tRIBS-Erosion: A parsimonious physically-based model for studying catchment hydro-geomorphic response

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A B S T R A C T

Our goal is to develop a model capable to discern the response of a watershed to different erosion mechanisms. We propose a framework that integrates a geomorphic component into the physically-based and spatially distributed TIN-based Real-time Integrated Basin Simulator (tRIBS) model. The coupled model simulates main erosive processes of hillslopes (raindrop impact detachment, overland flow entrainment, and diffusive processes) and channel (erosion and deposition due to the action of water flow). In addition to the spatially distributed, dynamic hyporheic variables, the model computes the sediment transport discharge and changes in elevation, which feedback to hydrological dynamics through local changes of terrain slope, aspect, and drainage network configuration. The model was calibrated for the Lucky Hills basin, a semi-arid watershed nested in the Walnut Gulch Experimental Watershed (Arizona, USA). It is demonstrated to be capable of reproducing main runoff and sediment yield events and accumulated volumes over the long term. The model was also used to study the response of two first-order synthetic basins representative of lowland and highland areas dominated by fluvial and diffusive erosion processes to a 100-year long stationary climate. The analysis of the resultant slope-contributing area relationships for the two synthetic basins illustrates that the model is consistent with assumed principles of behavior and capable of reproducing the main mechanisms of erosion.

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1. Introduction

The dynamics of sediment production and movement from sources (hillslopes and highlands) to sinks (lowland channels and water bodies) have tremendous consequences on the sustainability of terrestrial and aquatic environments and have direct impacts on planning and management of earth resources and human health. Erosion rates have gone up to unprecedented levels in large parts of the globe during the past several centuries (Wilkinson and McElroy, 2007). This increase has been attributed to growing climatic extremes (droughts and floods), urbanization, forest management practices, and in particular agricultural expansion. Cultivated lands globally lose soil at rates in the range of 200–1300 [mm/m2 per year] which surpass any plausible rate of soil development (Nearing et al., 1999; Pimentel et al., 1995; Wilkinson and McElroy, 2007). In forested terrain, human activities including road construction, fire suppression, and silvicultural practices have altered the hydrological response, and combined with wildfires lead to debris flows with magnitudes equivalent of hundreds to thousands of years of long-term erosion rates.

Understanding and predicting the basin hydro-geomorphic response and its relation to landscape morphology and hydrological states is critical for managing soil and water resources under global change and for studying the linkages between landforms, climate, and vegetation over geomorphic time scales (Slattery et al., 2002; Wainwright and Thornes, 2003). Process-based research in geomorphology provides numerous examples that illustrate tight linkages between the landscape form and dominant erosion processes (Ahnert, 1987; Gilbert, 1909; Kirkby, 1985; Kirkby and Chorley, 1967), and show how the type and frequency and magnitude of sediment fluxes vary across scales of time and space (Benda and Dunne, 1997a, 1997b; Bierman et al., 2005; Kirchner et al., 2001; Meyer et al., 2001; Wolman and Miller, 1960). The unique role of river channels and their morphologies in regulating sediment output from basins in different climates have also been discussed (Benda and Dunne, 1997b; Bull, 1997; Schumm, 1999). These and many other studies in the geomorphology literature provide invaluable insights for...
improved predictions of erosion rates and basin sediment yields at a range of scales.

The soil erosion prediction technology arguably started with the Universal Soil Loss Equation (USLE, Wischmeier and Smith, 1965, 1978) that was developed based on over 10,000 plot-years of runoff and erosion measurements on plots of 22.13 m length. The USLE equation relates erosion to a series of indices based on soil properties, conservation practices, vegetation cover condition, local slope, and mean climatology of the area. Because of its simplicity, the USLE equation has been widely used in practice and implemented, with some modifications, in various numerical models of farm and range-land erosion (Renard et al., 1994; Williams et al., 1984). However, these models only represent sediment transport capacity due to a combination of rainsplash and overland flow erosion, which are assumed to describe processes only at the hillslope scale.

With the pioneering work of Foster and Meyer (1972), erosion models began adapting shear stress and stream power based equations developed for sediment transport and erosion for river channels, and applying those at the scale of agricultural and range lands using various methods for flow routing. Their effort effectively extended the scale of erosion modeling from small, rainsplash-dominated field plots to small catchment scales. This idea has led to the development of more mechanistic erosion models that couple models of watershed hydrology (mostly based on infiltration excess runoff generation) with Foster and Meyer (1972) equations at the field-scale (100–1000 ha), such as CREAMS (Knisel, 1980) and SWRRB (Williams et al., 1985), and at the basin-scale (>1000 ha), such as ANSWERS (Beasley et al., 1980), AGNPS (Young et al., 1989), WEPP (Nearing et al., 1989), CAS2D (Ogden and Heilig, 2001), LSEM (De Roo et al., 1995), EROSION 3D (Schmidt et al., 1999), and KINEROS (Woolhiser et al., 1990). Digital elevation models were later adopted by most modeling groups to represent landscape topography and used to route flow and sediment and identify areas of net erosion and deposition (De Roo, 1998; Morgan and Quintron, 2001).

Most of the aforementioned existing erosion models were developed as short-term decision-making tools for agricultural and range-land management. Two notable limitations of these models are: they lack channel erosion processes, which could account for a significant portion of basin sediment yields; and, with the exception of a few research models (e.g., Mitas et al., 1997), they use a fixed landscape elevation field represented by a DEM, or cascading planes that collectively represent the shape and slope of a basin. As a result, there is no feedback between erosion processes and the landscape form. This presents a critical limitation to model predictions especially for basins with a well-defined valley network where episodic deposition and erosion events could occur.

At the other end of the spectrum, landscape evolution models (LEMs) were designed to examine the connections between geomorphic processes and resulting landforms (e.g., Tucker and Bras, 1998; Veneziano and Niemann, 2000; Willgoose et al., 1991), and study the topographic outcome of erosion-uptilt interplay over geologic, millennial time scales (Whipple and Tucker, 2002). Few LEM studies concentrated on the role of vegetation (Collins and Bras, 2008; Collins et al., 2004; Istanbululluoglu and Bras, 2005) and subsurface flow hydrology (Huang and Niemann, 2006); overall, LEMs often simplify or completely disregard the dynamics of environmental variables and hydrology over the time scales of years to decades (Martin and Church, 2004).

At the intermediate time scales, decadal to centennial, memory in basin hydrology (soil moisture, subsurface, and groundwater flow), geomorphology (e.g., drainage patterns), and vegetation can affect the watershed geomorphic response (e.g., Greenway, 1987). Many observational studies underscore the importance of accurate representation of basin hydrological response in predicting sediment fluxes: due to a strong nonlinear dependence of sediment entrainment and transport on runoff discharge, inaccuracies in runoff prediction can have an amplified influence on predicted erosion (Bennet, 1974; Rojas and Woolhiser, 2000; Wainwright and Parsons, 1998). Only few LEMs operating at the intermediate time scales have been adapted to integrate a more sophisticated representation of hydrological processes (Hancock et al., 2011). For example, a grid-based model of Coulthard et al. (2002, 2005), CAESAR, introduced a topography-driven, quasi-steady-state description of the hydrological response. Operating at a fine spatial scale, the model has the capability to include effects of different vegetation covers and soil types.

Climatic extremes (both floods and droughts) have increased significantly in the past century (Seager et al., 2007). Future climate predictions further suggest even more extreme climates in some parts of the globe (Bates et al., 2008; Christensen et al., 2007). Prediction of erosion rates in transient climate conditions will require realistic representations of the coupled geomorphic, hydrologic, and ecologic responses that are relevant to the time and space scales of a river basin. The objective of this study is to present a framework that seamlessly integrates a geomorphic component to a hydrologic model. The framework is formulated to be suitable for addressing the coupled response both at the short-term and intermediate time scales, i.e., hours to sub-millennial time intervals. The hydrologic model is the physically-based, spatially distributed TIN-based Real-time Integrated Basin Simulator (tRIBS) (Ivanov et al., 2004a). The geomorphic component is mainly based on CHILD, the Channel-Hillslope Integrated Landscape Development model (Tucker et al., 2001a). Model calibration has been carried out using streamflow, sediment yield, and meteorological data for the Lucky Hills basin, a semi-arid watershed nested in the Walnut Gulch Experimental Watershed (Arizona, USA).

The paper is organized as follows. Section 1 provides a brief introduction of the main aspects concerning the modeling of soil erosion. Section 2 describes the new geomorphic component embedded into the tRIBS model. Section 3 introduces the Lucky Hills basin and describes the data used for the model calibration (Section 4). Section 5 presents a first application to two synthetic domains and its results, which are discussed in Section 6.

2. Model formulation

2.1. Overview of hydrological model

This paper presents an integration of a geomorphic component into an existing spatially-distributed physically-based hydrological model, the TIN (Triangulated Irregular Network)-based Real-time Integrated Basin Simulator, tRIBS (Ivanov et al., 2004a, 2004b). tRIBS reproduces essential hydrologic processes over complex topography of a river basin. The model explicitly considers spatial variability in precipitation fields and land-surface descriptors and the corresponding moisture dynamics. The model stresses the role of topography in lateral soil moisture redistribution by accounting for the effects of heterogeneous and anisotropic soil. It has been recently used as framework to develop a simulation of rainfall-triggered landslides (Arnone et al., 2011). A brief outline of the implemented process parameterizations is described in the following.

1) For simulating the precipitation interception, the Rutter canopy water balance model (Rutter et al., 1971, 1975) is used at the hourly time step. Canopy water dynamics are formulated to be species dependent, such that model parameters vary for different vegetation types.
2) Surface energy budget is simulated at each computational element at the hourly scale; shortwave and longwave radiation components are obtained accounting for a geographic location, time of year, aspect and slope of the element surface, and atmospheric conditions (e.g., Bras, 1990). The Penman–Monteith equation (Monteith, 1965; Penman, 1948), the gradient method (Entekhabi, 2000), and
the force–restore (Hu and Islam, 1995; Lin, 1980) methods are used to estimate the latent, sensible, and ground heat fluxes at the land surface. An optimum surface temperature is sought that leads to the energy balance closure. A species-dependent parameterization of stomatal conductance allows for diurnal variation of transpiration flux.

3) Latent heat flux is partitioned into evaporation from wet canopy, vegetation transpiration, and bare soil evaporation; the latter two are limited by available moisture in the soil zone, depending on vegetation fractional coverage of an element and canopy state.

4) For simulating the process of infiltration (minute resolution), an assumption of gravity-dominated flow in a sloped, vertically heterogeneous, anisotropic soil is used. The evolution of the wetting and top fronts in an element may lead to unsaturated, perched, surface, and completely saturated states. The unsaturated and the saturated zones are coupled, accounting for the interaction of the moving infiltration front with a variable water table. Topography and soil control the magnitude of the lateral moisture transfer in the unsaturated zone. Continuous soil moisture permits handling both storm and interstorm periods, thus allowing long-term simulations over a range of hydrometeorological forcings.

5) For computing the groundwater dynamics, a model based on the Boussinesq’s equation under the Dupuit–Forchheimer assumptions is used, allowing for a lateral water redistribution in the saturated zone and its dynamic interactions with the unsaturated zone. The lateral exchanges between contiguous elements are calculated by using the depth-averaged aquifer transmissivity, the flow width and the local water table slope.

6) A snowpack dynamic model has been recently added (Rinehart et al., 2008) that permits the simulation of energy and mass budgets of snow-covered areas.

7) Runoff generation is made possible via four mechanisms: saturation excess (Dunne and Black, 1970), infiltration excess (Horton; 1933; Loague et al., 2010), perched subsurface stormflow (Weyman, 1970), and groundwater exfiltration (Hursh and Brater, 1941). The runoff is obtained by tracking the infiltration fronts, water table fluctuations, and lateral moisture fluxes in the unsaturated and saturated zones. The runoff computed in subsurface module at the minute scale is used as input to the geomorphic model.

In order to represent basin characteristics, the model uses a multiple-resolution approach based on a TIN. The primary motivation for the using TINs is the multiple resolutions offered by the irregularly spaced nodes. This offers a flexible computational structure that reduces the number of computational elements without a significant loss of information (Vivoni et al., 2004) and translates to computational savings. Another advantage is that linear features can be precisely preserved in the mesh which allows mimicking of terrain breaklines, stream networks, and boundaries between heterogeneous regions. The computational elements of the coupled model are regions derived as the dual representation of the triangular mesh, i.e., in the form of Voronoi cells. Basin hydrological response can be simulated at very fine temporal (minutes to hour) and spatial (10–100 m) resolutions. Further details of model computational basis, structure, and descriptions of process parameterizations are presented in Ivanov et al. (2004a). A discussion of the model performance for mid-sized basins is provided in Ivanov et al. (2004b).

2.2. Coupling of hydrological and geomorphic dynamics

tRIBS calculates physical processes such as interception, infiltration, evapotranspiration as well as runoff using the Voronoi polygon network (VPN). It is assumed that Voronoi-based quantities that serve as input to the geomorphic model, i.e., precipitation, throughfall, canopy drainage (Section 2.3), and runoff rates (Section 2.4) are uniformly distributed inside each Voronoi cell. Possible subgrid variability of fluxes is implemented by using “mosaic” areal fractions of, for example, vegetation, throughfall, and bare soil (Section 2.3).

The VPN-based, locally produced runoff follows edges of a watershed TIN in accordance with pre-determined drainage directions (boundaries between Voronoi regions serve as interfaces between adjoining cells, see also Section 2.5). Unlike many other erosion models used in practice, the presented framework explicitly evolves landscape elevations through time (see Section 2.5), presenting a dynamic feedback loop between hydrology and geomorphology. For example, changes in elevation will impact local slopes, which may lead to formation of a different drainage graph and thus a different pattern of flow accumulation in a basin; locally modified site exposure (i.e., aspect) will imply changes in surface irradiance and therefore energy and latent heat partition; deepening of gullies may result in stronger connection of groundwater dynamics to surface processes, etc. Currently, runoff is assumed to propagate downstream within the step of 1 h, which limits the scale of application of the model to headwater basins.

Using hydrological quantities averaged at the hourly scale as input, the geomorphic component formulates flux equations of rainfall detachment and sheet erosion entrainment combined with the continuity of mass to compute sediment dynamics across the landscape. The rate of change in landscape elevations, \( \frac{\partial z}{\partial t} \), is predicted by the smaller of local erosion capacity or divergence of sediment flux (i.e., the excess sediment transport capacity, Tucker and Slingerland, 1997; Tucker et al., 2001b):

\[
\frac{\partial z}{\partial t} = -\text{Min}[D_q, \nabla q_s],
\]

where \( \text{Min} \) is the minimum operator, \( D_q \) [L/T] is the detachment/entrainment capacity (i.e., this work describes sediment detachment by rainfall and entrainment by flowing water), \( q_s \) [L^2/T] is the sediment load of overland flow and \( \nabla \) is the divergence operator, i.e., the sum of incoming sediment flux minus the sediment transport capacity, \( q_s \), at a given location. The term \( D_q \) represents a combination of the sediment detached due to raindrop impact (rainsplash) and the sediment entrained in the sheet overland flow. In the application of this equation, similar to runoff, sediment is routed on a cell-by-cell basis using the steepest descend flow direction. Detached/entrained sediment of an upstream Voronoi element (or multiple elements) enters a downstream cell and is used in calculating the divergence of the sediment flux. In a continuous application of the geomorphic model to a watershed domain that has a spatially variable structure of hydrological states, the transported sediment accumulates or moves downstream and landscape elevations consequently change as a result of these processes (Section 2.5). The exact geomorphological functionality introduced in the geomorphic model by considering the two mechanisms, detachment by raindrop impact and entrainment by sheet flow erosion, is presented in the following.

2.3. Raindrop impact detachment

Raindrops break cohesive bonds between soil particles detaching them from soil and making them available for transport. This phenomenon is influenced by rainfall and soil characteristics, ground and canopy cover, and water flow depth over soil. In order to develop model with the properties of parsimony and robustness, a well-tested approach is implemented here, as compared to more complex approaches that have only been tested for idealized conditions (e.g., Kinnell; 2005; 2006). In particular, the conceptual approach of Wicks and Bathurst (1996) is used to assess the rate of soil detachment by raindrop, \( D_{\text{DR}} \) [M L^{-2} T^{-1}]

\[
D_{\text{DR}} = k_F \bar{w}_W [C_{\text{DR}} M_{\text{DR}} + C_p M_p],
\]

where
where \( k \) [\( \text{L}^{-1} \)] is the raindrop soil erodibility and \( F_w \) is the shield effect of surface water. Rain action is split between the action due to the direct raindrop impact and the effect of leaf drip. The terms \( M_8 [\text{M}^2\text{T}^{-2}] \) and \( M_9 [\text{M}^2\text{T}^{-3}] \) are the rainfall squared momentum and the leaf drop squared momentum, respectively (Salles et al., 2000; Styczen and Hogh-Schmidt, 1988), which cause the splash of the soil particles into the air. The variables \( C_R \) and \( C_D \) are weighted areal fractions that quantify the direct raindrop and leaf drip impacts on rainsplash detachment respectively.

The fraction of the canopy protected against drop erosion, \( V \) is the vegetative fraction of a Voronoi cell, and \( p \) is the throughfall coefficient for the vegetation fraction, i.e., the fraction of rainfall over canopy not intercepted by vegetation. In Eq. (3), the terms \( (1 - C_R) \) and \( (1 - C_V) \) represent the splash erosion in the bare soil and in the vegetated soil fraction of the Voronoi cell, respectively.

The fraction \( C_R \) is modeled according to:

\[
C_R = (1-C_R)/(1-V) + pV, \tag{3}
\]

where \( C_R \) is the fraction of the Voronoi cell protected against drop erosion, \( V \) is the vegetative fraction of a Voronoi cell, and \( p \) is the throughfall coefficient for the vegetation fraction, i.e., the fraction of rainfall over canopy not intercepted by vegetation. In Eq. (3), the terms \( (1 - C_R) \) and \( (1 - C_V) \) represent the splash erosion in the bare soil and in the vegetated soil fraction of the Voronoi cell, respectively.

The fraction \( C_D \) is modeled according to:

\[
C_D = F_l [1-C_R]/(1-p)V, \tag{4}
\]

where \( F_l [0÷1] \) is the fraction of rainfall intercepted by canopy, \( (1 - p) \), that reaches the soil in the form of leaf drip. The variable \( F_l \) is defined as the fraction of intercepted rainfall over canopy not intercepted by vegetation. In Eq. (4), the terms \( (1 - C_R) \) and \( (1 - C_V) \) represent the splash erosion in the bare soil and in the vegetated soil fraction of the Voronoi cell, respectively.

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2.4. Overland and channel transport and deposition

The model uses the shear stress-based formulations for the entrainment and transport of sediment by runoff discharge (Nearing et al., 1999; Yang, 1996), making no distinction between overland flow and channel flow as the form of the shear-stress based formulations for both cases are nearly identical. This assumption is consistent with field observations that in most landscapes overland flow often quickly forms rills once an entrainment threshold is exceeded. In addition, at the size of elements typically used in simulations, it is difficult to distinguish rills from overland flow features. No subgrid parameterization to represent rill erosion is used at this time. Therefore we calculate flow per unit width of a Voronoi edge, which more realistically represents overland flow according to the accepted assumption of sheet flow.

The effective boundary shear stress, \( \tau \) [\( \text{ML}^{-2}\text{T}^{-2} \)], is calculated according to the well known power function of local discharge and slope:

\[
\tau = k_q q^{n_v} S^{n_s}, \tag{5}
\]

where \( q \) [\( \text{L}^3\text{L}^{-1} \)] is the local discharge per unit width of Voronoi edge (for any given Voronoi cell, it is obtained by summing the outfluxes of topographically upstream elements that contribute to that cell following the surface topographic gradient divided by the width), \( S \) [\( \text{rad} \)] is the local slope, \( m_q \) and \( n_q \) are empirical parameters (Istanbulluoglu et al., 2004; Willgoose et al., 1991). Assuming locally uniform overland flow conditions and using Manning's equation for the flow velocity, \( m_q \) and \( n_q \) are equal to 0.6 and 0.7, respectively (Tucker et al., 2001a). The variable \( k_q \) [\( \text{ML}^{-2}\text{T}^{-2} \)] depends on the Manning's coefficient \( n_q \) through the expression (Istanbulluoglu et al., 2005; Simons and Senturk, 1977; Tucker et al., 2001b):

\[
k_q = \rho_w g n_q^{1.5} (n_s + n_q)^{0.5}, \tag{6}
\]

where \( \rho_w \) is the water density [\( \text{M}/\text{L}^3 \)], \( g \) is the acceleration of gravity [\( \text{L}/\text{T}^2 \)], \( n_q \) is the Manning's roughness coefficient for soil, and \( n_v \) is the Manning's roughness coefficient for vegetation calculated following Istanbulluoglu et al. (2005):

\[
n_v = n_{vCF} \left( \frac{V}{V_R} \right)^{\alpha}, \tag{7}
\]

where \( n_{vCF} \) and \( V_R \) are the Manning's roughness coefficient and the vegetation fraction for a reference cover condition of a selected vegetation type, respectively, and \( \alpha \) is a parameter. The variables \( V_R \) and \( \alpha \) here are assumed to be equal to 0.95 and 0.5, respectively for understory vegetation such as grass or shrub. Istanbulluoglu et al. (2005). Eq. (6) assumes site roughness \( n_q \) to be composed of two components related to the bare soil and vegetated fractions. The addition of vegetation roughness in the numerator as an additive term implies the partitioning of total shear stress on the bare soil: as \( n_v \) increases, \( \tau \) would reduce as a result of vegetation taking up a higher fraction of the total shear stress. For a detailed formulation, see Istanbulluoglu and Bras (2005). The spatial variation of \( n_{vCF} \) and \( n_v \) inside the domain can therefore be taken into account (Fig. 2).
The flow entrainment capacity, $D_{f}[L/T]$, is calculated as:

$$D_f = k_b(\tau - \tau_c)^{p_b},$$

(8)

where $k_b$ is the soil erosion efficiency coefficient [L/T (ML$^{-1}$ T$^{-2}$)$^p_b$], $\tau_c$ [M/LT$^2$] is a threshold stress for particle entrainment (i.e., "the critical shear stress"), and $p_b$ is an empirical parameter equal to 2.3 (Nearing et al., 1999).

The transport capacity, $D_T[L^3/T]$, is calculated as:

$$D_T = W k_f (\tau - \tau_c)^{p_f},$$

(9)

where $W$ [L] is the channel width, considered to be equal to the width of Voronoi edge (Tucker et al., 2001b, Fig. 7), $p_f$ is a parameter equal to 2.5, and $k_f$ is a coefficient considered for the transport of a single sediment size-fraction and calculated using the following expression (Simon and Senturk, 1977; Yalin, 1977):

$$k_f = k_s \sqrt{\frac{g(s-1)d^3}{(s-1)\gamma d^2}},$$

(10)

where $s$ is the ratio of sediment density to water density (taken as 2.65) and $d$ is the dominant grain size [L], often taken as $d_{50}$, i.e., the diameter corresponding to the 50% value of the granulometric distribution curve, and $k_s$ is a calibration coefficient. The values of $k_s$ are reported to be in the range of 4–40 in different studies (Yalin, 1977).

### 2.5. Computational steps

An outline of a numerical procedure for the solution of Eq. (1) as implemented in the model is presented below. Similar to locally produced runoff (Ivanov et al., 2004a), eroded material is assumed to follow the TIN edges in accordance with the consecutive drainage directions. The boundaries between Voronoi regions define the interfaces between adjoining cells. When a mass moves into a neighboring cell, the length of a given interface is used as the width in the flux computation (Ivanov et al., 2004a; Tucker et al., 2001a, 2001b). Both runoff and sediment are assumed to propagate downstream within the step of 1 h, limiting the scale of application of the model to headwater basins.

Calculation starts at the Voronoi cell with the highest elevation and proceeds downstream to the basin outlet cell:

1. For each computational element the rate of soil detachment by raindrop $D_R$ and the entrainment capacity rate $D_T$ are calculated using Eqs. (2) and (8), respectively.
2. The transport capacity rate $D_T$ is calculated with Eq. (9).
3. For each cell a potential rate of transport-limited erosion is calculated using the control volume approach. The continuity of mass equation can be written for any cell $i$ as

$$\frac{\Delta z_i}{\Delta t} = \frac{1}{(1-\eta)} \frac{\partial D_T}{\partial x}$$

where $\eta$ is the bed sediment porosity and $z_i$ is the elevation at node location $i$. Assuming $\frac{\partial z_i}{\partial x} \approx \frac{\Delta z_i}{\Delta x} \Delta t$ is solved using the forward differencing in time:

$$\frac{\Delta z_{i,\text{pot}}}{\Delta t} = \frac{1}{1-\eta} \left( \sum_j^n q_{i,j} \right) - D_{T,i},$$

(11)

where $q_{i,j}$ is the incoming sediment flux for the cell $i$, $m$ is the number of upstream nodes that flow directly to node $i$ and $A_i$ is the Voronoi area of node $i$. The resulting elevation change, $\Delta z_{i,\text{pot}}$, corresponds to the potential change. It is evaluated for each cell.

4. For each cell, the rate of detachment/entrainment-limited erosion is calculated based on the sum of rates $D_R$ and $D_T$. This gives the maximum rate of sediment, $\frac{\Delta z_{i,\text{ava}}}{\Delta t}$, that can be removed from the cell, conditioned by the local soil erodibility and the actual shear stress, even under infinite transport capacity. The potential rate of detachment/entrainment-limited erosion is:

$$\frac{\Delta z_{i,\text{ava}}}{\Delta t} = - \left( \frac{D_R + D_T}{\rho_s} \right).$$

(12)

where $\rho_s$ [M/L$^3$] is the soil density. $\Delta z_{i,\text{ava}}$ is always smaller than zero.

5. Finally the two rates are compared with the following logical statements:

a. If $\Delta z_{i,\text{pot}} > 0$, then a deposition occurs and $\Delta z = \Delta z_{i,\text{pot}}$;

b. If $\Delta z_{i,\text{ava}} < \Delta z_{i,\text{pot}} < 0$ (erosion), then $\Delta z = \Delta z_{i,\text{pot}}$ and a transport-limited erosion occurs;

c. If $\Delta z_{i,\text{pot}} < \Delta z_{i,\text{ava}} < 0$ (erosion), then $\Delta z = \Delta z_{i,\text{ava}}$ and a detachment/entrainment-limited erosion occurs.

d. In the event that $|\Delta z_{i,\text{pot}}| > |\Delta z_{i,\text{ava}}|$, the transport capacity rate of the flow is not at the maximum value (i.e., detachment/entrainment-limited) and upon calculation of the actual amount of erosion/deposition in the cell, the outgoing sediment flux from the cell into a downstream cell is re-defined as

$$D_{T,i} = \left( \sum_j^n q_{i,j} \right) - A_i(1-\eta) \frac{\Delta z_{i,\text{ava}}}{\Delta t}.$$  

(13)

6. At the hourly scale, the model updates the elevation of the entire VPN, as well as re-sorts nodes following the topography-dictated network order. The latter is determined based on local maximum surface slopes (Ivanov et al., 2004a) and thus leads to a continuously updated drainage pattern. An update of all of the above terrain elements contributes to feedback from geomorphic processes of erosion and deposition to the watershed hydrological dynamics.

### 3. Lucky Hills watershed description and data

Data used in this study were collected in the Lucky Hills watershed #103, which is nested within the Walnut Gulch Experimental Watershed (WGEW) in southeastern Arizona, USA (31° 43’ N, 110° 41’ W) (Renard et al., 1993). The watershed, located near the town of Tombstone, is approximately 3.7 ha (0.037 km$^2$) in size and its elevation is between 1364 and 1375 m above sea level (Fig. 3). Average annual precipitation is approximately 300 mm, 70% of which falls during the summer monsoon between the months of July and September (Tiscareno-Lopez, 1991); much of the remaining precipitation is concentrated in the winter months of December through February (Nichols et al., 2002). The typical storms tend to be very localized and of short duration. The mean annual temperature is 18°C with the average monthly maximum temperatures of 35°C in June, and the average monthly minimum temperatures of 2°C in December. Vegetation in the area is dominated by desert shrub and semi-arid rangeland plants. The dominant vegetation types are Creosote bush (Larrea tridentata, shrub) at 2 to 5 m spacing (26% cover), and Whitethorn (Acacia constricta, shrub), with lesser populations of Desert Zinnia (Zinnia acerosa, shrub), Tarbush (Flourensia cernua, shrub), and Black Grama (Bouteloua eriopoda, grass). Larger Creosote shrubs are about 1 m tall and can be characterized by a spatially averaged leaf-area index value of 0.4 (Flerchinger et al., 1998). Canopy cover during the rainy season is approximately 25%. Approximately 50% of the ground area is covered with rock, an alluvial outwash consisting predominantly of gravel and cobble sized material, and the remaining 25% is bare soil. The dominant soil type is the
McNeal Gravelly Sandy Loam (60% sand, 25% silt, and 15% clay) with approximately 25% of rock fragments in the surface layer. The rock content of the soil profile is 28% by volume between 0 and 5 cm and then decreases with depth (Hymer et al., 2000). Table 1 reports a summary of the principal characteristics of the watershed (Nearing et al., 2005).

The basin morphology consists of a knick-point type drop (~10 m) in the middle of the flow path (Fig. 3). Runoff and sediment are measured with calibrated, Santa-Rita type supercritical flumes and traversing slot sediment samplers at the outlet of the watershed. Sediment is sampled by an automatic traversing slot sampler, which takes a depth-integrated sample at specified time intervals within a flow event (Renard et al., 1986). Runoff and sediment yield result almost exclusively from convective storms during the summer season. Event runoff and sediment discharge from 1999 to 2009 were collected at the location FL103 (Fig. 3). Rainfall data from 1999 to 2009 were collected at the rain gage RG83, which is located inside the watershed (Fig. 3). The rain gage RG83 is a part of dense network of weighing-type recording rain gages used to capture the temporal and spatial variability of precipitation inside the Walnut Gulch watershed. Goodrich et al. (2008) describe the instrumentation and long-term precipitation database. Data were downloaded at http://www.tucson.ars.ag.gov/dap/.

4. Model setup and calibration

An essential component in evaluating reliability of hydrological and geomorphic models is validation against observed data. Historically, the model performance has been measured through comparisons with discharge and sediment yield time series at the catchment outlet (Flanagan et al., 1998; Ivanov et al., 2004a). There have been only few examples of model validation based on the interior catchment information because of the typical lack of information and the limitations of most models to use such information (Houser et al., 1998; Ivanov et al., 2004a, 2004b; Refsgaard, 1997; Senarath et al., 2000).

4.1. Data

Climate, hydrology, digital elevation model (DEM), and land use data necessary for calibrating the model, were collected for the Lucky Hills watershed #103 (Nearing et al., 2005). The calibration period extends from mid-July, 1999, to mid-August, 2009. Rainfall is assumed to be uniformly distributed over the entire basin. Data for the rain gage RG83 (Fig. 3) were used. Original rainfall data collected at the resolution of 1 min were aggregated to 15-minute resolution data (Fig. 4). Fig. 4 also illustrates average daily cycles of other climatic data used in simulations and collected at an hourly time step from a flux tower near the Lucky Hills basin (station BREB in Fig. 3). They were assumed to be representative for the entire basin area as well. Finally, runoff volumes and sediment yield for each event occurred during the calibration period were observed at the outlet section FL103 (Fig. 3) of the watershed.

Spatially uniform soil and land use characteristics were assumed, i.e., the vegetation fraction, \( v \), the saturation soil moisture content \( \theta_s \), the residual soil moisture content \( \theta_r \), the saturated hydraulic conductivity, \( K_s \), and the conductivity decay parameter, \( f \). The soil hydraulic parameter values were inferred from measurements of Schaal and Shouse (2004). Some of these parameters were subsequently calibrated as described in the following and in Francipane (2010).

Table 1

<table>
<thead>
<tr>
<th>Characteristic</th>
<th>Lucky Hills watershed</th>
</tr>
</thead>
<tbody>
<tr>
<td>Area [km²]</td>
<td>0.037</td>
</tr>
<tr>
<td>Annual rainfall [mm]</td>
<td>300</td>
</tr>
<tr>
<td>Land use/plant community</td>
<td>Shrub dominated rangeland</td>
</tr>
<tr>
<td>Plant cover [%]</td>
<td>25%</td>
</tr>
<tr>
<td>Soil type</td>
<td>gravelly sandy loam</td>
</tr>
</tbody>
</table>

4.2. Spatial–temporal scales of calibration

Since the Lucky Hills watershed is small (3.7 ha) and has no nested gages, calibration was carried out for the entire basin area using data for the outlet FL103 (Fig. 3). Manual streamflow and sediment yield calibration were carried out. It was a stepwise approach that included the analysis of a number of variables considered at different spatial and temporal scales.

According to Yapo et al. (1998), when the objective of a geomorphological model is to study the landscape morphology over the time scales of hundreds of years, it is important to ensure that the model reproduces correctly the runoff and sediment yield volumes.
at large and continuous scales, rather than only at the event scale. Yapo et al. (1996) suggest that a calibration period of at least 8 years is needed to produce results that are independent of the selected period. Calibration over a 10-year period (mid-July, 1999 to mid-August, 2009) was carried out in this study. A previous calibration effort of the model at the event scale was done by Arnone et al. (2008).

4.3. Calibration

The approach used in calibration targeted satisfactory model performance over the long-term rather than over short-term time scales. Two metrics of the system were used in calibration in a sequential mode: streamflow and sediment yield.

4.3.1. Streamflow

Runoff regime of the Lucky Hills watershed is typical of many semiarid regions where channels are dry for most time of the year. Typically, runoff occurs as a result of intense thunderstorm rainfall. The flood peak arrives very quickly after the start of generation, and runoff duration is short (Keppel and Renard, 1962). It is primarily produced by the infiltration-excess component (Chorley, 1978). TRIBS can reproduce such a mechanism, if hydrometeorological conditions are conducive to runoff generation. The initial water table position, which affects the initial soil moisture by capillary rise, has been arbitrarily set to the depth of 50 m. This assumption is consistent with observations of very deep groundwater in the region.

In this study, the calibration of the model was carried out with respect to several parameters, such as $K_s$, $\theta_s$, $\theta_r$, and $f$. These are the most sensitive model parameters (Ivanov et al., 2004b). The mass curves of observed and simulated runoff volumes for the entire simulation period are shown in Fig. 5(a). The model is capable of reproducing main runoff events, even though capturing each individual event accurately may be problematic in some instances. As Fig. 5(a) shows, over the long-term, the fit between the observed and simulated accumulated volumes is “good”, according to the criteria provided by...
4.3.2. Sediment yield

The calibration of the sediment production/deposition model was carried out after the calibration of hydrological processes. Overall, the following parameters were adjusted during calibration: the soil erodibility coefficient, \( k_s \), the critical shear stress, \( \tau_c \), the Manning’s roughness coefficient for the soil, \( n_s \), and the Manning’s roughness coefficient for a vegetation of reference, \( n_{vR} \).

Some of the most sensitive parameters are related to the formulation of shear stress exerted by sheet flow on soil. Tiscareno-Lopez et al. (1994) define the effective shear stress, \( \tau_f \), acting on the bare soil surface as a fraction of the total shear stress, \( \tau \), due to shear stress partitioning between soil and vegetation:

\[
\tau_f = \tau \left( \frac{n_s}{n_s + n_{vR}} \right)^{0.5}.
\]

The value of \( n_s \) can be related to the median sediment diameter, \( d_{50} \), using empirical relationships that have the following general form (Yen, 1992)

\[
n_s = k \cdot d_{50}^{0.5},
\]

where \( k \) is a constant, considered to be equal to 0.0474 for sandy soils (Yen, 1992), and \( p \) is reported to be equal to 1/6. Considering \( d_{50} \) to be within the range of 0.5 mm and 50 mm, typical of sand and gravel, \( n_s \) can be assumed ranging between 0.013 and 0.03 m\(^{-1/3}\)/s. Ranges of vegetation roughness, \( n_v \), can be estimated from the lookup tables of the Manning’s roughness coefficient \( n (= n_s + n_v) \) for overland flow for a variety of land use conditions (Engman, 1986; Woolhiser et al., 1990). For a typical range of \( n_v \), for medium to dense brush, \( n_v \) can be assumed to be between 0.03 and 0.05 m\(^{-1/3}\)/s. In this study, values of \( k_s \) and \( p \) provided by Nearing et al. (1999) were used as the initial values in the entrainment capacity Eq. (8) when calibration was carried out for the Lucky Hills basin.

Accumulated sediment yield volumes for the entire simulation period are shown in Fig. 5(b). The model is capable of reproducing main events but, as can be seen in the figure, for some, it does not reproduce the volumes perfectly. Fig. 5(b) also shows that the fit between the observed and simulated data is acceptable and representative of the long-term process. The \( R^2 \) and \( E_r \) indicate that the fit is “satisfactory” to “good” (Table 3) for daily values, according to the criteria provided by Moriasi et al. (2007). The mass balance error, \( M_b \), suggests that on average the observed sediment yield is slightly overestimated by the model (Table 3).

5. Response of a catchment to different morphologies and predominant mechanisms of erosion

5.1. Geomorphic initialization conditions

The patterns of erosion and deposition in the landscape are largely controlled by the topographic structure and the frequency and magnitude of rainfall events. The performance of the model was further examined for two synthetic catchments with different hillslope and channel characteristics. The basins had been previously constructed with the CHILD landscape evolution model (Tucker et al., 2001a) using either fluvial erosion or diffusive transport (e.g., soil creep), as predominant geomorphic processes. To simulate these two domains in CHILD, only the hillslope transport component was altered. All else being equal, using a smaller hillslope transport coefficient leads to longer channels and smaller and steeper hillslopes (Fig. 6), while using a higher hillslope transport coefficient leads to smaller channels and longer and gentler hillslopes (Fig. 7). These domains, first used in an ecohydrological study by Ivanov et al. (2008b), are of rectangular shape with an outlet in the southwest corner. In the following, the two synthetic basins will be referred to as the “concave”, CV, and
"convex", CX, domains, respectively. In total, 2,400 computational elements are used to represent each of the synthetic topographies that cover a wide range of slope magnitudes and aspects. The dimensions of the basic computational element, the hexagonal Voronoi element, are approximately 40 m×40 m.

The dominant forms of sediment transport process for the two basins can be inferred from the slope–contributing area (S–A) diagram (Fig. 8) (Yetemen et al., 2010). They characterize the initialization states of these watersheds for simulations carried out in this study. Specifically, two characteristic regions can be identified: region I with a positive S–A gradient that corresponds to hillslopes with lower drainage areas where diffusive erosion is predominant; region II is mainly dominated by the fluvial transport (Yetemen et al., 2010). In region II, channels appear to be identical but longer in the fluvial-erosion dominated domain (CV) than in the diffusion–erosion dominated domain (CX). The steepest slopes in the landscapes are located at the transition between regions I and II, where diffusive processes give way to fluvial erosion, marking the location of the valley head. According to the S–A relation in the figure, valleys begin with a smaller contributing area (the location is indicated by a gray vertical line) in the CV basin and with a higher contributing area (the location is indicated by a black vertical line) in the CX basin.

5.2. Objectives and design of experiments

Long-term simulations were designed using the CV and CX domains as geomorphic initializations. The objective was to test the model at larger, first-order catchment scale as well as assess the discriminative power of the model with respect to predominant geomorphic processes that are characteristic of these geometries. The same parameter values obtained in the calibration for the Lucky Hills basin were used, except for the soil erodibility (see the formulation of entrainment capacity (8) and Tables 4 and 5). Specifically, the synthetic domains were developed for climatic conditions (in terms of precipitation) nearly identical to that of the Lucky Hills site; however, a smaller erodibility was used. When the erodibility parameter obtained in the calibration was used in the model, the erosion response of these domains was very strong and leads to widespread

Table 5

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Description</th>
<th>Initial values</th>
<th>Source</th>
<th>Final values</th>
</tr>
</thead>
<tbody>
<tr>
<td>0</td>
<td>ID</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1</td>
<td>Ks</td>
<td>Saturated hydraulic conductivity</td>
<td>32.81 mm/h</td>
<td>C</td>
</tr>
<tr>
<td>2</td>
<td>θs</td>
<td>Soil moisture at saturation</td>
<td>0.46</td>
<td>C</td>
</tr>
<tr>
<td>3</td>
<td>θr</td>
<td>Residual soil moisture</td>
<td>0.0429</td>
<td>C</td>
</tr>
<tr>
<td>4</td>
<td>λ</td>
<td>Pore-size distribution index</td>
<td>0.3813</td>
<td>L</td>
</tr>
<tr>
<td>5</td>
<td>φb</td>
<td>Air entry bubbling pressure</td>
<td>–63 mm</td>
<td>L</td>
</tr>
<tr>
<td>6</td>
<td>f</td>
<td>Conductivity decay parameter</td>
<td>0.032 mm⁻¹</td>
<td>L</td>
</tr>
<tr>
<td>7</td>
<td>A1</td>
<td>Anisotropy ratio for saturated zone</td>
<td>1</td>
<td>L</td>
</tr>
<tr>
<td>8</td>
<td>An</td>
<td>Anisotropy ratio for unsaturated zone</td>
<td>1</td>
<td>L</td>
</tr>
<tr>
<td>9</td>
<td>n</td>
<td>Porosity</td>
<td>0.46</td>
<td>L</td>
</tr>
<tr>
<td>10</td>
<td>k</td>
<td>Volumetric heat conductivity</td>
<td>0.214 J/m·s·K</td>
<td>L</td>
</tr>
<tr>
<td>11</td>
<td>c</td>
<td>Soil heat capacity</td>
<td>1.209,573 J/m³·K</td>
<td>L</td>
</tr>
<tr>
<td>12</td>
<td>c0</td>
<td>Soil cohesion parameter</td>
<td>1.706 N/m²</td>
<td>L</td>
</tr>
<tr>
<td>13</td>
<td>c1</td>
<td>Root cohesion parameter</td>
<td>1000 N/m²</td>
<td>L</td>
</tr>
<tr>
<td>14</td>
<td>c2</td>
<td>Soil internal friction angle</td>
<td>20°</td>
<td>L</td>
</tr>
<tr>
<td>15</td>
<td>K0</td>
<td>Erodibility coefficient</td>
<td>3.77E−08 m/(kg m⁻¹ s⁻²)</td>
<td>C</td>
</tr>
<tr>
<td>16</td>
<td>c3</td>
<td>Critical shear stress</td>
<td>1.5 Pa</td>
<td>C</td>
</tr>
<tr>
<td>17</td>
<td>k</td>
<td>Coefficient k6 to calculate k</td>
<td>20</td>
<td>L</td>
</tr>
<tr>
<td>18</td>
<td>k8</td>
<td>Coefficient k6 to calculate k</td>
<td>20</td>
<td>L</td>
</tr>
<tr>
<td>19</td>
<td>k9</td>
<td>Diffusive coefficient</td>
<td>3.17E−10 m²/s</td>
<td>L</td>
</tr>
<tr>
<td>20</td>
<td>n0</td>
<td>Manning’s roughness coefficient for soil</td>
<td>0.025 m⁻¹/³/s</td>
<td>C</td>
</tr>
<tr>
<td>21</td>
<td>d</td>
<td>Dominant grain size</td>
<td>0.0005 m</td>
<td>L</td>
</tr>
</tbody>
</table>

L = from literature; C = from calibration.

Fig. 6. A map of vegetation fractional cover overlaid on surface topography of the CV domain.
erosion, indicating non-equilibrium conditions. In order to alleviate this effect of initial conditions, a smaller erodibility value consistent with the one used previously in the CHILD model was used.

Note that the model can distinguish the influence of vegetation on both overland flow and rainsplash erosion. In order to assure a more realistic effect of vegetation cover on the geomorphic dynamics of the two synthetic basins “placed” in a semiarid setting, a 30-year long average spatial distribution of vegetation cover fraction was obtained with the tRIBS+VEGGIE model (Ivanov et al., 2008a, 2008b). tRIBS+VEGGIE had been previously calibrated for the Lucky Hills site, so as to ensure the distribution of biomass consistent with the basin morphology and climate. Fig. 9 shows the time series of simulated soil moisture, as well as simulated and observed leaf area index (LAI) and biomass at the site. Figs. 6 and 7 illustrate the simulated distributions. The CV domain exhibits the spatial variation of vegetation with a fractional cover ranging between ~36% and ~44%. The CX domain has a vegetation fractional cover in the range of ~41.5÷44% over the entire domain. North facing hillslopes exhibit higher values of vegetation fractional cover for both domains. Such spatial patterns can be attributed to the local effects of shortwave irradiance and rainfall (Ivanov et al., 2008b). Vegetation initialized in this fashion was considered to be constant throughout the entire simulation period and not prone to the effect of erosion, i.e., vegetation fraction remained constant in time.

In order to robustly infer the metrics of local dynamics characterizing predominant geomorphic processes (such as local erosion/deposition rates), a long-term simulation is required since it would allow filtering out the effects of individual geomorphic events. However, observational records with data suitable for forcing a hydrological model are typically short. A scenario of climate forcing at the computationally feasible, intermediate time scale of 100 years was therefore generated using the Advanced WEather GENerator (AWE-GEN, Fatichi, 2010; Fatichi et al., 2011), an hourly-scale weather generator capable of reproducing low and high frequency characteristics of hydro-climatic variables (e.g., precipitation, air temperature, solar radiation, cloud cover, relative humidity, vapor pressure, wind speed, and atmospheric pressure) and their essential statistical properties. The climate scenario was based on meteorological variables observed at the Tucson airport over the period of 1960–2000. The generated realization assumes stationary climate corresponding to the observational period (see...
Fattichi et al., 2011 for details). Fig. 10 shows the statistics of the generated climate scenario.

5.3. Results

Using the generated climate, the developed model was run for 100 years and erosion patterns were compared for the two basins. The net change in the landscape elevations of the two simulated domains was inferred as the difference between the elevation at the beginning and the end (i.e., after 100 years) of the simulations (Figs. 11 and 12). Positive values indicate net erosion, while negative values indicate deposition. Fig. 11 shows erosion activity mainly concentrated in the stream network of the CV domain. This agrees with the more heterogeneous terrain of the initial state of the basin that was obtained by emphasizing fluvial erosion as the predominant process. The CX domain shows higher erosion activity in the hillslopes (Fig. 12). The erosion rates that can be inferred from Figs. 11 and 12 are consistent with the values reported in literature (Jha and Paudelb, 2010). They range between 0 and ~0.15 mm/year of deposition and between 0 and ~0.40 mm/year of erosion for the fluvial-erosion dominated domain (CV); between 0 and ~0.58 mm/year of deposition and between 0 and ~0.60 mm/year of erosion for the diffusion-erosion dominated domain (CX). Following the classification of Table 6 (Jha and Paudelb, 2010), almost 85% of the CX domain can be characterized by “very low” and “very high” erosion rates, with the remaining part of the basin corresponding to “low” to “high”...

Fig. 9. Calibration of TRIBS + VEGGIE for the Lucky Hills basin. The figure shows the time series of simulated soil moisture (top panel), leaf area index (LAI), and biomass for the period of 1997–2008. “MODIS” refers to the inferences of LAI based on data from the Moderate Resolution Imaging Spectroradiometer. The circle point at the bottom plot refers to the actual observations in the Lucky Hills basin (the data were obtained at http://www.tucson.ars.ag.gov/dap/).

Fig. 10. A statistical representation of the 100-year scenario of stationary climate simulated with the Advanced WEather GENerator (AWE-GEN) model of Fattichi et al. (2011): precipitation interannual variability (top), the probability of exceedance of hourly rainfall (middle row, left subplot), and the diurnal cycles of air temperature, relative humidity, wind speed, incoming longwave, and shortwave fluxes.
erosion; almost 95% of the CV domain can be described by “very low” erosion, with the remaining 5% equally attributed to “low” and “moderate” erosion rates. Only few locations in the CV domain are characterized by “high” and “very high” erosion.

The net effects of different erosive patterns are illustrated in Fig. 13. The figure shows the net change in elevation with respect to slope (left panel) and contributing area (right panel) for each computational element of the domain. The contributing area that corresponds to the maximum slope magnitude in the S–A plot (Fig. 8) designates the location of the channel head. Such locations can be identified in Fig. 13. Considering the values of contributing area where valleys begin (Fig. 8), it is possible to identify regions dominated either by diffusive processes (I) or by fluvial erosion (II) (Fig. 13). Hillslope erosion is always lower with respect to erosion in the stream network. Despite the smaller range of slopes, the CX domain exhibits a higher range of surface elevation changes. This indicates an evident dependence of erosion on slope (Fig. 13, left panel), typical of catchments with a strong diffusive component. On the contrary, the CV domain has a low variability of elevation changes corresponding to higher slope magnitudes, which signifies a low dependence of erosion on diffusive mechanism. The same behavior is reflected in Fig. 13 (right panel). The CX domain exhibits a higher variability of erosion in the region I, as compared to the CV domain, and both basins exhibit an almost log–log linear relationship of elevation change with contributing area for region II. One may notice that the region I exhibits different values of changes in elevation for the same value of slope in the CX domain case (e.g., red rectangle in Fig. 13, left panel) and contributing area (e.g., red rectangle in Fig. 13, right panel). This could be related to the influence of slope and aspect on the spatial distribution of vegetation in the basin (Fig. 7) and opens an interesting question about the influence of vegetation on the spatial distribution of erosion activity. Such inferences, however, are beyond the scope of this study.

The difference in initialization conditions and predominant erosion mechanisms affects the simulation results in terms of sediment yield. After 100 years of simulation, the CX domain leads to a
sediment yield that is about 20÷25% higher than that of the CV domain (Fig. 14). The higher sediment yield of the CX domain is likely a sign of the transient diffusive activity over the hillslopes that makes a larger amount of sediment available for transport.

6. Summary

A coupled hydro-geomorphic modeling approach to catchment evolution is presented in this study. The model integrates an existing physically-based, spatially-distributed hydrological model TIN-based Real-time Integrated Basin Simulator with the geomorphic functionality of a landscape evolution model. Besides hydrological processes, the coupled model simulates essential erosive processes of hillslopes, such as detachment due to raindrop impact and entrainment due to overland sheet flow, as well as channel processes, i.e., erosion and deposition due to the action of channelized flow. This work introduces a number of novel features, as compared to the current generation of erosion and landscape evolution models. They are summarized in the following.

- Although a proper description of hydrology, in particular, the runoff generation process, is crucial for estimating erosion/geomorphic response of a catchment, most erosion and landscape evolution models use over-simplified representations of hydrology. The model developed in this study enhances the traditional, discipline-focused approach of geomorphic simulation by merging it with a fully transient description of hydrological processes. The model has the capacity to improve predictability of the coupled hydro-geomorphic response associated with climate and environmental change.
- The representation of dynamic feedbacks between local erosion and elevation change in numerical models has been shown to improve erosion and deposition predictions. While this coupled phenomenon is represented in most landscape evolution models, such models are typically developed to address long-term landscape development questions, implying a number of simplifications. These involve coarse time steps with underlying gross integration of hydrological dynamics, an assumption of uniform rainfall rates, simplifications of environmental variability of land-surface properties, and limited capabilities of predicting the influence of geomorphic change on the watershed hydrologic response. The fine temporal and spatial resolutions allowing one to properly capture the consequences of complex rainfall events over heterogeneous surface, as well as the capability to incorporate geomorphic feedbacks into a hydrological simulation through changing elevation, slope, aspect, and drainage pattern, represent a unique feature of the developed framework. This is an important step towards, for example, modeling future climate consequences in soil and water resources management.
- From a technical perspective, the coupled model uses variable mesh resolution associated with TINs, which are not frequently employed in geoscience modeling. Both advantages and disadvantages of TINs have been discussed in numerous previous publications (e.g., Tucker et al., 2001a; Vivoni et al., 2004, among others). For example, TINs offer the flexibility in representing terrain and landuse variability, terrain breaklines, stream networks, and boundaries between heterogeneous regions but feature more cumbersome ways of storing representation information and
mapping procedures, such as precipitation mapping on VPN. Within the framework of the presented model, using a TIN presents interesting possibilities of dynamic re-meshing based on hydrologic and geomorphic dynamics, which might lead to further computational savings.

With an increased number of processes described in the coupled model, one apparent drawback is that the computational efficiency becomes much lower than that of the existing erosion and landscape evolution models. For example, while sub-millennium (hundreds of years) simulation time scales are feasible in a serial implementation with a modern-era personal computer, carrying out multi-millennial runs will require the capabilities of a parallelized architecture (e.g., Vivoni et al., 2011) and high-performance computing facilities.

The presented model was calibrated for the Lucky Hills basin, a sub-basin of the Walnut Gulch Experimental Watershed (Arizona, USA). The calibration procedure demonstrates a good capability of the model to reproduce main runoff events and the accumulated streamflow and sediment yield volumes over a long-term period. While the observed and simulated volumes do not perfectly match for every individual event, the fit over the long period is good and consistent in terms of temporal dynamics. The model thus represents a potentially robust tool for long-term assessments of change of geomorphic dynamics as a result of land use alterations in headwater basins, climate or vegetation change.

Two synthetic basins characterized by different morphological characteristics and initially generated so as to illustrate the predominance of different erosion mechanisms (fluvial and diffusive) were used in the evaluation simulations for a 100-year period of stationary climate. The changes in surface elevation demonstrate a consistent performance of the model, which is capable of identifying regions dominated by diffusive processes and fluvial erosion. These are also confirmed based on an analysis of the “change in elevation” vs. slope and contributing area diagrams.

The model application to the case study and to the synthetic domains has highlighted the following aspects:

- The developed model parsimoniously reproduces essential processes that regulate the phenomena of erosion and deposition; the model performance is consistent with assumed principles of behavior.
- The case study simulations are consistent with values reported in literature in terms of mean annual soil erosion rates; the fluvial erosion is the more efficient geomorphic process for the considered environment.
- The model is capable of differentiating between stream network and hillslopes and reproduces the relevant geomorphic processes, i.e., fluvial and diffusive erosion, when applied to first-order catchments developed under these conditions.
- The model can parsimoniously mimic the effect of vegetation on erosion.

Furthermore, the results obtained in simulations suggest a topographically differential effect (i.e., the dependence on site aspect and slope) of vegetation on erosion activity in a semiarid environment. In order to address these preliminary inferences, further efforts will enhance the model by introducing modeling of ecological processes that mimics growth and death of vegetation with erosion/deposition events.

Acknowledgment

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Appendix A

A.1. Soil erodibility coefficient, $k_r$

The values of soil erodibility coefficient, $k_r$ [J$^{-1}$], can be found in literature as a function of soil texture (Table A.1).

<table>
<thead>
<tr>
<th>Data source</th>
<th>Mean $k_r$ coefficient [J$^{-1}$]</th>
</tr>
</thead>
<tbody>
<tr>
<td>Clay</td>
<td>19.0</td>
</tr>
<tr>
<td>Silt clay</td>
<td>18.2</td>
</tr>
<tr>
<td>Silty clay</td>
<td>16.2</td>
</tr>
<tr>
<td>Silt loam</td>
<td>29.8</td>
</tr>
<tr>
<td>Silt clay</td>
<td>29.8</td>
</tr>
<tr>
<td>Loam</td>
<td>39.8</td>
</tr>
<tr>
<td>Sandy loam</td>
<td>28.2</td>
</tr>
<tr>
<td>Sand</td>
<td>32.0</td>
</tr>
</tbody>
</table>

A.2. Shield effect of surface water, $F_w$

The shield effect factor, $F_w$, is evaluated as a function of the surface water depth $h$ [L] and the median raindrop diameter $D_m$ [L] as follows:

$$F_w = \begin{cases} \exp(1-h/D_m), & \text{if } h > D_m \\ 1, & \text{if } h \leq D_m. \end{cases}$$

The median raindrop diameter $D_m$ is based on an empirical relationship of Laws and Parson (1943): $D_m = a \cdot I_c$, where $a = 0.00124$ and $c = 0.182$ are the empirical coefficients.

A.3. Leaf drip squared momentum, $M_{lp}$

The term $M_{lp}$ can be expressed as a function of rainfall intensity $I$ [L/T]:

$$M_{lp} = \alpha \cdot I^\beta.$$

Using the Marshall and Palmer (1948) drop-size distribution, the values of the two coefficients were found through a regression for four different rainfall intensity ranges, as shown in Table A.2 (Wicks and Bathurst, 1996):

<table>
<thead>
<tr>
<th>Rainfall intensity $I$ [mm/h]</th>
<th>$\alpha$</th>
<th>$\beta$</th>
</tr>
</thead>
<tbody>
<tr>
<td>0 $\leq I &lt; 10$</td>
<td>$2.69 \times 10^{-8}$</td>
<td>1.6896</td>
</tr>
<tr>
<td>10 $\leq I &lt; 50$</td>
<td>$3.75 \times 10^{-8}$</td>
<td>1.5545</td>
</tr>
<tr>
<td>50 $\leq I &lt; 100$</td>
<td>$6.12 \times 10^{-8}$</td>
<td>1.4242</td>
</tr>
<tr>
<td>100 $\leq I &lt; 250$</td>
<td>$11.75 \times 10^{-8}$</td>
<td>1.2821</td>
</tr>
</tbody>
</table>

A.4. Leaf drip squared momentum, $M_D$

The momentum $M_D$ can be calculated as:

$$M_D = \left(\frac{\text{mean drag force}}{b}\right)^2 \cdot F_1 \cdot D.$$
where \( v[L/T] \) is the raindrop velocity estimated with an approximation of \( \text{Epema and Riezebos (1983)} \), \( \rho [\text{ML}^{-3}] \) is the density of water, \( D_L [\text{L}] \) is the leaf drip diameter (typically assumed to be 5–6 mm), \( l_f \) is the canopy drainage fraction that reaches soil as leaf drip, and \( D [L/T] \) is the total canopy drainage (Ivanov et al., 2004a).


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