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Modeling wind erosion flux and its seasonality from a cultivated sahelian surface: A case study in Niger

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ABSTRACT

Wind erosion can strongly affect the cultivated areas in semi-arid regions, through soil losses and/or decrease in nutrient contents. Additionally, dust emitted by aeolian erosion affects both the biogeochemical cycles and the Earth radiation budget. Modeling wind erosion and dust emission remains complex especially in semi-arid regions where vegetation interacts with the wind field and may act as a protection of the soil. An existing and widely used wind erosion model is tested in the present study to check the capacity to reproduce observations collected over a millet field and a neighboring bare plot in southwestern Niger during a three-year period. Observations of sediment horizontal fluxes and of vegetation growth and decay show that most of the eroded mass is due to major events occurring at the end of the dry season and at the beginning of the rainy season for the millet field, while erosion also occurs during the dry season for the bare soil plot. Horizontal erosion fluxes were computed with and without obstacles and compared to the measurements. Simulations were found in a good agreement with erosion measurements for both bare and millet plots, in terms of temporal dynamics, reproduction of the major events and annual quantities.

Accumulated horizontal fluxes over the millet plot were found to be much lower than over the bare plot for both observations and simulations (respectively 235 to 565 kg m^{-1} y^{-1} and 332 to 526 kg m^{-1} y^{-1} over the millet plot; and 773 to 2692 kg m^{-1} y^{-1} and 1003 to 1986 kg m^{-1} y^{-1} for the bare plot). Total values over the 3-year period also show a much stronger (by a factor 3 to 4) wind erosion on the bare plot (5365 kg m^{-1} y^{-1} for observations, 4548 kg m^{-1} y^{-1} for simulations) than on the cropped one (1357 kg m^{-1} y^{-1} for observations, 1398 kg m^{-1} y^{-1} for simulations). Modeling is therefore able to represent anthropogenic impacts on wind erosion over typical Sahelian surfaces.

1. Introduction

In semi-arid areas, the soils support crop production for local population and/or fodder production for livestock. Yet, they can undergo huge mass losses due to wind erosion (see Bielders et al., 2004 for a review). Wind erosion decreases nutrient content (Okin et al., 2006) since nutrients are generally concentrated in the surface layer of the cultivated soils (Bielders et al., 2002). Annual nutrient losses can reach quantities of the same order of magnitude than those consumed for annual millet production (Sterk and Stein, 1997). Most of the eroded sediment redeposit close to the erosion area; for instance in fallows (Bielders et al., 2004) or even in specific areas within a cropped field, creating islands of fertility (Wezel et al., 2000). However, the finest part of these sediments can be transported over long distances. When deposited, these mineral particles play a decisive role on the availability of some nutrients such as iron or phosphorus for large oceanic and continental regions (e.g. Jickells et al., 2005; Swap et al., 1992). During their atmospheric transport, these particles contribute on average to 20% of the total aerosol optical thickness. Consequently, they affect the radiation balance of the Earth and thus its climate (e.g. Tegen et al., 1997).

Over the last years, significant progresses have been made in the understanding and modeling of wind erosion processes. In particular, specific models based on original and explicit process parameterizations have been developed (Gillette and Passi, 1988; Marticorena and Bergametti, 1995; Shao et al., 1996). These models helped prioritizing factors controlling wind erosion, such as aerodynamic conditions and soil properties (e.g. Marticorena et al., 1997) or soil moisture (Fécan et al., 1999). However, when the global emissions of mineral aerosols are considered, the respective contributions of bioclimatic changes and human management remain particularly difficult to distinguish.

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30 to 50% of the total atmospheric dust load would be due to climatic changes and human activities according to Tegen and Fung (1995). More recently, Tegen et al. (2004) estimated that dust emitted from the 'land use source' contributes up to 10% to the global dust load against 0% to 50% for Mahowald et al. (2004), who used the same data but different model and methodology. Yoshioka et al. (2005) proposed a contribution of 'new' desert areas and cultivation sources of 0 to 25% based on comparison between simulations and the Absorbing Aerosol Indices (AAIs) derived from Total Ozone Mapping Spectrometer (TOMS) measurements.

Recently, simulations of Sahelian dust emissions combining a Sahelian vegetation model with a dust emission model were performed (Pierre et al., 2012). These simulations aimed at estimating the Sahelian contribution of wind erosion to the atmospheric dust content in the theoretical case of a Sahel without crops and livestock. These simulations imply a very low contribution to dust emissions of such a "non-perturbated" Sahel, in agreement with available observations (Pierre et al., 2012). However, local wind erosion measurements performed in Sahelian cultivated areas are commonly reported (under about 500 mm of precipitation), strongly suggesting that the contribution of cropped areas to local wind erosion and dust emissions should be significant in the present climatic conditions (Abdourhamane Touré et al., 2011; Bielders et al., 2002; Rajot, 2001; Valentin et al., 2004).

Thus, the objective of the present study is to investigate the capacity of wind erosion models to quantitatively simulate the wind erosion over a bare soil and a cropped plot along a seasonal cycle. A 3-year experimental dataset of surface properties and wind erosion is used to discuss the quality of the simulations.

2. Material and methods

2.1. Physical principles

Wind erosion is a power function of the wind speed, but it occurs only when a threshold in wind speed is exceeded (Bagnold, 1941). This threshold is usually expressed in terms of friction velocity, which is proportional to the wind shear stress on the surface. The threshold depends on the top soil layer characteristics and on the presence of non-erodible elements on its surface. These non-erodible elements drive the partition of the energy provided by the wind on the surface (Raupach, 1992). They absorb part of the energy while the rest of it is exerted on the intervening surface, thus modifying the threshold wind friction velocity (Gillette, 1979). The frequency of wind erosion events depends on the number of times the wind speed reaches this threshold and the intensity of wind erosion fluxes depends on the amount by which the threshold is exceeded. Thus the description of the part of the wind energy transmitted to the soil surface is a key parameter to quantify dust emissions accurately.

In order to estimate the wind erosion threshold over rough surfaces, different models of drag partition between obstacles and erodible surfaces have been proposed. These drag partition models describe how the energy provided by the wind splits between the obstacles and the bare soil surface. Generally, the models use as input data either the aerodynamic roughness length of the surface (Marticorena and Bergametti, 1995), or the obstacle drag coefficients and the roughness density (ratio of the sum of obstacle frontal surfaces, face to wind, per unit surface at the soil) (Raupach et al., 1993). These schemes have been shown to be particularly relevant for relatively low roughness density and uniformly distributed obstacles (Darmenova et al., 2009).

In order to extent the *Marticorena and Bergametti's* drag partition scheme to typical desert vegetation types, MacKinnon et al. (2004) adjusted it to in situ measurements performed in Central Mojave Desert, USA. The new expression derived is thus adapted for higher roughness lengths.

The *Raupach*'s model requires several parameters that are difficult to determine in the case of a real field (e.g. β: the ratio of obstacle to surface drag coefficient, and m: the spatiotemporal variability of the wind shear stress on the surface around the obstacle). Since the aerodynamic roughness length is an integrative value for a given surface, *Marticorena and Bergametti*'s model will be considered here as more relevant for the present study. Its skills to represent wind erosion from a bare surface and from a cultivated plot will be investigated.

2.2. Parametrizations to be tested

For a given particle diameter D_p for a surface with obstacles, the Marticorena and Bergametti (1995) drag partition expresses the threshold wind friction velocity $U_t{}^*$ as the ratio of the threshold wind friction velocity of the bare surface $U_{ts}{}^*$ for the same particle diameter on an efficient fraction f_{eff}:

$$U_t*\left(D_p, Z_0, z_{0s}\right)=U_{ts}*\left(D_p, z_{0s}\right) / f_{eff}(Z_0, z_{0s}) \quad (1)$$

where

$$f_{eff}(Z_0, z_{0s})=1\text{-}(ln(Z_0/z_{0s})/ln(\delta/z_{0s})\ ) \quad (2)$$

and, according to Elliot (1958):

$$\delta/z_{0s} = a(X/z_{0s})^p \quad (3)$$

where Z_0 is the aerodynamic roughness length of the surface with the obstacles and z_{0s} the aerodynamic roughness length of the bare soil between obstacles; δ is the height of the Internal Boundary Layer (IBL) that is assumed to develop between the roughness elements, and X is related to the distance downstream of a given obstacle.

Marticorena and Bergametti (1995) validated expression *(1)* by comparison to portable wind-tunnel measurements on various non-vegetated erodible sites of the United States (Gillette et al., 1982; Nickling and Gillies, 1989).

In expression *(3)*, p can be taken equal to 0.8 according to Elliot (1958) and Marticorena and Bergametti (1995). These last authors also determined a mean value of δ by computing the efficient fraction for various size and density of obstacles. They also concluded that a constant value of $X = 10$ cm and $a = 0.7$ (following the corrected version of this parameterization, as in Darmenova et al., 2009) gave satisfying agreement with Marshall's (1971) measurements. Finally:

$$f_{eff}=1\text{-}\left(\left(ln(Z_0/z_{0s})/ln\left(0{,}7*(10cm/z_{0s})^{0.8}\right)\right.\right. \quad (4)$$

with Z_0 and z_{0s} in cm.

As mentioned above, MacKinnon et al. (2004) proposed an adaptation of this parameterization to extend its use to rougher vegetated surfaces. This study is based on measurements from vegetated sites in the Mojave Desert (USA), where porous vegetation can be dense without preventing the wind shear stress to reach the surface (Wolfe, 1993). Whereas the largest value of the aerodynamic surface roughness considered by Marticorena and Bergametti (1995) and Marticorena et al. (1997) was about 0.2 cm, values reported by Wolfe (1993) reach 3.8 cm and even 7 cm for one of the 11 study sites of MacKinnon et al. (2004). These authors have adjusted the drag partition scheme by choosing a lower and an upper bound of threshold friction velocities encompassing both *McKinnon*'s and *Marticorena et al.* values. These bounds correspond respectively to a smooth surface of the most erodible particle size (60 to 120 μm in diameter, $U_t^* = 21.7$ cm s^{-1}), and to a threshold friction velocity of 100 cm s^{-1} for an aerodynamic surface roughness of 10 cm. They finally obtained a value of 12,255 cm for parameter X:

$$f_{eff}=1\text{-}\left(ln(Z_0/z_{0s})/\ ln\left(0{,}7*(12255\ cm/z_{0s})^{0.8}\right)\right. \quad (4')$$