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Original Research

Multiple effects of sediment transport and geomorphic processes within flood events: Modelling and understanding

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ABSTRACT

Flood events can induce considerable sediment transport which in turn influences flow dynamics. This study investigates the multiple effects of sediment transport in floods through modelling a series of hydraulic scenarios, including small-scale experimental cases and a full-scale glacial outburst flood. A non-uniform, layer-based morphodynamic model is presented which is composed of a combination of three modules: a hydrodynamic model governed by the two-dimensional shallow water equations involving sediment effects; a sediment transport model controlling the mass conservation of sediment; and a bed deformation model for updating the bed elevation. The model is solved by a second-order Godunov-type numerical scheme. Through the modelling of the selected sediment-laden flow events, the interactions of flow and sediment transport and geomorphic processes within flood events are elucidated. It is found that the inclusion of sediment transport increases peak flow discharge, water level and water depth in dam-break flows over a flat bed. For a partial dam breach, sediment material has a blockage effect on the flood dynamics. In comparison with the 'sudden collapse' of a dam, a gradual dam breach significantly delays the arrival time of peak flow, and the flow hydrograph is changed similarly. Considerable bed erosion and deposition occur within the rapid outburst flood, which scours the river channel severely. It is noted that the flood propagation is accelerated after the incorporation of sediment transport, and the water level in most areas of the channel is reduced.

1. Introduction

Floods are one of the most catastrophic natural hazards for people and infrastructure, including floods resulting from intense rainfall, dam break, and the sudden release of meltwater from ice sheets caused by volcanic activity, etc. (Alho et al., 2005; Carrivick et al., 2010; Carrivick & Rushmer, 2006; Manville et al., 1999). Such high-magnitude sudden onset floods generally comprise an advancing intense kinematic water wave which can induce considerable erosion and sediment loads, thereby causing rapid geomorphic change. Morphological changes during flood events can in turn have significant implications on flow dynamics. Therefore, how flow dynamics interact with morphological change is a topic of growing interest in the research community.

Sediment-laden flows involve a complex flow–sediment interaction process. To date, the understanding of flow–sediment transport interactions is limited. A variety of small-scale experiments such as dam-break flows over a movable bed and breach formation have been conducted in recent studies (Carrivick et al., 2011; Spinewine & Capart, 2013; Zech & Soares-Frazao, 2007). The studies have reported the geomorphic impacts of rapid and transient dam-break flows and the implications of sediment transport on flow dynamics. However, these experiments are only small-scale and the effects of sediment transport are only considered for a specific event. The insights obtained are useful but inevitably have limitations. In recent decades, efforts have been made to model extreme flood events to demonstrate both flow dynamics and geomorphic processes (Cao et al., 2004; Capart & Young, 1998; Fraccarollo & Capart, 2002; Guan et al., 2014; Li & Duffy, 2011; Li et al., 2014; Simpson & Castelltort, 2006; Wong et al., 2014; Wu & Wang, 2007; Xia et al., 2010). Existing numerical work mostly focused on modelling of small-scale flow events or analysis of idealised dam-break hydraulics. These studies provide fundamental insights on the complex flow–sediment interactions. The implications of morphological changes on flow dynamics must be investigated through testing at both small-scale and large-scale with various scenarios. To extend the knowledge on the effects of sediment transport and geomorphic processes within floods, this paper specifically adopts a 2D hydro-morphodynamic model to

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simulate several types of flow events with and without the inclusion of sediment transport. The selected flood events include not only a dam-break flow over a movable bed, a small-scale, partial dam breach due to overtopping, but also a large-scale, glacial outburst flood. The layer-based morphodynamic model developed by Guan et al. (2014) is extended to a non-uniform model, where hiding/exposure effects, and bed slope effects are considered.

This paper is organised as follows: the extension of the layer-based morphodynamic model is implemented in Section 2; in Section 3, the numerical model solution is described; Section 4 presents the results and discussion of the three selected flow events; conclusions are drawn in Section 5.

2. Hydro-morphodynamic model

A bedload dominant sheet flow model has been previously presented and validated by the authors (Guan et al., 2014). This model is extended to model non-uniform sediment transport in this paper. The framework for the layer-based model includes three modules:

- a hydrodynamic model governed by the two-dimensional shallow water equations with sediment effects;
- a sediment transport model controlling the mass conservation of sediment, and
- a bed deformation model to update the bed elevation in response to erosion and deposition.

A model can never represent all the features of flow and sediment transport. The major assumptions of the present model include: (1) the model is for bed material load, (2) sediment material is assumed to be non-cohesive, (3) the collision effects between particles are ignored, (4) sediment mixtures are transported with a same velocity, and (5) within one time step the bed evolution is ignored, but the bed is updated at every time step.

2.1. Hydrodynamic model

The hydrodynamic model involves the mass and momentum exchange between flow and sediment. The governing equations can be written as

$$\frac{\partial \eta}{\partial t}+\frac{\partial hu}{\partial x}+\frac{\partial hv}{\partial y}=0 \tag{1}$$

$$\frac{\partial hu}{\partial t}+\frac{\partial}{\partial x}\left(hu^2+\frac{1}{2}gh^2\right)+\frac{\partial huv}{\partial y}=gh\left(-\frac{\partial z_b}{\partial x}-S_{fx}\right)+\frac{\Delta\rho u}{\rho}\frac{\partial z_b}{\partial t}\left(\frac{1-p}{\beta}-C\right)-\frac{\Delta\rho g h^2}{2\rho}\frac{\partial C}{\partial x}-S_A \tag{2a}$$

$$\frac{\partial hv}{\partial t}+\frac{\partial huv}{\partial x}+\frac{\partial}{\partial y}\left(hv^2+\frac{1}{2}gh^2\right)=gh\left(-\frac{\partial z_b}{\partial y}-S_{fy}\right)+\frac{\Delta\rho u}{\rho}\frac{\partial z_b}{\partial t}\left(\frac{1-p}{\beta}-C\right)-\frac{\Delta\rho g h^2}{2\rho}\frac{\partial C}{\partial y}-S_B \tag{2b}$$

where t=time; g=gravity acceleration; h=flow depth; z_b=bed elevation; $\eta=h+z_b$ water surface; u, v=average flow velocity in the x and y directions respectively; p=sediment porosity; C=volumetric concentration in flow depth; ρ_s, ρ_w=density of sediment and water respectively; $\Delta\rho=\rho_s-\rho_w$; ρ=density of sediment–flow mixture; S_{fx}, S_{fy}=Manning's n based frictional slopes velocity in the x and y directions, respectively; $\beta=u/u_s$=flow-to-sediment velocity ratio determined by the equation proposed by Greimann et al. (2008); and S_A, S_B are the additional terms related to the velocity ratio β.

$$S_{A,B}=\frac{\Delta\rho V}{\rho}\left(1-\frac{1}{\beta}\right)\left[\left(C\frac{\partial hu}{\partial x}+C\frac{\partial hv}{\partial y}\right)-\left(hu\frac{\partial C}{\partial x}+hv\frac{\partial C}{\partial y}\right)\right] \tag{3}$$

$$\beta=\frac{U}{U_s}=\frac{U}{U_*}\frac{\sqrt{\theta}}{1.1\left(\theta/\theta_c\right)^{0.17}\left[1-\exp\left(-5\theta/\theta_c\right)\right]}$$

where $V=u$ for x direction; $V=v$ for y direction. $U=\sqrt{u^2+v^2}$, $U_s=\sqrt{u_s^2+v_s^2}$ are the total flow velocity and the total sediment velocity in the sheet flow layer, U_* is the shear velocity; θ=average dimensionless bed shear stress; and θ_c=critical dimensionless bed shear stress.

2.2. Sediment transport model

For transport of non-uniform sediment, the mass equation of the ith sediment class in a sheet flow layer is given by

$$\frac{\partial hC_i}{\partial t}+\frac{1}{\beta}\frac{\partial huC_i}{\partial x}+\frac{1}{\beta}\frac{\partial hvC_i}{\partial y}=-\frac{1}{\beta}\frac{(q_{bi}-F_iq_{b*i})}{L_i} \tag{4}$$

where F_i=the proportion of the ith size class; C_i and L_i=volumetric concentration, and non-equilibrium adaptation length of the ith size class, respectively; and $C=\sum_{i=1}^{N}C_i$; $q_{bi}=hC_i\sqrt{u^2+v^2}$, q_{b*i}=actual sediment transport rate, total sediment transport capacity for the ith size class respectively. No universal sediment transport equation is available, and each empirical formula has its own application range. The commonly used equations depending on the bed slopes are selected: the Meyer-Peter & Müller equation (MPM) (Meyer-Peter & Müller, 1948) and Smart and Jäggi equation (SJ) (Smart & Jäggi, 1983). The MPM equation is derived for bed load transport based on the experimental data for bed slope from 0.0004 to 0.02 and dimensionless bed shear stress smaller than 0.25. Therefore, in this study, the approach taken by others (Abderrezzak & Paquier, 2011; Nielsen, 1992; Zech et al., 2008) is followed and the MPM is modified by incorporating a calibrated coefficient. The modified MPM equation is used for bed slopes smaller than 0.03. For steep slopes greater than 0.03, the SJ equation derived by expanding the database of MPM to the steep slopes range up to 0.03–0.20 and high bed shear stress is applied with a limitation of maximum S_o at 0.20.

$$q_{b*i}=\psi 8\left(\theta_i-\theta_{c,i}\right)^{1.5}\sqrt{sgd_i^3}\quad 0\le S_o<0.03 \tag{5}$$

$$q_{b*i}=4.2\frac{h^{1/6}}{n\sqrt{g}}\min\left(S_o,0.2\right)^{0.6}\theta_i^{0.5}\left(\theta_i-\theta_{c,i}\right)\sqrt{sgd_i^3}\quad S_o\ge 0.03 \tag{6}$$

where ψ=calibrated coefficient; $s=\rho_s/\rho_w-1$=submerged specific gravity of sediment; S_o=bed slope; θ_i=dimensionless bed shear stress of ith fraction; θ_{ci}=critical dimensionless bed shear stress of ith fraction; n=Manning's roughness; and d_i is the diameter of the ith class size. The non-equilibrium adaptation length has been widely investigated either by theoretical analysis or laboratory experiments (Armanini & Di Silvio, 1988; Fang & Wang, 2000; van Rijn, 1987; Wu, 2004). Therein, the relationship, $hU/\gamma\omega$, was chosen in this study. Thus, the right term of Eq. (4) excluding the velocity ratio could be converted to $\gamma\omega(C-C_{b*})$ which is a commonly used concept in the research community (Fang & Wang, 2000; Xia et al., 2010), where $C_{b*}=q_{b*}/hU$. L_i for the ith class size is given as follows:

$$L_i=\frac{h\sqrt{u^2+v^2}}{\gamma\omega_{f,i}}\text{ with }\gamma=\min\left(\frac{1}{\beta}\frac{h}{h_b},\frac{1-p}{C}\right) \tag{7}$$

where γ is the ratio of depth-averaged sediment concentration and near-bed sediment concentration within sheet flow layer; h_b is the thickness of the sheet flow layer; $\omega_{f,i}$ is the effective settling velocity of the ith sediment particle size. In high concentration mixtures, the settling velocity of a single particle is reduced due to the presence of other particles. Considering the hindered settling