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Aeolian process-induced hyper-concentrated flow in a desert watershed

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SUMMARY

Ephemeral desert channels are characterized by very high rates of sediment transport during infrequent flood events. Here we show that aeolian process-induced hyper-concentrated (AHC) flows occur in the Sudalaer desert watershed in the Ordos Plateau of China, which primarily transport 0.08–0.25 mm non-cohesive aeolian sand and have a peak suspended sediment concentration of $1.1–1.4 \times 10^6$ mg l^{-3}. Aeolian sand supply and storage in the channel play a crucial role in causing hyper-concentrated flow. Our results indicate that non-cohesive aeolian sand can be entrained from the bed and suspended in the turbulent flow when the channel bed slope exceeds a critical threshold (0.0003). We also show that if the frequency ratio of wind-blown sandstorms to rainstorms T_w/T_p exceeds $\beta(\gamma - \gamma_0)/\alpha\ (P/V^3)\ (A/L)$ (where α is the wind-blown sand transport coefficient, β is the runoff coefficient, $\gamma - \gamma_0$ is the increase in suspension concentration caused by addition of aeolian sands, P is the density of rainstorms, V is the wind speed of sandstorms, A is the runoff-generating area, L is the aeolian sand-filled channel length), an AHC flow occurs during the passage of a flood in a desert channel. Since high-frequency aeolian processes provide an adequate quantity of transportable sediment and promote AHC flow, most of the infrequent rainfall-induced floods occurring in arid zones can develop as AHC flows.

1. Introduction

In arid zones, wind erosion and water erosion are basic geomorphic processes whose interaction controls sediment transport and yield and shapes the landscape (Langford, 1989; Tooth, 2000; Bullard and Livingstone, 2002; Breshears et al., 2003; Bullard and McTainsh, 2003; Field et al., 2009). There is evidence to suggest that many active aeolian dunes terminate at stream channels and transport large volumes of aeolian sand into them (Smith and Smith, 1984; Thomas et al., 1997; Bullard and McTainsh, 2003; Nanson et al., 1995; Ta et al., 2008), leading to narrowing, aggrading, and even damming (Marker, 1977; Teller and Lancaster, 1986; Anderson and Anderson, 1990; Jones and Blakey, 1997; Mason et al., 1997) of the heavily aeolian sand-filled channels; for instance, Smith and Smith (1984) showed that abrupt addition of aeolian sand led to a 40-fold increase of bed load in the William river that crosses the Athabasca dunefields. Knapp (1938), Bagnold (1962) and Parker et al. (1986) argued that when the bed slope exceeds the ratio of the sediment fall velocity to the mean down-channel flow velocity, non-cohesive sediment can be entrained from the bed by turbulence and lead to high-speed and erosive hyper-concentrated flow. We therefore propose that if a particular threshold of channel bed slope is exceeded, the aeolian sand supply may result in very high suspended sediment loads during desert stream channel flooding. Even though aeolian sand-filled channels show very high rates of sediment transport during rainstorm-induced flooding, resulting in deleterious effects on the downstream river system and ecology (Zhi and Shi, 2002; Ta et al., 2011), there are very few empirical data on this aeolian process-induced hyper-concentrated flow in desert channels.

In order to further understand these processes, here we present an analysis of aeolian and fluvial processes and their related hyper-concentrated flows in the Sudalaer desert watershed in the Ordos plateau of China, which is a typical watershed characterized by aeolian–fluvial interplay. Our main objective is to clarify how aeolian processes regulate the development of hyper-concentrated (AHC) flow in order to provide criteria for predicting AHC occurrence in arid and semiarid watersheds.

2. Theoretical analysis

2.1. Wind-blown sand transport and related channel fill

When a sandstorm occurs in a sand dune desert and moves across a stream channel at angle θ, where θ is defined as the anticlockwise angle from the wind direction to the channel section

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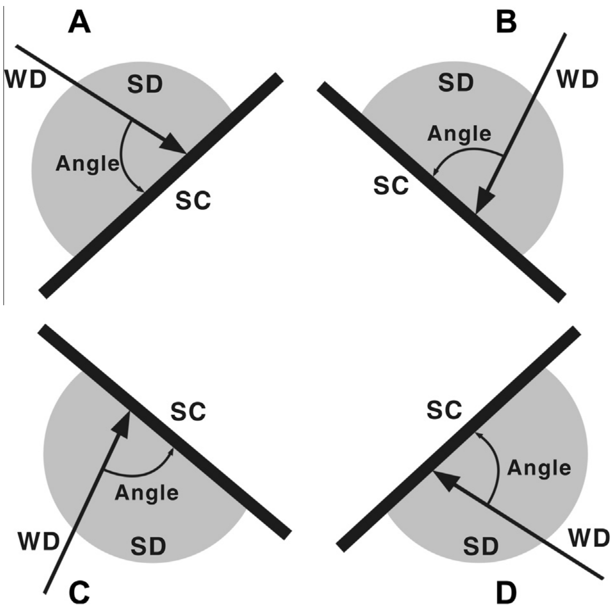

Fig. 1. Schematic of the direction of wind-blown sand entering a stream channel transported by sandstorms, relative to different spatial distributions of sand dunes and stream channels. WD is the wind direction of sandstorms, SD is the sand-dune desert, and SC is the stream channel; the angle is defined as the anticlockwise angle from the wind direction to the channel section.

($0 < \theta < 180°$) (Fig. 1), a recirculating air flow immediately develops in the opposite direction in the channel. This leads to separation and deposition of coarse aeolian sands on the lee side of the channel regardless of whether the wind direction is normal for that local channel section (Fig. 2a). It follows that if strong winds do not cross the sand dune desert, no aeolian sand can be transported into stream channels. The more powerful the aeolian process, the narrower the channel becomes due to increased deposition of aeolian sand. In general, the wind-blown sand transport rate is (Bagnold, 1941):

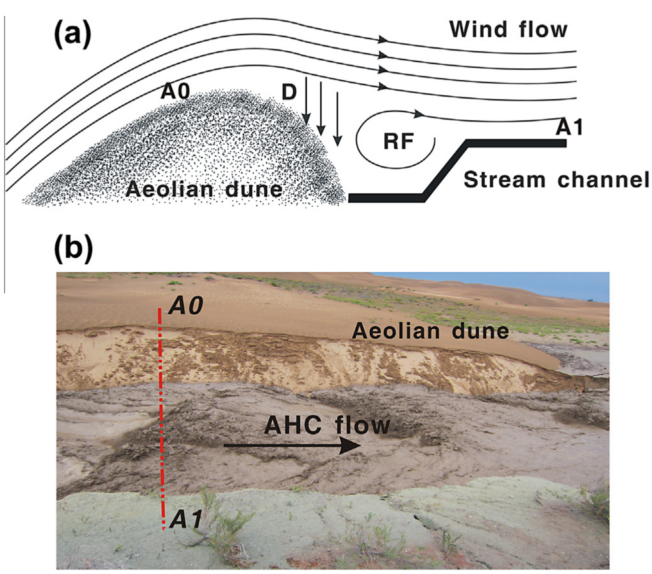

Fig. 2. Schematic of the development of an AHC flow via the interplay between aeolian and fluvial processes. (a) Illustration of the separation of aeolian sand from wind-blown sandstorms due to channel-induced recirculating air flow, leading to deposition of aeolian sand in stream channels (A_0A_1 is the channel cross-section in Fig. 1b); D denotes the deposition of aeolian sand; RF denotes a recirculating air flow. (b) The development of an AHC flow through entrainment of aeolian sands in the aeolian sand-filled channel.

$$q_w = C_1\sqrt{\frac{d}{D}}\frac{\rho_a}{g}u_*^3 \quad (1)$$

where q_w is the wind-blown sand transport rate, C_1 is an empirical dimensionless constant (1.8 for the naturally graded sands occurring in dunes), d is the grain diameter of the aeolian sand, D is the grain diameter of a standard 0.25 mm sediment, g is the gravitational acceleration, u_* is the wind shear velocity during saltation, and ρ_a is the air density. The shear velocity in Eq. (1) can be expressed as (Sauermann et al., 2001):

$$u_* = \frac{v(z)\kappa}{\ln\frac{z}{z_0}} \quad (2)$$

where $v(z)$ is the wind velocity at a height z over the sand surface, κ is the von Kármán constant, and z_0 is the roughness length of the sand surface. From Eqs. (1) and (2), the wind-blown sand transport rate q_w is:

$$q_w = \alpha v^3(z) \quad (3a)$$

$$\alpha = C_1\sqrt{\frac{d}{D}}\frac{\rho_a\kappa^3}{g\left(\ln\frac{z}{z_0}\right)^3} \quad (3b)$$

where $z_0 = 2.5 \times 10^{-5}$ m, $\kappa = 0.4$, $\rho = 1.225$ kg m^{-3}, $g = 9.8$ m s^{-2}, $C_1 = 1.8$, and $D = 0.25$ mm. If $z = 10$ m, α can equal 5×10^{-6} kg s^2 m^{-4}.

Assuming the probability distribution function of wind speed v_i of wind-blown sandstorms in the θ direction with respect to the channel section ($0 < \theta < 180°$) is f_{wi} over the course of a year, the annual deposition of aeolian sands in a channel with a length L can be expressed as:

$$Q_d = TL\alpha\sum_i f_{wi}v_i^3 \quad (4a)$$

$$f_{wi} = T_{vi}/T \quad (4b)$$

$$v_i = v_0 + \Delta v(i-1), (i = 1, 2, 3, \ldots) \quad (4c)$$

where Q_d is the annual deposition of aeolian sands in the channel, T is time (one year or 3.1536×10^7 s), T_{vi} is the duration of the sandstorm with an average wind speed v_i in the θ direction with respect to the channel, v_i is the graded wind speed at an interval Δv, and v_0 is the critical entrainment wind speed. In this way, the aeolian sand supply, stored temporarily in the channel during the windy seasons, can provide adequate sediment for transport during flooding.

2.2. Flood passage and the development of AHC flow in an aeolian sand-filled channel

When a flood flushes downstream in an aeolian sand-filled channel with bed slope S, significant channel erosion and collapse of dune slips can occur. Such collapses create a dense particulate flow that travels down the dune slips to enter the channel bed. Flood turbulence entrains aeolian sands from the channel bed, which increases and maintains the suspended sediment concentration and suspension, and maintains the flow of the water–sediment mixture against the bed shear stress, leading to an erosive hyper-concentrated flow (Fig. 2b). Following Pantin (1979), Parker et al. (1986), Winterwerp et al. (1992), and Akiyama and Stefan (1985), we describe the mass, moment, and energy balance equations for the AHC flow as follows:

$$\frac{\partial\gamma h}{\partial t} + \frac{\partial u\gamma h}{\partial x} = \sigma(F_e - F_d) \quad (5a)$$

$$\frac{\partial Uh}{\partial t} + \frac{\partial U^2 h}{\partial x} = \frac{\sigma - \rho}{\sigma}g\gamma hS - v_*^2 \quad (5b)$$