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Fully-coupled hydrologic processes for modeling landscape evolution

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ABSTRACT

Although current landscape evolution models can predict landscapes with specific concave-convex slopes, regolith thicknesses, drainage densities and relief, these models rarely include realistic groundwater and overland flows, and channel-hillslope interactions. To overcome the potential drawbacks, this study couples hydrologic processes with hillslope and channel sediment transport processes to form a new hydrologic-morphodynamic model (LE-PIHM) for regolith formation and landscape evolution. Two scenarios with and without groundwater flow are presented to demonstrate the importance of this coupling. Comparison of the steady state landforms indicates that hillslopes are steeper and relief is higher with groundwater flow. The sensitivity of the solution to mesh geometry is tested and it is shown that model simulations maintain the characteristic features of a landscape over a reasonable range of maximum area and minimum interior angle. To predict long-term landscape change, a morphological acceleration technique is presented and a method for choosing an optimal morphological scale factor is introduced.

1. Introduction

Landscape evolution models (LEMs) aim to quantitatively predict the evolution of landscapes and their detailed spatial characteristics. In general LEMs are based on solving a system of equations for the continuity of mass, geomorphic transport functions that describe the generation and movement of sediment and (to a lesser extent) solutes on hillslopes, a representation of runoff generation and the routing of water across the landscape, geomorphic transport functions for erosion and transport of water-sediment mixtures in channels, and rock particle motions due to tectonics (for a review see Coulthard, 2001; Tucker and Hancock, 2010). Under relatively constant forcing, these models predict landscapes with specific concave-convex slopes and spatially variable regolith thicknesses, drainage densities, bedrock elevation, and relief (Beaumont et al., 2000; Bishop, 2007; Braun and Sambridge, 1997; Coulthard et al., 2000; Howard, 1994; Istanbulluoglu and Bras, 2005; Paik, 2012; Tucker et al., 2001; Tucker and Hancock, 2010; Tucker and Slingerland, 1994; Willgoose et al., 1991). Thus LEMs offer the prospect of testing the fitness of various quantitative laws of diffusion (Martin and Church, 2004; Rempe and Dietrich, 2014; Roering et al., 1999; Tucker and Slingerland, 1994), advection (Howard, 1994; Tucker and Bras, 1998; Tucker and Slingerland, 1994), and soil production processes (Heimsath et al., 1997, 2009; Roering, 2008). LEMS also allow us to generalize locally measured observations and fluxes to watershed and larger scales (Roering and Gerber, 2005; West et al., 2013), and reveal non-intuitive interactions between morphological processes and the resulting landforms (Perron et al., 2008, 2012; Willett et al., 2014).

The ever-growing interest in understanding the co-evolution of subsurface zone promotes the development of LEMs to consider more details of subsurface hydrological processes than heretofore included. This is needed, for example, in landscape-pedogenesis modeling which simulates soil evolution as a function of erosion and pedogenic processes because it shows a strong link between soil particle weathering and soil moisture (Cohen et al., 2010; Minasny et al., 2015). Also, Critical Zone (CZ) science, which studies the environmental gradient from atmosphere to bedrock at different spatial and temporal scales, sees surface and subsurface hydrological processes as vital at the air-soil and soil-bedrock interface (Anderson et al., 2008; Brantley et al., 2007). Moreover, where the infiltration capacity is high enough (Abrams et al., 2009; Higgins, 1982; Howard, 1988; Kochel et al., 1985; Laity and Malin, 1985; Lamb et al., 2008; Lobkovsky et al., 2007; Petroff et al., 2011, 2012, 2013; Schumm et al., 1995), the very nature of the landscape is different. Channels are formed by groundwater sapping such that the channels are bounded by steep walls and

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Nomenclature

z	Ground surface elevation (m)
e	Bedrock surface elevation e (m)
h	Regolith thickness in vertical (m)
H	Slope-normal thickness (m)
θ	Landscape surface slope in radians
U	Bedrock uplift rate (m/yr)
B_w	Bedrock weathering rate (m/yr)
q_{c1}	Lateral volumetric regolith flux by creep (m^2/yr)
q_{c2}	Lateral volumetric regolith flux by tree throw (m^2/yr)
q_s	Surface flux of regolith sediment by overland flow (m^2/yr)
σ_{re}	Bulk density of regolith (kg/m3)
σ_{ro}	Bulk density of bedrock (kg/m3)
P_o	Maximum bedrock weathering rate (m/yr)
α	Fitting coefficient for bedrock weathering equation (m^{-1})
K_1	Morphological diffusivity in linear creep equation (m^2/yr)
K_2	Morphological diffusivity in nonlinear creep equation (m^2/yr)
S_c	Critical gradient of slope
c_1	Volume of tree root plat per tree throw event ($m^3/event$)
c_2	Net downslope distance per event ($m/event$)
c_3	Density of tree throw event ($event/m^2$)
c_4	Frequency of tree throw event ($event/yr$)
W	Width of the root plat (m)
D	Pit depth (m)
K_3	Morphological diffusivity by three-trow (m^2/yr)
q_s^*	Dimensionless Einstein number
τ^*	Shields stress
τ_c^*	Critical Shields stress
D_{50}	Median grain diameter (m)
R	Submerged specific gravity of sediment
τ_0	Shear stress (kg/mS^2)
C_f	Drag coefficient
V	Vertical average velocity of overland flow (m/s)
R_{eD}	Renolds number
A_0	Weathering rate for bare bedrock (m/yr)
b	Weathering rate constant
ν	Kinematic viscosity (m/s^2)
u_*	Fluid shear velocity (m/s)
Ψ_{canopy}	Canopy water storage (m)
Ψ_{snow}	Snow (m)
Ψ_{surf}	Surface water depth (m)
Ψ_{unsat}	Water storage in the unsaturated zone (m)
Ψ_{sat}	Water storage in the saturated zone (m)
$vFrac$	fraction of vegetation coverage
f_s	Fraction of snow
P	Precipitation rate (m/day)
E_c	Evaporation on canopy (m/day)
TF	Water through fall (m/day)
SM	Snow melt (m/day)
q_{sw}	Volumetric overland flow per unit width (m^2/day)
P_{net}	Water reaching ground surface (m/day)
I	Infiltration rate (m/day)
E_s	Evaporation from surface water (m/day)
E_g	Evaporation from unsaturated zone (m/day)
E_{sat}	Evaporation from saturated zone (m/day)
R	Recharge rate (m/day)
E_{gt}	Transpiration from unsaturated zone (m/day)
E_{tsat}	Transpiration from saturated zone (m/day)
q_{sw}	Volumetric overland flow per unit width (m^2/day)
q_{gw}	Volumetric lateral groundwater flow per unit width (m^2/day)
Γ_{surf}	Conductivity of overland flow (m/day)
n_s	Gauckler-Manning coefficient ($day/m^{1/3}$)
Γ_{sat}	Horizontal hydraulic conductivity (m^2/day)
K_{eff}	Effective hydraulic conductivity (m^2/day)
Q_{c_ij}	Volumetric sediment flux by creep and tree-throw of the ith control volume in the jth direction (m^3/day)
Q_{s_ij}	Volumetric sediment flux by overland flow of the ith control volume in the jth direction (m^3/day)
Q_{sw_ij}	Lateral overland flow from element i to its jth neighbor (m^3/day)
Q_{gw_ij}	Lateral groundwater flow from element i to its jth neighbor (m^3/day)
f_{MSF}	Morphological Scale factor
[S]	Level of saturation
A_1	Maximum weathering rate at the critical depth (m/yr)
h^*	Critical depth where maximum weathering rate occurs (m)

Software availability

Version 1.0 This version of LE-PIHM can be made available upon request. A public version of LE-PIHM will be available soon in LE-PIHM@Github.com.

terminate in "theater-like" box canyons, and their hydraulic geometries are influenced by seepage erosion and bank collapse (Fox et al., 2007). Even if the infiltration rate is not high enough to preclude overland flow, it is also very common that subsurface flows change the regolith moisture and infiltration rate, and consequently change the timing and magnitude of surface runoff and discharge.

To date, most models neglect or simplify groundwater processes by focusing on landscapes in which the infiltration rate is thought to be low relative to overland flow. A few studies have modeled the interaction between surface and subsurface water (Barkwith et al., 2015; Francipane et al., 2012; Tucker et al., 2001; Tucker and Bras, 1998; Willgoose et al., 1991). For example, Tucker and Bras (1998) discussed the influence of saturation thresholds on drainage basin morphology. Precipitation was simply partitioned to surface runoff and subsurface flow as a function of drainage area, soil transmissivity and surface slope without really simulating hydrological processes (e.g. infiltration and surface water routing). Later, Tucker et al. (2001) improved the generation of runoff by infiltration-excess or saturation-excess mechanisms, but did not include base flow (portion of streamflow that comes from the sum of deep subsurface flow and delayed shallow subsurface flow). Francipane et al. (2012) improved the hillslope transport component of CHILD landscape evolution model (Tucker et al., 2001) by considering vegetation interception, evaporation, subsurface flow and snow. Barkwith et al. (2015) developed landscape evolution model by emphasizing the influence of subsurface flow on soil moisture storage and sediment transport. But these two models focus on landscape evolution at decades to a few hundreds of years without considering the effects of tectonic and bedrock weathering