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Estimation of PM20 Emissions by Wind Erosion: Main Sources of Uncertainties

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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, thay 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*t) 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*t 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_i) 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 (Fv) and its size distribution. The saltation part of the model also provides Fh. Thus, the model also yields □ 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 Fv 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 □, Fv can be expressed as:

$$Fv = □(D, u^*, Z0, e_i) Fh(D, u^*, Z0, h, C) \quad (1),$$

where □ 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, Fh also depends on u*, D, Z0, 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 □ to its parameters have shown (Alfaro and Gomes, 2001) that the effect of Z0 is to increase saltation, and hence sandblasting, thresholds. The size distribution of the soil aggregates has the most important effect on the □ 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, □ becomes approximately independent of u*. As a consequence, well above threshold, Fv becomes more or less proportional to Fh, 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 Fv increases less rapidly than in the previous case. It is found (Fig. 1) that Fv approximately increases as $u*^2$.

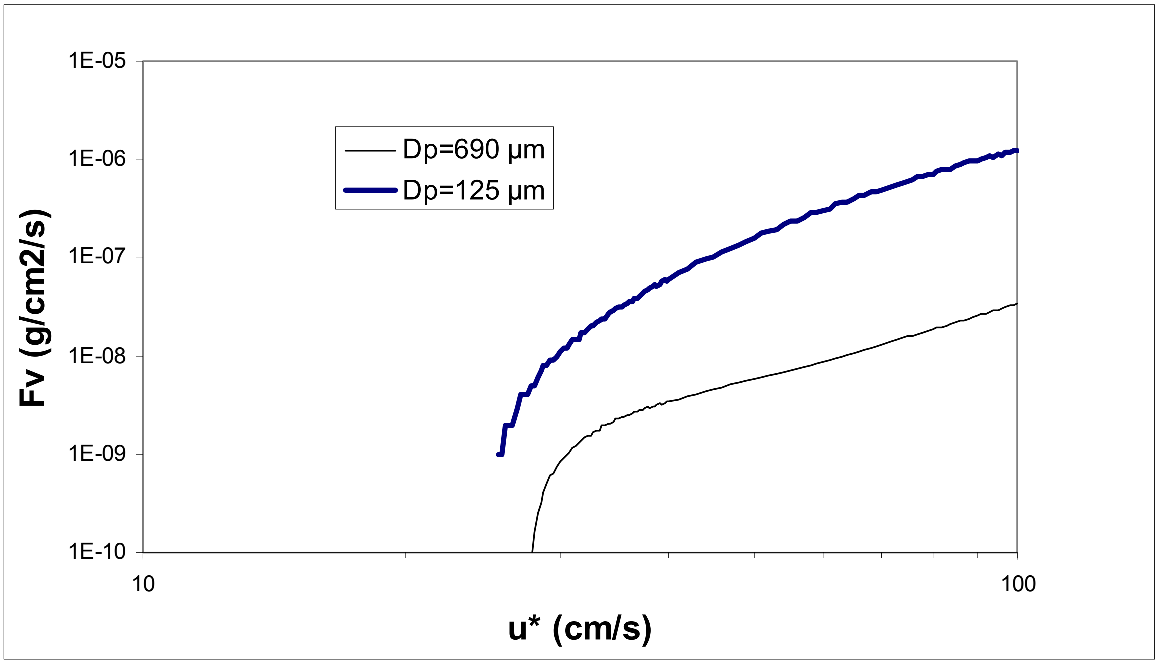

Fig. 1: PM20 fluxes computed by the DPM for two soils made of fine aggregates (125 µm, bold line), or coarse aggregates (690 µm, thin line) populations. Soils are considered as deprived of nonerodible elements.

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 Fv increases with u*. In all cases, increases quite rapidly with u* just above threshold and equation (1) implies that Fv then increases more rapidly than Fh, that is much more rapidly than $u*^3$. At higher friction speeds the rate of increase of Fv with u* progressively goes down. Though this is not mathematically true, Fv 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, Fv cannot be considered as a power function over a wide u* range explains in part the large scatter obtained when trying to plot Fv 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 micro-topography, 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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