

Introduction

The loess-paleosol sequences of the Chinese Loess Plateau record a high-resolution record of Plio-Pleistocene climate change. Of the several proxies used for paleoclimatic analysis in the Chinese Loess Plateau, magnetic susceptibility and grain size have been particularly key. Magnetic susceptibility increases reflect pedogenesis during interglacials, driven by increased precipitation and temperature (e.g. Maher and Thompson, 1992). In addition, grain-size analysis of the quartz fraction in particular (Porter et al., 1995; Xiao et al., 1995) has been used as a proxy for monsoonal wind strengths. To date, such analyses have been applied only to Plio-Pleistocene loess successions. Here, we attempt to extend these techniques to very ancient loess successions. Similar to the Quaternary, the Late Paleozoic is well known as a time of significant glaciation, marked at high (paleo)latitudes by widespread and unequivocal evidence for continental glaciers, and at low latitudes by pervasive and classic Pennsylvanian “cyclothems” typically consisting of intercalated marine and continental strata that record the repeated waxing and waning of the Gondwanan ice sheets (Crowell, 1978; Veevers and Powell, 1987). The glacial-interglacial fluctuations that drove glacioeustasy were also recorded in low latitudes of Pangea as fundamental climatic shifts, from relatively arid glacials to more humid interglacials (e.g., Soreghan, 1994, 1997), recorded within loessitic sections as alterations between loess and paleosol, respectively (G. Soreghan et al., 2002). Within the southeastern Paradox basin (southeast Utah), the lower Cutler beds consist of approximately 250 m of lithified loess (eolian silt) with interbedded paleosols, and accumulated in western equatorial Pangea during Late Pennsylvanian-Early Permian time (ca. 300 Ma). Pedologic, geochemical, and magnetic studies of this loess-paleosol sequence indicate that the system behaves much like that of the Chinese Loess Plateau, wherein magnetic susceptibility can be related to pedogenic alteration (cf. Maher and Thompson, 1992), and reflects probable glacial-interglacial fluctuations that operated in western equatorial Pangea (G. Soreghan et al., 1997, 2002). In addition to “icehouse” conditions, however, the paleogeography of late Paleozoic Pangea has led many to suggest extreme seasonality and mega-monsoonal circulation patterns (e.g. Parrish, 1993). M. Soreghan et al. (2002) employed provenance analysis of upper Paleozoic loessite deposits of the western U.S. to document apparent monsoonal circulation in western equatorial Pangea. Accordingly, loess systems of late Paleozoic western Pangea record the influence of both icehouse and monsoonal conditions, analogous to today. In this contribution, we present preliminary data on quartz grain-size trends through loessite-paleosol couplets of the lower Cutler beds, in order to assess their temporal variation, and possible relation to atmospheric circulation (wind strength and variation) in western equatorial Pangea during late Paleozoic time.
Methods

The lower Cutler beds loessite is well lithified, and not easily disaggregated to allow standard grain size analysis. Accordingly, we employed image analysis of backscattered electron (BSE) microprobe images to measure a proxy for grain size. For this study, we chose nine loessite-paleosol couplets, three each from the base, middle and top of the section, in order to examine both short- and long-term trends in relative grain size. We prepared polished microprobe rounds of 54 samples from the nine couplets and acquired several (8-12) BSE images of each sample (Fig. 1a). Each image was then digitally analyzed and filtered in Adobe Photoshop to highlight grain outlines of approximately 800 quartz grains per sample. We focused on quartz alone, owing to its usefulness as a wind strength proxy and its resistance to chemical weathering that may have occurred in both pedogenic and diagenetic environments (cf. Porter and An, 1995; Xiao et al., 1995). In BSE mode in the microprobe, quartz grains can be identified as uniformly gray grains with smooth surfaces and their grain boundaries are distinct. A series of filtering steps, including noise reduction, edge finding, and thresholding were employed to isolate and highlight the quartz grains (Fig. 1b). Although these image filtering and processing steps worked well in general, problems and ambiguities arose in samples containing, for example, minor authigenic silica. Accordingly, it was sometimes necessary to manually eliminate non-quartz grains, or segment grains.

Following image processing, we used the National Institute of Health’s (NIH) freeware to measure the grain area, perimeter, and major and minor axes of the imaged quartz grains in order to determine the variation in apparent quartz grain sizes among the different samples in a single profile, and among different profiles.

Results and Discussion

In general, within loess-paleosol couplets, the median grain area of quartz commonly fines from the loess into the paleosol (Fig. 2). We follow Porter and An (1995) and Xiao et al. (1995) in using quartz as a good proxy for wind strength, and thus infer that the general fining-up trend reflects decreasing wind strengths from time of significant loess accumulation (glacials) to times of pedogenesis (interglacials). Further, integration of the apparent quartz grain size data with previously collected data on magnetic susceptibility through the section reveals a strong inverse correlation \( r^2 = \)
0.7885) wherein smaller median grain area values correlate to higher magnetic susceptibility values, similar to results reported by Porter et al (2001) for the Chinese Loess Plateau. Because magnetic susceptibility appears to reflect a primary climate control in this system (G. Soreghan et al., 2002), we infer that this correlation supports our inference of using quartz grain size as a proxy for wind strength in this very ancient system.

**Conclusions**

(1) Grain-size analysis on lithified loessite is possible using digital image analysis.
(2) In the upper Paleozoic lower Cutler beds loessite, apparent quartz grain sizes generally decrease upward within a loess-paleosol couplet, which we infer to reflect decreasing wind strengths associated with the transition from glacial to interglacial conditions.
(3) Quartz grain size inversely correlates with magnetic susceptibility, which we infer to suggest a primary climatic control on both parameters.
(4) Our study suggests that techniques applied for paleoclimatic analyses in Plio-Pleistocene loessite may also be applicable to very ancient loess successions.
(5) Loess-Paleosol successions of the late Paleozoic preserve high-frequency climatic cyclicity related to glacial-interglacial fluctuations, analogous to Plio-Pleistocene successions.

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Climate Change in the 21st Century and the Impact on Dunefield Mobility in the Kalahari

M. Knight, Department of Geography, University of Sheffield, U.K. S10 2TN
(Email: ggp00mk@sheffield.ac.uk)

D.S.G Thomas, Department of Geography, University of Sheffield, U.K. S10 2TN
(Email: d.s.thomas@sheffield.ac.uk)

G.F.S Wiggs, Department of Geography, University of Sheffield, U.K. S10 2TN
(Email: g.wiggs@sheffield.ac.uk)

Introduction

The Mega Kalahari is over 2,500,000km², fringing Angola and Zambia in the north and Botswana and southern Africa in the south (Grove, 1969). Sand dunes are considered vegetated and degraded in the northeastern reaches (Thomas et al, 2000) whilst episodic activity has been witnessed in the drier southwest. Here the climate and therefore dune activity has been dynamic over the past few decades, resulting in increased dune mobility throughout the droughts of the 1980s and 1990s (Bullard et al, 1997).

If this warming trend continues over the next one hundred years it may lead to some change in the geomorphology of southern Africa. Bridgman (1998) warns that the predicted northward shift in the Inter-Tropical Convergence Zone (ITCZ) and the strengthened El Nino Southern Oscillation (ENSO) phenomenon could lead to such a scenario, with the associated drying effect being a result of a 10 to 20% reduction in precipitation. Even wind speeds may reflect fluctuating ITCZ boundaries making it possible for increased aridity and enhanced windiness in areas where dunes are presently bordering the critical threshold for sand movement. However Joubert et al (1996) describe a wetter southern Africa where vegetation could flourish, which would provide the potential to stabilise all dunes in the Kalahari region.

In addition to the work completed in north America by Muhs and Maat (1993) and Wolfe (1997), it will be important to see what effect climate change may have on dune activity in a southern hemisphere dunefield, the spatial extent of such changes, its magnitude and any seasonal alterations in the timing of peak activity. It is also important to investigate the methods used to predict the extent of dunefield activity by highlighting the problems and uncertainties of using Global Climate Models (GCMs).

Methods

Dune mobility for each calendar month is modelled using the simple indices of Lancaster (1988) and Talbot (1984), which consider both erosivity (wind power) and erodibility (soil moisture). These are given in equations 1 and 2 respectively:

\[
\text{Lancaster} = \frac{W}{(P:PE)} \quad \text{(Eq. 1)}
\]

\[
\text{Talbot} = \frac{V}{M_o^2} \quad \text{(Eq. 2)}
\]

Where \( W \) is the percentage of time the wind is above the threshold for sediment movement, \( P \) is precipitation, \( PE \) is potential evaporation, \( V \) is wind speed and \( M_o \) is the Thornthwaite moisture index.

Future predictions for climate are gained from the results of four GCMs that consider a doubling of atmospheric carbon dioxide and an increase of carbon dioxide that is combined with an increase in cooling sulphates.
Challenges

Three main areas of uncertainty have been identified; what GCM results provide for those assessing climate change impacts on the environment; what data are available for dune mobility index calibration; and how equations one and two represent monthly dune activity when the original intention was to produce an annual potential (Table 1). Two subsets of these problems are particularly relevant to aeolian studies: one questions the integration of GCM output with Lancaster's and Talbot's mobility indices and the other concerns the calculation of dune mobility for each calendar month.

Table 1. Challenges of predicting future dune mobility using climate predictions made by GCMs and simple mobility indices.

<table>
<thead>
<tr>
<th>Challenges</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>GCM climate output</strong></td>
</tr>
<tr>
<td>Coarse spatial resolution of output (finest is 2.5° lat. x 3.5° long.)</td>
</tr>
<tr>
<td>Data integration problems with impact assessment models (mobility indices)</td>
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<tr>
<td>Lack of natural inter-annual variability in climate predictions made for the next 100 years</td>
</tr>
<tr>
<td>Uncertainty surrounding GCM performance and accuracy</td>
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<tr>
<td><strong>Calibration of monthly mobility values</strong></td>
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<tr>
<td>Lack of ground truth data or measurements of dune mobility for each calendar month</td>
</tr>
<tr>
<td>Problems of different grid resolutions when validating GCM output</td>
</tr>
<tr>
<td><strong>Dune mobility indices</strong></td>
</tr>
<tr>
<td>Often very simple</td>
</tr>
<tr>
<td>Zero precipitation yields high mobility values and there are no lag times considered</td>
</tr>
<tr>
<td>High values do not fit into the original mobility categories</td>
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<tr>
<td>Output describes only one of four activity states</td>
</tr>
</tbody>
</table>

Data integration

There are problems with the temporal and spatial resolution of available climate data, the number of variables simulated and the format in which they are provided. For example some GCMs do not provide predictions for wind power. When they do, data are not ideal for input into the mobility indices as one mean monthly value is given whereas the extreme speeds or range would be much more useful. In addition, for the Talbot index, Thornthwaite moisture values are not modelled and some metric to imperial data conversion has to take place. Predictions are made for each grid cell but these cells can cover large areas incorporating places with differing levels of dune activity (Fig. 1). To solve some of these problems several alternatives have been investigated such as using climate generators to simulate a future wind environment, using Weibull distributions to calculate the probabilities of sand transporting events, or modification of the existing mobility indices to better incorporate what is available from the GCMs. Each have their own set of advantages and limitations but rarely do they overcome all the problems revealed.

Calibration of dune mobility values calculated for each month

Mobility values calculated per month do not fall into the original categories of dune activity created by Lancaster and Talbot that define crest activity, interdune movement and inactivity (fig. 2). Therefore values are produced which have not been related to what is observed in the field as there are very few records that can be used to compare what has been postulated by the GCMs to that which has been measured on a monthly basis. Zero precipitation yields infinite mobility values and the lack of lag times means that mobility output responds instantaneously to reduced rainfall. Smoothing procedures can incorporate artificial lag responses but these can average the temporal trend too severely. Remote sensing products may help to identify lag relationships between rainfall and vegetation dieback, which can then be incorporated into a more realistic model of intra-annual dune mobility.
Overview

Several areas of uncertainty have been revealed when attempting to predict future states of sand dune activity, which include the use of GCM results, the use of simple indices to calculate monthly mobility and the calibration of what is predicted to what has been observed. Despite these uncertainties, dune mobility indices may still be extremely useful when coupled with GCM output as their simplicity allows predictions to be made with very little input data. They are easy to use, they provide some indication as to the magnitude of dune mobility and they have been calibrated in several different dunefields across the world. However, there is great potential for discussion on the issues identified.

References


Subarctic aeolian reactivity in northern Finnish Lapland: a 10 ka year record

Mia M. Kotilainen, Department of Geology, P.O. Box 64, 00014 University of Helsinki, FINLAND (mia.kotilainen@helsinki.fi)

Introduction

Two post-glacial dune fields have been studied in the subarctic Finnish Lapland. Study areas are situated just below and beyond the Scots pine (Pinus sylvestris) forest limit. These dune fields were originally formed shortly after deglaciation along the same esker chain, which runs from Muddusjärvi dune field to northeast towards Iijärvi dune area in northern Finnish Lapland. The dune fields are situated 70 km apart. Aeolian activity has been reported widely in subarctic and arctic environments: in Alaska (e.g. Koster & Dijkmans 1988), Canada (e.g. Filion 1984, 1987; Filion et al. 1991), Greenland (e.g. Dijkmans 1990), Iceland (e.g. Arnalds 1990), and Fennoscandia (e.g. Bergqvist 1981, Seppälä 1995, Käyhkö 1997, Kotilainen 1991). Aeolian landforms are known to be sensitive indicators of environmental changes, especially in cold environments. The aeolian formations are in most cases derived from glaciofluvial or fluvial sediments and the reactivation processes can be related to climate and/or human impact. In this presentation the main focus is the dating of the aeolian reactivation events of these dune fields during the Holocene.

Methods

The stratigraphy of the dunes at both fields was studied with fine resolution sedimentology. In total 55 \(^{14}\)C datings from buried charcoal horizons at dune lee sides were obtained. Dune section was studied in 3 D and the charcoal horizons were carefully revealed by removing the upper sediment layers. Pieces of charcoal were then picked with tweezers from the table. All recent material (e.g. plant roots) was carefully removed from the pieces. Most of the samples were dated using conventional radiocarbon dating at the University of Helsinki Dating laboratory. Four samples were dated using AMS-technique.

Sedimentology and stratigraphy of the sandy units between charcoal horizons were studied and the most important units for dating purposes were determined. Those included sand layers under the lowest charcoal horizon and also between the charcoal horizons as well as layers covering a major erosional plane. The sand units were sampled for TL/OSL –dating using black plastic tube (diameter 10 cm), which was hammered into the dune section. Dating of the 15 TL/OSL -samples was carried out at the University of Helsinki Dating laboratory.

Results

Based on data of \(^{14}\)C datings from buried charcoal horizons on dune lee sides and TL/OSL dates of the aeolian sediments between them, the main periods of aeolian reactivation phases were estimated. The earliest reactivation phase (Phase I) appears to have taken place at around 8400 - 7900 cal years BP. The next dune building phase occurs at around 7300-5800 cal years BP (Phase II). At 4500-3850 cal years BP is Phase III and 3200-2400 cal years BP Phase IV. The last reactivation phase V occurs only at the sparsely vegetated Iijärvi dune field at around 500- cal years BP and is a continuing process.
Conclusions

The dune fields had been colonized by vegetation after the most intensive aeolian period just after deglaciation, and this oldest charcoal horizon represents the average age of this burned down pioneer forest. The second charcoal horizon indicates the average age of the next stabilizing vegetation at the dune field. The period between these charcoal horizons marks the reactivation phase. Forest fires are natural phenomena and do occur frequently throughout the forest history. However, the length of the reactivation phase is a signal of climate and/or human impact on the area.

This earliest aeolian reactivity phase can be compared to rapid environmental changes observed in many proxies around North Atlantic region. The rapid cooling around 8.2 ka has been recorded e.g. in Greenland ice-core proxies (O’Brien et al., 1995), North Atlantic deep-sea sediments (Bond et al., 1997) and lake sediments e.g. in Sweden and Finland (e.g. Korhola et al., 2001). Reactivation of the dune fields might be related to reorganization of the atmospheric and surface ocean circulation over Greenland and North Atlantic. It is possible that this “8.2 ka cooling event” resulted in surface water cooling and decrease in the North Atlantic Current northward transport of surface water causing cooling of the high latitudes. This mechanism might decrease precipitation over the subarctic Fennoscandia and thus trigger the onset of the aeolian reactivation in the region. These results suggest coupled ocean–atmosphere forcing of the whole Holocene aeolian history in the subarctic Finnish Lapland.

References


Linear dunes in the western Sahara, Mauritania: chronology and reconstruction of Late Pleistocene and Holocene wind regimes

Nicholas Lancaster, Desert Research Institute, 2215 Raggio Parkway, Reno, NV89512, USA. Nick@dri.edu

Gary Kocurek, Department of Geological Sciences, University of Texas, Austin, TX 78712, USA. garyk@mail.utexas.edu

Ashok Singhi and V. Pandey, Planetary and Geosciences Division, Physical Research Laboratory, Navrangpura, Ahmedabad, INDIA 380 009. singhvi@prl.ernet.in

Max Deynoux and Jean-Francois Ghienne, Centre de Géochimie de la Surface, EOST, CNRS-Université Lois Pasteur, 67084 Strasbourg-Cedex, France. mdeynoux@illite.u-strasbg.fr

Khalido Lô, Département de Géologie, Faculté des Sciences et Techniques, Nouakchott, Mauritania

Introduction

The western Sahara in Mauritania is dominated by extensive sand seas that consist largely of linear dunes (Kocurek et al., 1991) that form the western end of sand transport systems that originate in the northern Sahara of Algeria and terminate in the Atlantic Ocean. Previous workers in the region (Fryberger, 1980; Sarnthein and Diester-Haas, 1977) drew attention to the existence of crossing dune trends and superimposition of dunes on different alignments, suggesting that wind regimes in the region have changed over time from one generation of dunes to the next. We targeted one area in the western part of Mauritania between latitudes 18°30’ and 20°30’ N, in which several dune trends are visible on satellite images, and where three of the sand seas in the region (Azefal, Agneitir, Akchar) are in close proximity.

The sand seas and their component dune elements were mapped on Landsat TM images of the area using their distinctive spectral characteristics. We identified three dune-trend classes that decrease progressively in size: (1) a NE-SW-trending (045°) class of large, degraded linear ridges that appear brown in images, (2) a NNE-SSW-trending (020°) class of moderate-sized linear dunes, commonly with active crestal areas, which appear yellow in images, and (3) a class of N-S-trending small linear dunes that appear white in images.

Each of the dune-trend classes was selected for OSL dating. Dune sands were sampled by a combination of hand-dug pits (maximum depth 1.5 m) and augering (maximum depth of 5 m). The OSL ages cluster into 3 groups: (1) 24-15 ka, (2) 13-10 ka, and (3) after 5 ka. Each of these periods of eolian activity is associated with a distinct linear dune trend. The oldest ages (24-15 ka) are all associated with the NE-SW dune trend, and are centered around the Last Glacial Maximum. Dunes with an age of 13-10 ka trend NNE-SSW, and span the period of the Younger Dryas event. N-S oriented linear dunes overlies the prominent pedogenic surface developed during the Holocene African Humid Period. The late Holocene ages in our study are associated with local eolian activity in interdune areas, but appear to be representative of a much more widespread period of dune activity that is correlated with the N-S linear dunes and continues to the present day. The ages of these periods of dune construction closely parallel the marine record of increased dust transport in this region (deMenocal et al., 2000).
Assuming that eolian bedforms are aligned to maximize gross sediment transport normal to their crests (Rubin and Ikeda, 1990), it is possible to determine the dune trend that best satisfies the gross-bedform-normal rule in present day wind regimes, as well as to simulate the most likely combination of winds that produced dunes of different trends in the past. Simulations of the dune-forming winds using this approach indicate that wind regimes in this area during the Last Glacial Maximum and the Younger Dryas were characterized by enhanced easterly, northerly, and northwesterly winds. This suggests intensification of the seasonal high pressure cells during periods of dune construction.

References


Eolian Deflation of Holocene Playas and Formation of the White Sands Dune Field.

Richard P. Langford, Dept of Geological Sciences, University of Texas at El Paso, El Paso, Texas 79968-0555, USA, langford@geo.utep.edu

Introduction

White Sands National Monument is the largest field of gypsum dunes in the world (Figure). Deposition in many of the desert basins of the region ceased during the Mid-Pleistocene. In the central part of the Tularosa basin, deposition continued through the late Pleistocene and Holocene, where pluvial Lake Otero extended across the basin (Seager et al., 1987; Blair et al., 1990; Buck, 1996)). During the latest Pleistocene the last Lake Otero shoreline formed at 1,216 m elevation (Seager et al., 1987), and was followed by desiccation and the formation of the playa lakes and dune field.

The association of climatically constrained eolian deflation in the creation of the dunes is now being emphasized (Fryberger, personal communication, 2000; Langford, in press). The dunes lie downwind of a large deflation basin that has been incised into the sediments of Lake Otero (Figure). Modern, gypsum-crusted playa lakes, including Lake Lucero are found in the lowest parts of the deflation basin and are one source of gypsum sand for the dune field. Between the dunes and the playa lakes lies the Alkali Flat, a largely unvegetated expanse of blowing gypsum sand within salt flats that contain scattered Mesquite and salt brush.

Erosional Shorelines

An erosional shoreline surrounds modern Lake Lucero playa. Erosional shorelines are created by the erosive activity of waves when the playa is flooded. The base of an erosional shoreline marks the level of flooding of the playa. This shoreline is partially buried beneath aggrading playa mud and salt, but forms a well-defined escarpment around all except the northern end of the playa. The base of this erosional shoreline escarpment is the almost-horizontal playa surface. Over fifty measurements give escarpment slopes of 30° to 40° in poorly consolidated Lake Otero sediments.

Two similar erosional shorelines can be correlated on the slopes above the modern playa (Figure). The lowest of the shorelines occurs at approximately 1,191 m elevation (5.5 meters above the surface of Lake Lucero). A second shoreline is located at approximately 1,200 m of elevation (14.5 meters above the surface of Lake Lucero). These higher shorelines, mark elevations where playas formed between stages of deflation. The lower erosional shoreline (1191 m) is marked as the L-2 shoreline and the upper shoreline (1200 m), the L-1 shoreline. The L-2 shoreline is almost identical in morphology to the shoreline surrounding Lake Lucero. The upper, L-1 shoreline is a subtler feature that slopes 6° to 10° and is more gullied. Based on morphology the L-1 shoreline is older than the L-2 and Lake Lucero shorelines.

Older lacustrine sediments of Lake Otero are exposed by erosion within and between all the shorelines. These sediments, consisting of laminated carbonates and evaporites indicate deposition on the floor of a semi-permanent saline lake. Topographic profiles show the extent of deflation. Logged wells within the dunes indicate that the base of the eolian gypsum sand lies below the Lake Otero shoreline and slopes gently to the west. At the western edge of the dune field, the older lake muds are exposed, just above the L1 shoreline.

Source of the Dunes

The horizontal beds of lacustrine Lake Otero sediments are the key to understanding the history of the White Sands because they define a low-relief Pleistocene lake floor. Thus, the topography of the basin below the Lake Otero Shorelines is largely a product of Post-Otero erosion. Because the shorelines are preserved, deflation has deepened the basin, but not widened it. Most previous authors have noted that two sources of gypsum dune sand, Otero Sediments and Lake Lucero...
Allmendinger (1972; LeMone, 1987). Allmendinger (1972) described a gypsum crystal-bearing layer in alkali flat up to 9m above the surface of Lake Lucero and suggested that deflation of these Pleistocene sediments produced most of the dunes. The main mass of the White Sands dune field begins abruptly near the L1 shoreline. However, several patches of parabolic dunes are forming by deflation of Lake Lucero and extend downwind from Lake Lucero, partially burying the L1 and L2 shorelines. Eolian sand was collected along a transect beginning near the L1 shoreline at the edge of the dune field. Samples near the shoreline include many angular unfrosted blades, indicating little transport. While 4 km into the dune field, the sands are equant and frosted, indicating that the source of the White Sands is probably the older lake sediments near the L1 shoreline.

**Age of the Dunes and Association with Climatic Events.**

The shoreline history described above implies, that while today, the dune field is actively migrating and is receiving a limited sand supply from the Alkali Flat, the majority of the growth of the White Sands dune field must have occurred during three short-lived arid climatic events, when the deflation basin was being deepened. One probably predated the dune field and occurred at the end of the Pleistocene resulting in deflation to the L1 shoreline. Two more events, the earliest circa 7,000 years BP formed the dune field and deflated to the L2 and Lake Lucero shorelines. Because the Pleistocene, Lake Otero Gypsum beds lie below the L1 shoreline, it is unlikely that a large dune field had formed until the deflation event between the L1 and L2 shorelines. No datable materials have been found yet on the L1 shoreline. However, searches of the archeological database at White Sands National Monument shows Folsom culture sites above the elevation of the L1 shoreline, but not within the L1 deflation basin (Eidenbach, Personal Communication, 2000). This suggests that the L1 shoreline can be dated to younger than the 9,800 to 10,800 BP range of Folsom activity (Ingbar, 1992).

Initiation of the White Sands Dune field coincided with deflation to the L2. Radiocarbon dates collected from the sediments below Lost River playa provided dates of 1,750 and 5,840 years BP (Monger and Gallegos, 1997). The dated sediments filled a basin dammed by the dunes suggesting that the growth of the White Sands dune field, and incision to the L2 shoreline occurred after 9,000 and before 5,840 BP. Two deflation events, dated at 7,000 and 4,000 ybp are increasingly identified with playa basin and eolian sand generation in the region (e.g. Allen, 1991; Buck 1996).

Several recent authors have noted the influence on climate change on playa hydrology (Jacobson, 1988; Doremus et al., 1989; Rosen, 1994; Fan et al., 1997). These authors emphasize that deposition of evaporates on playas only occurs when there is a balance between an influx to the playa and discharge of dense brines plume beneath the playa. When inflow is greater than discharge a lake forms. When discharge exceeds inflow, the groundwater table beneath the playa subsides and deflation may occur. The implications of these studies to the White Sands are twofold. First, generation of gypsum from the playa surface will be episodic and associated with climate fluctuations. Second, the deflation events are probably associated with longer-term arid climatic event that result in a lowering of the water table beneath the floor of the deflation basin.

Figure. Composite aerial photograph of Lake Lucero and the White Sands dune field. Dashed lines show the Lake Lucero, L2, and L1 shorelines. Posted elevations mark sites of GPS measurements of shoreline elevation. (Aerial photograph supplied by White Sands National Monument and prepared by ESRI, Albuquerque, NM). Inset—Location map showing the relationship of the gypsum dunes to Lake Lucero and the alkali flat. Cross section A to A’ shows the elevations of the Lake Lucero, the L 1 and L 2 shorelines and the Lake Otero shoreline. Photos are of gypsum grains collected along the transect.

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Wind-Strength Variations Inferred from Quartz Grain-Size Trends in the Maroon Formation Loessite (Pennsylvanian-Permian, Colorado, U.S.A.)

K.M. Moreland, School of Geology and Geophysics, University of Oklahoma, 100 E. Boyd Street, Norman, OK 73019 (kmmorel@ou.edu)

M.J. Soreghan, School of Geology and Geophysics, University of Oklahoma, 100 E. Boyd Street, Norman, OK 73019 (msoreg@ou.edu)

G.S. Soreghan, School of Geology and Geophysics, University of Oklahoma, 100 E. Boyd Street, Norman, OK 73019 (lsoreg@ou.edu)

Introduction

Understanding past climates of the Earth is of fundamental importance in predicting and managing future climate change. Late Paleozoic Pangea, in particular, is of significant interest to paleoclimatologists because the configuration of this supercontinent is hypothesized to have greatly perturbed Earth’s climatic system. The pole-to-pole land distribution and cross-latitudinal orientation of Pangea have led many to suggest extreme monsoonal circulation and strong seasonality for the Pangean interval (e.g. Robinson, 1973; Parish, 1993). The initiation and temporal evolution of Pangea’s megamonsoon was also likely modulated by higher frequency glacial-interglacial climate fluctuations that are well preserved in the upper Paleozoic loess-paleosol sequences of the western U.S. (Soreghan et al., 2002). Previous studies have shown that similar sequences present in the Chinese Loess Plateau serve as high-resolution records of Plio-Pleistocene climate change. Grain-size analysis of the quartz fraction in particular (Porter et al., 1995; Xiao et al., 1995) has been used as a proxy indicator of monsoon wind strength, while increases in magnetic susceptibility values represent times of pedogenesis during interglacials (Maher and Thompson, 1992). This study attempts to extend such techniques to very ancient loess successions of the Late Paleozoic to assess atmospheric circulation patterns (wind strength and variation) in western equatorial Pangea.

The Pennsylvanian Maroon Formation loessite is located within the eastern Eagle basin (northwest Colorado) and consists of approximately 700 m of lithified loess (eolian silt) punctuated with numerous interbedded paleosols. It accumulated during the Late Pennsylvanian-Early Permian (ca. 300 Ma) in western equatorial Pangea at paleolatitudes of approximately 5°-10° N (Johnson, 1989). The loess-paleosol sequences exposed in this unit exhibit significant variations in magnetic susceptibility and record glacial-interglacial climate fluctuations of Pangea (Soreghan et al., 1997; Tramp et al., in review). The loess units represent relatively arid glacial periods marked by high dust influx, whereas paleosols record more humid interglacials with decreased silt influx and enhanced pedogenic activity. In this study, we present preliminary data on variations in quartz grain-size within several loessite-paleosol (glacial-interglacial) couplets and of the Maroon Formation, and we assess their possible paleoclimatic significance.

Methods

The loessite of the Pennsylvanian Maroon Formation is well lithified and not easily disaggregated, making standard grain-size analysis inappropriate. Consequently, we analyzed apparent grain-size using backscattered electron (BSE) microprobe images. For this study, cores of approximately 2.5 cm in diameter were drilled from the Maroon Formation using a portable gas-powered and water-cooled

Drill. Samples were collected from 3 consecutive loessite-paleosol couplets at the base, middle, and top of the 700 m section, for a total of 9 couplets and 69 samples. We prepared polished microprobe rounds and collected 10 BSE images for each sample (Fig. 1a); and, minerals were identified by their gray levels (intensities) and textures. The digital images were then imported into Adobe Photoshop 5.5 for image processing designed at isolating quartz grains, with a target of 800 quartz grains per sample. Owing to its resistance to chemical weathering, which is common in both pedogenic and diagenetic environments, quartz constitutes a more reliable proxy for paleowind strength (cf. Porter and An, 1995; Xiao et al., 1995). A series of filtering steps, including noise reduction, edge finding, and thresholding were employed to create binary images in which quartz grains were highlighted (Fig. 1b). Overall, this filtering routine worked relatively well but in some cases it was necessary to manually separate grain-to-grain contacts created by authigenic silica, and remove spurious pixels.

Each binary image was then imported into the National Institute of Health’s (NIH) freeware software, NIH Image v. 1.62, for grain-size analysis. This program measures characteristics of individual grains, such as area, perimeter, and major and minor axes. The software allows the user to predefine a minimum quartz grain area for analysis (8 \(\mu\m\)), eliminate grains touching edges, and fill in interior holes. Text file results were imported into Microsoft Excel for statistical analysis and, ultimately, for the determination of apparent (2D) quartz grain-size trends in single profiles as well as throughout the entire section.

Results and Discussion

In general, median quartz grain area decreases both up through the entire Maroon Formation and from parent loessite to pedogenically altered loessite within any given loess-paleosol couplet (Fig. 2). Following the findings of Porter and An (1995) and Xiao et al. (1995), we relate the upwardly fining trends within individual couplets to decreased wind speeds during interglacials. We are still investigating the long-term temporal variation in quartz grain-size throughout the entire Maroon Formation, but we suggest that it may reflect the initiation and evolution of the megamonsoon in western equatorial Pangea during the Late Paleozoic. Additionally, the grain-size data also inversely tracks \(r^2 = 0.916\) previously collected magnetic susceptibility values throughout the section, such that finer quartz grain-size corresponds to higher magnetic susceptibility values, a result similar to that reported by Porter et al. (2001) for the Chinese Loess Plateau. Magnetic susceptibility varies in synchrony with glacial-interglacial cycles in this study, which is a relationship implying a dominant climatic control (G. Soreghan et al., 2002). We feel that this correlation ultimately allows us to extend the use of quartz grain-size as a proxy index for wind strength to this very ancient system.
Conclusions

(1) Grain-size analysis of lithified loessite can be performed by applying filtering routines to 2D backscattered electron (BSE) images.

(2) In the Pennsylvanian Maroon Formation, apparent quartz grain-sizes decrease upward within individual loess-paleosol couplets, reflecting a decreased wind strength associated with the transition from glacial to interglacial conditions.

(3) Quartz grain-size and magnetic susceptibility are inversely correlated, suggesting a primary climatic control on both parameters.

(4) This study suggests that techniques utilized for Plio-Pleistocene paleoclimatic analysis are also applicable to very ancient loessite successions.

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The concept of mineralogical maturity and the origin and evolution of dune fields in the western United States

Daniel R. Muhs, U.S. Geological Survey, MS 980, Box 25046, Federal Center, Denver, Colorado 80225 USA (dmuhs@usgs.gov)

Introduction

Studies of dune fields in the western United States show that mineralogical maturity can provide new insights into the origin, evolution and long-term history of eolian sand bodies. Mineralogical maturity can be defined as a compositional state of a clastic sedimentary body wherein there is a dominance of quartz and an absence or minority of less-resistant particles such as feldspars, detrital carbonates or lithic fragments. Sandstones that meet this definition are classified as quartz arenites or orthoquartzites. Many of the world’s great sand seas in Africa, Asia and Australia are quartz-dominated and thus can be considered to be mineralogically mature.

In the western United States, the largest dune fields are found in the semiarid central and southern Great Plains. Smaller dune fields are found in the northern Great Plains, the Basin and Range province (including the Mojave and Sonoran Deserts), the Colorado Plateau, and intermontane basins of the Rocky Mountains. Studies conducted over the past 10 years show that these dune fields have a wide range in degree of mineralogical maturity. Major oxide analyses provide an indirect but quantitative estimate of mineralogy, where key oxides serve as proxies for common minerals found in dune sand. Examples include SiO₂ (quartz), K₂O (K-feldspar), Na₂O (plagioclase), Al₂O₃ (all feldspars) and CaO and MgO (carbonates).

Based on geochemical criteria (see Figure 1), dune fields of the western U.S. fall into three categories of mineralogical maturity: immature, intermediate, and mature. Examples of immature dune fields include many of those found within the Mojave Desert, such as the Cadiz and Danby dunes of California. Intermediate-maturity dune fields include several found in the central Great Plains (Fort Morgan and Wray dune fields of northeastern Colorado) and some in Rocky Mountain basins (Killpecker dunes, Wyoming). Mineralogically mature dune fields include larger sand bodies in the central and southern Great Plains (Nebraska Sand Hills and the Muleshoe and Monahans dunes of Texas and New Mexico), as well as smaller dune fields in the Sonoran Desert (Algodones, California and Parker, Arizona).

Mojave Desert dune fields may be mineralogically immature because they are not derived from major river systems but rather from small streams draining granitic mountain blocks. Furthermore, the dune fields are confined largely to structural basins adjacent to the source mountains themselves. Thus, sediments in these dunes have undergone little transport by either fluvial or eolian processes. In contrast, intermediate-maturity dune fields are those fed by major fluvial systems, such as the South Platte River. Sediments in these dunes underwent a considerable amount of fluvial transport prior to eolian transport, which concentrated quartz in the downstream reaches of the source rivers, where the dune fields are located. Intermediate-maturity dune fields are found in relatively flat, open landscapes, where there are fewer limitations on eolian transport compared to intermontane desert basins.
Figure 1. Plot of bulk SiO2 (reflective of quartz content) vs. Al2O3 + Na2O + K2O (reflective of feldspar content) in dune fields of the western United States. Data for the Cadiz and Danby dunes from Zimbelman and Williams (2002); data for the Killpecker dunes from Gibbons et al. (1990); data for all other dune fields from this study.

Some dune fields, such as the Algodones dunes of California and the Parker dunes of Arizona, may have achieved a degree of mineralogical maturity largely by inheritance. Both of these dune fields are derived from sediments of the lower Colorado River. Lower Colorado River sediments are themselves quartz-dominated, probably because they contain a large component of sand derived from mineralogically mature sandstones of the Colorado Plateau. The Algodones dunes show only slight enrichment of quartz (as measured by SiO2 content) compared to medium and fine-sand-sized Colorado River sediments.

In contrast, quartz-dominated dune fields of the central and southern Great Plains did not achieve mineralogical maturity solely by inheritance. It is true that the Nebraska Sand Hills and the Muleshoe and Monahans dune fields of Texas and New Mexico may have been fed, at least in part, by large river systems (the North Platte and Pecos Rivers) that concentrated quartz by fluvial processes. In addition, dunes in both areas probably inherited sand from older, quartz-rich eolian sheet sands (Pliocene sands in Nebraska and the Pleistocene Blackwater Draw Formation in Texas and New Mexico). Nonetheless, dunes in both regions are richer in quartz (higher in SiO2 and depleted in K2O, Na2O, and Al2O3) than any known source sediment or combination of source sediments. It follows from this that less-resistant particles, such as feldspars, have been depleted within the dune fields themselves, most likely by ballistic impacts from strong winds. Thus, mineralogically mature dune fields of the Great Plains reflect (1) relatively old ages for genesis of the dune fields, (2) extended periods of eolian activity as opposed to extended periods of stability, or (3) a combination of both of these factors.

References


Cenozoic soil sequences and paleoenvironments of West Texas, USA

C.G. OLSON, USDA-NRCS, 100 Centennial Mall, Lincoln, NE
S. M. CASBY-HORTON, USDA-NRCS, Temple, TX
B.L. ALLEN, Texas Tech University, Lubbock, TX
M.A. Cano-Garcia, INIFAP, Oaxaca, Mexico

Abstract

Since the Tertiary, material eroding from the backwearing of the eastern edge of the High Plains escarpment has been aggrading in a basin identified by Frye and Leonard (1964) as Pleistocene Lake Lomax. The presence of buried soils in exposures indicates episodic rather than continual rates of deposition. Throughout much of the year, evapotranspiration is high and many ground soils have accumulated calcium carbonate horizons. Several ground soils also contain petrocalcic horizons. These near-surface petrocalcic horizons are likely relict and in places appear to be degrading under current climatic conditions. In order to better understand present and earlier climatic conditions, paleoenvironmental variability can be inferred both from the clastic depositional sequences and the profile characteristics of buried soils. North of Big Springs, Texas, a hole drilled from the ground surface to the top of the Triassic beds penetrated more than 30 m of successive buried soils. Source materials include colluvium from eroded Ogallala, Cretaceous, and Tertiary clastic sediments. Few buried A horizons are preserved in the paleosols. Many paleosols are clearly welded. Commonly, there are alternating sequences of argillic and calcic horizons indicating fluctuations in precipitation intensity or in its occurrence. Some soils are separated by intervening colluvial and lake-bed sediments. The lacustrine deposits have a low bulk density that may indicate some ash influence. At least one buried soil located immediately below a lacustrine deposits exhibited gleyed colors indicative of reducing conditions. Following subaerial exposure, the gleyed colors rapidly became indistinguishable from the whitish colors of carbonate in the calcic and petrocalcic horizons of paleosols above and below. A sharp sedimentary boundary, marked by a very abrupt loss of carbonates at 29 m, is representative of a significant environmental change. A red, Triassic paleosol containing numerous root casts and insect burrows lies beneath and indicates a much higher level of biotic activity in this weathered shale than found in soils elsewhere in the sequence.

Reference

Spatial Variability of Aeolian Sands on Starczynow “Desert” (Eastern Part of Silesian Upland)

J. Pelka-Gosciniak, University of Silesia, 41-200 Sosnowiec, Bedzinska 60, Poland (pelka@us.edu.pl)

Introduction

The influence of wind on the substratum material is not only documented by changes in its features, but also in a form of segregation in space, i.e. territorial, according to wind direction, proper distribution of this material in respect of fraction (grain size), sorting, shape, degree of mechanical abrasion. Wind is characterized by a specific dynamic force, so the material should struggle with gravitation force during transport, even under the conditions of rather gentle airflow. Therefore, the clear differentiation of these deposits within whole sandy fields as well as in aeolian forms selected have been studied.

The aim of this paper is to know the spatial differentiation of textural features of sandy deposits in Starczynow desert in Silesian Upland in southern Poland.

Materials and Methods

To solve problem above-mentioned, a detailed field study was carried out during which 900 sand samples were collected from aeolian cover sands and selected dunes by using a series of knot points of a square net.

The whole material was subjected to laboratory investigation, which included standard analyses of grain size distribution according to equations of Folk and Ward (1957), quartz grain abrasion 1-0.8 mm, applying methods of morphoscopy by Cailleux (1942) and mechanical graniformametry by Krygowski (1964). Also the content of heavy minerals and feldspars was analyzed.

The statistics compilation of data allowed to map the textural parameters trends as well as spatial distribution. These data were also the base to make a complex estimation of these parameters anomalies. For a set of particular parameters smoothed by a movable mean, average values and their confidence intervals (at $\alpha =0.05$) were calculated. It was assumed, that parameters within this interval characterize the average conditions of the analyzed sand, while those beyond the intervals (positive or negative anomalies) indicate a relative predominance of deposition (accumulation) or redeposition (deflation) processes. It was estimated that redeposition is described by positive anomalies of Mz (mm), $\delta$, $K_G$ and negative anomalies of Sk, while the opposite anomalies of these parameters indicate deposition (Pelka-Gosciniak, 1999).

Area of investigation

The Starczynow “Desert” is situated in the eastern part of Silesian Upland (Southern Poland) and it occupies an area located backward of the Middle Triassic cuesta and in front of the Upper Jurassic cuesta. It is a compact area of aeolian sand, which till the 60s was intensively blown out. It has a flat surface, sloping from the east to the west, varied by numerous different dunes in form of clear parallel belts. Aeolian coversands show changing thickness within the whole desert. They form monotonous, slightly waved area of little inclination (in the majority 0-2°) with weakly marked slopes and relative heights. Their surface is varied by small hillocks and flat, rather irregular depressions without drainage, where in some places signs of rill- and sheet wash occur (Pelka-Gosciniak, 2000a).
This desert is not a typical climatic dry area, but it is a result of human impact on the natural environment. The wind activity in the area investigated in the Late Vistulian and Holocene caused the transformation of sand features and the formation of numerous accumulative forms. Till the early Middle Ages this area was covered with dense forests. Tree cutting and uncovering of sandy substratum was a consequence of wood demand to heat in the contemporary lead and silver ironworks, which were located near Olkusz. Due to uncovering of material substratum, the hitherto fixed sand was removed and the remodelling of the earlier relief followed. The formation of aeolian relief is still going on (Pelka-Gosciniak, 2000b; Szczypek, 1995; Szczypek, Wach, 1991; 1999). The origin of the sand of the Silesian Upland was subject of many investigations. The majority of authors connect the time of their formation with the Middle Polish Glaciation. Recent research connects the age of deposits of the Biala Przemsza valley with three glaciations: Oder, Warthe (Middle Polish) and Weichselian (Szczypek, Wach, 1989).

The genesis of sands in the eastern part of Silesian Upland can be defined as complicated, i.e. in different parts of this unit deposits of different origin occur: fluvioglacial, fluvial-proluvial, fluvial or proluvial-deluvial. It can be assumed that deposits located in the eastern part of sandy area of Starczynów “Desert” are of fluvial-proluvial origin (Pelka-Gosciniak, 1999, 2000a).

The anemological conditions of area investigated were analyzed by means of data supplied by the Institute of Meteorology and Water Management in Katowice from the nearest meteorological stations in Olewin, Zabkowice and Slawkow for the period 1961-1990. These data indicate that in the last 30 years winds from widely understood western sector (about 44% of cases) clearly predominate. Winds from these directions are also characterized by the largest mean velocities from 3 to 3.6 m/sec.

Taking into account the results of field measurements, i.e. observations of morphological axes and internal structure of aeolian forms selected and wind-shaped trees in the Starczynow “Desert” it can be also stated, that under real conditions the westerly winds predominated in area investigated (Pelka, 1994).

Results

Aeolian coversands in the Starczynow “Desert” exhibit a thickness of 1.5 m in the central and eastern parts to 2.5 m in the western part. Below thin series of structureless sands (30-50 cm), laminated sand deposits occur, where laminae dip generally in eastern or north-eastern direction at an angle of 6–8°. The substratum sand shows horizontal stratification. These sands are mostly built of medium-grained material (0.5-0.25 mm) (approximately to 56.4%). The content of the fraction>0.8 mm is rather small: 1.89%. The addition of dusty fraction is characterized by similar content and it amounts to 1.47%. The value of mean grain diameter Mz reaches 0.316 mm. The sorting of sands is moderate, the value of standard deviation δ equals 0.599. The deposits analyzed are characterized by good mechanical abrasion with the majority of well-rounded grains of γ type – on average 42.1%. Medium rounded grains of β type and angular grains of α type sum up relatively to 43.0% and 17.6%. Sands analyzed are characterized by the high share of mat-rounded grains of RM type – 44.7%, the value of abrasion degree parameter Wo amounts on average to 1376.

In the composition of aeolian covers 1.5-2% heavy minerals are present, among them garnet makes about 32% of the transparent fraction mass. The small amount of feldspars (about 2%) was also noticed.

To know the more important textural parameters of dune sands, material from dunes: barkhan-like, transverse and longitudinal was collected.

From the observation of the basic textural features of aeolian sands on Starczynow “Desert” results that materials which build aeolian coversands and dunes are very similar. It refers to grain size distribution as well as to quartz grain abrasion because the material, which builds dunes is medium-grained (on average 59%), the value of Mz=0.315 mm. The share of fraction>0.8 mm amounts to 1.52%. It contains about 2.5% of additional dusty fraction. Dune sands are characterized by moderate...
sorting ($\delta=0.56$), they are also well abraded, grains of $\gamma$ type is about 41.9%, $\beta$ and $\alpha$ - relatively 32.08%, 26%. Value of Wo amounts on average to 1311 (Pelka-Gosciniak, 1999, 2000a).

Conclusions

The analysis of the aeolian coversands and dunes allows to state that even in relatively small areas and within rather small forms, a clear spatial segregation of sand is observed. First of all, the influence of wind on the character of aeolian sands on Starczynow “Desert” is reflected in the coarsening of sandy material, evident by the increase in Mz (in mm) value and the decrease in Sk value, less sorting degree as the Upper Jurassic cuesta comes near (the increase in $\delta$ values), the decrease in values of $K_G$, which means the pulsatory changes in the environment energetics during deposition and the improvement of the abrasion degree by the increase in values of both Wo and the content of well-rounded grains of $\gamma$ type. The decrease of RM content together with direction of dune-forming wind is also typical of the analyzed area. The aeolian deposits are also characterized by the increasing share of garnets and the decreasing content of feldspars from west to east. Considering the tendencies of changes in the spatial distribution of dunes, it is evident to that the barkhan-like and longitudinal dunes are very similar, because their material becomes finer, better sorted and worse abraded. The transverse dune presents the opposite trend.

Therefore, the textural features connected with grain size distribution behave in a different way from the typical aeolian environment. The reason is the existence of an older relief in the form of the Upper Jurassic cuesta, which is a barrier on the route of wind. On one hand such a situation can be explained by the fact that wind, which blows over the obstacle decreases in its velocity, creating the conditions for coarser material to accumulate. But on the other hand it can be explained by the increase in velocity of air stream which overcomes the barrier and which is connected with the larger share of coarser material, transported by dragging. The size of grains decreases upward slope. Over the barrier only finer material is transported. It seems that the latter factor is of larger importance in the distribution of textural features of material in the vicinity of cuesta slope.

Another influence of older relief on sandy material textural features is observed in the vicinity of river valleys. At rivers which are natural barriers to the further development of wind transport as well sand translocation, accumulation prevails.

The second characteristic feature of aeolian sands occurring in the Starczynow “Desert”, which results from the influence of wind on the texture of the deposits, is the clear zonality of areas with deposits in the phase of redeposition (deflative areas) and deposition (accumulative ones). This zonality of textural features is observed in both grain size distribution and quartz grain abrasion. It is concluded that areas with sands in the phase of redeposition (deflative) are characterized by the occurrence of material of weaker degree of abrasion, but the areas with the tendency to deposition by better abraded one (Pelka-Gosciniak, 1999, 2000a).
References


The Developmental Significance of Dune Morphology in the Prairie Provinces of Canada

Z.K. Pfeiffer, Department of Geography, University of Guelph, Guelph, ON N1G 2W1
Email: zpfeiffe@uoguelph.ca
S.A. Wolfe, Geological Survey of Canada, 601 Booth St., Ottawa, ON K1A 0E8
Email: swolfe@gsc.nrcan.gc.ca

A recent inventory of sand dune occurrences across the Canadian Prairie Provinces identified 125 isolated dune fields; 6 in Manitoba, 43 in Saskatchewan, and 76 in Alberta, covering a total area of about 34,000 km². The orientation and morphology of these deposits are indicative of the changing geologic and climatic conditions across the provinces during the Holocene. For each occurrence, orientations and morphologies were interpreted and classified from air photographs ranging in scale from 1:20 000 to 1:80 000.

Classification results reveal four regions of similar development with 5 subsections determined by morphological differences. These regions reflect the spatial variations in sediment supply (amount and grain size of source material), sediment availability (through moisture and vegetation), and wind regime (direction, duration, and intensity). Within each region, temporal variations can be extrapolated using current knowledge of Holocene climatic change as well as the relative chronologies recorded in the dune morphology.

Temporal and spatial variations have been combined using sediment state theory, outlining the development of the aeolian landscape within each region. The sediment supply was generated during the early Holocene (12,000 to 8,000), and is of glaciofluvial, glaciolacustrine, or glacial deltaic origins. Estimations of wind regime (transport capacity) over time were based upon modern wind data, allowing for greater winds in the early Holocene as indicated by climatic models, as well as general morphological characteristics. Current levels of activity determined sediment availability, with historic variability based upon climatic inferences. Benchmarks of past activity indicated by dune morphology are currently being verified using existing radiocarbon and luminescence dates, reducing the uncertainties of these initial diagrams.
Late Quaternary Eolian History of the Needles Area of Canyonlands National Park, Utah: Dunes and Dust

Marith Reheis, U.S. Geol. Survey, MS-980 Federal Center, Denver, CO 80225 (mreheis@usgs.gov)

Richard Reynolds, U.S. Geol. Survey, MS-980 Federal Center, Denver, CO 80225 (rreynolds@usgs.gov)

James Yount, U.S. Geol. Survey, MS-980 Federal Center, Denver, CO 80225 (jyount@usgs.gov)

Helen Roberts, Luminescence Laboratory, Institute of Geography and Earth Sciences, University of Wales, Aberystwyth SY23 3DB, Wales, U.K. (hmr@aber.ac.uk)

Harland Goldstein, U.S. Geol. Survey, MS-980 Federal Center, Denver, CO 80225 (hgoldstein@usgs.gov)

Yarrow Axford, National Snow and Ice Data Center, University of Colorado, Boulder, CO 80309 (axford@nsidc.org)

Nancy Shearin, Bureau of Land Management, PO Box 7, Monticello, UT 84535 (nshearin@ut.blm.gov)

Introduction and Methods

In southern Canyonlands National Park (CNP), Utah, layers of locally derived eolian sand are separated by paleosols containing eolian dust. These deposits record cyclic, late Quaternary changes in land-surface stability. Eolian dust in soils and surficial deposits strongly influences landscape evolution and ecosystem dynamics of drylands. We are applying mineralogic, chemical, textural, biologic, monitoring, and dating studies to soils and surficial deposits of the semi-arid Colorado Plateau to (1) recognize eolian dust, (2) identify past and modern dust sources, (3) study the geomorphic history, (4) document the contribution of dust to nutrient uptake by plants, and (5) assess land-surface vulnerability to wind erosion in response to changes in climate and land use. Here, we present preliminary data on the depositional history, age, and paleoclimatic setting of eolian deposits in the Needles district of CNP.

Eolian sand is locally active in the CNP area, mainly in settings with an abundant sand supply and where stabilizing vegetation has been disturbed. An older depositional record is preserved in small relict dune fields and where sand has accumulated against topographic obstructions. Within the Park, studies were limited to hand-augered holes supplemented by arroyo cuts and by two hand-dug soil pits in a dune crest and a dune swale. Additional information was obtained from backhoe pits excavated in support of an archaeological investigation just outside the boundary of CNP (Shearin et al., 2000). Optically stimulated luminescence (OSL) ages were obtained using the single-aliquot regenerative-dose (SAR) technique for both the coarse-grained (90-125 µm) and fine-grained (4-11 µm) quartz fractions of several sand layers exposed in pits and arroyo cuts. Two different grain sizes were analyzed to test whether silt was deposited together with the dune sand or was infiltrated later as dust. In addition, one radiocarbon age was obtained on charcoal at the base of an alluvial deposit that truncated a paleosol formed on eolian sand. Palynological analyses were conducted on six sediment samples from the soil pit in the dune swale.
Results and Discussion

Stratigraphy and Ages

The main study area, located in the Grabens area of CNP, consists primarily of subdued, vegetated dune ridges and swales with a maximum relief of about 5 m. Auger holes showed that the sand mantle is at least 4 m thick in the middle of the area, thinning to a few cm at the edges where it overlies sandstone bedrock. All of the auger holes exposed multiple sequences of eolian sand separated by poorly to moderately developed, silt- and carbonate-enriched soils that formed in the sand layers (Fig. 1). Each soil probably reflects a few thousand years of surface stability during which erosion, weathering, and additions of eolian dust modified the dune surfaces in a setting similar to that of the present day.

Soil pits in a dune crest (VP-1) and a swale (VP-2) provide more detailed information on the near-surface soils and sediments (Fig. 1). Data from these pits combined with auger-hole data suggest that the upper 40-50 cm represent a modern soil forming in eolian sand that overlies a better-developed soil (consistently represented by a silty textural B horizon). The contact between these two soils is not an obvious unconformity, but is clearly indicated by trends in particle size and magnetic susceptibility. These relations suggest that a former stable land surface represented by the textural B horizon has been buried by a thin, relatively young sand; soil formation has partially blended or “welded” the two soils.

The coarse-grained quartz OSL ages of eolian sand units indicate three episodes of eolian activity (Fig. 1). Ages of 8.6 and 3.7 ka were obtained for the uppermost thin sand and ages of 13.1 and 7.2 ka for the first buried sand. The older ages for the two units were obtained on samples from the dune-crest pit, whereas the younger ages were from the dune-swale pit. We are uncertain whether the differing ages represent long spans of time (4,000-5,000 years) when the dunes were active, or whether the younger ages in the swale site reflect colluvial reworking of sediment derived from the dune crests (more likely). An OSL age of ~41 ka, obtained on a truncated dune sand exposed in an arroyo cut (site 8U-14), indicates an earlier period of eolian activity. Charcoal at the base of the alluvium overlying the dune sand gave a calibrated 14C age of ~5.3 ka. The alluvium gave OSL ages of ~12 ka for the coarse-grained quartz and ~8 ka for the fine-grained quartz. This age difference, with apparently older alluvium overlying younger charcoal, is probably caused by alluvial reworking of sediment from older dune sand.

The fine-grained quartz fraction of the dune sand yielded consistently younger OSL ages than the coarse-grained quartz (Fig. 1). Both coarse- and fine-grained OSL ages were of high precision, with analytical uncertainties of < 5%. Although these ages (with one exception), were indistinguishable within 1 or 2 σ analytical uncertainty, the slightly younger fine-grained ages may result from younger silt infiltrating the dune sands when they stabilized.

One deep arroyo cut exposes over 4 m of graben-fill deposits consisting of a thin cap of eolian sand overlying alluvial sand and pebbly sand (9U-21, Fig. 1). Three episodes of alluvial deposition are defined by three buried soils. The uppermost eolian sand is similar in character and degree of soil development to the modern soil and thin eolian sand in the nearby dune field; the youngest buried soil, formed in alluvium, is also similar to the youngest buried soil in the nearby dune field. Deposits at the base of the second alluvial unit gave very similar OSL ages (within 1 σ analytical uncertainty) of about 26 ka for the coarse- and fine-grained quartz.

At an archaeological site outside of CNP, stratigraphic relations among thin eolian sand units and paleosols (Shearin et al., 2000) suggest a geomorphic history similar to that of the main study area. These pits exposed two thin sheet sands overlying a buried soil formed on weathered bedrock, locally admixed with older eolian sand. The uppermost sheet sand is very young and in part historic based on the presence of a piece of glass at the base of the sand. A buried soil formed in the underlying sheet sand is weakly developed and similar in character to the surface soil in the main study area. The eroded paleosol formed on the underlying bedrock is moderately developed and has properties similar to those...
of the youngest buried soil in the dune field. A similar paleosol formed in alluvium was encountered beneath 145 cm of slopewash deposits deposited in a nearby small drainage.

![Figure 1. Stratigraphy and ages of dune sand and alluvium in study area from selected auger holes and outcrops.](image)

**Paleoclimate Data**

Pollen abundances in samples from the dune-swale pit suggest significant climate changes over the period of deposition. The lowermost sample (110-115 cm depth, or 20-25 cm below the coarse-grained OSL age of 7.2 ka) contains abundant grass and common arboreal pollen (13 %; Pinus and Juniperus,) and Artemisia (11 %). The middle part of the record, between about 45-95 cm depth, is characterized by very high Cheno-Am percentages, and sparse (<5 %) to absent arboreal pollen. The lowest arboreal pollen percentages and lowest Artemisia:Cheno-Am ratio (an indicator of summer precipitation, e.g., Shafer, 1989) are both coincident with the base of the surface soil, just below the coarse-grained OSL age of 3.7 ka. These data suggest that a Cheno-Am-rich flora reflecting aridity accompanied landscape destabilization and renewed eolian or slopewash deposition at this site. Most notably, Pinus and Juniperus expand greatly between 14 cm depth and the surface, accompanied by a doubling of the Artemisia:Cheno-Am ratio. The presence of a probable Tamarix (tamarisk) grain in the Bw horizon (5-14 cm) indicates that the uppermost sediments in the dune swale are less than 100 yr old.

These pollen data are consistent with pollen analyses from the archaeological site (Shearin et al., 2000). Here, the buried soil formed in sand and weathered bedrock as well as the soil buried beneath the slopewash contained relatively abundant Artemisia, grass, and arboreal pollen, whereas the soil formed on the pre-modern sand sheet and the slopewash deposits contained little Artemisia, grass, and arboreal pollen and abundant Cheno-Am pollen. Taken together, the two sets of pollen analyses and the OSL ages suggest a relatively wet period in the study area before 7-8 ka and drought during the middle to late Holocene, between about 7 and 3 ka.

**Conclusions**

The stratigraphy and ages of deposits in the dune field and associated alluvium combined with pollen data suggest these preliminary interpretations: (1) An episode of eolian activity at about 40 ka. (2) A wetter or more stormy period that produced alluvial deposition (graben fill) from some time prior to about 25 ka until before about 13 ka. (3) A second period of eolian deposition around 13 ka in the dune field. (4) A wetter period around 10 ka, during which soils formed relatively quickly by infiltration of eolian silt and clay, and trees and grass were common on the landscape. (5) A dry and
perhaps stormy period between about 9-8 ka and 4-3 ka, during which some parts of the dune field were reactivated, slope wash accumulated, alluvial erosion and deposition occurred, and more desert-like vegetation was dominant. (6) Increasing moisture after about 3 ka, during which dunes stabilized and trees and grass reoccupied favorable settings.

These paleoclimatic interpretations are consistent with conclusions from previous studies of pollen and packrat middens on the Colorado Plateau. The closest locality to the study area that lies at a similar altitude is Cowboy Cave, west of the study area. The record from this cave indicates increased moisture from an enhanced monsoon from about 13.5 to 8.9 ka, followed by inception of more xeric conditions (Spaulding and Peterson, 1980, cited in Shafer, 1989). The ratio of Artemisia to Cheno-Am pollen increased sharply during the enhanced monsoon period. The period of mid-Holocene aridity apparently ended with a rise in effective precipitation some time between 4 and 2 ka (e.g., Sharpe, 1991; Weng and Jackson, 1999).

**References**


Optical dating chronologies of dune reactivation in the south-eastern Arabian Peninsula

Stephen Stokes and Richard Bailey, Oxford Luminescence Research Group, School of Geography and the Environment, University of Oxford, Mansfield Road, Oxford OX1 3TB, UK

Abstract

The optically-stimulated luminescence (OSL) signal within quartz can be enhanced by thermal transfer during pre-heating. This may occur via a thermally-induced charge transfer from low temperature traps to the OSL trap. The effect, as empirically measured via recuperation tests is observed to be negligible for old samples but may be a significant problem for younger deposits. The prospect of thermal transfer remains a major concern in the dating of young samples as the process of thermal decay and transfers of geologically unstable traps (typically in the TL range 160-280ºC) may in result be incomplete. Upon pre-heating such a sample might undergo thermal transfer to the dating trap and result in a De overestimate. As a result, there has been a tendency for workers to adopt less rigorous pre-heats for young samples.

We have investigated the pre-heat dependence of numerous samples from various depositional environments at five temperature pre-heats (200ºC, 220ºC, 240ºC, 260ºC, 280ºC), employing a single aliquot regeneration (SAR) protocol. SAR De,s were also calculated for 45 modern/young samples of different depositional environments and compared with previous multiple aliquot additive dose (MAAD) data. Results demonstrate no significant De dependence upon pre-heat temperatures, indicating that thermal transfer is negligible. A close correspondence between MAAD data and the current SAR data for these samples is also illustrated, reinforcing the applicability of SAR for the determination of modern and young sediment De,s.

We describe the results of a detailed analysis of a suite of samples collected from Sahelian West Africa as a demonstration of the potential of optical dating for young (i.e. contemporary) aeolian sediments. The Sahelian Margin of West Africa is an area of recent environmental catastrophe and human suffering via the food shortage and land degradation consequences of prolonged drought. The propensity of this region to suffer drought has been related, using environmental data collected during the period of instrumental records, to a combination of low mean annual rainfall levels and significant departures from the mean which relate to sea surface temperature anomalies in the adjacent tropical Atlantic Ocean. The potential impact of increasing population density in the area has also been identified as a potential contributing factor to the desertification of the region which has occurred over the last c. 30 years. Despite the significant environmental and human consequences of such droughts there is a paucity of long term environmental data for this region. Aeolian dune reactivations in this area are a potentially highly useful environmental archive of past periods of extended drought conditions which may have resulted in localised or widespread dune reactivation. Here we describe the initial results from an ongoing programme of research which seeks to develop a detailed record of past dune reactivations in Mali and Nigeria. We find evidence for repeated Holocene dune reactivation events and a significant number of reactivations which commenced at the time
of onset of the last major drought cycle in the early 1970s. We obtain ages as young as 20-30 years for some significant dune units (thickness up to 1m) and describe the results of a series of experiments to establish the significance of pre-heat temperature on Single Aliquot Regeneration (SAR) equivalent dose determinations and recycling ratios. Optical dating of sand sized quartz could provide a very useful tool for palaeogeographical mapping of ancient and historical dune reactivations in this region and elsewhere.
Aeolian Geomorphology of the Salt Basin, West Texas

V.P. Tchakerian, Department of Geography and Geology & Geophysics, Texas A&M University, College Station, Texas 77843 (vatche@geog.tamu.edu)

P.R. Rindfleisch, Department of Geography, Texas A&M University, College Station, Texas 77843 (soilgeo18@hotmail.com)

J. Given, Department of Geography, Texas A&M University, College Station, Texas 77843 (jgiven@tamu.edu)

D.E. Wilkins, Department of Geosciences, Boise State University, Boise, Idaho 83725 (dwilkins@boisestate.edu)

Introduction

This paper examines the aeolian processes and landforms of the Salt Basin in West Texas. The Salt Basin, a northwest-southeast trending faulted block, covers an area of 22,000 km² and is located about 160 km east of the city of El Paso. Two major types of aeolian deposits mantle the study site: (1) quartz based sand dunes and sheets and (2) gypsum based dunes. The quartz based dunes and sands consist of two major sub-types: (a) older stabilized aeolian deposits (two to three depositional units) seen only in arroyo exposures and, (b) more recent nebkha dunes, interspersed with cryptobiotic soils. Based on stratigraphy and radiocarbon dating from a number of arroyo exposures and on geomorphic and granulometric studies from the nebkhas and the gypsum dunes, the oldest aeolian units were most likely deposited during the mid-Holocene (about 6400 yr BP) and rest on top of older carbonate-cemented, gypsum-sand aeolianites and fanglomerates. From the mid-Holocene through late-Holocene time (c. 400 BP), a number of major aeolian depositional pulses are recognized, terminating with the stabilization of the quartz-based dunes and the formation of nebkhas with cryptobiotic soils. Present aeolian activity is limited to the gypsum dunes and controlled primarily by evaporative pumping of groundwater and surface production of gypsum, as well as to the deflation of existing gypsum dunes owing to anthropogenic activities such as grazing and other agricultural activities. The highly episodic nature of aeolian activity seems to be controlled largely by sediment supply, availability, storage and transport capacity, and only to a lesser degree to periods of aridity (such as the Altithermal) and/or variations in wind regime.
Climatic Factors Affecting Mobility and Stability of Sand Dunes

H. Tsoar, Department of Geography and Environmental Development, Ben-Gurion University of the Negev, Beer Sheva 84105, Israel (tsoar@mail.bgu.ac.il)

Introduction

Sand dunes are known to be: (i) free of vegetation and active (ii) partly vegetated and active (iii) fully vegetated and fixed (Wilson, 1973). Low rainfall and high potential evaporation result in sparse or nonexistence vegetation (Thornthwaite, 1931) and hence active sand dunes. The amount of vegetation on sand dunes is also checked by sand transport when winds are strong enough. Several wind erosion or sand mobility indices have been developed for various parts of the world (Ash & Wasson, 1983; Talbot, 1980), all of which are based on these two factors that influence dune mobility. The first one deals with the degree of windiness (expressed as the average annual wind velocity to the third or fourth power, or as the annual percentage of days experiencing winds above the threshold for sand movement). Most dunes will be mobilized if windiness is increased. The second factor is the vegetation growth cover that is taken as a function of the ratio between the average precipitation (P) and evaporation (E).

Considering the above two factors, an equation was developed by (Lancaster, 1988) that is widely used by many geologists and geomorphologists, to determine whether sand dunes would be active or fixed and the expected effect of climate change.

\[
M = \frac{W}{P/E}\frac{P}{PET}
\]

where: \( M \) = sand mobility index, \( W \) = percent of days during the year with sand moving winds, \( P/E \) = the ratio of mean annual precipitation to mean annual potential evapotranspiration.

The critical values of equation 1 for Southern Africa are: \( M > 200 \) for fully active dunes with no vegetation and \( M < 50 \) for inactive vegetated dunes. These values are performed very well for the Great Plains sand dunes (Muhs & Maat, 1993) but they are not in accord with the results from many other sand dunes areas in the world. There are many examples of unvegetated active sand dunes in humid areas and of vegetated fixed dunes in arid regions (Cooper, 1958; Tsoar & Möller, 1986).

The physical properties of dune sand and its effect on vegetation

Precipitation and evaporation are not very effective on sand dune vegetation. Dune sand is devoid of runoff and is known to have high rates of infiltration because of its relatively big pore spaces. As a result, sand quickly reaches its field capacity, which is less than 5%, and with abundant rainfall water infiltrates to the groundwater where plants cannot reach. A profusion of rain will only leach the sand of its nutrients. Because easy and deep percolation occurs in dune sand, moisture is stored at depths where it is protected from evaporation during the long dry periods. Hence, precipitation and evaporation have a different effect on vegetation, which grows on sand, as opposed to on other soils.

Wind power is the most important factor in sand dune mobility because of the non-cohesiveness of the sand. Wind above a certain wind power can erode sand to such an extent that it prevents seeds from germinating in the sand and stabilizing it. According to the sand transport equations (Bagnold, 1941), the sand flux (q) is directly proportional to the cube of the wind. The wind factor in equation 2 only
refers to the percent of days during the year with sand moving winds and not to the wind magnitude. In addition it does not refer to the wind directionality.

A much better index for the wind magnitude is the drift potential (DP) of the wind (Fryberger, 1979), which refers to the sand transport equation:

\[ DP = \sum q = \frac{U^2(U - U_t)}{100} t \]  

where \( U \) is the wind velocity, measured at a height of 10 m, \( U_t \) is the threshold wind velocity and \( t \) is the time the wind blew (in percent).

DP is a parameter of the potential maximum amount of sand that could be eroded by the wind during a year. DP, being roughly proportional to the rate of sand transport and the time the transport happens, is a measure of the wind power. The index of the directional variability of the wind is the ratio of the resultant drift potential to the drift potential of the wind (RDP/DP).

By analyzing rainfall data it was found that there is no correlation between rainfall and dune mobility and stability. Because of the low field capacity of dune sand, rain is effective for plants only if it falls very frequently and in amounts small enough to bring the sand moisture to field capacity level most of the time. The frequency of the rainfall can be determined by the arithmetic mean of the monthly rainfall deviation from the monthly average. There is no link between this deviation and the amount of vegetation on sand dunes.

By process of elimination, it remains that the predominant factor affecting sand dune mobilization is wind erosion, as expressed by the drift potential (DP) and the annual wind direction variability (RDP/DP). Analysis of the relationship of DP versus RDP/DP for vegetated and unvegetated dunes from several sand dunes all over the world can highlight the effect of wind power on the vegetation cover of sand dunes (Fig. 1). When RDP/DP is low, wind energy is distributed on more than one slope of the dune and the energy exerted on each slope is lower. Sand dunes in areas where the annual average rainfall is \( \geq 50 \text{mm} \) are unvegetated and mobile under wind conditions in which \( M \) (according to Equation (3)) exceeds 1. 

![Figure 1. DP versus RPD for stations with and without vegetation.](image-url)
Other factors that influence the mobility and stability of sand dunes are related to human activity. There are many examples of the destruction of vegetation by grazing, trampling and wood gathering. These actions are known as processes of desertification. On the other hand, human are also making efforts to artificially stabilize sand dunes because of their apprehension of shifting sands. Most of the coastal sand dunes in Europe have been undergoing processes of fixation for the last 200 years (Favennec, 1996).

The relationship between wind power and vegetation cover can be recapitulated by a hysteresis curve (Fig. 2). When climate changes in the form of a decrease in wind power, vegetation will start covering the sand dunes in increasing numbers as the wind power decreases below 1000 DP. However, when this process is reversed, increase of wind power over vegetated dunes will not cause the extinction of vegetation when DP increases above 1000. There is a threshold for the destruction of vegetation by tempest winds but the value of the wind power for this occurrence is not known. The artificially stabilized sand dunes along the coasts of Europe have DP values above 1000 and they are in the upper reverse side of the hysteresis curve.

\[ M = \frac{DP}{1000 - \left( \frac{RDP}{DP} \right)} \]  \hspace{1cm} (3)

Figure 2. Hysteresis curve related to changes in wind power and vegetation cover.
References


Development and Morphology of Falling Dunes, Northeast Kuwait

A. Al-Enezi, KISR-EUD, Kuwait P.O. Box 24885 (E-mail: aenezi@safat.kisr.edu.kw)
K. Pye, RHBNC-SPME, Egham, Surrey, TW20 0EX (E-mail: k.pye@gl.rhbnc.ac.uk)
R. Misak, KISR-EUD, Kuwait P.O. Box 24885 (E-mail: rmisak@safat.kisr.edu.kw)

Introduction

The interaction of topography with sand-laden winds is responsible for the development of different forms of aeolian landforms. In the literature, these types of aeolian landforms have different terminology, being termed “sand shadow” or “sand drifts” by Bagnold (1941), “fixed” dunes” by Howard (1985) and Recently, “topographically anchored sand dunes” by Cooke et al. (1993) and topographical controlled sand dune by Lancaster and Tchakerian (1996).

In Kuwait, the continually transported sand across the surface under the influence of the prevailing northwesterly wind, result in the formation of different aeolian landforms including falling dunes that are the most common features in the northeast of the country.

The main objective of this study is to identify factors controlling the development and distribution of the falling dunes in Kuwait with emphasis on the influence of the topography on the morphology and the size of these dunes.

Methods

Aerial photographs of 1992, (scale 1:29000), and SPOT image of 1995 (scale 1: 100000) were analyzed to identify the distribution of the falling dunes along the escarpment and presented on a map of scale 1:50000.

The morphometric characteristic of selected falling dunes was identified using total station theodlite, where measurements of the dimensions of the dissecting wadis was through aerial photos study.

Results

Distribution of Falling Dunes

Falling dunes are the most common aeolian landforms in the northeast of Kuwait. They are associated with Al-Atraff, Al-Mutla, and Jal Az-Zor escarpment and occur at the northwestern corner of Umm Al-Rimmam Depression. Based on the aerial photographs mapping, there are four principal zones of falling dunes that can be distinguished along the escarpment of the study area. They are located downwind of regional aeolian sand pathways trending in a northwest – southeast direction with variable length and width. Their alignment is related mainly to microtopographical control and the prevailing northwesterly wind.
Falling Dune Morphometry

The falling dunes vary considerably in size. The length of the falling dunes along the escarpment ranges from 38 to 383 m and the width from 7 to 85 m, while they attain a maximum height of 12m. Despite of the importance of the continuous supply of sand, this range of falling dunes size is controlled chiefly by combination of the height of the cliff and in part the size of the wadis, where the falling dunes mostly confined (Fig. 1 & 2).

![Figure 1](image1.png)

**Figure 1.** Correlation between length of falling dunes and wadis along escarpment.

![Figure 2](image2.png)

**Figure 2.** Bivariate plots of cliff height Vs length of the falling dunes.

Falling dune morphology

Most falling dunes along the escarpment accumulate in a simple form, existing as a single finger like deposit oriented in the NW-SE direction. Within this general form, there is slight morphological variation. This variability can be explained by the factors related to the changes of the hilly terrain morphology (Fig. 3). There are three major geomorphologic association between topography and morphology of the falling dunes. These are: falling dune attached to cliff headland; (2) falling dunes confined within wadis of different orientations;
(3) falling dunes blocked by an isolated hill downwind of the escarpment.

![Diagram of falling dunes with different forms related to topography](image)

**Figure 3.** Different forms of falling dunes related to the effect of topography with emphasis on the airflow over the escarpment. (a) falling dunes with sinuous crest attached to cliff headland outside of a wadi. (b) falling dunes with straight slipface attached to headland close to the side of a wadi. (c) falling dune with curved slipface attached to headland close to the side of a wadi. (d) falling dunes without linear pattern located within wadi oriented parallel to the sand-laden wind and accumulated at the side wall of the wadi. (e) crescent-like falling dune within wadi oriented at angle to the sand-laden wind. (f) diverted falling dune around isolated hill downwind of the cliff headland.

**Discussion**

**Development of Falling Dunes**

The comparison of aerial photographs of 1972 and 1992, indicates the recent development of these dunes along the escarpment. This changes in a relatively short time period was chiefly as result of a combination of climatic conditions and increased availability of source by human activity under favorable topographic and aerodynamic conditions.

**Climatic Conditions**

The wind regime and rainfall are the most important climatic elements that influence
dune development in Kuwait. Aeolian processes were highly active during a dry period from 1981 to 1992. In that period the rainfall was irregular and below the average (110 mm/year). Strong prevailing northwesterly winds during this drought period exerted further influence in activating aeolian processes in Kuwait. This almost unidirectional nature of the effective wind in Kuwait is important in the development of falling dunes since they can not survive where there is a large variation in wind direction (Bagnold, 1941; & Cooke et al, 1993).

**Human Activities**

The development of falling dunes in Kuwait accompanied the drought period from 1981 to 1992, when the aeolian processes were highly active and the supply of sand was plentiful as result of land misuse by human activities including overgrazing, off-road traffic, desert camping and gravel quarrying. In addition, the 2nd Gulf War and it consequences accelerated the development of falling dunes through destruction of vegetation cover and natural surface sediment armor.

**Topography**

Falling dune in Kuwait developed as result of interaction of the prevailing northwesterly wind with Jal Az-Zor escarpment. Their development significantly effected by the orientation and slope angle of the escarpment. It can be observed that well developed falling dunes associate mainly with steep sided escarpment and there was no falling dunes were developed along the escarpment with gentle slope (less than 30°).

Along Jal Az-Zor escarpment most of the falling dune accumulate in the zone where the escarpment oriented normal to the prevailing northwesterly. However at the far northeast of the study area the escarpment swing to southeast direction running parallel to the prevailing wind there were no falling dunes developed except at Ras Al-Subiyah where the escarpment oriented slightly at angle with the prevailing wind.

**References**


Remobilisation of a parabolic dune in Kennemerland, the Netherlands

S.M. Arens, Bureau for Beach and Dune Research, Iwan Kantemanplein 30, 1060 RM Amsterdam, The Netherlands (E-mail: Arens@duinonderzoek.nl)

Q. Slings, nv PWN North-Holland Water Supply Company, van Oldebarneveldweg 40, 1901 KC Castricum, The Netherlands (E-mail: Rienk.Slings@pwn.nl)

C.N de Vries, nv PWN North-Holland Water Supply Company, van Oldebarneveldweg 40, 1901 KC Castricum, The Netherlands (E-mail: Cees.d.Vries@pwn.nl)

Introduction

Large parts of the Dutch coastal dunes consist of parabolic dunes, which developed in episodes between 800 and 1850 AD. It is not clear yet which conditions were responsible for the sudden development of these dunes. Klijn (1990) argues that coastal erosion of the barrier landscape, induced by sea level rise and increased storminess must be the main cause of the massive input of sand and subsequent transgressive dune formation. Due to stabilizing activities, but possibly also to climate change, all these dunes were stabilized in the past two hundred years. At present, aeolian activity in the coastal dunes is restricted to small-scale features like blowouts.

Parabolic dunes are believed to be features in a transitional landscape between mobility (transgressive dunes) and stability (vegetated dunes). Several authors describe transitions from stability to mobility and reverse due to human influence (grazing, sod and wood cutting; planting) and/or climate change (decreased/increased precipitation, increased/decreased windiness). Anthonsen et al. (1996) describe the transition of a dune in Denmark (Råbjerg Mile) from crescentic to parabolic, probably due to minor climate changes. Hesp (in press) describes transitions from parabolic dunes to transgressive dunes and vice versa in New Zealand, due to human activity. Tsoar and Blumberg (in press) ascribe the transition from barchanoid to parabolic dunes in Israel to a decrease of human pressure.

In the Netherlands several experiments are carried out to restore aeolian dynamics. In 1998 a complex parabolic dune in the Kennemerduinen, The Netherlands, was remobilised by removing vegetation (pine forest) and soil. The aim of the experiment is to investigate whether large-scale dunes can be mobile in the present climate and measures like these are successful for durable dune management, ensuring periodic rejuvenation of the landscape by natural processes.
Figure 1. Upwind side of the devegetated crest (left) and downwind side (right)

Figure 1 illustrates the dune, two years after the reactivation. The system represents two different situations. The northern part consists of a 150 m wide parabolic shape, which was completely devegetated, including crest and slipface (Figure 1). The southern part consists of a narrower, 50 m wide parabolic shape, of which only the stoss slope was devegetated. This paper discusses some of the results and the differences in development for these two situations.

Methods

Each month erosion and deposition is measured using erosion pins. These are established in 5 transects and on several significant places over the area. Absolute heights of the erosion pins and transects are measured with a lasertheodolite and with GPS equipment yearly. Results give insight in slope development and rates of processes. In 1999 and 2001, aerial photographs at a scale of 1:2500 were taken of the area. These were used to map vegetation establishment, activity of processes and geomorphological development. Wind data of the Royal Meteorological Office are used to get insight in the wind energy and to compare this to ‘average’ wind conditions.

Results

Wind data (KNMI, 2000; 2001) reveal that since the start of the experiment storminess was less than average. Yearly rainfall was considerably more than average, with total amounts of around 1000 mm in 1999, 2000 and 2001. Despite these ‘unfavourable’ conditions for sand transport, considerable changes occurred in the area and large volumes of sand were transported.

In the deflation plane most of the surface is erosive, although locally the surface has reached the groundwater table and vegetation is beginning to establish (mainly *Ammophila arenaria* in clumps, *Carex arenaria* and mosses, *Bryum spec.*, *Juncus articulatus*, *J. alpinarticulatus*, *Carex trinervis*). The average vegetation cover is less than 10%, but on the trailing ridges vegetation cover is increasing up to 50%. Cross sections 1 and 2, which cross the deflation plane, show mainly erosion, with a maximum of 0.33 m between 24/09/99 and 12/12/01.

The presence of a large, bare deflation plane in front of the dune results in a large input of sand onto the dune. Partly this is deposited in front of the dune, partly on the lee face, and, in the devegetated case, partly several 10s of meters behind the back of the dune.
In the completely devegetated case, the (bare) crest is eroding, while deposition takes place in front of the dune and at the back (Figure 2a). A maximum erosion of 2.55 m on the crest was recorded since the start of the experiment, with an average of 0.93 m, measured over 8 pins. The crest is irregular and several small blowouts have been developing. Because of the blowouts in the crest and 3-dimensional flow around the back of the dune, no clear slipface is developing. The characteristics of the dune change from parabolic into transgressive, dome shaped. Parts of the crest are still covered with old roots, which may reduce the erosion of the crest. There is hardly any vegetation development on the crest.

In the partly devegetated part, the crest has grown more than 1 m because of deposition in the vegetation on top (Figure 2b). The shape of the dune is more isolated and comparable to a large trough blowout. The lee slope is a clear slipface. Most of the sand that passes the crest is deposited on the slipface. Only during very strong winds, sand is moving in suspension (jettation, see Arens et al., in press) and is deposited in the vegetation further downwind. Apparently most of the sand that accumulates on the slipface is derived from the deflation plane. The slope has not changed, and mainly acts as a transport slope, over which sand that is eroded from the deflation plane is transported and finally deposited on the crest and in the slipface.

Figure 3 shows the mobility of the crest line and the (toe of the) leeface. The leeface moved between 5-12 m (northern part) and 0-7 m (southern part) in the period April 1999 to July 2001. From profile 3 it appears that the displacement of the crest and slipface is 2.5 m respectively 4.3 m between November 2000 and November 2001.
Discussion

Large-scale dynamics are possible in the present climate in the Netherlands, even when conditions are less windy and more rainy than average. The relatively limited extent of dynamic features in the Dutch coastal dunes is obviously related to the stabilising efforts of human beings. However, it is likely that also the large-scale dynamics in the past were the result of human action (see Hesp, in press), although Klijn (1990) argues that this cannot be the sole explanation for the large-scale mobility these dunes experienced in the past.

The parabolic dune adapts to a new situation. The dune probably developed in a situation with a vegetated deflation plane and a crest covered with marram. Most of the transport (erosion) probably occurred on the windward slope and sand was mainly deposited on the crest. Consequently the slope was steep and concave. In the new situation, with a large source of sand, and limited vegetation growth, the dune transforms into a transgressive dune, with a long and smooth stoss slope. Probably, the crest erodes until the ratio between stoss slope length and height is in equilibrium.

In a dune landscape, mobility is governed by sand supply (availability), wind energy and vegetation characteristics (e.g. Nishimori and Tanaka, 2001). There is some threshold for transition from mobile to stable dunes and the reverse. Apparently, this threshold reflects a range of conditions, in which parabolic dunes are characteristic features. Depending on local variations, particular spots are mobile or stable. Thus, in one and the same landscape, small-scale transgressive dunes, parabolic dunes and stabilised dunes can coexist.

Conclusions

The results prove that large-scale aeolian dynamics can be restored in the Netherlands, by removal of vegetation and soil, even with the limited wind energy and the relatively large amounts of rainfall that we experienced in the last 3 years. The type of dune that develops depends on the presence and growing capacity of vegetation. Based on three years of measurements, our preliminary conclusion is that within the same area, dune types where sand transport dominates over vegetation and dune types where vegetation dominated over sand transport can coexist.

The local configuration is important for the type of slope that develops: concave,
eroding slopes when there is no upwind sand source, and long, aerodynamically adapted slopes with an upwind source.

References


Patterns of Airflow in Namib Interdunes: their characteristics and significance

M.C. Baddock, School of Environmental Science, University College Northampton, Northampton, NN2 7AL, UK (Email: matthew.baddock@northampton.ac.uk).

I. Livingstone, School of Environmental Science, University College Northampton, Northampton, NN2 7AL, UK (Email: ian.livingstone@northampton.ac.uk).

G.F.S. Wiggs, Sheffield Centre for International Drylands Research, Department of Geography, University of Sheffield, Sheffield S10 2TN, UK (Email: giles.wiggs@sheffield.ac.uk).

Interdunes are spatially important areas in many regions of dunes, and yet in relation to the amount of research that has been conducted on the dynamics of sand dunes, interdunes have been relatively ignored. It is only quite recently that airflow and sediment transport in the lee of dunes have been studied (e.g. Sweet and Kocurek, 1990; Frank and Kocurek, 1996; Walker, 1999; Nickling et al., 2002). The lack of research attention attracted by interdune dynamics is in spite of the considerable importance that their study may have for our understanding of sand dune systems as a whole.

In particular, through studying the dynamics of interdunes, our understanding of their role in controlling dune spacing may be improved. A range of theories exists to explain the interdune spacing exhibited between dunes, and each theory incorporates a different view concerning the role of the dynamics in interdunes. One theory is that wind flow patterns within interdunes act as a fundamental control on dune spacing. A contrasting idea holds that interdune dynamics have no significance for dune geomorphology, and interdunes are considered ‘dynamically neutral’ in terms of dune development.

To investigate the relative significance of interdune dynamics, the operation of key geomorphological processes within interdunes was observed in a period of fieldwork in the Skeleton Coast dunefield and northern Namib Desert in Namibia, southern Africa. Data were gathered for the patterns of airflow, turbulence and sediment transport as observed across relatively simple transverse dune interdunes, with a range of different interdune situations being investigated. This paper will present data on airflow patterns in interdunes as collected from the fieldwork, and will discuss the significance of the results in terms of the dynamics of the sand dune system as a whole.

References


Combining ground penetrating radar (GPR) surveys and luminescence (OSL) dating to determine dune migration

Bristow, C.S., School of Earth Sciences, Birkbeck University of London, Malet Street, London WC1E 7HX c.bristow@ucl.ac.uk

Bailey, S.D., School of Earth Sciences, Birkbeck University of London, Malet Street, London WC1E 7HX bsimon@blueyonder.co.uk

Duller, G.A.T., Institute of Geography and Earth Sciences, University of Wales, Aberystwyth, Ceredigion SY23 3DB geoff.duller@aber.ac.uk

Wintle A.G., Institute of Geography and Earth Sciences, University of Wales, Aberystwyth, Ceredigion SY23 3DB aqw@aber.ac.uk

Ground penetrating radar (GPR) is a geophysical technique that produces high resolution images of the shallow subsurface. Aeolian sands and sand dunes usually have a high resistivity giving good penetration to GPR signals. They also contain large structures that can be resolved in the subsurface making aeolian sand suitable targets for GPR surveys. In aeolian sand and sand dune deposits GPR reflections can be related to bedding, sets of cross-stratification, palaeosols and bounding surfaces. Radar stratigraphy and radar facies analysis can be used to interpret GPR profiles. Horizontal and dipping reflections and cross-cutting relationships, toplap, downlap, offlap and truncation of reflections are used to establish a relative chronology. In addition, the extent of bounding surfaces and depositional units can be mapped in the subsurface. Having established the dune stratigraphy and relative chronology of depositional units, sampling for luminescence dating can be accurately targeted at stratigraphic horizons defined by GPR surveys.

Examples of GPR profiles across linear dunes in the Namib Sand Sea are used to illustrate the application of GPR to dune stratigraphy. A combination of radar facies analysis and radar stratigraphy have been used to identify large sets of cross-stratification deposited when the dune was most active and bounding surfaces formed during periods of stabilisation, non-deposition or erosion. A drilling and dating campaign has been designed on the basis of the dune stratigraphy as defined by a GPR survey. Sampling is targeted at large sets of cross-stratification formed when the dunes were most active, and to bracket bounding surfaces formed when the dunes were stable or even eroded. This will yield a well constrained chronology of dune activity and stabilisation to be integrated with palaeoclimate data.

At Aberffraw, a transgressive coastal dune field on the island of Anglesey, U.K., a GPR profile has been used to define 6 stratigraphic units at the inland edge of the dune field. The younger units have been sampled using auger boreholes. Optical dates based on measurements of the optically stimulated luminescence (OSL) signal from quartz from the younger units show that sand accumulated between 1465 and 1745 AD. Dates within stratigraphic units show close agreement and allow the chronology of dune emplacement to be accurately reconstructed.
The OSL technique is used to date the movement of aeolian sand grains by calculating the time elapsed since their last exposure to sunlight. The OSL age represents a period of sand activity and mobility, as opposed to the alternative method of radiocarbon dating organic horizons in dunes which represent periods of stability.

Combining these geophysical and geochronological methods improves interpretation of dune stratigraphy, the chronology of sand accumulation, stabilisation surfaces, erosion surfaces and the history of dune mobility.
Movement of a small slipfaceless dune in Namibia

Bristow, C.S.,
School of Earth Sciences, Birkbeck, University of London, Malet Street, London WC1E 7HX
c.bristow@ucl.ac.uk

Lancaster, N.,
Desert Research Institute, Reno, Nevada nick@dri.edu

In 1977 Ed McKee marked the location of some small dunes on the northern edge of the Namib Sand Sea near Gobabeb in Namibia. We revisited two locations in 1999, 22 years later and found that one dune had changed shape and moved almost 200m while another small slipfaceless dome dune appears to have become detached from the northern end of a sinuous linear dune and moved 90m towards the ENE.

A topographic survey of both dunes was conducted using a Sokkia Set 5 Total Station. Using the field measurements it is possible to calculate the migration direction, distance and the volume of the dome dune. The dome dune has a diameter of 45m is 1m high in the middle and has a volume of 551m³, it appears to have moved 90m towards the ENE. A linear migration rate of around 4 m/yr. This compares favourably with migration rates for small dunes measured in previous work by Besler (1975) indicates around 15m movement towards the NE over 4 years between 1969 and 1973, a rate of 3.75 m/yr, further monitoring between 1973 and 1978 in Besler (1980) showed that movement slowed to 2 m yr. The migration direction is also close to the calculated resultant for potential sand transport, roughly 10° west of south, calculated by Livingstone (1989) and 213° (Lancaster 1985) calculated from adjacent interdune areas.

Taking a bulk density for dry sand of 2200kg/m³ then the dune will have a mass of 1,212,200kg, approximately 1,200 tonnes. Assuming that the dune volume has been conservative, ie. It has not increased or decreased greatly in volume over nineteen years we can attempt to estimate sand transport in kgm⁻¹s⁻¹. Where the mass of the dune is 1,212,200kg, the width of the dune is 45m, and the time in seconds is 693,992,000, this gives 0.000039 kgm⁻¹s⁻¹, or 1.2 tonnes/m/yr. This compares favourably with the measurements of dune encroachment by Ward and von Brunn (1985) who give values between 0.08 to 0.69 m³/m/yr (Ward and von Brunn 1985) which is equivalent to 0.000006 to 0.000048 kg/m/s.

This represents the net migration, in reality the gross sand transport is probably much greater. The vector sum (resultant) potential sand transport estimates for the northern Namib calculated by Lancaster (1985) are 63 tonnes/m/yr, equivalent to 0.002 kg/m/s. These values are two or three orders of magnitude greater than the values based on measurements of dune migration indicating that dune migration is only a small part of total potential sand transport.

References


Besler, H., 1980. Die dunen Namib: Entstehung und dynamik eines Ergs. Stuttgart 1980. (pg 136 Fig.7)


Evolution of Fertility and configuration of Aeolian soils in the processes of shifting sand fixation in Tengger desert

Duan Zhenghu*,** Xiao Honglang * Wang Gang**

* Shapotou Desert Experimental Research Station, Cold and Arid Regions Environmental and Engineering Research Institute, Chinese Academy of Sciences; 260 Donggang West Road, Lanzhou, 730000, P.R. of China; and ** State Key Laboratory of Arid Agroecology, Lanzhou University, 298 Tianshui Road, Lanzhou, 730000, P.R. of China

Abstract

With the succession from artificial plants to natural vegetation, the environment of soil formation, circulation of soil material were influenced in the processes of shifting sand fixation and the mean of soil particle size changed from 0.2 mm to 0.08-0.14 mm. The capacity of available soil water enlarges five times. Infiltration of soil water came to a close because of the increase of soil water capacity and the change of redistribution of soil water in profiles. Soil microorganisms grow out of nothing and evolved from simple to complex. Interaction of all mentioned-above processes obviously brought about accumulation of soil fertility, evolution of soil profiles and development of the profiles towards zonal characteristic. The difference of micro-topography is closely relative to redistribution of material and energy in the soil formation.
Networks dunes in southeastern Tengger Desert: morphology, sediment, and dynamics

Ha Si† China Center of Desert Research at Beijing Normal University, Beijing, P.R.China, 100875 (E-mail: hsed@bnu.edu.cn)  
†Key Laboratory of Environmental Change and Natural Disaster, the Ministry of Education of China, Beijing Normal University, Beijing, P.R.China, 100875

Introduction

Networks dunes, which are one of the commonest dune types in world desert, are characterized by network pattern resulted from the intersection of longitudinal and transverse elements. In contrast to other major dunes such as transverse (Howard et al, 1978;), longitudinal (Tsoar, 1983; Livingstone, 1986) and star dunes (Lancaster, 1989), whose mechanism and advance are well known, the networks dunes still pose many unanswered questions. In present research, the systematical field measurements on the networks dunes at southeastern Tengger Desert conducted in order to understand its morphology and dynamics. They included: measurements of wind direction and velocity for the regional wind regime and of surface airflow especially secondary wind currents as the result of the interaction of primary wind with the dune forms, analysis of grain size and sorting of surface sediment and spatial variation, identification of aeolian cross stratification types in natural exposure and trenching part.

Dunes morphology and morphometry

Networks dunes at the southeastern fringe of the Tengger Desert consist of SW-NE trending main ridges and nearly vertical secondary ridges (SE-NW orientation). Main ridges’ slopes on both sides are asymmetric, northwest-facing slopes are long and gently (6-12°), southeast-facing slopes are short and steep (28-32°), dune height varies between 3-20m, and interdune depressions are less noticeable, with a spacing ranging from 30m to 170m. Their secondary ridges are superimposed on the windward slope of the main ridges, both the east and west-facing slopes are gentle in lower part (4-14°) but steep in upper part (19-30°), slipface are not well developed and restricted to upper portion of the ridges, 1-6m in height and 20-70m in spacing. According to analytical data, there is significant correlation between the heights and the spacing of their respective primary and secondary ridges (Ha Si, 1995). From the comparisons of aerial photographs of three different periods since 1950s, it can be seen that over the past 40 years the shapes of networks dunes in the region, as a whole, remained
stable. From this it follows that the networks dunes in the region are in dynamic equilibrium with modern wind regime and sand supply.

**Regional and surface wind flow over the dunes**

Networks dunes in the southeastern Tengger Desert are formed under the low-energy and bidirectional wind regime. Both the prevailing wind direction and annual and seasonal resultant sand transport directions (fig. 1) are perpendicular or oblique at a larger angle (>50°) to main ridge crest line (235°, 55°) but are oblique to the secondary ridge (155°, 335°) at a smaller angle (<40°). According to dune forms and wind regime. The annual resultant sand transport direction in the study area is perpendicular to the primary ridges and oblique to the secondary ridges with a small angle. Primary ridges are transverse dunes, while secondary ridges are longitudinal dunes. Viewed from the spatial assemblage, the secondary ridges are superimposed on the windward slopes of the primary ridge, therefore the networks dune can be regarded as complex transverse dune.

Surface airflow over the dune is the main factors controlling the erosion deposition patterns and the morphology of sand dunes. The main and secondary ridges of the networks dunes have different surface airflow patterns although they are in the same regional wind conditions. This shows that they have different mechanism of movement. The primary ridges are under the influence of transverse airflow, which occurs separation after crossing the dune crest and causes sand deposition there due to sharp reduction in wind velocity (fig. 3a), finally forms the large scale slip face and leads to the sand dune migration in the prevailing wind direction or resultant sand transport direction. Hence, the primary ridges possess the dynamical characteristics of transverse dunes. The secondary ridges are under the influences of oblique airflow (fig.3.b, fig.3.c), which occurs attached deflection on the lee slope and forms longitudinal airflow parallel to the dune crest line, in such case, the sand eroded from the windward slop is not deposited all on the lee slope (except for the upper narrow long zone) but continuous to move along the dune crest line direction with the deflected wind. Since the wind velocity over the lee slope increases with the decreasing incident angle of the primary wind, coupling with the lower threshold velocity for sand movement on a down sloping surface than on a flat surface, wind erosion takes place on the lee slope of the secondary ridges and finally leads to extension of the ridges. However, the crest lines of the secondary ridges also have lateral shifts due to the alternating influences of bidirectional winds. The wind erosion caused by oblique winds on both sides of the sand dunes and the dune extension are the crucial elements for elucidating the dynamical processes of longitudinal dune movement and this has been demonstrated by the field observations and experiments (Tsoar, 1983; Livingstone, 1986). Thus the secondary ridges have the dynamical characteristics of longitudinal dunes.
Internal structure

Internal structures of sand dunes are the direct evidences of dune deposition and development processes. There is a good correspondence between the external form, internal structures and the surface processes of the primary and secondary ridges of the dunes. The sedimentary structure dominated by the monoclonal high angle tabular cross-bedding of the main ridge is the typical transverse dune feature formed under the influences of unidirectional winds. From this it follows that primary ridges are formed under the influences of dominant wind (northwest wind). The bipolar azimuths and bimodal dip distribution tabular-wedge cross strata with a chevron-like pattern of the secondary ridges is the typical structure of longitudinal dunes. The alternating reactivation surfaces existed in the secondary ridges representing third-order bounding surface their inclination indicate the extent of erosion between two deposition events; low-angle climbing ripple lamination and upper concave divergent bedding are mainly caused by the deflection of oblique wind over the lee slope. Therefore, the secondary ridges are formed under the alternating influences of dominant wind (northwest wind) and subdominant wind (northeast wind) and they maintain their morphology under the effects of bidirectional oblique winds. The second order bounding surface separated the main ridge deposits from the secondary ridge deposits shows that the secondary ridges are the secondary dunes formed on the basis of main ridges. From this one might conclude that the networks dunes are initiated and developed as a result of the modification of transverse dunes as they extend or migrated into area of seasonally varied bidirectional wind regime through a series of form-flow interactions.

Conclusion

Networks dunes are very common in the deserts of the world but have been seldom researched. Through the studies on the regional wind, surface airflow of the dune field, and internal structure of networks dunes. It was suggested that the main ridges of the networks dunes were formed by the prevailing northwest winds, while the secondary ridges were
formed by the alternating actions of the prevailing wind secondary wind (northeast wind) on the basis of the primary ridges. Viewed from morphodynamic types, they belong to the complex dunes formed by longitudinal dunes superimposed on the transverse dunes.

![Figure 4](image)

Figure 4 Sedimentary structure of networks dunes, (a) primary ridge and (b) secondary ridge

**References**


Exceptionally coarse-grained wind ripples in the Wright Valley, Antarctica

Nicholas Lancaster, Division of Earth and Ecosystem Sciences, Desert Research Institute, 2215 Raggio Parkway, Reno, Nevada, 89512, USA. nick@dri.edu

William G. Nickling, Department of Geography, University of Guelph, Guelph, Ontario, Canada, N1G 2W1. nickling@uoguelph.ca

John A. Gillies, Division of Atmospheric Sciences, Desert Research Institute, 2215 Raggio Parkway, Reno, Nevada, 89512, USA. jackg@dri.edu

Ripples, with a surface composed of 1-2 cm diameter gravel, have been described from several localities in the McMurdo Dry Valleys of Antarctica (Ackert, 1989; Henderson et al., 2002; Selby et al., 1974), but the conditions under which they form have remained a matter of controversy.

We studied the morphology and morphometry of ripples near Bull Pass in the lower Wright Valley (77°31’ S; 161° 50’ E) and documented their sedimentary characteristics. Wind ripple height ranges from 0.06 to 0.09 m, with ripple wavelength varying between 1.7 and 3.2 m. Ripple height increases somewhat with wavelength. The mean ripple index ranges between 49 and 88, much higher than for any wind ripples reported previously. Crest strike azimuth directions of these ripples are typically NW-SE, indicating formation by winds from the southwest (down valley).

The ripples are composed of a mixture of medium–coarse sand and medium to fine gravel, with weak bedding dipping to the north-east. The surface particle size of the ripple crests is dominated by medium gravel (9.6 mm) fining slightly to the northeast to fine gravel (6.8 mm). Ripple height and wavelength appear to increase and the ripples become better-defined and more regularly spaced as grain size decreases.

Although these ripples are clearly eolian in origin, the exact mechanism by which the coarse surface particles are moved is still not known. It appears that impacts by single saltating sand grains have insufficient force to move the larger particles, and that alternative mechanisms, perhaps including multiple saltation impacts may be required.

References


Dune Sand Transport as Influenced by Direction, Magnitude and Frequency of the Erosive Winds, Ordos Plateau, China

L.Y. Liu, China Center of Desert Research, Beijing Normal University, Beijing 100875, China (E-mails: lianyou@public.lz.gs.cn; lianyou@weru.ksu.edu)
E. Skidmore, USDA-ARS, Wind Erosion Research Unit, Throckmorton Hall, Kansas State University, Manhattan, KS 66506, USA (skidmore@weru.ksu.edu)
L. Wagner, USDA-ARS, Wind Erosion Research Unit, Throckmorton Hall, Kansas State University, Manhattan, KS 66506, USA (wagner@weru.ksu.edu)
J. Tatarko, USDA-ARS, Wind Erosion Research Unit, Throckmorton Hall, Kansas State University, Manhattan, KS 66506, USA (jt@weru.ksu.edu)

Ordos Plateau is a region with extensive wind erosion, severe desertification and various aeolian sand hazards in China. In order to determine aeolian sand transport in this region, the relationship between sand transport rate and wind speed at 10min frequencies was established by field observation in both Qubqi Sand Desert and Mu Us Sandy Land. With an instantaneous spinning-cup anemometer, threshold wind speeds (2m above the ground) on mobile, semi-fixed and fixed dune surfaces were estimated in the field. Wind speeds at 0.5 and 2m heights above the ground were measured with standard spinning-cup anemometers. Synchronous with the wind speed measurement, sand transport in every 2-cm segment up to 40-cm height was measured by two step-like sand traps. High-resolution meteorological 10min average wind velocity data (10m above the ground) from the local weather stations were collected and converted to the height of 2m for calculation of sand transport potential. Aeolian sand quantity transported by the erosive winds was calculated for all speed levels in 16 directions, and annual quantities of sand transport on different dune surface types were determined by both vector operation and vector diagram techniques.

Threshold wind speed was 5-6 m s⁻¹ on shifting dune and 6-8 m s⁻¹ on semi-fixed and fixed dune surfaces. Sand transport rate increased radically with the increase of surface shifting mobility and near-bed wind speed. The sand transport rate on the shifting dune surface was higher by approximately an order of magnitude than the semi-fixed dune surface, and sand transport rate on the semi-fixed dune surface, in turn, was higher by an order of magnitude than the fixed dune. Three specific parameters - wind speed, blowing time and wind direction, were identified to be decisive for sand transport. The quantity of sand transport was affected directly by wind speed and duration, while the overall sand movement depended upon directions of the erosive winds. In the study area, erosive winds and aeolian sand transport mainly took place in springtime. The prevailing erosive wind directions were W, WNW and NW, with frequency of more than 60%, and sand transported in these three directions made up more than 70% of the total in all the 16 directions. The overall direction of sand transport was determined by the prevailing erosive winds with azimuth angles from 288.7 to 303.6°, indicating a general southeastward encroachment of aeolian sand. Wind frequency decreased as the negative power with the increasing wind speed. High magnitude strong winds had a low frequency, but they could play a dominant role in aeolian sand transport.
Toward a Genetic Classification of Aeolian Sand Dunes

Kevin R. Mulligan, Department of Economics and Geography, Texas Tech University, Lubbock, Texas, USA 79409-1014

Vatche P. Tchakerian, Department of Geography, Texas A&M University. College Station, Texas, USA

Aeolian sand dunes occur in a wide variety of forms in many different environmental settings. Although several notable attempts have been made to classify dune forms, the problem of dune classification is complicated by the diverse terminology used in the literature. More importantly, most dune classification systems fail to emphasize the genetic linkage between different dune types. In many situations dune morphologies can be represented as part of a continuum from one dune type to another.

The purpose of this paper is to describe a simple dune classification system that stresses the genetic linkage between the different types of dunes controlled by 1) autogenic processes, 2) vegetation and 3) topography. In the first case, dunes controlled by autogenic processes reflect bedform self-organization and the nature of dune morphology is largely a function of the wind regime, sand supply and time. In the second case, vegetation is considered to be an important controlling variable and dunes are classified as part of a continuum reflecting the degree of sand accumulation or deflation. Lastly, dunes are classified in relation to topography, expressed as either sloped terrain or cliffed terrain.
Numerical modelling of flow structures over transverse aeolian dunes

Parsons, D.R. Department of Geography, University of Sheffield, Western Bank, Sheffield, S10 2TN, UK (E-mail: d.parsons@sheffield.ac.uk)

Wiggs, G.F.S. Department of Geography, University of Sheffield, Western Bank, Sheffield, S10 2TN, UK (E-mail: g.wiggs@sheffield.ac.uk)

Walker, I.J. Department of Geography, University of Victoria, Victoria, British Columbia, V8W3P5, Canada (E-mail: ijwalker@uvic.ca)

Introduction

Numerical flow models have been widely applied in engineering disciplines for many years. In the last few years, there has been a proliferation of the use of Computational Fluid Dynamics (CFD) in the fields of geomorphology and hydrology (see Bates and Lane, 1998). These models enable an improved simulation of important processes providing prediction fields that allow considerable insight into the spatial distribution of these processes. CFD modelling has offered a new methodology that is complementary to traditional field and laboratory approaches. Indeed, the models can provide details of the flow field that are often difficult to measure and offer controlled conditions in which certain aspects of the experimental set up can be varied rapidly. This paper applies a CFD model to flow over transverse desert dunes and describes the sensitivity of different elements of the flow field to variations in geomorphic parameters. The model used is capable of simulating the highly turbulent reverse flow vortex in the lee of the dune and so is able to provide an acceptable solution of the downwind distance to flow re-attachment given variations in dune height, windward slope length and aspect ratio.

Methods

This paper employs the code PHOENICS™ 3.4, which is one of several commercially available CFD programs. The hybrid-upwind scheme applied in the model is only first order accurate and can suffer from numerical diffusion when flow is highly skewed relative to the grid. Nevertheless, it is more stable than higher order schemes and investigations analogous to this present one have indicated that errors due to the interpolation scheme and not likely to be significant. In this study, the two-equation k-ε model, modified by renormalization group theory (Yakhot et al., 1992), is applied. This turbulence model is recommended for simulating flows with significant mean strain and shear. For example, it has been shown to perform better in the prediction of sheared and re-circulating flows over backward facing steps (e.g. Bradbrook et al., 1998).

In order to ascertain the capabilities of the model it was initially used to predict the measured flow velocities for the wind tunnel experiment of Walker and Nickling (in press). This successful validation procedure is described in more detail in Parsons et al. (2002). Based on this validation it was deemed appropriate to use the model to test the effect of simple dune geometry variations on certain elements of the flow field. The flow elements chosen for study were:
Streamwise velocity at dune crest
Streamwise velocity at dune toe
Lee-side separation zone length

The latter of these parameters was of particular interest due to the ability of the model to predict reverse flow in the highly turbulent lee-side eddy. A Cartesian structured finite-volume approach was adopted and the dune geometry was represented in the model thorough a ‘cut-cell’ technique where the intersections of the geometry with the grid lines were determined and the areas and volumes of partially blocked cells were calculated to a high degree of accuracy. The equation formulation was modified to account for the local non-orthogonal intersection, resulting in significantly enhanced predictions. Model runs were undertaken with varying dune height, stoss slope length and dune aspect ratio.

Details of the differing dune geometries used in the model runs are shown in Table 1. All units are in centimetres and degrees (allowing testing against the wind tunnel data of Walker and Nickling, in press).

Table 1: Dune experimental details

<table>
<thead>
<tr>
<th>Experiment Number</th>
<th>Lee base length</th>
<th>Stoss base length</th>
<th>Dune height</th>
<th>Aspect ratio</th>
<th>Lee slope angle</th>
<th>Stoss angle</th>
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</table>

Results

Data presented in Figure 1 confirm the view that an increase in dune height (i.e. an increase in aspect ratio) causes a corresponding increase in the length of the lee-side separation vortex (i.e. flow re-attachment occurs further downwind from the dune crest) (Walker & Nickling, 2002). The data also show, however, that a similar result occurs if dune height is maintained at a constant but stoss slope length is decreased (i.e. an increase in aspect ratio). The different slopes of the lines in Figure 1 for changing stoss length and changing dune height indicate that the length of the separation zone is more sensitive to the latter. However, the fact that variations in these different elements of dune geomorphology produce differing results also indicates that the use of the aspect ratio to describe the structure of the lee-side airflow is flawed.
Investigation of flow velocity at the dune crest and toe indicates that both acceleration at the crest and deceleration at the toe are sensitive to changes in dune height (Figure 2). Results not shown here show a similar but less intense relationship between flow velocity and dune stoss length at the toe, with increasing stoss slope length leading to an increase in velocity.
Both these results are expected given that stoss slope angle is more sensitive to a change in dune height than a change in stoss slope length. A steepening of this windward angle leads to both an increase in crestal acceleration and deceleration in velocity in the toe region (Wiggs et al. 1996).

**Conclusion**

Preliminary analysis of CFD-derived flow structures over a transverse desert dune has shown the potential to quickly test hypotheses relating to wind flow and dune geometry. The use of the aspect ratio to describe lee-side flow structures has been highlighted as an area of concern and the differing sensitivity of the flow to changing dune height and length has been described. Further detailed analyses of shear stress and turbulent momentum are planned as well as the incorporation of more realistic dune geometries consisting of concave-convex windward slopes.

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The Effect of Vegetation Increase on the Morphology and Dynamics of the Israeli Coastal Dunes

H. Tsoar, Department of Geography and Environmental Development, Ben-Gurion University of the Negev, Beer Sheva 84105, Israel (E-mail: tsoar@mail.bgu.ac.il)

D. G. Blumberg, Department of Geography and Environmental Development, Ben-Gurion University of the Negev, Beer Sheva 84105, Israel (E-mail: blumberg@mail.bgu.ac.il)

Introduction

The southeastern Mediterranean coastal dunes, formed about 1000 years ago (Tsoar, 1990b), experienced intensive human impact until the second half of the 20th century. As a result, natural vegetation could not grow and flourish therefore, active barchan and transverse dunes were formed (Fig. 1A). Since the 1950s, however, a sporadic recovery of the vegetation brought about a change in the morphology and dynamics of the sand dunes.

Figure 1. Aerial photographs of part of the research area south of Ashdod. A. 1944. B. 1995

The aim of this study is to monitor the effect of vegetation recovery on the morphology and dynamics of the coastal sand dunes. Research was carried out in the Ashdod – Nizzanim Sand Park. The average annual rainfall in the research area is 500 mm and the average relative
deviation is 29%. The rainy season starts in October and ends in May. Winter and spring are the only seasons with strong winds above the threshold of sand transport in the southeastern Mediterranean, mostly coming from the SW - W. In summer the wind is milder and originates as a sea breeze from the NW that is much more consistent than those of the winter and spring storms. Fig. 2 shows the sand rose of the DP values. The research area is categorized as a low-energy wind environment (Fryberger, 1979), which usually fosters vegetated sand dunes (Tsoar & Illenberger, 1998).

Methods

Aerial photographs were analyzed for 12 dates from 1944, 1956, 1959, 1960, 1966, 1970, 1974, 1980, 1985, 1986, 1990, and 1995. The oldest available aerial photographs of the area were taken in 1917 and 1918. There is no observable change in the dunes and vegetation cover between 1917 and 1944. Therefore, the 1944 photograph that was used in this work represents the dune and vegetation conditions during the first half of the 20th century and probably prior to that time as well.

The analysis of these aerial photographs was done digitally, using geographic information technologies in which all of the photographs were brought into one geographic framework. All of the images were geocoded into a UTM projection using the WGS 1984 ellipsoid description of Earth. This projection resulted in metric coordinates that were easy to use for quantitative analysis. Overall, the resulting images had a ground resolution of 1 to 2 meters per pixel.

Vegetation is easily discerned in the aerial photographs. The change in the amount of area with vegetation cover was measured by looking at the brightness value of the pixels in a 560 acre area that appeared in three aerial photographs. The rate of advance for 15 dunes that were identified in all of the aerial photographs was found by determining the advance of the brink line. This was done by averaging the distance of 15 perpendicular lines drawn between every pair of brink lines. Some of the brink lines change during this period from crescent shapes opening into the downwind direction (typical to barchan-transverse dunes), to crescents opening into the upwind direction (typical to parabolic dunes).
Results and Discussion

In 1944, the dunes were of transverse/barchan type with no vegetation cover (Fig. 1A). The stable interdune areas were used for agriculture. As a result of the cessation of agricultural and grazing land-use, vegetation started to recover, first in the interdune areas (the agricultural plots) during the 1950s and then on the dune crests during the 1960s. There were few changes in the 1970s, mainly in the appearance of some parabolic forms, though slip-faces were distinct. Vegetation sprang up rapidly during the 1980s and even more so in the early 1990s, covering the slip-faces and changing the form of the transverse dunes into parabolic forms (Fig. 1B).

Results show an increase in vegetation cover from 4.3% in 1944 (most of it in the agricultural plots in the interdunes areas) to 8.4% in 1974 and to 17% in 1995 (most of it on the crest and lee sides of the dunes). The average rate of advance of the 15 dunes is shown in Fig. 3. There was a discernible decrease in the rate of advance except during 1966-1974. It is assumed that the decrease between 1956-1966, 1974-1980 and 1980-1990 is due to the increase in vegetation cover. The decrease in the rate of advance occurred simultaneously with the gradual process of transformation of barchan and transverse dunes into parabolic dunes.

Since the limiting factor for vegetation on dune sand is wind erosion (Tsoar, 1990a), vegetation would be able to germinate and sprout on those areas of the dune that have little or no erosion. According to the profile of barchan or transverse dunes, erosion on the windward slope of the dune diminishes gradually toward the crest, which is an area of neither erosion nor deposition (Fig. 4). Hence, vegetation recovers on the barchan and transverse dune crests only (Fig. 5), which starts the process of transforming these dunes into parabolic ones. The dynamics of the transverse and barchan dunes is modified once vegetation is established on the crest. Some of the sand eroded on the windward slope will be trapped by the vegetation on the crest and will create several nebikhas there. The consequent reduction in the amount of sand deposited on the slip face will cause vegetation to establish itself there as well. Only the eroded windward slope will be devoid of vegetation (Fig. 5).

The continuous erosion will change the windward slope profile from convex to concave. The now parabolic dune will advance by erosion of sand on the windward slope, which is trapped by vegetation on the crest and the lee slope. The strong bed scour on the upper windward slope undercuts the shrubs, exposes their roots and destroys the plants. Hence, the parabolic dune advances by undermining the frontal row of vegetation on the crest that is facing the wind. Shrubs located downwind trap the sand eroded by this process.

Figure 3. The average rate of advance and advance direction for 15 dunes.
Figure 4. Profiles of a barchan dune (before and after) advance at the rate of c. Note that the crest is the only area with neither erosion nor deposition.

This mechanism of advance is different from the one described in the literature for parabolic dunes (Livingstone & Warren, 1996), which refers to the anchoring of the trailing arms by vegetation and the relatively high forward advance of the central apex.

Figure 5. The coastal dunes in 1995. Note the vegetated crest and the bare windward slope

References


Advances in research on dune-airflow-sand transport dynamics: incorporating secondary flow and sand transport processes

I. J. Walker, Department of Geography, University of Victoria, P.O. Box 3050 Station CSC, Victoria, BC, V8W 3P5 (E-mail: ijwalker@uvic.ca)

Introduction

The interactions between airflow, dune form and sediment transport are complex and vary over several spatial and temporal scales. Where vegetation is absent, these interactions control dune form, spacing and alignment. To date, research on windward slope airflow and sand transport dynamics is extensive (e.g., Lancaster 1985, Mulligan 1988, Frank & Kocurek 1996a, Lancaster et al. 1996, Wiggs et al. 1996, McKenna Neuman et al. 1997, 2000) and is reviewed elsewhere (Nickling & McKenna Neuman 1999, Wiggs 2001). Though widely cited in aeolian literature, secondary lee-side flow patterns (e.g., flow separation, reversal, deflection, shear layers, and internal boundary layer redevelopment) are poorly understood with regard to their role in transport mechanics and dune maintenance. This uncertainty relates to the complexity of lee-side flow that often precludes use of the traditional wind profile approach (i.e., the Prandtl–von Kármán equation) to predict surface shear stress and resultant sand transport. The other key impedance to progress is the lack of appropriate instrumentation for precise measurement of multi-directional and sediment-laden airflows (see McKenna Neuman, this volume).

This paper reviews recent research that characterizes airflow and sand transport over unvegetated, flow-transverse aeolian dunes based on extensive field, wind tunnel and flow visualization studies (see Walker & Nickling 2002). Several new empirical models are presented explaining the behaviour and sedimentological significance of various secondary flow phenomena including: i) macroturbulent flow regions in the lee of isolated and closely spaced dunes, ii) surface shear stress variations over idealized model dunes measured using Irwin-style pressure sensors and, iii) development and sedimentological implications of three-dimensional lee-side flow structures for dune sediment budgets and migration. Recent efforts to measure lee-side sand transport (grainfall and deflected saltation) are also discussed (e.g., Walker 1999, Nickling et al. 2002) and areas for future research are identified.

Discussion

To date our understanding of airflow and sand transport dynamics over transverse dunes is limited largely to the windward slope resulting in an incomplete picture of dune sediment budgets, morphodynamics, and migration. Recent research indicates however that form-generated secondary airflow patterns (e.g., streamline compression and curvature, flow acceleration, separation, reversal, deflection, turbulent shear zones) are not merely a passive consequence of flow-form interaction. Rather, they play an active role in dune dynamics and may control dune spacing and migration. Figures 1 and 2 show the two dimensional structure of secondary lee-side flow over sharp-crested dunes (Frank & Kocurek 1996b, Walker 2000). Flow is characterized by a separation cell (E) that extends 4-10h that may extend laterally to form roller or helical vortices. A wake region of less organized turbulence extends above the...
separation cell and dissipates downstream. The upper wake (C) consists of small vortices shed from separation that generate a steep velocity gradient overlying a slower, lower wake region (D). Detailed wind tunnel measurements (not shown) show a sharp, s-shaped profile with the boundary between C and D marked by an inflection point just above dune height. This identifies a thin shear layer that enlarges over the interdune to a shear zone (H) above flow re-attachment that may be the result of Kármán vortex streets (Walker 2000). Though the magnitude of Reynolds stress varies, the extent of H is independent of incident speed, and hence, Reynolds number. Turbulence statistics (skewness) show a balance in turbulent mixing and thus, a momentum defect level dissecting this zone caused by form drag. Vertical velocity fields show a prevalence of updrafts in the near crest region capable of suspending grains beyond typical saltation trajectories (Nickling et al. 2002) while downdrafts prevail within H. Over both isolated and closely spaced dunes, flow re-attachment and internal boundary layer (I) re-development are controlled by the location of the turbulent stress maximum (G) and related downdrafts in H. Because dune size determines the location and extent of E, G, H and hence, I, it is perhaps the most important control on dune spacing where sand supply is not a limiting factor.

Fig. 1: Secondary lee-side flow patterns over transverse dunes (Walker & Nickling 2002).

Fig. 2: Model of lee-side flow regions over closely spaced dunes (Walker & Nickling 2002).

Surface shear stress (SS) responsible for sand transport over dunes is topographically-controlled by streamline curvature (which either enhances or dampens SS if streamlines are concave or convex respectively by controlling turbulent fluctuations in the flow) and flow acceleration effects (Wiggs et al. 1996). For instance, concave streamline curvature at the toe conveys turbulent structures (i.e., additional turbulent stresses) toward the bed; this despite an apparent drop in profile-derived estimates of SS at this location. Convex curvature at the crest suppresses turbulence by damping vertical motions. Though the significance of curvature effects on dune dynamics is debated, for most dunes (where h/L > 0.05), the effects cannot be assumed negligible (Van Boxel et al. 1999). Figure 3 shows a model of SS over idealized dunes based on measurements using Irwin-style pressure sensors (Irwin 1981). SS declines upwind of the dune and drops rapidly at the toe due to an adverse pressure gradient, flow deceleration, and an abrupt change in flow angle. Though this implies reduced sand transport competence, a high SS variability (CVss) indicates turbulent conditions perhaps sufficient to inhibit deposition at the toe. Thus, turbulent stresses contribute to a greater and more variable SS than is apparent from time-averaged streamwise estimates alone (Wiggs et al. 1996, Walker 2000). Up the stoss, flow accelerates and SS rises to a maximum at the crest. Flow becomes steadier as streamlines compress and streamwise
accelerations dominate the flow. Flow unsteadiness and concave curvature contribute less to SS generation with distance up the stoss and with incident windspeed. In the lee, SS drops significantly then increases rapidly 1-2h upwind of re-attachment; this despite flow expansion and deceleration in the separation cell. Flow visualization shows this is a result of strongly reversed surface flow. A peak in CVss at re-attachment indicates turbulent gustiness generated by separation-shed eddies impacting the surface; this causes the re-attachment point to wander by 0.5h and generates intermittent sand transport in this region (Walker 1999, McKenna Neuman et al. 2000).

As the IBL redevelops SS increases rapidly to 12h and CVss decreases, both approaching upwind values by 25h.

**Fig. 3**: Variations in SS over idealized transverse dunes (Walker 2000).

Most models of dune-airflow dynamics view the system as two-dimensional. However, secondary flows generate 3-d (reversed, deflected, lateral) mass and energy transfers that must be considered for interpretation of dune sediment budgets, dynamics and migration. Figure 4 shows lee-side flow response at different heights below the dune crest for various incident angles. This deflection mechanism explains the development of helical vortices that transport sand intermittently in saltation along the interdune corridor (Walker 1999). Deflection is greatest in the zone of maximum flow expansion and deceleration upwind of re-attachment causing the crest-parallel component to deflect flow vectors parallel to the crest differentially with height within the separation cell. As flow accelerates beyond re-attachment flow vectors and sand transport deflect back toward crest-normal. This simple mechanism explains, in part, why longitudinal flows are observed in the lee (e.g., Sharp 1966, Tsoar et al. 1985) and may promote oblique migration of transverse dunes under relatively transverse incident flows.

**Fig. 4**: Flow deflection mechanism based on detailed field measurements and flow visualization in lee of a transverse ridge (Walker 2000).

**Conclusions**

Recent research on airflow and sand transport over transverse dunes indicates that secondary airflow and sand transport patterns (streamline curvature, flow separation and reversal, helical vortices, shear layers, lee-side deflection) may play a significant role in dune
morphodynamics. Empirical models presented here explain relations between dune form, secondary flow and sand transport that are key to dune maintenance and migration. Further validation and refinement of these models is underway to better define the significance of these process-response relations using new, higher-frequency turbulence and sand transport instrumentation. Other related areas in need of research include: more detailed field characterization of the lee-side flow field using turbulence instrumentation; further research on effects of dune size and spacing on secondary flow and sand transport; a comprehensive study of sand transport over and in the lee of dunes using more precise and higher frequency measurement of airflow saltation and grainfall; research on the effects of incident flow angle on both stoss and lee flow fields and sand transport (i.e., the ‘fetch-effect’) on sand transport into and over dunes.

References

Late Holocene Episodic Aeolian Activity and Landscape Development in the Cimarron River Valley; Western Oklahoma

K. Lepper, Luminescence Geochronology Lab, MS J495, EES-10, Los Alamos National Laboratory, Los Alamos, NM 87545 (E-mail: lepper@lanl.gov)

G.F. Scott, USDA-NRCS, PO Box 846, Langston, OK 73050 (E-mail: greg.scott@ok.usda.gov)

Introduction

The major rivers of Western and Central Oklahoma are shallow sandy streams that flow generally northwest to southeast. Throughout the Quaternary these rivers have been migrating down a shallow regional slope towards the southwest (Madole et al., 1991). In the process the rivers leave a sequence of terraces to the northeast while reworking older terrace deposits of antecedent systems. Overprinted on this fluvial staircase is an aeolian record of dune formation and migration that reflects regional changes in sediment supply and drought. The resulting landscape is geomorphically dynamic and geochronologically complex.

The Cimarron River valley is a major feature of western Oklahoma. It enters the state from Kansas and flows to its junction with the Arkansas River. In the study area, the valley averages 40 to 50 kilometers wide, but only about 50 meters in elevation from valley floor to the north divide. On the north side of the river, ten fluvial terraces mantle the valley wall from the flood plain to the divide between the Cimarron River and the Salt Fork of the Arkansas River.

Aeolian activity within the valley of the Cimarron River in western Oklahoma has been an important factor in the development of the present landscape. Aeolian landforms such as dune fields and sand sheets are extensive on the north and east sides of the river where they mantle all or part of each of the terraces. Previous work established a variety of ages for dunes in the area and showed the consistency of the prevailing winds since the end of the Pleistocene, but did not establish a chronology of aeolian activity or study the episodic development of the aeolian landforms on the landscape.

Since significant aeolian activity is negligible in the current climate, we assume that the aeolian features are a product of past arid episodes. Historical records are lacking in Oklahoma, but records from Kansas indicate intense arid episodes in the 1800s (Muhs and Holliday, 1995). A large body of work exists that shows aridity as a recurring feature of the Holocene.

In Southeastern Major and Northwestern Kingfisher counties in Oklahoma a sequence of Quaternary terraces of the Cimarron River has been identified which includes eight distinct terrace levels (Scott, 1999). Field observations in this area have indicated that the ridge dune on the second terrace level above the flood plain of the river (Q2) contains a record of middle to late Holocene environmental change (Scott, 1999).

This paper uses a multidisciplinary approach including geomorphic surface mapping, soil/stratigraphic analysis, radiocarbon dating and OSL dating to investigate the chronology of soil formation and aeolian activation periods recorded in the Q2 ridge dune deposits as well as the rates of sediment accumulation and the spatial variability of aeolian processes during active aeolian episodes. These studies show that aeolian landforms superimposed on fluvial terraces adjacent to the Cimarron River in Major County are the product of distinct climatic conditions, in which periods of dune activity are episodic and accumulation of sediments is relatively rapid. Soil forming processes operated on the stabilized dunes during intervening wetter/cooler periods. We also demonstrate that the synthesis of geomorphic techniques, soil stratigraphy, 14C dating, and advanced OSL dating can provide high-resolution pictures of aeolian activity in the Southern Plains.
Methods

Geomorphic Methods

Surface Mapping: The geomorphic surfaces of 400 km² in eastern Major County were mapped in the field at 1:24,000 scale. Field observation, aerial photography, USDA-NRCS soil survey maps, and USGS topographic maps were used as resources to identify terraces, dune fields, and sand sheets. The mapped area extends from the Cimarron river flood plain (T0) up to the Qt8 terrace level. Two sites, Hanor and Hajek farms, were selected for intensive investigation. These two sites of about 40 ha are about 5 km apart; both are on the ridge dune complex that sits on fluvial terrace Qt2.

Dune morphology: In the study area dune type, size, slope, weathering, soil development, and vegetation vary in a progressive manner with distance from the river. Parallel ridge dunes, mostly <3m high, occur on the flood plain. On the Qt1 terrace, the small dunes are consolidated into a large, steep ridge dune, with occasional blowouts transformed into parabolic dunes. Dunes on the Qt2 terrace surface have been modified by northwest winter winds into large parabolic dunes, however, prominent ridge dune complex remains on the escarpment rising to the Qt2 terrace. Dunes on the Qt3 to Qt7 terraces are progressively more complex, with gentler slopes and more strongly developed soils.

Soil Stratigraphy

Twenty-one soil profiles were selected for sampling to represent soils from all the landforms and geomorphic positions in the study area. A truck-mounted Giddings probe was used to extract the soil cores. Soils were described to the standards of Schoenberger et al. (1998). Buried surfaces, truncated soil sequms, indicators of age and weathering, and depositional features were of particular interest in developing the soil stratigraphy. The cores were also used to identify profiles and horizons for radiocarbon and OSL dating.

Geochronologic Methods

Relative chronology techniques: In the study area, several characteristics associated with the dunes provide information to rank the dune formations from youngest to oldest. Vegetation, depth of weathering, development of the soil profile, dune pattern, position, distance to the sand source, and dune morphology all give clues to the relative age and relationships of the major dune fields. During preparation of the geomorphic surface map, these characteristics were inventoried and used to support the correlation. The types of dunes, abundance of each type, slope, shape, and height were also noted. Percentage of ground cover and percentage of the dune field cultivated were estimated from aerial photos and inventoried.

Radiocarbon Dating: Three samples were selected for radiocarbon dating. At the Hanor site a ¹⁴C sample was taken from a buried soil horizon developed in fluvial deposits of the Qt2 terrace to provide a limiting age for the start of aeolian deposition on the terrace. Two additional ¹⁴C samples from higher stratigraphic positions within the Qt2 ridge dune complex, representing a later Holocene soil forming period, were also collected – one at the Hanor site and one at the Hajek site.

About 500 grams of each sample (bulk SOM) was submitted to Beta Analytic labs in Miami, Florida. After pretreatment the remaining carbon was reduced to graphite and dated by accelerator-mass-spectrometer (AMS) ¹⁴C measurement. Analysis (by Beta Analytic) included calendar calibration and isotopic correction.

OSL dating: Seven samples for OSL dating were collected from backhoe pits in four dunes in the Qt2 terrace ridge dune complex. Samples were taken from C soil horizons with intact primary sedimentary structure (low angle cross bedding). All sample preparations were carried out under subdued lighting. Dating measurements were conducted on surface-etched quartz grains from the 125-150µm fraction using a Riso DA-15 automated OSL/TL reader. Both green-light (526±30 nm) and blue-light (470±30 nm) stimulation were used. The resulting OSL signal was measured in the UV emission range (340±80 nm).
Single aliquot regeneration (SAR) procedures were used to determine a set of equivalent doses ($D_e$) for each sample (Murray & Wintle, 2000). $D_e$ data sets ranged in size from 48 to >100 determinations per sample. In addition to the standard SAR calibration procedures, an additional “check dose” was used with each aliquot to monitor the success of the method in recovering a known dose and to represent the integrated error associated with the measurement and calibration process (Lepper, 2001).

Individual dose rates for the samples were calculated from the concentration of the radioisotopes of K, Rb, Th and U and their daughters in each horizon plus the cosmic ray dose at the sample depth. All dose rate inputs were adjusted for average water content which was taken to be 4 ± 1 %. Elemental concentrations of K, Rb, Th and U were determined at The Ohio State University Research Reactor by instrumental neutron activation analysis (INAA).

**Synopsis of Results**

The results of this investigation, which will be presented in much greater detail at this meeting (ICAR5, Lubbock TX, July 2002) and in a forthcoming paper, are summarized below:

- The buried soil formed in fluvial materials at the base of the Hanor section represents an extended late Pleistocene / early Holocene soil forming period ($^{14}$C date: 12.8-11.5 ka BP) and can be correlated to a regionally extensive paleo-surface known as the Brady soil.

- The Brady surface is overlain by aeolian deposits from at least one middle Holocene event dating to ~3.3 ka BP, the oldest OSL age at the Hajek site.

- $^{14}$C dates from the higher stratigraphic positions within the dunes at Hajek and Hanor farm indicate dune stability and soil formation between 1200 and 1600 yr BP.

- The latest period of aeolian activity began after 1000 yr BP and ended after 750 yr BP (abundant OSL ages between 775-865 yr BP)

- The modern soil formed on the Qt2 dune complex is consistent with formation times less than 500-1000 years, which is in agreement with the OSL dates for the most recent remobilization period.

- OSL dates from the Hajek site indicate a truncation in the profile that was not apparent in the soil stratigraphy -- highlighting a potential added benefit of OSL dating for aeolian field studies.

- At these sites, advanced OSL dating methods produced stratigraphically consistent ages without exception, permitting "net" depositional rates to be calculated (1.1-1.4 cm/yr).

- The color of soils developed in aeolian sands has been used in some locations as an indicator of age. In this area, soil color (variation in redness) was not an effective temporal indicator. The Buckminster dune was virtually the same age as the dunes at the Hajek site, but much redder.

**Conclusions**

The chronology of events recorded in the aeolian deposit on the Qt2 terrace of the Cimarron River in West-central Oklahoma is consistent with regional paleoclimate variations in the Osage Plains reported by Hall (1988). The aeolian activity period between 750-900 yr BP appears to be consistent with observations throughout the Great Plains (Arbogast, 1996; Muhs et al., 1996, Stokes and Swinehart, 1997, and others). Results of this investigation also suggest that the drought reflected in these deposits started earlier in Southern Plains (OK, KS) than in the Northern Plains (NB).

At these study sites $^{14}$C and advanced OSL dating methods produced coherent and complimentary results at a greater temporal resolution than radiocarbon dating by itself could have achieved. Combining geomorphic methods, soil/stratigraphic descriptions, radiocarbon and OSL dating methods enabled us to achieve a high-resolution picture of aeolian activity at these sites.

The success of the integrated methods and multidisciplinary approach used in this investigation opens the door for extended study in this area. An analysis and sampling transect across progressively
higher and older terrace dune complexes could help decipher migration and entrenchment rates for the Cimarron River in Western Oklahoma, as well as reveal other locally and regionally significant climatic events.

References


The Simulation of Dust Storm in China

Song Zhenxin, National Meteorological Centre, Beijing, 100081

Abstract

Wind erosion occurs in many arid, semiarid and agricultural areas of the world. The desert areas of China, which occupy approximately 13% of China’s total surface area, are major sources of Asian dust. The major wind-erosion areas are the sandy lands in western and northwestern China together with the extensive region regions of The Gobi desert in northern and northeastern China, especially along the basin of the Yellow River. In this paper, we analysis the geographical distribution of duststorm which was simulated by using an integrated numerical modeling system.

The purpose of simulation on dust storm is to produce quantitative predictions of wind erosion on scales from paddock to global. Our integrated wind erosion modeling system coupled the following three major components: (1) An atmospheric-prediction model, together with a land-surface model; (2) a wind-erosion model and (3)a geographic information database. The atmospheric model provides the necessary input data for the wind erosion scheme, including wind speed and precipitation. It also provide input data for the land-surface model which produces predictions for soil moisture. Dust transport and deposition are also considered in the atmospheric model. The wind-erosion model predicts streamwise saltation and dust emission rate for given atmospheric, soil and land surface conditions. The geographic information database provides spatially distributed parameters, such as soil type and vegetation coverage, for the atmospheric, land surface and wind erosion models.

Dust storms in China occur mainly in spring and winter, but most frequently in April. In spring, surface soils frozen in the previous winter become especially loose, creating a favorable condition for wind erosion. As a example, the severe dust storms of 20 March and 6-7 April 2002 were simulated. The results show the integrated model system can simulated the main characteristics of the two cases. Wind erosion model produced estimates of wind erosion intensity and patterns which are in reasonable agreement with observation. An integrated wind-erosion modeling system offers the possibility of determining wind erosion patterns on broad scales with high spatial resolution, as well as dust transport and deposition.