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Impact of hydrologically driven hillslope erosion and landslide occurrence on soil organic carbon dynamics in tropical watersheds

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Abstract

The dynamics of soil organic carbon (SOC) in tropical forests play an important role in the global carbon (C) cycle. Past attempts to quantify the net C exchange with the atmosphere in regional and global budgets do not systematically account for dynamic feedbacks among linked hydrological, geomorphological, and biogeochemical processes, which control the fate of SOC. Here we quantify effects of geomorphic perturbations on SOC oxidation and accumulation in two adjacent wet tropical forest watersheds underlain by contrasting lithology (volcaniclastic rock and quartz diorite) in the Luquillo Critical Zone Observatory. This study uses the spatially explicit and physically based model of SOC dynamics tRIBS-ECO (Triangulated Irregular Network-based Real-time Integrated Basin Simulator-Erosion and Carbon Oxidation) and measurements of SOC profiles and oxidation rates. Our results suggest that hillslope erosion at the two watersheds may drive C sequestration or CO₂ release to the atmosphere, depending on the forest type and land use. The net erosion-induced C exchange with the atmosphere was controlled by the spatial distribution of forest types. The two watersheds were characterized by significant erosion and dynamic replacement of upland SOC stocks. Results suggest that the landscape underlain by volcaniclastic rock has reached a state close to geomorphic equilibrium, and the landscape underlain by quartz diorite is characterized by greater rates of denudation. These findings highlight the importance of the spatially explicit and physical representation of C erosion driven by local variation in lithological and geomorphological characteristics and in forest cover.

1 Introduction
Hillslope erosion and deposition are closely linked to the carbon (C) cycle. The erosion and burial of soil organic C (SOC) have important consequences for C exchange with the atmosphere, and have the potential to influence the global climate [Doney and Schimel, 2007; Hilton et al., 2015; Lal, 2004]. The role of geomorphic processes in the redistribution and storage of SOC is poorly understood and highly debated [Billings et al., 2010; Stallard, 1995]. Different assumptions invoked to describe fundamental operating processes [Van Oost et al., 2007] have led to a wide range of contrasting global estimates of the effect of erosion on the global C cycle [Harden et al., 1999; Jacinthe and Lal, 2001; Lal, 2004; Stallard, 1998]. Some researchers estimate that soil erosion is an important source of C losses to the atmosphere [Jacinthe and Lal, 2001; Lal, 2004]. Others argue that deposition and burial of eroded C, and replacement of eroded C by C sequestration at the site of erosion lead to a net atmospheric C sink [Berhe et al., 2007; Harden et al., 1999; Van Oost et al., 2007].

Previous efforts to model SOC erosion [Billings et al., 2010; Coleman et al., 1997; Harden et al., 1999; Liu et al., 2003; Rosenbloom et al., 2006; Stallard, 1998; Van Oost et al., 2005, 2012; Yoo et al., 2005] included limited feedbacks among complex physical and biogeochemical processes. The episodic character of SOC erosion has not been explicitly accounted for at the watershed scale, but constitutes an important feature of SOC dynamics in eroding soils [Dialynas et al., 2016]. The dynamic representation of linkages among the underlying natural mechanisms that drive SOC transport and decomposition can be crucial when estimating the erosion-induced C exchange with the atmosphere [Berhe and Kleber, 2013; Liu et al., 2003; Van Oost et al., 2007]. Moreover, the influence of land use change and management practices [Dlugoß et al., 2012] on the capacity of eroding soil profiles to accumulate and oxidize SOC has been poorly quantified [Billings et al., 2010; Dialynas et al., 2016].

Our knowledge of the fate of eroded SOC at the landscape scale is limited. The fate of eroded SOC depends on a multitude of factors including topography and forest cover [Dialynas et al., 2016], ecohydrological variables [Quijano et al., 2013], and soil aggregate effects [Berhe et al., 2012; Papanicolaou et al., 2015]. Eroding soils are complex, dynamic systems [Doetterl et al., 2012]. The challenges inherent in simulating C fluxes of eroding soils led previous studies to invoke the simplifying assumptions that the entire amount of eroded SOC is oxidized [Lal, 1995], or that it is fully protected by burial at depositional sites [Smith et al., 2001]. Systematically tracking the dynamics of eroded SOC is crucial for improving our understanding on the net erosion-induced C soil-atmosphere exchange [Fiener et al., 2015; Liu et al., 2003]. The impact of soil erosion on the redistribution of SOC continues to be studied [Billings et
The use of physically based spatially and depth-explicit models that quantify the fate of eroded SOC may provide important insights on estimating the net erosion-induced C exchange with the atmosphere at different ecosystems [Chappell et al., 2016; Doetterl et al., 2016; Hu and Kuhn, 2014].

Tropical forests play an important role in the global C cycle through high rates of net primary production and long-term storage in biomass and soils [Ciais et al., 2013]. The humid tropics presently occupy about 25% of the Earth's land surface [Stallard, 2012]. Forty five percent to 52% of the global terrestrial biomass C, and 11–14% of the global soil C is located in tropical forests [Prentice et al., 2001]. Intense hydrometeorological phenomena in the humid tropics have the potential to trigger events of rapid sediment and C transport [Heartsill Scalley et al., 2012; Hilton et al., 2008; Larsen and Torres-Sánchez, 1992; Ramos Scharrón et al., 2012; West et al., 2011; Wohl and Ogden, 2013]. The propensity for shallow landslide occurrence in Montane tropical ecosystems can be significantly influenced by land uses and human disturbance [Gellis et al., 2006; Guns and Vanacker, 2014; Larsen, 2012]. Moreover, tropical vegetation, and warm and humid conditions favor high rates of net primary productivity (NPP) and decomposition, leading to highly dynamic SOC in time and space [Stallard, 2012]. Tropical forests are generally a net atmospheric C sink, through plant uptake and C burial [Ciais et al., 2013]. At the same time, land use and land cover change in the tropics have contributed large CO₂ emissions to the atmosphere [Houghton, 2012]. For these reasons, tropical ecosystems are important settings for studying hillslope erosion and the associated effects on C exchange with the atmosphere.

The Luquillo Critical Zone Observatory (CZO) is a tropical site of particular hydrological and geomorphological interest characterized by diverse topography and different underlying lithologies [Murphy et al., 2012]. Past land use has influenced the soils and the density of forest cover at the Luquillo CZO [Foster et al., 1999]. The history of Puerto Rico is characterized by rapid rates of tectonic uplift which, followed by gradual erosion, have shaped the Luquillo Mountains into morphologically different domains [Brocard et al., 2015]. Longitudinal river profiles are segmented by knickpoints, deviating from smooth concave-upward profiles [Pike et al., 2010]. The hydrogeomorphic behavior of different watersheds under relatively similar climatic conditions is strongly dependent on local topographic and lithological characteristics and on the exact geographic location relative to the Luquillo Mountains [Murphy and Stallard, 2012]. This is the case for the morphologically diverse adjacent watersheds of Rio Mameyes and Rio Icacos in the Luquillo CZO (section 3), which are characterized by comparable mean annual precipitation (MAP) and by contrasting lithology [Buss and
The Icacos watershed is underlain by the Rio Blanco quartz diorite, while the Mameyes is for the most part underlain by volcaniclastic rock [Dosseto et al., 2014]. This setting provides the unique opportunity to test for the effects of underlying parent material and different forest types on erosion rates and on the fate of C under comparable climatic conditions.

In this study, we developed and used a novel spatially explicit framework to quantify the impact of erosion and landslide occurrence on the redistribution of SOC and on the associated C exchange with the atmosphere in the Mameyes and Icacos watersheds. The model accounts for dynamic feedbacks among hydrological, geomorphological, and biogeochemical processes in a physically based manner at the watershed scale. This work uses measurements of SOC content across a range of depths and catena locations [Johnson et al., 2015], and stresses the role of different forest types on the erosion-induced C exchange with the atmosphere.

2 Methods

2.1 Spatially Explicit Model of SOC Dynamics

A modified version of the tRIBS-ECO (Triangulated Irregular Network-based Real-time Integrated Basin Simulator-Erosion and Carbon Oxidation) model [Dialynas et al., 2016] was developed to quantify the impact of upland SOC erosion and sedimentation on C exchange with the atmosphere at the Luquillo CZO. The tRIBS-ECO is a spatially and depth-explicit model of SOC dynamics based on coupled and physical representations of hydrological, geomorphological, and biogeochemical processes. The version of tRIBS-ECO introduced in this study accounts for the redistribution of SOC by soil erosion and by landslide occurrence, at the watershed scale.

A mass balance equation for SOC sources and losses is implemented at each soil profile. At each time step (daily), the model estimates the SOC erosion and deposition resulting from erosional processes and shallow landslide occurrence. Depth-dependent biogeochemical parameters such as SOC content, oxidation rates, and production rates, are expressed as continuous functions of depth, and are integrated over the soil thickness. The modeled biogeochemical properties vary in time in response to geomorphic perturbations. The following SOC mass balance equation for each cell and at each time step (for a unit area) [Dialynas et al., 2015] was implemented:

\[
\frac{\Delta SOC}{\Delta t} = \int_{0}^{H_t} l_t(z)dz - \int_{0}^{H_t} k_t(z)C_t(z)dz - \frac{1}{\Delta t} \int_{0}^{H_t} C_t(z)dz \bigg|_{out} + \frac{1}{\Delta t} \sum_{i=1}^{n} \int_{0}^{h_{t,j}} C_{t,j}(z)dz \bigg|_{in} \tag{1}
\]
where $t$ is time step [$T$], $z$ is depth [$L$], $H_t$ is temporally variant soil thickness [$L$], $SOC$ is total soil organic C storage in the soil profile [$ML^{-1}$], $\Delta t$ is discrete time interval, $I(z)$ is SOC production rate [$ML^{-1}T^{-1}$], $k(z)$ is SOC oxidation rate [$T^{-1}$], $C(z)$ is depth-dependent SOC content corresponding to the cell equation 1 is applied to [$ML^{-1}$], $n$ denotes the number of upstream cells contributing to lateral SOC influx, and $h_t$ is the eroded soil layer [$L$] (section 2.3). The depth-dependence of $C(z)$, $k(z)$, $I(z)$ at different soil profiles in the Luquillo CZO is discussed in section 4.1. The net C exchange with the atmosphere is represented by the sum of the first two terms of the right-hand side of equation 1. The third term corresponds to lateral SOC loss to erosion, and the fourth term represents SOC incoming from $n$ upstream eroding cells. In tRIBS-ECO, SOC is transferred with eroded sediment, and it is oxidized upon transport, or it is preserved at depositional sites [Lal, 2003]. The spatially explicit model quantifies the net erosion-induced contribution to atmospheric CO$_2$ of diverse soil profiles with different forest covers.

The biogeochemical characteristics such as SOC content, oxidation rate, and the rate of SOC production vary with soil depth. Eroding soil profiles can experience severe loss of topsoil, and landslide sites can be characterized by the removal of most of the soil profile along with the associated organic matter [Stallard, 2012]. As a result, depth-dependent biogeochemical properties such as SOC oxidation and production can be significantly altered by geomorphic disturbances. Land management may moderate this alteration at eroding soil profiles and at landslide sites. The effect of land uses on altered SOC oxidation and production are represented (Figure 1) by coefficients $a_k$, and $a_I$, respectively, in the model [Dialynas et al., 2016]. The framework is clarified by the example illustrated in Figure 1. In this soil profile, the SOC oxidation rate ($k(z)$) decreases exponentially with depth. Assume that by time $t_1$ removal of the topsoil leads to the new soil surface (dashed line) having an oxidation rate $k_t(z = 0)$, which is significantly different from the initial one, $k_{t_1}(z = 0)$. Management practices, such as fertilization and enhancement of productivity, may restore the SOC oxidation of the initial soil profile at different depths, at a rate $a_k$ (Figure 1). This framework is also applied to the production rate of SOC (using coefficient $a_I$), which can be altered by geomorphic perturbations [Billings et al., 2010].
Effect of land uses on altered oxidation rate of soil organic carbon ($k(z)$) at an eroding soil profile. Assume that by time $t$, erosion of topsoil leads to the new surface (dashed line) having an oxidation rate $k(z = 0)$, which can be significantly altered from $k(z = 0)$ (depending on soil types and management practices) [Dialynas et al., 2016]. Land use may have a moderating effect on altered oxidation at a rate $a$, (section 2.1).
Land management practices may regulate topsoil erosion and associated lateral SOC losses [Harden et al., 1999]. In tRIBS-ECO, any enhancement of SOC generation in eroding soils [Berhe et al., 2007] is parameterized using the scheme discussed in the previous paragraph, independent of the particular management strategy. High values of $a_I$ lead to enhancement of SOC production (see section 4.2). Soils rich in SOC may also be managed so as to result in accelerated rates of SOC oxidation (higher $a_k$) [Dialynas et al., 2016]. In contrast, degraded and poorly managed soils can be characterized by increased fractions of hard to degrade organic matter, and by lower rates of labile SOC generation [Billings et al., 2010], parameterized by lower values of $a_k$ and $a_I$.

The tRIBS-ECO has substantial advantages and novel characteristics over previous efforts [e.g., Billings et al., 2010; Rosenbloom et al., 2006; Yoo et al., 2005] that simulate soil-atmosphere C exchange induced by erosion. Advances of the proposed physically based approach include the episodic representation of SOC erosion which is based on the coupling of hydrologic, geomorphic, and biogeochemical processes at the watershed scale (sections 2.2 and 2.3); tracking the potential of eroded SOC to undergo mineralization or be stabilized based on local topographic variation [Dialynas et al., 2016]; and systematically accounting for effects of land uses on altered SOC oxidation and production at eroding soil profiles.

2.2 Physically Based Modeling of Hydrological Processes

The tRIBS (Triangulated Irregular Network (TIN)-based Real-time Integrated Basin Simulator) models hydrological processes in a physically based manner at the watershed scale [Ivanov et al., 2004a, 2004b; Vivoni et al., 2004]. The spatially explicit model accounts for spatial variability in precipitation fields and land-surface descriptors (e.g., forest types) and stresses the role of topography in the dynamics of soil moisture. The model's computational elements within the TIN are Voronoi polygons [Vivoni et al., 2004]. The irregular representation of topography aims to reduce computational cost by removing computational elements with little loss of information. The watershed's hydrologic response is simulated at hourly scale and at different (e.g., 10–100 m) spatial scales. The tRIBS model represents essential hydrological processes at basin scale including canopy interception, surface energy balance and evapotranspiration, runoff routing, water infiltration, and lateral redistribution of soil moisture in the unsaturated and saturated zones.

Rainfall interception is modeled using a canopy water balance method [Rutter et al., 1975, 1971]. The dynamics of intercepted water vary with different vegetation species. Water
infiltration is simulated by assuming gravity-dominated flow in heterogeneous, anisotropic soil ([Garrote and Bras, 1995](#)). The dynamics of a wetting front and a top front lead to different states of the vadose zone, described by unsaturated, perched saturated, surface saturated, and completely saturated conditions. The relative position of the groundwater table at each time step feeds back to the evolution of the wetting front in the unsaturated zone. The soil moisture is laterally redistributed during storms and interstorm periods. This process is controlled by topography, soil heterogeneity, and by the hydraulic conductivity of the anisotropic soil in the parallel to the surface and normal soil directions. A quasi three-dimensional model is used for the simulation of groundwater flow, which represents the dynamic interaction with the unsaturated zone, in addition to the lateral redistribution in the saturated zone. Different mechanisms contribute to runoff generation: saturation excess, infiltration excess, perched subsurface stormflow, and groundwater exfiltration. Simulation of surface runoff is based on the dynamic interaction of water table depth, infiltration fronts, and lateral soil moisture redistribution.

### 2.3 Physical Representation of Soil Erosion and Landslide Occurrence

To simulate hydrological feedbacks on landscape evolution in a spatially explicit manner, tRIBS was coupled with a geomorphic model (tRIBS-Erosion) ([Francipane et al., 2015, 2012](#)). The coupled hydrogeomorphic model calculates the local sediment discharge resulting from raindrop detachment and from sheet erosion at Voronoi polygons located in hillslopes and channels. The estimated elevation changes from soil erosion and deposition are applied across the watershed at each time step. Local changes in topographic characteristics (i.e., slope and aspect) and in the drainage network configuration dynamically feedback into the watershed's hydrology.

The hydrogeomorphic model represents raindrop impact erosion both from direct rainsplash impact and from leaf drip, which break soil aggregates, inducing transport of fine sediment grains. The spatial heterogeneity of rainsplash erosion is explicitly accounted for, and it depends on different factors and surface descriptors, including soil types (section 4.6), ground and canopy cover, rainfall intensity, and overland flow. The model accounts for the subgrid variability within Voronoi cells through fractions of vegetation, and bare soil, as well as variability of processes such as throughfall. Also modeled is erosion by overland flow at hillslopes and channels.

Overland flow erosion is represented by shear stress-based empirical and physical formulations, which describe initiation of motion, sediment entrainment, and transport capacity. At each Voronoi polygon, the combined sediment discharge by the two-erosional processes is calculated.

The estimated erosion potential is either entrainment-limited or transport-limited, and is
controlled by local topographic gradients. The underlying formulations of the hydrogeomorphic model are described by [Francipane et al., 2012].

The model also accounts for sediment transported in landslides, which is the dominant component of hillslope erosion in the Luquillo Mountains [Larsen, 2012]. To quantify the associated impacts on the dynamics of SOC, a slope stability module was included in the proposed framework [Dialynas et al., 2015]. Previous landslide modeling efforts at the Mameyes basin [Arnone et al., 2011, 2014, 2016; Lepore et al., 2013] focused on landslide occurrence in response to tropical storms at fine (e.g., hourly) time scales, and also used versions of the tRIBS framework [Ivanov et al., 2008a, 2008b, 2004a]. The slope stability component is based on the infinite slope model [Arnone et al., 2011], which assumes that the plane of failure is parallel to the soil surface. The level of stability is typically assessed by evaluating the factor of safety (FS), which expresses the extent to which destabilizing forces exceed in magnitude forces that favor slope stability. At each time step and at each computational element the FS is estimated by Arnone [2011]:

$$FS_t = \frac{c}{h_s \rho_s \sin \alpha_t} + \left(1 - \frac{\left(\frac{\theta_t - \theta_s}{\theta_t - \theta_r}\right) \rho_w}{\rho_s}\right) \frac{\tan \phi}{\tan \alpha_t}$$

where $c$ is the combined effect of root cohesion and effective soil cohesion, $\phi$ is the friction angle, $h_s$ is the thickness of the landslide soil mass, $\rho_s$ and $\rho_w$ are the soil and water densities, respectively, $\alpha_t$ is the time variant local slope angle, $\theta_s$ is the average soil moisture content in the landslide control volume at each time step, $\theta_t$ is the saturated volumetric water content, and $\theta_r$ the residual volumetric water content. A detailed description of the limit equilibrium method of the stability module is given by Arnone et al. [2011]. The landslide module explicitly accounts for the spatial heterogeneity of factors that control slope stability, such as the mechanical and hydrological characteristics of different soil types, local terrain characteristics, and the time varying soil moisture content at each Voronoi polygon.

The model estimates the possible landslide deposition path based on the concept of run-out distance [Bathurst et al., 1997], and alters the topographic characteristics of the landscape to account for elevation changes caused by erosion or deposition. The length of the maximum run-out distance can be estimated as a proportion (e.g., 40%) [Arnone et al., 2011] of the elevation difference between the landslide head and the deposition starting point [Vandre, 1985].

Furthermore, soil deposition depends on slope morphology [Burton and Bathurst, 1998]. For steep slopes (typically greater than 10°–15°) landslides move downhill unconditionally, and for gentle slopes (usually less than 4°–5°) the detached material halts unconditionally. For intermediate slopes, the movement of the landslide material is either limited by the maximum run-out distance, or by reaching gentler slopes along the landslide path, as discussed above [Arnone et al., 2011]. Moreover, in the implemented framework the deposited landslide mass at
each cell of the postfailure movement is inversely proportional to the local slope. Smaller sediment volumes are being deposited at relatively steeper slopes, and larger ones are being deposited at downhill gentler slopes, respectively.

Landslides and severe rainsplash and sheet erosion are processes that both depend on, and alter the landscape morphology. Therefore, the geomorphic processes are interconnected by dynamic feedbacks. More precisely, at each time step landslides and soil erosion and deposition alter the slope morphology at the watershed scale. The updated slopes are estimated across the watershed, which in turn control the soil erosional potential of the landscape.

3 Study Area

The hydrogeomorphic response of the Mameyes and Icacos watersheds in the Luquillo CZO (northeastern Puerto Rico) was simulated in terms of erosion of upland SOC and soil-atmosphere C exchange. The Rio Mameyes drains into the Atlantic Ocean on the north side of Puerto Rico. The Rio Icacos is a tributary of the Rio Blanco, which discharges into the Caribbean Sea on the southeast side of the island [Dosseto et al., 2014]. The elevation at the Mameyes watershed (Figure 2a) ranges from 104 to 1046 m. The elevation at the Icacos watershed (Figure 3a) ranges from 615 to 845 m [Larsen, 2012]. The two watersheds have been focal points of several hillslope erosion and bedrock weathering studies [Arnone et al., 2014, 2016; Buss and White, 2012; Chabaux et al., 2013; Dosseto et al., 2012, 2014; Larsen, 2012; Lepore et al., 2013; Stallard, 2012] because of their particular geomorphological interest. On average, the Mameyes watershed is characterized by steeper slopes compared to the Icacos watershed (mean slope of 21° for Mameyes, and 13° for Icacos, respectively) [Larsen, 1997]. The two watersheds are characterized by frequent rainfall-triggered landslides. According to Larsen [2012], shallow landslides constitute 93% and 98% of the total hillslope erosion for Mameyes and Icacos watersheds, respectively. The Rio Icacos is characterized by fine bed sediment which can be mobilized by moderate runoff, while the Rio Mameyes bed material is significantly coarser. The spatial distribution of different soil textures is described in Figures 2b, 3b and section 4.3. The unweathered volcaniclastic bedrock is located at the depth of 16 m [Dosseto et al., 2012]. Soil erosion and the rate of regolith production control the soil thickness. The contrasting lithology in the study area contributes to different morphology and sediment yield at the two watersheds, with direct implications on the hydrological behavior of the Mameyes and Icacos watersheds. Murphy and Stallard [2012] discussed the climatological and hydrological characteristics of the two watersheds. The mean annual temperature at the Mameyes and Icacos watersheds are 22.8°C and 21.4°C, respectively. The spatially averaged MAP at the Icacos watershed, which is characterized by a mean elevation of 686 m [Murphy et al., 2012], is 4150
mm yr\(^{-1}\). The spatially averaged MAP at the Mameyes watershed, with a mean elevation of 508 m, is 3760 mm yr\(^{-1}\). The MAP at the Mameyes watershed, which is characterized by greater relief, exhibits higher spatial variation than the Icacos watershed [Murphy and Stallard, 2012]. Rio Icacos has a mean annual discharge of 3760 mm yr\(^{-1}\) with a drainage area of 3.26 km\(^2\). The mean annual runoff of Rio Mameyes is 2750 mm yr\(^{-1}\), and the drainage area is equal to 17.8 km\(^2\).

Figure 2

(a) Digital Elevation Model of the Mameyes watershed; site B corresponds to the location of the Bisley tower (section 4.4), (b) spatial distribution of soil textures, and (c) spatial distribution of forest types.
The Luquillo CZO is characterized by four forest life zones [Ewel and Whitmore, 1973; Holdridge, 1967]: subtropical wet forest, subtropical rain forest (below 600 m of elevation), lower Montane wet forest, and lower Montane rain forest (above 600 m). Based on species composition, the vegetation can be classified into: tabonuco forest (dominated by *Dacryodes excelsa*), Colorado forest (dominated by *Cyrilla racemiflora*), palm forest (dominated by *Prestoea montana*), and dwarf (cloud) forest (with *Tabebuia rigid* as a common species) [Ewel and Whitmore, 1973; Waide et al., 1998; Weaver and Murphy, 1990]. The tabonuco forest dominates lower slopes up to elevations of around 650 m, while the Colorado forest typically occupies higher elevations at the lower Montane life zones, up to 900 m [Lepore et al., 2013]. Palm stands are common in most life zones throughout the Luquillo CZO, and they
may occupy steep and poorly drained locations, while dwarf forest is located at higher ridges, at
poorly drained soils, and corresponds to less than 10% of the vegetation of Luquillo CZO
[Johnson et al., 2015; Waide et al., 1998]. The spatial distribution of forest cover for the
Mameyes and Icacos watersheds is illustrated in Figures 2c and 3c, respectively [Helmer et
al., 2002; Puerto Rico GAP Analysis Project (PRGAP), 2006]. The Mameyes watershed is
dominated by tabonuco forest. Higher elevation sites at the watershed are dominated by
Colorado and palm forest. The Icacos watershed is primarily covered by Colorado forest with
sites of palm forest at the western slopes.

The patterns and density of vegetation at the Luquillo CZO have been altered by human land use
change during the recent history of the island. Forest cutting in Puerto Rico started in the 16th
century for pasture, cropland, timber, and fuelwood [Wadsworth, 1950], and peaked around 1900
[Foster et al., 1999], with almost complete deforestation of the island in the early 1900s
[Larsen, 2012; Murphy et al., 2012]. Sites at lower elevations of the Luquillo mountains have
been subject to intense human land use, primarily for pasture and farming. Isolated selective
timber harvesting and cutting of tabonuco and Colorado forests for fuelwood and charcoal took
place through 1940 [Foster et al., 1999]. Human land use influenced the forest density at various
locations within the area [Thomlinson et al., 1996], which was followed by reforestation after the
1930s [Foster et al., 1999].

4 Input Data and Parameters

4.1 Biogeochemical Parameters

Soil biogeochemical properties were obtained from previous studies at the site [Johnson et
al., 2015; Wang et al., 2003; Weaver and Murphy, 1990]. Johnson et al. [2015] measured SOC
content in the Luquillo CZO at various catena positions and from different soil horizons reaching
deep soils (140 cm). The analysis included SOC measurements at different depths in Colorado
and palm forests at the Icacos watershed, and in tabonuco, Colorado, and palm forest at the
Mameyes watershed (three replicates were obtained per topographic position at each site). SOC
depth profiles exhibited significant variation among forest types (Figure 4). Average SOC stocks
at Colorado, palm, and tabonuco soils were 21, 18, and 14 kg m⁻² respectively. Moreover, on
average Johnson et al. [2015] reported a 37.5% increase of total SOC content at the root zone
(top 80 cm) in valleys compared to ridges.
Observed organic carbon content [Johnson et al., 2015] and model inputs (section 4.1) at different topographic locations corresponding to (a) Colorado, (b) palm, and (c) tabonuco soils at the Mameyes watershed, and (d) Colorado and (e) palm soils at the Icacos watershed, respectively.

Caption
In this model the depth-dependence of SOC content is represented by the following exponential expression:

\[ C_t(z) = C_a,t \left( e^{C_{u}z} + e^{C_{c}z} \right) \]

where \( C_u, C_a, \) and \( C_c \) are parameters (see also section 2.1 for notation). Equation 3 corresponds to dynamic depth-profiles of SOC content. At time \( t = 0 \), we represented the initial SOC content at different catena locations within the two watersheds by fitting equation 3 to the SOC profiles reported by Johnson et al. [2015] (Figure 4). Equation 3 roughly approximated the available
SOC measurements at multiple soil profiles. At each time step, the depth-variation of SOC in the parsimonious model is estimated by calculating $C_a$, based on equations 1 and 2 (the associated calculation steps are given in Dialynas et al. [2016]).

Hillslope erosion leads to the redistribution of the initial SOC content across the landscape. Mechanisms that drive advection-diffusion phenomena (e.g., bioturbation, tillage, soil creep) can lead to the mixing of the SOC content estimated by equation 1 across different horizons within the soil column [Chaopricha and Marín-Spiotta, 2014; Dialynas et al., 2016]. Mixing of the calculated SOC content (equation 1) in the soil profile is assumed on an annual basis. The associated SOC depth-dependence is estimated based on equation 3. This simplifying assumption can reasonably represent the significant mixing [Koven et al., 2013] that may characterize the SOC-rich surficial horizons.

The depth-dependence of SOC oxidation and production rates is described by the means of exponential functions [Dialynas et al., 2016; Wang et al., 2015; Yoo et al., 2006] (see section 2.1 for notation):

$$k_z(t) = k_{a,t} e^{k_{a,z}}$$

$$l_z(t) = l_{a,t} e^{l_{a,z}}$$

where $k_z$, $k_{a,t}$, $l_z$, and $l_{a,t}$ are parameters (subscripts $a$ and $b$ correspond to soil parameters expressing topsoil values, and depth-variation, respectively). The topsoil values of SOC oxidation and production rates ($k_{a,t}$ and $l_{a,t}$, respectively) are time variant in eroding soils (section 2.1). The parameters $k_a$ and $l_a$ express the depth-dependence of SOC oxidation and production, respectively. $l_a$ was initialized based on measured values of aboveground NPP [Wang et al., 2003; Weaver and Murphy, 1990] for different forest types in the Luquillo CZO (Table 1). For each soil profile, $l_a$ was constrained such that at time $t = 0$ the total SOC production over depth (equation 5) is equal to the belowground NPP in each forest type (Table 1).

Table 1. Aboveground NPP (ANPP) and Belowground NPP (BNPP) Values for Different Forest Types Across the Luquillo CZOa

<table>
<thead>
<tr>
<th></th>
<th>Tabonuco</th>
<th>Palm</th>
<th>Colorado</th>
<th>Cloud</th>
</tr>
</thead>
<tbody>
<tr>
<td>ANPP (g m⁻² yr⁻¹)</td>
<td>1404</td>
<td>1268</td>
<td>527</td>
<td>540</td>
</tr>
<tr>
<td>BNPP (g m⁻² yr⁻¹)</td>
<td>324</td>
<td>293</td>
<td>122</td>
<td>170</td>
</tr>
</tbody>
</table>
The ANPP values were derived from observations [Weaver and Murphy, 1990]. The BNPP values were reported by Wang et al. [2003], based on belowground and aboveground biomass estimates for the entire Luquillo CZO.

We initialized topsoil SOC oxidation rates \( k_{a, t=0} \) for each vegetation type based on the SOC turnover characteristics reported in Cusack et al. [2010], who performed radiocarbon measurements at shallow soils at the Luquillo CZO. They analyzed soils covered by Colorado forest at the Icacos watershed (640 m of elevation) and soils from tabonuco-type forest located at the Bisley Experimental watersheds (260 m), which are part of the Rio Mameyes drainage system [Scatena, 1989]. We applied the oxidation rates to the corresponding forest types in the Icacos and Mameyes watersheds. Results from the upper-elevation site \( k_{a, t=0} \) of 0.095 yr\(^{-1}\) were applied to sites with higher altitude Colorado and cloud forests, which are also characterized by similar NPP (Table 1). Estimates from the lower elevation site \( k_{a, t=0} \) of 0.085 yr\(^{-1}\) were applied to tabonuco and palm forest soils (the NPP of which is also comparable).

This study assumes steady state conditions at time \( t = 0 \) with zero net soil-atmosphere C exchange [Jenny, 1941; Van Oost et al., 2007] to quantify the SOC redistribution driven by geomorphic perturbations. At the beginning of the simulation, we have \( \frac{\partial \text{SOC}}{\partial t} = 0 \), and equation 1 yields:

\[
\int_{0}^{H_t} l_t(z)dz \bigg|_{t=0}^{H_t} = \int_{0}^{H_t} k_t(z)C_t(z)dz \bigg|_{t=0}^{H_t}
\]  

(6)

At time \( t = 0 \), the depth-dependence of SOC oxidation rate (parameter \( k_a \)) was constrained according to equation 6 for each soil profile.

### 4.2 Effects of Land Uses on Soil Organic Carbon Fluxes

Land use plays a key role in the interaction between terrestrial sediment transport and soil-atmosphere C exchange [Billings et al., 2010]. In tropical forests, land use and land use change may have an important effect on C sequestration and on CO\(_2\) emissions to the atmosphere [Ciais et al., 2013]. Our work explores a variety of parameterized land use scenarios and their effect on anthropogenic erosion at the Luquillo CZO. To define potential land use scenarios, we conducted a sensitivity analysis on \( a \) and \( a_I \) at single eroding sites at the Icacos and the Mameyes watersheds by using equation 1 in 100 year simulations of eroding soil profiles, at an erosion rate of 1 mm yr\(^{-1}\) [Larsen, 2012]. The assumed erosion rate is consistent with estimates of average hillslope sediment flux at the two watersheds (see section 6.1). We considered a soil profile in
Colorado forest and in tabonuco forest for the case of Icacos and Mameyes watersheds, respectively.

The results of the sensitivity analysis for the two watersheds are illustrated in Figure 5. The total net difference in SOC content (ΔSOC) at the soil profile for each set of $a_k$ and $a_I$ values expresses the erosion-induced net C exchange with the atmosphere. Relatively high values of $a_k$ represent a significant effect of land use on altered oxidation, while low values reflect little effect on eroding soil profiles. Similar is the effect of $a_I$ to SOC production, respectively [Dialynas et al., 2016]. According to Figure 5, a net erosion-induced C sink can result from relatively low values of $a_k$ and high values of $a_I$. A net decrease in SOC storage can result from higher values of $a_k$ and lower $a_I$ as soil erosion proceeds. To assess the range of ΔSOC resulting from different values of $a_k$ and $a_I$, three scenarios of C fluxes were considered: (1) a maximum source scenario; (2) an intermediate scenario; and (3) a maximum sink scenario (Table 2). At the maximum sink scenario $a_k$ is minimized, and $a_I$ values were selected above which the ΔSOC in Figure 5 does not significantly change. Similarly, at the maximum source scenario $a_k$ was minimized and $a_I$ was maximized, accordingly. At the intermediate scenario, we selected moderate values of $a_k$ and $a_I$, based on Figure 5. The three scenarios of C fluxes were defined based on the aforementioned simplifying assumptions (e.g., constant erosion rates) by applying equation 1 to eroding soil profiles, and they can roughly represent erosion-induced C fluxes at eroding soil profiles in the two watersheds. The ranges of $a_k$ and $a_I$ reflect plausible states of the Luquillo CZO ecosystem under different land uses (section 6.5). The likelihood of the assumed scenarios is studied in section 6.3.
Figure 5

(a) Sensitivity analysis on the influence of land uses to the total soil organic carbon storage difference (ΔSOC) for the Icacos and (b) the Mameyes watersheds (section 4.2). ΔSOC results from the net effect of soil organic carbon (SOC) production, oxidation, and SOC loss to erosion, as the effects of land uses to SOC oxidation and production (α_k and α_l, respectively) vary [Dialynas et al., 2016]. Positive values of ΔSOC indicate net increase of SOC storage, while
negative values represent net SOC loss at the eroding site (color variation corresponds to vertical axis (ΔSOC)).

**Caption**

**Table 2.** Values of $a_k$ and $a_I$ for the Mameyes and the Icacos Watersheds.

<table>
<thead>
<tr>
<th></th>
<th>Mameyes</th>
<th></th>
<th></th>
<th>Icacos</th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>I</td>
<td>II</td>
<td>III</td>
<td>I</td>
<td>II</td>
<td>III</td>
</tr>
<tr>
<td>$a_k$ (yr$^{-1}$)</td>
<td>0.00008</td>
<td>0.00004</td>
<td>0</td>
<td>0.0008</td>
<td>0.0004</td>
<td>0</td>
</tr>
<tr>
<td>$a_I$ (g m$^{-2}$ yr$^{-1}$)</td>
<td>0</td>
<td>4</td>
<td>8</td>
<td>0</td>
<td>2.25</td>
<td>4.5</td>
</tr>
</tbody>
</table>

- $a_I$, II, and III correspond to the maximum source, the intermediate, and the maximum sink scenarios, respectively (section 4.2).

### 4.3 Topography and Soil Textural Classes

This study uses a recent fine resolution (10 m) Digital Elevation Model (DEM) obtained from the US Geological Survey’s National Elevation Dataset [United States Geological Survey (USGS), 2009]. To reduce the computational cost of the simulations, a 30 m mesh (Figure 3) of computational elements was obtained for the Icacos watershed (3.26 km$^2$), and a 50 m mesh (Figure 2) for the significantly larger Mameyes watershed (17.8 km$^2$) [Arnone et al., 2013; Ivanov et al., 2004a; Lepore et al., 2013; Vivoni et al., 2004]. Soil classification maps of the two watersheds were obtained from local soil surveys (http://websoilsurvey.nrcs.usda.gov/) conducted by the U.S. Department of Agriculture (USDA). According to the U.S. Department of Agriculture (USDA) [1951] classification system, the soil textural classes of the Mameyes and Icacos watersheds were grouped into four soil types, i.e., clay-loam, sandy-loam, silty-clay, and clay (Figures 2b and 3b).

### 4.4 Hydrometeorological Forcing

We studied the redistribution of sediment and SOC at the two diverse watersheds using a spatially explicit simulation of hydrogeomorphic and biogeochemical processes in tRIBS-ECO.
driven by a 100 year hydrometeorological forcing, which was obtained from a stochastic weather generator (e.g., AWE-GEN, Faticchi et al. [2011], Ivanov et al., [2007], RainSim, Burton et al. [2008], Castalia, Efstratiadis et al., [2014]). This study uses the Advanced Weather Generator (AWE-GEN) [Faticchi et al., 2011; Ivanov et al., 2007]. The AWE-GEN produces hourly time series of hydrometeorological variables including precipitation, air temperature, atmospheric pressure, relative humidity, solar radiation, cloud cover, wind speed, and it preserves statistical characteristics at fine time scales, in addition to the interannual variability of precipitation and temperature. We used daily rainfall series (1973–2006) from the Pico Del Este station (NOAA, station ID: 666992, lat. 18.27, long. 65.76) as inputs to AWE-GEN, in addition to daily meteorological data (air temperature, wind speed, relative humidity (1993–2010)) from the Bisley tower (USGS, station ID: 50065549, lat. 18.31, long. 65.74), and to hourly time series of atmospheric pressure and cloud cover (1993–2010) from the San Juan airport weather station (NOAA, station ID: 668812, lat. 18.44 long. 66.0), to parameterize the weather generator.

4.5 Hydrologic Model Calibration

We calibrated the hydrologic model of the Mameyes and Icacos watersheds by comparing the simulated watershed response to observed river discharge. Daily rainfall from the Bisley tower and the meteorological data discussed in section 4.4 were used as model inputs. The corresponding rainfall time series is illustrated in Figure 6. Daily river discharge records provided by the USGS at the outlets of the two watersheds (station ID 50065500 for Rio Mameyes, and 50075000 for Rio Icacos, respectively) during 2001 were used (Figure 6). The calibration period (1 January 2001 to 24 June 2001) was selected based on the availability of fine hydroclimatic data and river discharge observations for the two watersheds.
Figure 6

Open in figure viewer
PowerPoint

Hydrologic model calibration in terms of river discharge at the (a) Icacos and the (b) Mameyes watersheds. The observed rainfall intensity during the calibration period (1 January 2001 to 24 June 2001) is also provided.

Caption

The most essential parameters controlling the dynamics of hydrologic processes such as water infiltration and lateral redistribution of soil moisture at the computational element scale are the saturated conductivity \( (K_s) \), the saturated volumetric water content \( (\theta_s) \), the residual volumetric water content \( (\theta_r) \), the air entry bubbling pressure \( (\psi_b) \), the pore distribution index \( (m) \), in addition to the conductivity depth decay parameter \( (f) \), the saturated soil anisotropy \( (A_s) \), and the unsaturated soil anisotropy \( (A_u) \) [Ivanov et al., 2004a]. The anisotropy coefficient is defined as the ratio of the saturated conductivities parallel to soil surface over the one normal to soil surface. \( f \) represents the exponential decay of the surface saturated conductivity [Ivanov et al., 2004a]. Significant variation of these parameters across different soil types has been reported in the literature [Rawls et al., 1982]. A detailed description of the role of each parameter on the watershed's hydrologic response is given by Ivanov et al. [2004a, 2004b]. In a modeling effort
based on tRIBS, Lepore et al. [2013] tuned the hydrological parameters of the model for the Mameyes watershed using series of observed soil moisture content. As stated previously, we tuned the response of Mameyes and Icacos watersheds with recorded river discharge. The values of $K_s$, $\theta_s$, $\theta_r$, $\psi_b$, and $m$ reported in Lepore et al. [2013] for each soil type were used (Table 3). $A_s$, $A_r$, and $f$ were calibrated, which are key parameters affecting the hydrological dynamics at the watershed scale (Table 3). Values from the literature [Bras, 1990; Ivanov et al., 2004a, 2004b; Lepore et al., 2013; Rutter et al., 1975; Schellekens, 2000; Weaver and Murphy, 1990] were used for plant properties [Ivanov et al., 2004a] controlling hydrologic processes such as evapotranspiration, rainfall interception, and canopy storage.

**Table 3. Soil Hydrological and Mechanical Parameters**

<table>
<thead>
<tr>
<th></th>
<th>Clay Loam</th>
<th>Sandy Loam</th>
<th>Silty Clay</th>
<th>Clay</th>
</tr>
</thead>
<tbody>
<tr>
<td>$K_s$ (mm h$^{-1}$)</td>
<td>50</td>
<td>50</td>
<td>50</td>
<td>10</td>
</tr>
<tr>
<td>$\theta_s$ (mm$^3$ mm$^{-3}$)</td>
<td>0.56</td>
<td>0.55</td>
<td>0.55</td>
<td>0.53</td>
</tr>
<tr>
<td>$\theta_r$ (mm$^3$ mm$^{-3}$)</td>
<td>0.075</td>
<td>0.041</td>
<td>0.051</td>
<td>0.090</td>
</tr>
<tr>
<td>$M$</td>
<td>0.200</td>
<td>0.322</td>
<td>0.127</td>
<td>0.130</td>
</tr>
<tr>
<td>$\psi_b$ (mm)</td>
<td>-250</td>
<td>-150</td>
<td>-340</td>
<td>-370</td>
</tr>
<tr>
<td>$f$ (mm$^{-1}$)</td>
<td>0.002 (0.001)</td>
<td>0.002 (0.001)</td>
<td>0.002 (0.001)</td>
<td>0.002 (0.001)</td>
</tr>
<tr>
<td>$A_s$ (–)</td>
<td>200 (100)</td>
<td>200 (100)</td>
<td>200 (100)</td>
<td>200 (100)</td>
</tr>
<tr>
<td>$A_r$ (–)</td>
<td>200 (100)</td>
<td>200 (100)</td>
<td>200 (100)</td>
<td>200 (100)</td>
</tr>
</tbody>
</table>
Two sets of values of shear stress based soil erodibility ($K_b$) are given, corresponding to the Mameyes and to the Icacos (in parenthesis) watersheds, respectively. The model reasonably reproduced the hydrologic response of the two watersheds (Figure 6). The model simulation satisfactorily reproduced both base flow and hydrograph recession limbs, two factors that characterize the watershed's hydrological behavior. Discrepancies in the Icacos watershed were attributed to the spatial gradient of rainfall between the watershed and the location of Bisley tower (Figure 2a). The Pearson's squared correlation ($r^2$) among the observed and simulated discharge for the period of calibration, and the total water mass balance error (i.e., percentile difference of the simulated from the observed total water mass) are given in Table 4. Because $r^2$ can be sensitive to extreme events, the Nash-Sutcliffe efficiency (NSE) was estimated, which quantifies the variance of residuals normalized by the variance of observations [American Society of Civil Engineers (ASCE), 1993]. Acceptable NSE values range between 0 and 1, and $r^2$ values greater than 0.5 are considered as generally accepted levels of performance [Bennett et al., 2013]. The associated values for Mameyes and Icacos watersheds are included in the acceptable ranges, and the percentile differences of total water volume suggest that the performance of the calibration procedure is satisfactory.

Table 4. Metrics of Hydrologic and Geomorphic Validation Performances
<table>
<thead>
<tr>
<th></th>
<th>River Discharge</th>
<th>Sediment Yield</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Mameyes</td>
<td>Icacos</td>
</tr>
<tr>
<td>Mass balance error</td>
<td>−4.7%</td>
<td>5.4</td>
</tr>
<tr>
<td>$r^2$</td>
<td>0.68</td>
<td>0.62</td>
</tr>
<tr>
<td>NSE</td>
<td>0.62</td>
<td>0.61</td>
</tr>
</tbody>
</table>

4.6 Geomorphologic Model Calibration

The hydrogeomorphic model was calibrated by reproducing events and accumulation of sediment at the outlets of the two watersheds. We used daily rainfall records from Pico Del Este station (section 4.4), which cover the length of the calibration period (January 1995 to November 1999). Daily series of observed sediment yield were used from the USGS stations discussed in section 4.5. Mechanical properties of each soil type controlling landslide occurrence are the cohesion ($c$) and friction angle ($\phi$). Soil properties primarily affecting erosion driven by overland flow are the shear stress-based soil erodibility ($K_s$), and the critical shear stress ($\theta_c$), while the raindrop detachment soil erodibility ($K_r$) influences rainsplash erosion. Raindrop detachment also depends on the spatially variant vegetation fraction ($v$), and on the drip coefficient ($F_d$), which characterizes the amount of intercepted rainfall resulting in leaf drip. A detailed description of the biogeomorphic properties and the spatially explicit representation of erosion processes and shallow landslide occurrence is given by Francipane et al. [2012] and by Arnone et al. [2011], respectively.

The geomorphic model was calibrated in terms of the observed sediment yield (Figure 7). $c$ was calibrated for each soil texture, and $K_s$, $\theta_c$, and $K_r$ were tuned starting from literature values [Francipane et al., 2012; Meyer and Harmon, 1984; Yalin, 1977]. Friction angle values for each soil type were selected from the literature [Bjerrum and Simons, 1960; Lumb, 1966, 1970]. The soil mechanical parameters for each soil type are given in Table 3. The values of vegetation
parameters used in this study are: $v$ equal to 0.7 and $F_r$ equal to 0.6. The simulated accumulated volumes of sediment yield for the two watersheds are illustrated in Figure 7. Furthermore, we assessed the calibration of the geomorphic model using the metrics described in section 4.5 (Table 4). For both watersheds, the estimates were included in the ranges of acceptable performance (i.e., from 0 to 1 for NSE, and from 0.5 to 1 for $r^2$), and the total sediment mass balance error was acceptable [Moriasi et al., 2007]. This suggests that the model efficiently reproduced the observed accumulated sediment yield for the period of calibration at the Mameyes and Icacos watersheds. The simulation of geomorphic processes in the proposed framework depends on the parameterization of the hydrologic model for the two watersheds.
Calibration of the geomorphic model in terms of accumulated sediment yield for the (a) Icacos and the (b) Mameyes watersheds. The calibration period starts in January, 1995 and ends in November, 1999.
5.1 Hillslope Erosion and Deposition

The 100-year simulated landscape evolution at the Mameyes and Icacos watersheds is presented in Figure 8. Landslides, rainsplash erosion, and overland flow erosion drive the redistribution of sediment. Significant loss of topsoil is evident at the eroding hillslopes of the two watersheds. A substantial part of the eroded material was deposited across the stream network. Rates of total hillslope erosion at the Mameyes and Icacos watersheds were equal to 937 t km\(^{-2}\) yr\(^{-1}\), and 1123 t km\(^{-2}\) yr\(^{-1}\), respectively. These estimates correspond to the total simulated hillslope erosion (i.e., topsoil erosion and landslide occurrence) at the two watersheds over the 100 year period.
Figure 8

Heavily eroding hillslopes (yellow to red) and landslide sites (red) at the (a) Icacos and the (b) Mameyes watersheds. Depositional sites across river floodplains and at landslide run out tracks are illustrated in blue.

Caption
The propensity for landslide occurrence was higher on relatively steeper slopes at the two watersheds. In the Icacos watershed, slope instability occurred mainly at the north-western and
south-western parts of the watershed, but there is also landslide occurrence in southern and south-eastern areas. In the Mameyes watershed there was significant landslide occurrence in the north-western part of the watershed, and at the south-east side of the relatively steep south-western ridge. Also, landslides were triggered at the steep slopes in the proximity of Rio Mameyes in the northern part of the watershed. Landslide sediment was deposited along the run-out path. The simulated landslide locations and potential deposition paths are in agreement with previous studies reporting observed landslide scars [Larsen, 1997, 2012] (see supporting information) and assessing landslide risk [Arnone et al., 2014, 2016; Lepore et al., 2013] in the two watersheds.

5.2 Watershed-Integrated Carbon Exchange With the Atmosphere

We quantitatively estimated the hydrogeomorphic response of the two watersheds to the imposed 100 year hydrometeorological forcings in terms of SOC redistribution and C exchange with the atmosphere. Watershed-integrated results for the maximum source, maximum sink, and intermediate scenarios are illustrated in Figures 9 and 10 for Mameyes and Icacos watersheds, respectively. The simulated watershed-integrated C exchange with the atmosphere for the Mameyes yielded a source of 18.3 g m⁻² yr⁻¹ C for the maximum source scenario, a sink of 21.5 g m⁻² yr⁻¹ C for the maximum sink scenario, and a sink of 6.0 g m⁻² yr⁻¹ C for the intermediate scenario. The corresponding results for the Icacos watershed yielded a source of 14.9 g m⁻² yr⁻¹ C for the maximum source scenario, a sink of 17.1 g m⁻² yr⁻¹ C for the maximum sink scenario, and a sink of 3.3 g m⁻² yr⁻¹ C for the intermediate scenario.
Figure 9

Spatially explicit representation of the redistribution of soil organic carbon at the Mameyes watershed. (a) Total difference in soil organic carbon storage ($\Delta$SOC) at the watershed scale for the maximum sink scenario (net sink strength of 21.5 g C m$^{-2}$ yr$^{-1}$), (b) the intermediate scenario (net sink strength of 6.0 g C m$^{-2}$ yr$^{-1}$), and (c) the maximum source scenario (net source strength of 18.3 g C m$^{-2}$ yr$^{-1}$) (section 5.2).
Figure 10

Spatially explicit representation of the redistribution of soil organic carbon at the Icacos watershed. (a) Total difference in soil organic carbon storage (ΔSOC) at the watershed scale for the maximum sink scenario (net sink strength of 21.5 g C m⁻² yr⁻¹), (b) the intermediate scenario (net sink strength of 3.3 g C m⁻² yr⁻¹), and (c) the maximum source scenario (net source strength of 14.9 g C m⁻² yr⁻¹) (section 5.2).

Caption
5.3 Erosion-Induced Carbon Fluxes in Different Forest Types

This work simulated SOC losses with mobilized sediment, and the influence of erosion on the soil's capacity to produce SOC at tropical sites characterized by different forest cover. The temporal variation of SOC storage at eroding soil profiles covered by tabonuco, Colorado, and palm vegetation at the Mameyes watershed, and by Colorado and palm vegetation at the Icacos watershed, respectively, for the maximum source and maximum sink scenarios, are illustrated in Figure 11. The simulated erosion rates corresponding to the five soil profiles were comparable, ranging from 4.0 to 6.0 mm yr\(^{-1}\) (Table 5). The series corresponding to the two extreme scenarios in Figure 11 form the envelope of possible erosion-induced C fluxes for the soil profiles under study.
Figure 11

Total difference in soil organic carbon storage ($\Delta$SOC) at eroding sites with different forest cover at the (a) Icacos and (b) Mameyes watersheds. Positive and negative values of $\Delta$SOC represent a net C sink and source, respectively. Maximum sink and maximum source scenarios are illustrated in positive and negative axes, respectively. The effect of erosion on SOC production and oxidation at tabonuco and palm soils is more significant compared to Colorado soils.
Table 5. Total Difference in Soil Organic Carbon Storage (ΔSOC) in Eroding Soil Profiles at Different Forest Cover in the Mameyes (M) and Icacos (I) Watersheds

<table>
<thead>
<tr>
<th>Forest Type</th>
<th>Erosion Rate (mm yr⁻¹)</th>
<th>ΔSOC (kg m⁻²)</th>
<th>% ΔSOC</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>Max. Sink</td>
<td>Max. Source</td>
</tr>
<tr>
<td>Colorado (M)</td>
<td>5.0</td>
<td>0.51</td>
<td>-3.19</td>
</tr>
<tr>
<td>Palm (M)</td>
<td>5.2</td>
<td>1.05</td>
<td>-7.56</td>
</tr>
<tr>
<td>Tabonuco (M)</td>
<td>4.0</td>
<td>1.28</td>
<td>-8.15</td>
</tr>
<tr>
<td>Colorado (I)</td>
<td>5.1</td>
<td>0.41</td>
<td>-3.97</td>
</tr>
<tr>
<td>Palm (I)</td>
<td>6.2</td>
<td>0.76</td>
<td>-8.74</td>
</tr>
</tbody>
</table>

In the Icacos watershed, the palm forest soil profile exhibited an 85% increase in SOC storage for the maximum sink scenario relative to the soil profile covered by Colorado forest. For the maximum source scenario the palm soil profile experienced significant erosion-induced SOC loss, approximately 120% greater than the Colorado forest soil profile. In the Mameyes watershed, significantly greater SOC loss was simulated under tabonuco and palm than in the Colorado forest for the maximum source scenario (i.e., higher loss by 155% and 137%, respectively). The SOC production was also greater (by 151% and 106%, respectively) for tabonuco and palm soils for the maximum sink scenario, compared to the soil profile covered by Colorado forest. The percent of total SOC losses due to erosion was somewhat larger for tabonuco (47%) and palm (44%) soils at the Mameyes watersheds, and for the palm (51%) soil profile at the Icacos watershed. These percentile losses are lower compared to estimates of potential SOC loss to erosion reported in the literature [Harden et al., 1999]. The Colorado soil profiles experienced relatively lower SOC loss (Table 5).
5.4 Soil Organic Carbon Burial in Alluvial Sediments

We calculated the total simulated amount of lateral SOC influx to alluvial sediments, to quantify the rate of SOC burial in the floodplains of the two rivers. The associated SOC flux estimates for different scenarios, and the corresponding SOC concentrations in deposited sediment are presented in Table 6.

Table 6. Soil Organic Carbon (SOC) Deposition at the Floodplains of Rio Mameyes and Rio Icacos

<table>
<thead>
<tr>
<th></th>
<th>Lateral SOC Influx to River Floodplains (t km⁻² yr⁻¹)</th>
<th>SOC Concentration in Sediment Influx to River Floodplains (%)</th>
<th>Rates of Long-Term SOC Burial at River Floodplains (t km⁻² yr⁻¹)</th>
<th>Observed Total Organic C Yield (t km⁻² yr⁻¹)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>I</td>
<td>II</td>
<td>III</td>
<td>I</td>
</tr>
<tr>
<td></td>
<td>Mameyes</td>
<td>14.9</td>
<td>25.3</td>
<td>39.2</td>
</tr>
<tr>
<td></td>
<td>Icacos</td>
<td>32.4</td>
<td>40.1</td>
<td>52.1</td>
</tr>
</tbody>
</table>

- a I, II, and III correspond to the maximum source, the intermediate, and the maximum sink scenarios, respectively.

Stallard [2012] reported total organic C (TOC) yields approximately equal to 13 t km⁻² yr⁻¹ for Rio Mameyes, and 32 t km⁻² yr⁻¹ for Rio Icacos, respectively. Our model estimated that the approximate portion of the total SOC influx to depositional sites that undergoes long-term burial for the intermediate scenario was 49% and 20% of SOC for Rio Mameyes and Rio Icacos, respectively (see Table 6). The associated rates of C burial in alluvial sediments at the two rivers were significant at the maximum sink scenario. At the maximum source scenario, most of the
deposited SOC was eventually exported fluvially in the form of particulate or dissolved organic C.

6 Discussion

In this study, we used a coupled and physically based representation of watershed hydrology, and erosion and landslide processes to simulate the hydrogeomorphic behavior of two morphologically diverse watersheds in Puerto Rico. The proposed framework reproduced the natural spatial variability of sediment transport at the watershed scale and the dynamics of mobilized SOC, which is an important advance to conceptual approaches that are based on simpler assumptions on the fate of eroded SOC [Billings et al., 2010; Schlesinger, 1995; Smith et al., 2001].

6.1 Hillslope Erosion and Landscape Equilibrium

This study modeled dynamic feedbacks of landslide occurrence and topsoil erosion to hydrologic and geomorphic processes at the watershed scale. Both landslides and topsoil erosion alter local geomorphological characteristics of hillslopes, such as the slope and curvature, which control the rate of soil erosion, landslide activity, and the deposition of detached soil. Exposed landslide scars and depositional sites characterized by heterogeneous unconsolidated soil can be susceptible to runoff-driven erosion [Stark and Passalacqua, 2014] at higher rates compared to undisturbed soils. Severe soil erosion and deposition at steep hillslopes may alter the limit equilibrium that controls shallow landslide occurrence, and may therefore feed back to slope stability. The complexity that characterizes this dynamic interrelation can be significant, given the large small-scale variability of the natural processes that drive erosion and shallow landslides at the watershed scale [Bras, 2015; Dialynas et al., 2016; Kim et al., 2016].

The calculated hillslope erosion rates of Mameyes and Icacos watersheds (937 t km⁻² yr⁻¹ and 1123 t km⁻² yr⁻¹, respectively) derive from a physically based representation of the hydrogeomorphic behavior of the two watersheds. We used observations of a key set of surface descriptors and calibrated our approach for the watersheds under study (section 4.6). Larsen [2012] reported a range of total hillslope erosion for the Mameyes watershed equal to 523 to 2143 t km⁻² yr⁻¹, and a hillslope erosion estimate for the Icacos watershed equal to 750 t km⁻² yr⁻¹. These rates are in acceptable agreement with our results, yet they include relatively large inherent uncertainties in the measuring and in the estimation procedures (e.g., uncertainty on landslide scars dating).
The soil thickness in the two tropical watersheds can be determined by the relative balance between erosion and bedrock conversion to soil. This modeling study assessed whether the landscape at the two watersheds, which are characterized by contrasting lithology (quartz diorite in Icacos, and volcaniclastic rock in Mameyes watersheds, respectively) had reached a state of equilibrium. The estimated magnitudes of total hillslope erosion at the two watersheds were compared with the results of recent studies, which used Uranium-series isotopes to quantify weathering rates leading to regolith production. Chabaux et al. [2013] reported that the quartz diorite at the Icacos watershed is characterized by a weathering rate of around 45 mm kyr⁻¹. Dosseto et al. [2012] reported a markedly larger weathering rate for the volcaniclastic rock of the Mameyes watershed, equal to 335 mm kyr⁻¹. Our simulated rates of total hillslope erosion are equivalent to 1123 mm kyr⁻¹ for Icacos, and 937 mm kyr⁻¹ for Mameyes watersheds, respectively (assuming an average soil density of 1 t m⁻³, as suggested by Larsen [2012]). The erosion rate in Icacos exceeds regolith production by a factor of 25. High estimates of sediment export for Rio Icacos were also reported by McDowell and Asbury [1994], Shanley et al. [2011], and Stallard and Murphy [2012]. The comparison of our simulated results with observations suggests that there is a significant rate of landscape denudation at the landscape underlain by quartz diorite. For the Mameyes watershed, soil erosion is in the same order of magnitude with the rate of bedrock weathering. The landscape underlain by volcaniclastic rock may therefore be closer to a state of equilibrium. This is in agreement with the findings of Stallard [2012]. Results for the Icacos watershed contrast the findings of Brown et al. [1995] and Brocard et al. [2015] who used cosmogenic ¹⁰Be to estimate landscape denudation rates of 43 mm kyr⁻¹ and up to 100 mm kyr⁻¹, respectively. Our erosion estimate of 1123 mm kyr⁻¹ suggests that cosmogenic rates underestimate the erosional potential of the current state of the system. This may be attributed to the relatively larger time scales (tens of thousands of years) ¹⁰Be derived denudation rates are averaged over, which may not reflect short-term fluctuations of soil denudation rates driven by climatic oscillations, vegetation dynamics, and land use [Brocard et al., 2015].

The net effect of hillslope erosion and saprolite deepening controls long-term equilibrium conditions of soil thickness across the landscape. According to our simulations, the substantial rates of divergent sediment transport exceed local regolith production in the steep slopes of the Icacos and Mameyes watersheds (Figures 8a and 8b), with the potential to induce local exposure of deeper soil horizons and of emerging bedrock. At the same time, soil thickness gradually increases on lower slopes and in valleys with substantial deposition rates of colluvial and alluvial sediments. Low erosion rates at relatively flat interfluves combined with bedrock conversion to soil may lead to temporary states of local soil thickness equilibrium [Dietrich et al., 1995].
6.2 Soil Organic Carbon Redistribution Across the Landscape

*Dialynas et al.* [2016] demonstrated the importance of the episodic nature of erosion on the estimation of the associated C exchange with the atmosphere. According to our results, the amount of mobilized organic material by landslides and by episodic erosion exhibits significant topographic variation (Figures 9 and 10). This variation is controlled by slope morphology, forest cover, and by the depth-dependence of soil biogeochemical properties.

This work quantitatively estimated the amount of SOC that is transported with sediment from eroding hillslopes and landslide sites across the watershed. For the time scales under study, the length of possible SOC transport paths depended on local geomorphological characteristics and surface descriptors (e.g., forest cover). The SOC transport path can be relatively short in the case of slope failure, where SOC is rapidly transferred from unstable slopes to depositional sites, located directly downhill. Longer paths are associated with soil aggregates traveling from upland eroding sites to river floodplains. During transport, SOC may experience oxidation which depends on the length of the transport path and on the stability of soil aggregates [Lal, 1995].

According to our simulations, relatively flat ridge tops did not have significant landslide occurrences (Figure 8). Geomorphically stable ridges favor SOC accumulation, and the average biomass turnover time associated with geomorphic perturbations was significantly higher compared to lower catena positions [Scatena and Lugo, 1995]. Moreover, the residence time of buried SOC at depositional sites depended on local rates of sediment transport, and on the extent to which SOC burial effectively mitigates oxidation [Chaopricha and Marín-Spiotta, 2014; Van Oost et al., 2012]. Therefore, this study emphasizes that systematically tracking the dynamics of mobilized SOC at the watershed scale is crucial for estimating the combined effects of hillslope erosion [Dialynas et al., 2016; Harden et al., 1999], and landslide activity [Stallard, 2012] on soil-atmosphere C exchange.

6.3 Assessing the Likelihood of Modeled Scenarios

To evaluate the extent to which modeled scenarios represent the erosion-induced SOC redistribution across the two watersheds, the deposition rate of upland eroded SOC was quantitatively estimated at the floodplains of Rio Mameyes and Rio Icacos (section 5.4). The relatively large simulated estimates of total SOC deposition in alluvial sediments (Table 6) suggest that the two watersheds are characterized by significant rates of upland SOC erosion. Soils that have been forming uplands for hundreds to thousands of years may erode at rates greater than their rate of development [Stallard, 2012]. Hillslope erosion at relatively steep slopes and interfluves may lead to depletion of upland SOC stocks. Our results suggest that the
dynamics of soil thickness (section 6.1) exert a strong control on the soil's capacity to produce and store SOC.

The SOC concentration in the surficial mineral soils of the Luquillo CZO roughly ranges from around 1% to 3% [Beinroth et al., 1992; Johnson et al., 2015]. The preferential mobilization of soil aggregates that contain SOC can lead to the enrichment of the eroded sediment in SOC [Berhe et al., 2012; Wilson et al., 2009]. The enrichment factor that expresses the associated increase in SOC concentration has been estimated to be equal to 1.7 or lower [Kuhn et al., 2009; Polyakov and Lal, 2004; Rumpel et al., 2006]. Thus, it is reasonable that SOC concentrations in mobilized sediments at the Luquillo CZO are greater than 1% and up to around 5%. This range encompasses the simulated SOC concentration in eroded soils for the scenarios explored here (Table 6), and indicates that the three scenarios may reasonably quantify the SOC concentration in eroded sediment that is transferred to alluvial sites.

6.4 Influence of Forest Types on the Erosion-Induced Carbon Fluxes

The most important C export from the tropical ecosystems at the Luquillo CZO is SOC oxidation to atmospheric CO [Stallard, 2012]. Here we demonstrate the importance of forest type on the net erosion-induced C exchange with the atmosphere at the study site (section 5.3). Severely eroding soil profiles under different forest types at the Mameyes and Icacos watersheds were compared, for the simulated maximum sink and maximum source scenarios (Figure 11). In the maximum source scenario, erosion of surficial soil horizons reduced the potential of the remaining soil system to produce SOC. In the maximum sink scenario soil erosion may significantly alter the depth-dependent rate of SOC oxidation in the soil profile.

Tabonuco and palm forests are characterized by markedly greater (more than double) NPP values compared to Colorado forests (Table 1). Despite differences in NPP, the depth-dependence of SOC production rate can be similar for soil profiles across different forest types in the Luquillo CZO [Wang et al., 2003]. Lateral removal of topsoil may alter the rate of SOC production in tabonuco and palm soils more significantly, compared to Colorado soils. This was reflected in the considerably higher C loss to erosion at tabonuco and palm soils, simulated in the maximum source scenario (Figure 11 and Table 5). In the maximum sink scenario the erosion-induced alteration of the SOC oxidation rate was more significant in tabonuco and palm soils, while SOC production was maintained at high levels. This led to a greater net increase of SOC storage at tabonuco and palm soil profiles. Thus, the range of possible erosion-induced C fluxes illustrated in Figure 11 for the maximum sink and maximum source scenarios was greater for palm and
tabonuco soils. As a result, in the extreme scenarios modeled here, sediment transport had a stronger effect on the C exchange with the atmosphere at palm and tabonuco soils, compared to Colorado soils. Our results highlight that the spatial distribution of forest types is a key factor that controls the simulated erosion-induced soil-atmosphere C exchange at the watershed scale.

6.5 Net Atmospheric Carbon Sink or Source

The hydrogeomorphic response of the sites under study was quantitatively estimated in terms of SOC erosion and C exchange with the atmosphere at the watershed scale. SOC loss to erosion at the maximum source scenario was significant across the two watersheds. The maximum sink scenario is characterized by rapid replacement of eroded C by atmospheric C sequestration. Our results ranged from a maximum source strength of 18.3 g C m⁻² yr⁻¹, to a maximum sink strength of 21.5 g C m⁻² yr⁻¹ for Mameyes, and from a maximum source strength of 14.9 g C m⁻² yr⁻¹, to a maximum sink strength of 17.1 g C m⁻² yr⁻¹ for Icacos watersheds. The inferred range encompasses the previous estimates of erosion-induced C sink strengths by Harden et al. [1999] (10–20 g C m⁻² yr⁻¹), Van Oost et al. [2005] (3–10 g C m⁻² yr⁻¹), and Yoo et al. [2005] (up to 2.8 g C m⁻² yr⁻¹). The simulated soil-atmosphere C exchange for the intermediate scenario (6.0 g m⁻² yr⁻¹ C sink for Mameyes, and 3.3 g m⁻² yr⁻¹ C sink for Icacos, respectively) is in agreement with the C sink strength estimates reported by Van Oost et al. [2005], and Yoo et al. [2005]. The different scenarios may reflect changes in the effect of hillslope erosion on the soil-atmosphere C exchange as a result of different perturbations at the Luquillo CZO [Stallard, 2012]. Changes in land use and extreme hydroclimatic phenomena [Larsen and Torres-Sánchez, 1992; Scatena and Larsen, 1991] may lead to a transition between a net C source and a net C sink over time at the watershed scale in tropical ecosystems [Ciais et al., 2013; Harden et al., 1999; Van Oost et al., 2007].

This work studied how the spatial distribution of forest cover is inferred to control the ranges of potential erosion-induced C exchange with the atmosphere at the watershed scale. Colorado forest primarily (around 86% of the area) covers the Icacos watershed (Figure 3c), while the Mameyes watershed is for the most part (around 87% of the area) covered by tabonuco and palm trees (Figure 2c). As discussed in section 6.4, sediment transport can have a stronger impact on soil-atmosphere C exchange in tabonuco and palm soils, which are characterized by greater maximum C sink and C source strengths, compared to soils covered with Colorado forest. Therefore, the watershed-integrated response to geomorphic perturbations at the Icacos watershed resulted in a narrower range of potential C fluxes with the atmosphere, compared to the Mameyes watershed.
6.6 Soil Organic Carbon Replacement at Landslide Scars

According to our findings, landslides removed surficial and deeper soil layers and associated SOC. Part of the eroded C at fresh scars may be rapidly replaced by atmospheric C sequestration during succession [Stallard, 1998, 2012; Zarin, 1993; Zarin and Johnson, 1995]. For the maximum sink scenario, C sequestration led to rapid SOC production at fresh landslide scars. At the intermediate scenario, a substantial part of the eroded SOC was rapidly replaced at landslide sites. The rate of SOC replacement at landslide sites decreases with time [Stallard, 2012; Zarin and Johnson, 1995]. At the maximum source scenario, C replacement was limited, and the net erosion-induced SOC loss was higher at landslide sites (Figures 9 and 10).

Our simulations suggest that on the average, 62% of eroded SOC at landslide scars has been replaced by atmospheric CO$_2$ sequestration in 100 years at the Mameyes watershed. The corresponding SOC replacement in the Icacos watershed was estimated equal to 67%. The rapid rates of dynamic C replacement at landslide scars highlight the crucial role of landslide occurrence on C erosion and accumulation in tropical watersheds. The estimated rates of C replacement are in agreement with the ones reported by Stallard [2012], who used measurements from Zarin [1993] in a simple single-site mass balance model, to quantify the rate of C replacement at landslide scars in the Luquillo CZO. The data set characterized the regeneration of soils in chronosequences of landslide scars in the Mameyes and Icacos watersheds, in addition to other sites [Zarin and Johnson, 1995]. Stallard [2012] estimated that about half the eroded SOC can be replaced in approximately 80 years, and that replacement of the entire SOC loss occurs over 200 years. The congruence of our results with Stallard's [2012] analysis is notable, as the two estimates derive from different approaches.

6.7 Potential Limitations of the Approach

Different simplifying assumptions were invoked in the coupled spatially explicit framework. While we used SOC observations for the two watersheds under study, we represented SOC content for entire soil profiles by the means of continuous functions with depth (equation 3), which roughly approximated the initial SOC content from diverse soil profiles (Figure 4). The soil properties discussed in sections 4.5 and 4.6 were assumed time invariant and constant with depth for each soil textural class (except the depth-dependent hydraulic conductivity, Ivanov et al. [2004a]), and the spatial variation of forest types and parameters $F_i$ and $v$ are constant in time [Francipane et al., 2012; Lepore et al., 2013]. The initial rates of SOC production in surficial soil horizons were assumed spatially uniform (section 4.1) within each forest type [Dialynas et al., 2016]. The proposed approach does not account for potential feedbacks of SOC content on
soil erodibility, which was calibrated (section 4.6) and assumed time invariant over the simulation period. Also, the tRIBS-ECO models hydrologically induced fluvial transport of organic material at the two watersheds, yet it does not explicitly represent sources and sinks of particulate organic C (POC) and dissolved organic C (DOC) in the stream network [Shanley et al., 2011; Stallard, 2012; Stallard and Murphy, 2012].

In the future, model validation could be improved using spatially and temporally explicit measurements of SOC, observations of CO₂ emissions and of fluvial C export from these and other watersheds. The physically based model can be used to estimate sediment and C fluxes triggered by tropical hurricanes, which could potentially lead to important insights on how extreme hydroclimatic phenomena perturb tropical ecosystems. Also, the effect of land use and land use change on SOC decomposition and production rates may vary among differently managed sites [Bras et al., 2015; Doetterl et al., 2016]. Systematically accounting for this spatial variation may further constrain estimates of soil-atmosphere C exchange at the Luquillo CZO.

7 Conclusions

1. Using a spatially explicit physical representation of SOC erosion we simulated the capacity of tropical watersheds to serve as a net C sink or a C source in response to hydrogeomorphic perturbations.

2. Forest type is inferred to control the erosion-induced C exchange with the atmosphere in these Montane tropical watersheds. Hillslope erosion altered SOC production and decomposition rates in tabonuco and palm forests, with less significant effects in Colorado forests.

3. The hillslope erosion and landslide occurrence at the diverse watersheds of Rio Icacos and Rio Mameyes were quantitatively estimated by systematically accounting for dynamic feedbacks among linked hydrogeomorphic processes. Comparison with observations of bedrock weathering suggests that the landscape underlain by quartz diorite is characterized by a higher rate of denudation, while the one underlain by volcaniclastic rock may have reached a state closer to landscape equilibrium.

4. According to our simulations, frequent events of sediment transport lead to significant SOC erosion across the Luquillo CZO. Soil profiles at landslide sites are inferred to undergo rapid C replacement by atmospheric C sequestration.

5. We recommend that future studies attempting to assess the net soil-atmosphere C exchange in regional and global C budgets systematically represent hydrogeomorphic controls of C erosion driven by local variation in forest characteristics and climatic conditions.

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The biogeochemical data used in section 4.1 are available at https://www.sas.upenn.edu/lczodata/. The elevation data we used in section 4.3 are available at http://nationalmap.gov. The hydrometeorological data we used in section 4.4 are available at http://gis.ncdc.noaa.gov/map/viewer and https://www.sas.upenn.edu/lczodata/. The river discharge and sediment yield data we used in sections 4.5 and 4.6 are available at http://waterdata.usgs.gov/nwis/. Figure 1 has been reproduced with permission from John Wiley & Sons (Copyright 2016). The authors would like to thank three anonymous reviewers for their constructive comments, which led to significant improvement of the manuscript.