Erosion of unconsolidated beds by turbidity currents

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Turbidity currents are gravity flows of fluids with suspended, denser sediment, which remains aloft due to turbulence generated by the current motion itself. To remain active, turbidity currents must have an ability to entrain material from their base to counteract the sedimentation of particles from the current to the base. A number of decades ago, Bagnold, Engelund, and Fredsøe proposed a physical picture for erosion as a function of the overall velocity of the turbidity current (bed stress). Recently, it has been argued that the high-velocity form of this law is critical in determining the overall mechanics of turbidity currents, particularly their predeliction to erode or deposit sediment in different locations. This letter reexamines the Bagnold-Engelund-Fredsøe picture, and determines the corresponding erosion law in a way that is consistent with turbidity current mechanics, and has a high-velocity plateau that determines the qualitative features of turbidity current deposition and erosion. I also address the differential role of fluid and grain stress transmission in determining erosion; provided the grain stress transmission is less effective than the fluid stress transmission in eroding sediment, there will continue to be a high-velocity plateau in the erosion rate.

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I. INTRODUCTION

Turbidity currents are one of the predominant mechanisms by which sediment is moved from the continents to the deep oceans, in the process creating sand accumulations that can, over time, be transformed to deep water oil and gas reservoirs [1]. The exploitation of these reservoirs has been a major stimulus to the oil and gas industry, and the world economy, over the last two decades. The economic importance of these reservoirs has contributed to a revival of interest in the mechanics of sub-aqueous turbidity currents and other gravity flows; researchers hope that better understanding of the physics of these flows might lead to an improved ability to determine, from limited data, the stratigraphic structure and evolution of deep water sedimentary deposits.

The driving force for turbidity currents is gravity; the sediment suspended in these flows increases their density, so they move downslope amidst the lighter ambient fluid. The energy generated by this motion is partially converted into turbulence within the flows, which acts to keep the suspended grains aloft contrary to their natural inclination to settle. The balance between gravitational force and hydraulic drag, and between settling and the turbulent mixing of the particles, influences the (still poorly understood, see [2]) mechanics of these currents. An additional feature, and the primary focus of this work, is the ability of these currents both to erode or entrain an underlying unconsolidated granular layer, as well as add grains to this layer through settling of the turbidity current sediment. The balance between these two phenomena will determine the sediment loading in the current.

The modern era in the understanding of these flows properly begins with a work by Parker, Fukushima, and Pantin, which established, from conservation laws and hydrodynamic arguments,

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a system of alternatively three or four differential equations describing the evolution of turbidity current height, depth-averaged momentum, sediment load, and (in the four-equation version) depth-averaged turbulent kinetic energy [3]. Parker *et al.* were primarily concerned with determining the conditions under which turbidity currents auto-ignited, i.e. in which their erosion of underlying sediment compensated or overcompensated for the settling of sediment out of the current onto the underlying seafloor, leading to stable or growing turbidity currents downslope. Other authors, including [4], have used these equations, or modified versions thereof, to address a broader range of issues in the dynamics of turbidity currents, notably attempting to understand the hydrodynamic and topographical conditions under which turbidity currents are net erosional, net depositional, or bypass (neither eroding nor depositing).

II. EROSION

An illuminating feature of this latter work is the role of the sediment entrainment or erosion at the base of the flow in controlling overall flow dynamics. Although the turbulence generated by the flow itself is responsible for keeping aloft the (relatively heavier) sediment in the flow, net settling can still cause the sediment to accumulate in the bed at the base of the flow, over time extinguishing the flow, unless the flow has a compensating ability to erode or entrain unconsolidated material from that bed. The erosion rate from an unconsolidated sediment is a function of the bed stress $u_*^2 = \tau/\rho_f$, where τ is the shear stress (or drag) across the base of the flow, and ρ_f is the fluid density. Often u_* is nondimensionalized by the particle settling velocity v_s, given in its Stokesian form by

$$v_s = \frac{1}{18}\frac{Rg}{\nu_f}d^2, \tag{1}$$

where ν_f is the kinematic viscosity of the suspending fluid (water), d the particle diameter, g the gravitational acceleration, and R ($\approx$1.6 for silica) the excess relative density of the sediment compared to water.

There are various turbidity current erosion laws given in the literature as functions of the dimensionless tractive stress Z, defined by

$$Z = \sqrt{\mathrm{Re}_p}\frac{u_*}{v_s}, \tag{2}$$

with the particle Reynolds number given by

$$\mathrm{Re}_p = \frac{\sqrt{Rg}d^{3/2}}{\nu_f}. \tag{3}$$

One characteristic form is given by Garcia and Parker in [5], in terms of a slightly modified tractive stress Z',

$$Z' = \mathrm{Re}_p^{0.6}\frac{u_*}{v_s}, \tag{4}$$

as

$$E_S = v_s\frac{A(Z')^5}{1 + \frac{A}{0.3}(Z')^5}, \tag{5}$$

with $A = 1.3 \times 10^{-7}$. This gives the full or "bare" erosion rate, which can be adjusted by deposition arising from sedimentation of particles near the surface to give net erosion. Equation (5), with the tractive stress defined by Eq. (2) or Eq. (4), can be contrasted with the normal description of the threshold of erosion in fluvial settings (the "Shields curve" [6]), which gives this threshold as a function of the scaled stress,

$$\hat{\tau} = \frac{u_*^2}{Rgd}, \tag{6}$$

and a boundary layer Reynolds number

$$\mathrm{Re}_b = \frac{u_* d}{\nu_f}. \tag{7}$$

Some simple algebra allows us to rewrite Z or Z' in terms of $\hat{\tau}$ and Re_b. Let us first generalize

$$Z_\gamma \equiv (\mathrm{Re}_p)^\gamma \frac{u_*}{v_s}. \tag{8}$$

It is straightforward to show that

$$\frac{Z_\gamma}{18} = \frac{\hat{\tau}^{1-\gamma/2}}{\mathrm{Re}_b^{1-\gamma}}, \tag{9}$$

so that the claim that the erosion rate is determined by Z_γ implies that, in the Shields diagram, the erosion threshold (however defined) appears on a locus

$$\hat{\tau} \propto \mathrm{Re}_b^{\frac{1-\gamma}{1-\gamma/2}}. \tag{10}$$

Since the Shields curve is nonmonotonic, this cannot, in general, be true. Thus, the result of Garcia and Parker should be viewed as a fit to a relatively small region of the full Shields curve.

It is important to recognize that Eq. (5) was originally advanced as an empirical fit to observed erosion data. It is a feature both of this Garcia and Parker form [5] and of alternative erosion relations, such as that in [3], that the erosion rate goes to a constant for large values of the bed stress u_*. In these works, the data shown do not actually support the appearance of an erosion plateau, since these data are predominantly available only for tractive stresses well below that at which the plateau (as specified in the empirical fit) would emerge. The strongest argument for the appearance of the plateau is instead given by Engelund and Fredsøe in theoretical work from the related field of fluvial erosion [7], an argument which I repeat below.

Halsey, Kumar, and Perillo [4] have argued that this "plateau form" for the erosion rate is critical in determining the qualitative features of the turbidity flows, since balancing an erosion rate that is insensitive to bed stress against a sedimentation rate that depends on sediment load, or concentration in the flow, will tend to drive flows towards a fixed, "universal" value of sediment concentration. By contrast, these authors argue that the precise details of the erosion rate, e.g., the power law appearing in Eq. (5), do not dramatically affect the turbidity current physics; these details will not concern us here.

Given the importance of this plateau in controlling the dynamics of turbidity currents, it is disquieting that its existence is supported by theoretical arguments rather than by experimental observations. In this paper I will reexamine the motivation for ascribing a plateau to the erosion rate, as in Eq. (5). Although the existence of the plateau is not strongly corroborated by the experiments to which we have referred, the dependence of the erosion rate upon Z' (or, approximately, Z) is supported by this work. This work has nothing to add to the experimental basis for the erosion law, but focuses instead on clarifying and deepening the theoretical arguments for its existence, which are more subtle than perhaps expected.

III. BAGNOLD-ENGELUND-FREDSØE PICTURE

The most complete discussion of the theoretical rationale for a plateau-type erosion form is found in the work by Engelund and Fredsøe [7], although they attribute the physical reasoning to Bagnold [8,9]. The basic observation of Engelund and Fredsøe is that while the transverse stress (the bed stress) is constant as one passes through the bedload layer and into the underlying packed bed, the part of that stress mediated by the fluid versus the part mediated by grain interactions will in general change as one passes through this layer.

The kinetics of the fluid-mediated stress, which combines an overall average hydrodynamic stress with turbulent bursts and fluctuations, differs from that of the grain-mediated stress in a dispersed or suspended state, which will be communicated between layers parallel to the average flow either by particle motion between layers or by sharp, intermittent impulses when a mobile grain strikes another grain. If only the fluid-mediated force is effective in eroding (or mobilizing) the particles, then if this fluid-mediated force on the mobile layer above the bed exceeds that on the immobile underlying bed (because a component of the stress has been transmitted to the grains, which are now transmitting their portion to the underlying immobile bed), then it is possible for the "effective" bed stress, that part of the stress available to induce erosion of the immobile bed, to be at or below the Shields threshold, while the overall stress exceeds the Shields threshold.

The claim that the grain-transmitted portion of the stress is not effective in eroding or mobilizing immobile bed particles is perhaps surprising. In subaerial situations, it is well known that mobilized grains can dislodge multiple additional grains when striking the surface of a sand dune [8]. Of course, in the subaqueous case the relative excess density of sand particles is $R \approx 1.6$, while in the subaerial case it is $R \approx 750$, so the impact dynamics will clearly be quite different in the two cases, due to differences in the roles of inertial vs viscous forces. While it is thus conceivable that the Bagnold-Engelund-Fredsøe approach is correct, the physics involved is rather complex, and its elucidation by experiment will be more convincing than theoretical explanations. I will initially assume that only the fluid-transmitted shear stress is effective in mobilizing bed particles; I will return to the more general case below. Note that I am not claiming that only the time-averaged fluid-transmitted stress is effective in mobilizing the particles; it is well known that under certain circumstances turbulent bursts or eddies can play a significant role in erosion, which is why the classical Shields curve depends on Reynolds number as well as bed stress [6]. It remains an open question why the effect of grain-grain collisions (which will be controlled by elastic time scales) should be less important than the effect of fluctuations in fluid-mediated forces.

The overall typology of granular flows beneath a driven suspension flow was most recently demonstrated in work of Allen and Kudrolli [10]. An underlying jammed bed, at or near a random close-packed volume fraction $\phi \approx 0.6$, is overlain by a creeping flow region [11] with $0.45 < \phi < 0.6$ (the transition values should be viewed as approximate), which is in turn overlain by less dense bedload and ultimately suspended flows, which continuously transition to the sediment suspension in the turbulent fluid well above the bed. Some authors [12] have argued that at intermediate values of volume fraction the "dense granular flow" rheology proposed by Pouliquen and co-workers [13] may pertain.

Throughout this entire region, in steady state, the shear stress τ will obey

$$\partial_z \tau = \rho(z) g \sin\theta, \tag{11}$$

where $\rho(z)$ is the full density as a function of the coordinate z perpendicular to the bed and θ is the overall bed inclination to the vertical. Thus the shear stress varies slowly enough with z that we can regard it as constant through the boundary between the suspended and the creeping flow granular states. In the solid and creeping flow layers, the shear stress will be predominantly transmitted by grain-grain collisions and interactions; we call this the "condensed" region [10]. Towards the top of the this region, the shear stress is communicated by both the fluid (through its viscosity and strain rate gradient) and by the grain interactions, while, in the suspended region, the shear stress will be mostly fluid-transmitted. At the boundary between the suspended and condensed granular regions, we follow Bagnold-Engelund-Fredsøe and write

$$\tau = \tau_F + \tau_G, \tag{12}$$

decomposing the full shear stress into its fluid-transmitted and grain-transmitted components. The overall situation is illustrated in Fig. 1.

The equilibrium in particle exchange between the condensed and suspended flows will be determined by the kinetics at the boundary between these flows. Particles within the condensed layer will experience lift from the hydrodynamic forces communicating (at least part of) the bed

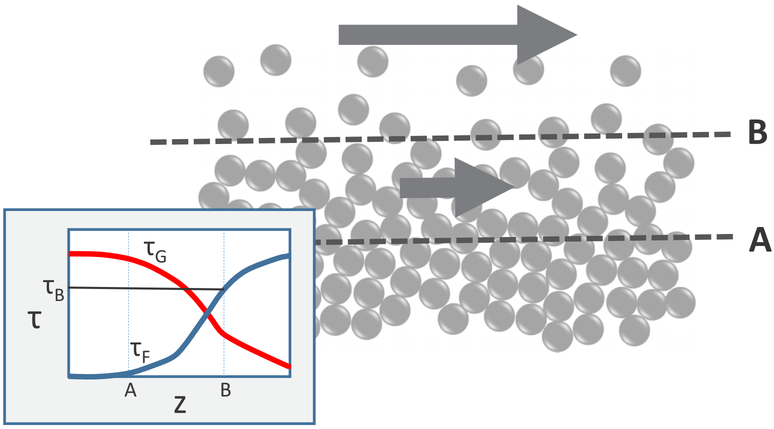

FIG. 1. The granular base underneath a turbidity current is a jammed solid (below A), which transitions to a creeping flow (between A and B) and ultimately a bedload and suspended load flow (above B). The inset displays the fluid-mediated (τ_F) and grain-mediated (τ_G) components of the overall stress. The stress across A is primarily transmitted by granular interactions, while that across B is transmitted both by fluid forces and by granular collisions. Note that the fluid component of the stress at B (τ_B) is less than the total stress.

stress, and may be advected into the suspended region; while particles within that region will settle back onto the condensed layer under the influence of gravity.

Since the fluid turbulence will not penetrate deeply into the condensed layer, at the base of the suspended layer the sedimentation of the particles into the dense granular layer can be approximated by Stokesian settling, as hypothesized by Parker *et al.* [3], or, in a different context, in [14]. If the concentration at the base of the suspended layer is ϕ_0, the rate at which particles settle into the dense granular layer will be

$$\mathcal{D} = v_s \tilde{\nu} \phi_0, \tag{13}$$

with the settling velocity v_s given by Eq. (1), and the factor $\tilde{\nu}$ accounting for the change in the one-particle settling rate due to the finite concentration of particles near the bed ("hindered settling"). It has been shown [10] that the Krieger-Dougherty form [15] for the dependence of the viscosity on concentration seems to be adequate near a sheared granular bed, so we use this form to write

$$\tilde{\nu} = \left(1 - \frac{\phi_0}{\phi_{\rm CP}}\right)^{1.5}, \tag{14}$$

where $\phi_{\rm CP} = 0.6$ is the close-packing concentration at which the granular viscosity diverges.

For the erosion or entrainment rate of particles from the dense granular layer into the suspended layer, we rely on a phenomenological approach. Let us presume that the dimensionless tractive stress Z (which is approximately equal to Z') controls the threshold of particle entrainment in the low Reynolds number limit that pertains to the hydraulics of typical turbidity currents.

Since we are assuming that only the fluid-mediated stress contributes to entrainment, we introduce the fluid-mediated stress parameter Z_F, defined by

$$Z_F(\tau_F) = \sqrt{\mathrm{Re}_p}\,\frac{\sqrt{\tau_F/\rho_f}}{v_s}. \tag{15}$$

For values of Z_F beyond the threshold of entrainment, we expect the entrainment rate to increase rapidly, given the activated nature of the process. It is simplest to choose the classic activation form for the entrainment rate $\mathcal{E}$ as a function of Z_F,

$$\mathcal{E} = v_s \exp(Z_F - Z_0), \tag{16}$$

where we can choose the set point Z_0 to fix the prefactor, arbitrarily, to equal v_s. Based on [3] and [5] we choose $Z_0 = 15$ for illustrative purposes in this work.

Note that we could as easily have chosen an activated process in which the argument of the exponential was the stress τ_F rather than Z_F, which would have the advantage of being more easily generalized to cases beyond that in which Z_F encodes the correct Reynolds number dependence of erosion. Qualitatively, the results of such an approach would be similar, so we continue to use Z_F for simplicity and for ease of comparison to the turbidity current literature.

To close this system of equations, it is sufficient to identify the dependence of τ_G on the concentration ϕ_0 and on u_* (or τ). This is a classic problem in the kinetic theory of granular systems. While there are some modern calculations of the shear stress corresponding to a granular flow, notably [16] and [17], unfortunately such calculations are performed in a vacuum, in which particle inelasticity is the only dissipation mechanism. Since we are in search of semiquantitative results only, it is sufficient to use the original formula determined by [9], as adapted by [7],

$$\tau_G = 0.027\rho_f(1+R)\lambda^2 u_*^2, \tag{17}$$

with the "linear concentration" $\lambda(\phi_0)$ set by

$$\phi_0 = \frac{\phi_{\mathrm{CP}}}{(1+\lambda^{-1})^3}, \tag{18}$$

where ϕ_{CP}, as above, is the close packed concentration. We can now solve for τ_F, using $\tau = \rho_f u_*^2$ and Eq. (12):

$$\tau_F(\tau, \phi_0) = \tau[1 - 0.027(1+R)\lambda^2] \tag{19}$$

and, hence,

$$Z_F = Z\sqrt{1 - 0.027(1+R)\lambda^2}. \tag{20}$$

Our purpose now is to determine the erosion rate as a function of Z. The net erosion is

$$\mathcal{E} - \mathcal{D} = v_s[\exp(Z_F - Z_0) - \tilde{\nu}(\phi_0)\phi_0]. \tag{21}$$

Setting this equal to zero, we obtain

$$Z = \frac{Z_0 + \ln\tilde{\nu} + \ln\phi_0}{\sqrt{1 - 0.027(1+R)\lambda^2}}, \tag{22}$$

which is an implicit equation for the volume fraction ϕ_0 as a function of tractive stress Z. Since in equilibrium $\mathcal{E} = \mathcal{D} = v_s\nu(\phi_0)\phi_0$, this is equivalent to a relation between $\mathcal{E}$ and Z. Note, however, that it is a feature of the Krieger-Dougherty form Eq. (14) that the deposition flux is actually maximum for $\phi_0 = 0.24$, and declines beyond this value.

The erosion rate $\mathcal{E}$ is plotted against the tractive stress Z in Fig. 2 for the case $R = 1.6$, $\phi_{\mathrm{CP}} = 0.6$, and $Z_0 = 15$. Note that, at higher bed stress, the bare erosion rate does saturate.

IV. SEDIMENT EXCHANGE WITH BULK CURRENT

Of course, ϕ_0 is not necessarily the same as the depth-averaged concentration of the turbidity current, $\bar{\phi}$. It has been found that in steady state, approximating $\phi_0 = r_0\bar{\phi}$ with $r_0 = 1.6$, a constant, appears to be adequate [18]. The question remains of whether it is the equilibration at the top of the condensed granular layer or rather the equilibration of ϕ_0 with $\bar{\phi}$ (driven by the turbulent mixing of the overall turbidity current) that will be the rate limiting step in the dynamics. Suppose that the turbidity current has an overall height h, while the region near the surface in which the concentration is ϕ_0 is of height h'. Then writing the advective derivative (applied to an arbitrary function ψ)

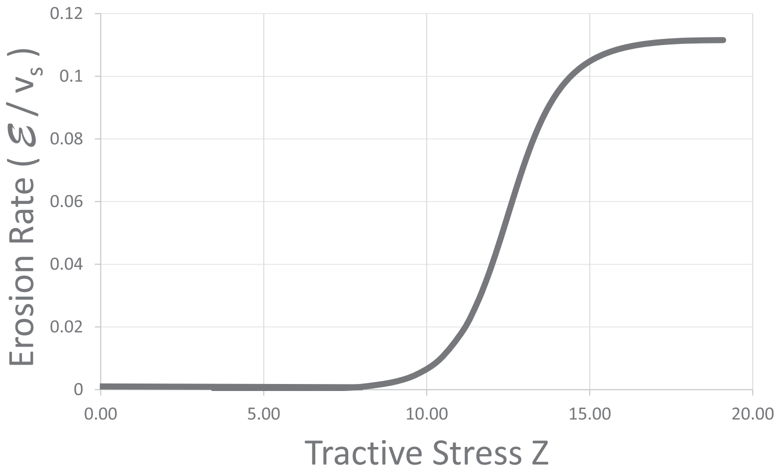

FIG. 2. The “bare” erosion rate $\mathcal{E}/v_s$, from Eq. (16), vs the tractive stress Z, as determined from Eq. (22), for $\phi_0 < 0.25$. As hypothesized by Garcia and Parker [5], a clear saturation is evident at larger values of tractive stress.

$D_t\psi = \partial_t\psi + \nabla(\vec{u}\psi)$, we see that we can write the phenomenological equations

$$D_t(h\bar{\phi}) = \Upsilon(\phi_0 - r_0\bar{\phi}), \quad (23)$$

$$D_t(h'\phi_0) = -\Upsilon(\phi_0 - r_0\bar{\phi}) + (\mathcal{E} - \mathcal{D}), \quad (24)$$

with Υ a constant determined by the details of the turbulent mixing; we expect that if the depth-averaged velocity in the bulk flow is u, then $\Upsilon \sim u$.

V. EFFECT OF GRAIN-TRANSMITTED STRESS

Finally, we return to the question of the role of the grain-transmitted shear stress τ_G. A more general approach than the one used thus far would acknowledge that the grain stress can play a role in moving particles from the dense granular to the suspended layer, but with an effectiveness that is perhaps different from that of the fluid stress τ_F. If we suppose that the relative effectiveness of the grain stress is reduced by a factor α, then we can write, defining $Z_\alpha = \sqrt{\mathrm{Re}_p}\sqrt{\tau_F + \alpha\tau_G/\rho_f}/v_s$,

$$\mathcal{E} = v_s \exp[Z_\alpha - Z_0)]. \quad (25)$$

Repeating the derivation above, and noticing that $\tau_F + \alpha\tau_G = \tau - (1-\alpha)\tau_G$, we find that

$$\tau_F + \alpha\tau_G = \rho_f(u_*)^2[1 - 0.027(1-\alpha)R\lambda^2]. \quad (26)$$

Thus the overall effect of setting $\alpha \neq 0$ is identical to one obtainable simply by changing the density of the particles. The qualitative form of the erosion rate will be the same, although the limiting value of the erosion at high Z will be increased with respect to the $\alpha = 0$ result. Obviously this will break down for larger values of α.

VI. DISCUSSION

I have demonstrated that, provided that the erosion or entrainment phenomenon is dominated by the fluid-mediated stress, and not the full stress, then the standard Shields curve picture of particle erosion is consistent with the appearance of a plateau in erosion rate beyond a threshold in bed stress. Although I took advantage of an equilibrium condition to compute the "bare" erosion rate evincing this plateau, my ambition is to apply this computed bare erosion rate in nonequilbrium circumstances as well, thereby connecting erosion-deposition dynamics to the formation or destruction of sedimentary bodies by turbidity current processes. Studies of turbidity current dynamics have now reached a stage where the existence (or nonexistence) of such a plateau is certainly relevant in connecting modeling results [4] to observations [19]. Although my arguments were general, and physically motivated, the example (mentioned above) of the role of granular collisions in subaerial flows suggests that we should be cautious in assuming that qualitatively new phenomena will not arise from a proper treatment of granular collisions, which is not only beyond the scope of this study, but is also a potentially fruitful area for experimental research.

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[1] E. Meiburg and B. Kneller, Turbidity currents and their deposits, Annu. Rev. Fluid Mech. **42**, 135 (2010).
[2] J. D. Parsons, C. T. Friedrichs, P. A. Traykovski, D. Mohrig, J. Imran, J. P. Syvitski, G. Parker, P. Puig, J. L. Buttles, M. H. Garcia *et al.*, The mechanics of marine sediment gravity flows, in *Continental Margin Sedimentation: From Sediment Transport to Sequence Stratigraphy*, edited by C. Nittrouer, J. Austin, M. Field, J. Syvitski, and P. Wiberg (Blackwell, Oxford, 2007), p. 275.
[3] G. Parker, Y. Fukushima, and H. M. Pantin, Self-accelerating turbidity currents, J. Fluid Mech. **171**, 145 (1986).
[4] T. C. Halsey, A. Kumar, and M. M. Perillo, Sedimentological regimes for turbidity currents: Depth-averaged theory, J. Geophys. Res.: Oceans **122**, 5260 (2017).
[5] M. Garcia and G. Parker, Entrainment of bed sediment into suspension, J. Hydraul. Eng. **117**, 414 (1991).
[6] V. A. Vanoni, *Sedimentation Engineering* (American Society of Civil Engineers, Reston, VA, 2006).
[7] F. Engelund and J. Fredsøe, A sediment transport model for straight alluvial channels, Hydrol. Res. **7**, 293 (1976).
[8] R. A. Bagnold, The movement of a cohesionless granular bed by fluid flow over it, Br. J. Appl. Phys. **2**, 29 (1951).
[9] R. Bagnold, Experiments on a gravity-free dispersion of large solid spheres in a Newtonian fluid under shear, Proc. R. Soc. London Ser. A **225**, 49 (1954).
[10] B. Allen and A. Kudrolli, Depth resolved granular transport driven by shearing fluid flow, Phys. Rev. Fluids **2**, 024304 (2017).
[11] M. Houssais, C. P. Ortiz, D. J. Durian, and D. J. Jerolmack, Onset of sediment transport is a continuous transition driven by fluid shear and granular creep, Nat. Commun. **6**, 6527 (2015).
[12] F. Boyer, É. Guazzelli, and O. Pouliquen, Unifying Suspension and Granular Rheology, Phys. Rev. Lett. **107**, 188301 (2011).
[13] Y. Forterre and O. Pouliquen, Flows of dense granular media, Annu. Rev. Fluid Mech. **40**, 1 (2008).
[14] D. Martin and R. Nokes, Crystal settling in a vigorously convecting magma chamber, Nature (London) **332**, 534 (1988).
[15] I. M. Krieger and T. J. Dougherty, A mechanism for non-Newtonian flow in suspensions of rigid spheres Trans. Soc. Rheol. **3**, 137 (1959).
[16] S. Savage and D. Jeffrey, The stress tensor in a granular flow at high shear rates, J. Fluid Mech. **110**, 255 (1981).
[17] K. A. Reddy and V. Kumaran, Applicability of constitutive relations from kinetic theory for dense granular flows, Phys. Rev. E **76**, 061305 (2007).

[18] M. Garcia, Experimental study of turbidity currents, Master's thesis, Department of Civil and Environmental Engineering, University of Minnesota, Minneapolis, MN, 1985.

[19] J. de Leeuw, J. T. Eggenhuisen, Y. T. Spychala, M. S. Heijnen, F. Pohl, and M. J. Cartigny, Sediment volume and grain-size partitioning between submarine channel-levee systems and lobes: An experimental study, J. Sediment. Res. **88**, 777 (2018).