On the non-uniqueness of the hydro-geomorphic responses in a zero-order catchment with respect to soil moisture

Jongho Kim^{a,b}, M. Chase Dwelle^a, Stephanie K. Kampf^c, Simone Fatichi^d, Valeriy Y. Ivanov^{a,*}

^a*Department of Civil and Environmental Engineering, University of Michigan*

^b*Department of Civil and Environmental Engineering, Sejong University, Seoul, 143–147, Republic of Korea*

^c*Natural Resources Ecology Laboratory, Colorado State University*

^d*Institute of Environmental Engineering, ETH Zürich*

Abstract

This study advances mechanistic interpretation of predictability challenges in hydro-geomorphology related to the role of soil moisture spatial variability. Using model formulations describing the physics of overland flow, variably saturated subsurface flow, and erosion and sediment transport, this study explores (1) why a basin with the same mean soil moisture can exhibit distinctly different spatial moisture distributions, (2) whether these varying distributions lead to non-unique hydro-geomorphic responses, and (3) what controls non-uniqueness in relation to the response type. Two sets of numerical experiments are carried out with two physically-based models HYDRUS and tRIBS+VEGGIE+FEaST, and their outputs are analyzed with respect to pre-storm moisture state. The results demonstrate that distinct spatial moisture distributions for the same mean wetness arise because near-surface soil moisture dynamics exhibit different degrees of coupling with deeper-soil moisture and the process of subsurface drainage. The consequences of such variations are different depending on the type of hydrological response. Specifically, if the predominant runoff response is of infiltration excess type, the degree of non-uniqueness is related to the spatial distribution of near-surface moisture. If runoff is governed by subsurface stormflow, the extent of deep moisture contributing area and its “readiness to drain” determine the response characteristics. Because the processes of erosion and sediment transport superimpose additional controls over factors governing runoff generation and overland flow, non-uniqueness of the geomorphic response can be highly dampened or enhanced. The explanation is that sediment composed by multi-size particles can alternate states of mobilization or surface shielding the transient behavior that is inherently intertwined with the availability of mobile particles. We conclude that complex nonlinear dynamics of hydro-geomorphic processes are inherent expressions of physical interactions. As complete knowledge of watershed properties, states, or forcings will always present the ultimate, if ever resolvable, challenge, deterministic predictability will remain handicapped. Coupling of uncertainty quantification methods and space-time physics-based approaches will need to evolve to facilitate mechanistic interpretations and informed practical applications.

* (Corresponding author), ivanov@umich.edu, (734) 763–5068
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1. Introduction

The superimposed responses of headwater catchments, such as zero-order basins, to a hydrometeorological event underlies the hydrological response characteristics of basins of higher orders [4, 72, 74, 75, 80]. Inferences about the hydrological behavior made at smaller scales are therefore important for understanding the hydrologic response of larger watersheds that feature similar geomorphoclimatic conditions. A number of field experiments have offered interesting insights on the features of hydrologic response of zero-order basins in various geographic settings. For example, in their work in Hitachi Ohta Experimental Watershed, Japan, Sidle et al. [67, 68] contributed to the development of the concept of “threshold hydrological response” that has been the subject of many subsequent studies [e.g., 23, 35, 48, 70, 81, 84], leading to useful conceptualizations [e.g., 32, 73] and an increased interest in identifying, understanding, and quantifying the notion of hydrologic “connectivity” [e.g., 51]. Other studies provided specific evidence that there is compartmentalization of different sources of water, e.g., highlighting that in watersheds with predominantly subsurface contributions, the chemical signature of event runoff can be quite different than that of rainfall and hillslope storage [12], and that event hydrographs are due to the prompt release of stored, “old” water [3, 44, 43, 64]. The identification of complexities in the hillslope/watershed hydrologic response has led to a search of a uniform platform for interpretation of space-time process interactions [49]. While the large variability of vital hydrologic characteristics (surface topography, bedrock geometry, soil properties, and vegetative cover) has made generalizations of field experiments difficult, “virtual laboratories” [e.g., 13, 17, 24, 29, 63, 80, 83, 59] have been important

in testing hypotheses and development of new ideas that should be subsequently tested with experimentation in the field. Explorations of internal hydrologic, ecohydrologic, and geomorphic dynamics in various settings highlighted the principal and secondary effects in the simulated behaviors and their relations [e.g., 19, 20, 22, 32, 31, 38, 39, 40, 52, 53, 60]. Importantly, the consensus of both empirical and modeling research has been that emerging complexity of hydro-geomorphic dynamics need to be explicitly linked to observable/quantifiable governing processes. Patterns, interdependencies, and time scales of relevant physical mechanisms should thus be appropriately determined. From a practical perspective, the enhancement as well as *limit identification* of predictive skill of operational models remain key issues [5, 11, 45].

Soil moisture is one of the the fundamental hydrologic states, whose space-time variations mediate most of the observed complexities and nonlinearities of the physical behavior of watershed systems. In both field and modeling studies, antecedent moisture, specifically, its magnitude and distributions have been shown to exert a strong control on the short-term runoff response to rainfall [e.g., 9, 14, 37, 57, 58, 66, 86]. Thus, intensive field campaigns have been dedicated to measuring spatial distributions of near-surface moisture and its relation to the integrated runoff response of a hillslope or catchment [e.g., 6, 10, 82]. However, mechanistic relations between soil moisture spatial distributions (especially for deep-profile water) and runoff response are not always readily evident from field measurements alone. Results from multiple locations have suggested that finer-scale organization of moisture within a hillslope or catchment, and not simply the mean, can affect the nature of the hydrologic response. For example, several studies have highlighted “preferred moisture states” (partitioned according to existence or absence of lateral sub-

surface exchange) that can be associated with distinct event responses to rainfall [e.g., 26, 55, 67]. Comparing soil moisture measurements in several catchment locations to event runoff, Zehe et al. [87] found that wetness levels in the lower part of a hillslope were more important than the overall catchment wetness in determining event runoff, suggesting that moisture spatial organization can have a distinct effect on the hydrologic response.

The studies of Ivanov et al. [34] and Fatichi et al. [20] numerically addressed aspects of soil moisture spatial variability in a hypothetical zero-order basin, as affected by topography and vegetation. Ivanov et al. [34] argued for the existence of the “attractor space” for domain soil moisture states, i.e., a set of spatial patterns (in relation to the mean basin wetness) that characterized majority of soil moisture spatial distributions exhibited by the domain. They also identified and illustrated hysteresis of spatial variability of depth-integrated soil moisture with respect to the mean catchment wetness. Fatichi et al. [20] combined an analytical approach and modeling to demonstrate the contributions and interactions of biotic and abiotic factors in determining the soil moisture spatial variability. In both studies, key reasons leading to the multiple moisture distributions were the imposed conditions of shallow soil over regular bedrock replicating surface topography. The explanation was that (1) while some storms lead to infiltration fronts that arrest at shallow depths and, ultimately, are depleted by the evapotranspiration process, (2) for certain wetting conditions, the high conductivity soil column allows for percolation of event moisture down to the impervious bottom, resulting in the saturated (or nearly-saturated) conditions that lead to efficient subsurface lateral exchange. These conclusions highlighted a non-trivial dependency of the catchment hydrological dynam-

ics on “hydraulic connectivity” between soil mantle shallow and deep moisture, implying that both first- and higher-order moment properties of soil water spatial distribution may control runoff response. Neither Ivanov et al. [34] nor Fatichi et al. [20] considered how the hydrologic response can be affected by the observed hysteresis in soil moisture spatial variability. A mechanistic attribution of the emerging non-uniqueness of runoff response thus remains to be explored.

Runoff is the main surface transport agent for a variety of admixtures, from chemicals to soil particulate matter. Complex physical dynamics arising in rainfall-runoff transformation thus must be characteristic of these other transport related processes and, likely, additionally superimposed by the features of the transient behavior of these admixtures. For example, a non-unique behavior in the geomorphic response (sediment loss and yield) had been observed at the plot and watershed scales [7, 56, 69]. The studies of Kim and Ivanov [38] and Kim et al. [40] demonstrated that potential clues to the non-uniqueness of sediment loss with respect to runoff can be elucidated by analyzing antecedent conditions of soil particle distributions: a preceding rainfall event can result in a deposited soil layer that can exhibit two conflicting roles; it can both increase and decrease soil erosion of the following event for the *same magnitude and timing* of overland flow. However, how spatial variations of watershed antecedent “hydraulic connectivity” can impact soil loss has not yet been explored.

This study advances the thread of research on why hydro-geomorphic variability caused by boundary conditions, domain properties, or variation of internal states really matters and can lead to response properties that are difficult, if ever possible, to address using a deterministic mindset. We aim to demonstrate why hydro-geomorphic response is in-

herently non-unique with respect to the essential hydrologic state, soil moisture. We also highlight that first-principle physics of variably saturated flow and sediment transport with spatial interactions can lead to complex nonlinear dynamics that are consistent with observed behaviors in experimental basins. Specifically, we investigate the effect of pre-storm (antecedent) soil moisture spatial distributions (three-dimensional—3D) on the hydro-geomorphic response—the drainage outflow, overland flow, and sediment transport from a zero-order catchment.

The study builds on the hypotheses that (1) a domain with the same mean soil moisture can have distinctly different spatial distributions of near-surface and deep-subsurface soil moisture, (2) the hydro-geomorphic response with respect to the mean domain wetness is non-unique, (3) the degrees of non-uniqueness are significantly different for the various response types addressed. The response characteristics of interest are the peak and volume of hydrograph and seepage face outflow, time rates of ascending and recession limbs, sediment flux peak, and total yield of sedigraph. To test the hypotheses and explore the mechanistic controls on hysteretic hydro-geomorphic responses, two models are used.

The first model, tRIBS+VEGGIE+FEaST [33, 34, 42, 41] calculates two-dimensional, spatiotemporal coupled dynamics of surface overland flow, erosion, and sediment transport relying on a quasi-three-dimensional representation of subsurface exchange. The addition of HYDRUS [78] that resolves variably saturated subsurface flow using a fully three-dimensional formulation, allows one to test whether hydrologic inferences made with tRIBS+VEGGIE+FEaST are consistent with those that emerge from a model that explicitly simulates three-dimensional soil moisture distributions. In two sets of numerical experiments, we explore mechanisms underlying the dynamics of hysteretic hydro-

geomorphic responses, and conditions under which these mechanisms become pronounced.

2. Methods and study design

The first part of experimental framework of this study follows many aspects of the design of Ivanov et al. [34], who used a mechanistic ecohydrological model to study vegetation effects on soil moisture spatial variability in a synthetic zero-order catchment. For the second part of experiments on hydro-geomorphic response, the design, approximations, and parameterization follow that of Kim and Ivanov [38]. The reader is referred to these studies for details. Only essential design elements and new aspects are discussed here.

2.1. Models

The hypothesized effect of non-uniqueness of the below surface hydrological response strongly depends on reliable modeling of subsurface flows. The models used in this study, tRIBS+VEGGIE+FEaST and HYDRUS3D, have different underlying structures and types of representation of domain characteristics. Using both models achieves a cross-benchmarking capability that makes inferences of this study less model-specific and ensures that the same results can be obtained with simpler models that properly represent the subsurface moisture exchange (see also Hopp et al. [30]). Another reason is that the convergence of HYDRUS numerical solver exhibits certain sensitivity to highly stratified moisture conditions in the subsurface (i.e., from dry to wet), while tRIBS+VEGGIE+FEaST could be used over a wider spectrum of conditions.

HYDRUS [78] is a finite-element model that solves the Richards equation for variably saturated subsurface flow. The three-dimensional version of the model is applied in this

study to simulate spatial and temporal patterns in soil moisture in response to sequences of atmospheric forcing variables. The model represents interactions with the atmosphere using an atmospheric (flux) boundary condition through which precipitation, potential evaporation, and/or potential transpiration can be applied. Transpiration is represented using a root water uptake function [21] that specifies how transpiration rates vary with pressure head.

tRIBS+VEGGIE [33, 34] mimics principal water and energy processes over the complex topography of a river basin. Each computational element exhibits two “big-leaf” representations of canopy (sunlit and shaded) coupled to a multi-layer soil model that computes soil moisture and heat transport, and root water uptake. The model solves the one-dimensional Richards equation for flow that is in the direction normal to the element’s surface. Gravity-driven flow for the unsaturated lateral exchange is assumed and the Dupuit-Forchheimer approximation [2] for the saturated lateral exchange is implemented. Subsurface flows are routed using the D- ∞ flow routing algorithm [71], and the flow directions change dynamically for the saturated zone. In this study, vegetation dynamics were not simulated. Only the biochemical model of photosynthesis and canopy stomatal behavior [16, 18, 46, 47] was used to simulate the response of latent heat flux to environmental conditions. The amount of leaf area as well as other structural characteristics of vegetation were imposed as pre-determined model input (see Section 2.2.1).

tRIBS+VEGGIE+FEaST [41] couples the models of Ivanov et al. [33] and Kim et al. [42], augmenting the hydrologic component with the capability to represent space-time dynamics of overland flow, erosion, and sediment transport, offering a holistic, multi-scale framework [39]. The hydrodynamic formulation of the overland flow model solves the two-

dimensional, depth-averaged, Saint-Venant equation using the finite volume method on an unstructured triangular mesh based on the framework of Roe’s approximate Riemann solver [65]. The construction for overland flow enables partially-wetted cells and the calculation of bottom slope, friction, and flux through reconstruction of variables on the mixed domain containing triangular and Voronoi polygons [42]. The erosion and sediment transport component of the model couples the overland flow formulations with advection-driven transport equations for sediments of multiple sizes using the Hairsine-Rose formulation [27, 28]. It differentiates the soil bed into an “intact” original soil layer and a “loose,” deposited layer.

2.2. *Simulation setup*

The design includes two specific modeling scenarios with different climate conditions and soil types (Table 1). The design of the first scenario is driven by the objective to obtain hydrologic response dominated by the subsurface flow over the long-term (1 year) and therefore a forcing representative of a humid maritime climate is selected. Infiltration excess overland flow is thus rare in this scenario, and the only type of runoff is seepage face flow. We compare the performance of the two models that were run in a continuous fashion.

The second, more extensive set of simulations is done to test the effects of soil moisture spatial patterns on the event response using meteorological forcing representative of a semi-arid climate. Two soil types and several precipitation forcings are used in this modeling scenario designed to generate both surface and subsurface flows that will also drive the dependent dynamics of soil erosion and transport.

2.2.1. Representation of simulation domain

The simulation domain represents a zero-order basin that contains an unchannellized hollow (Figure 1), which was previously detailed in Hopp et al. [31], Ivanov et al. [34], and Kim and Ivanov [38]. The total area is 450 [m²]; the thickness of the soil slab is 1 [m] in the direction perpendicular to the surface. To represent the subsurface domain, the models used 18 irregularly resolved mesh layers with thickness increasing from 0.02 [m], for the top layer, to 0.1 [m], for the bottom layer. In the horizontal plane, both models are based on unstructured grid system. HYDRUS used a non-uniform resolution with average spacing between the mesh nodes being 0.3 [m] (Figure 1); the total number of computational nodes was 25,110. tRIBS+VEGGIE+FEaST used 1 [m] by 1 [m] uniform horizontal spacing resulting in 8,100 computational nodes [33]. For surface flow and sediment transport, tRIBS+VEGGIE+FEaST adopts a triangular-based finite volume solver and used 900 triangular elements with areas of 0.5 [m²], where runoff computed in 450 Voronoi cells is transferred to triangular cells through a mapping procedure [42].

Table 1: Summary of simulation scenarios.

	First scenario	Second scenario
Modeling focus	Subsurface seepage flow	Surface overland flow Erosion Sediment transport
Climate type	Humid maritime	Semi-arid monsoon
Climate data	“Real” observations	“Representative” observations
Soil type	Loamy sand	Loamy sand / Clayey loam
Initial soil moisture	Uniformly distributed	8 ‘wetting’ and 8 ‘draining’ distributions (Fig. 4)
Simulation period	1 year / event	8 hours for overland flow 900 hours for subsurface flow
Model used	HYDRUS tRIBS+VEGGIE+FEaST	tRIBS+VEGGIE+FEaST

2.2.2. Soil type and land cover

The first soil type (Table 1) was assumed to be loamy sand parameterized with the van Genuchten-Mualem hydraulic model [25]. The same parameter set as that of Hopp et al. [31] and Ivanov et al. [34] was used in the first scenario. The loamy sand water retention characteristics favor the vertical propagation of a wetting front through the surface soils resulting in negligible surface runoff. For the second set of numerical experimentation, loamy sand and clayey loam were assumed and parameterized with the same van Genuchten-Mualem model. The parameterization for these two soil types are provided in Table S1 in Supplementary Materials. In the clayey loam soil, infiltration is unlikely to be pronounced for storms of a typical magnitude, thus this soil produces higher fractions of surface runoff and generates negligible seepage flow.

In order to simulate evapotranspiration in a similar manner in all models, simplifying assumptions with regards to the representation of surface land cover were necessary. Grass

Figure 1: An illustration of the zero-order basin representation in Hydrus-3D. The triangular mesh on the surface and subsurface mesh represent the computational domain. The white circles are the “Upslope”, “Midslope”, and “Downslope” locations for which the time series of soil moisture are analyzed in Section 3.1.

was assumed to cover the entire domain at the same spatially-uniform density. Roots were uniformly distributed over the one-meter soil depth. In tRIBS+VEGGIE+FEaST, constant leaf area index (LAI) was assumed to be equal to 2.0 and used throughout the entire period of simulation. This artificial parameterization of vegetation, which serves as a simple moisture uptake function, certainly affects the simulated dynamics. However, for the objective of this study and the virtual nature of experiments, such simplifications have only minor implications on the study inferences.

With regards to water uptake, the following parameters [21] were used in the sub-surface simulations: $\Psi_0 = -0.1$ [m] (pressure head below which roots can uptake water from the soil), $\Psi_{opt} = -0.25$ [m] (pressure head below which roots uptake water at the potential rate), $\Psi^* = -5$ [m] (pressure head corresponding to a threshold where water uptake suppression begins), $\Psi_w = -80$ [m] (pressure head at which root water uptake ceases). tRIBS+VEGGIE+FEaST explicitly computes energy partition by searching surface temperatures that solve the surface energy budget assuming energy conservation; the model only uses Ψ^* and Ψ_w [33].

The parameterization and calibration of tRIBS+VEGGIE+FEaST for flow and erosion are discussed in detail in Kim et al. [41] and Kim and Ivanov [38], and specific values of parameters to represent rainfall and flow induced erosion can be found in Kim and Ivanov [38]. Initially intact (cohesive) soil substrate condition was assumed. Three particle sizes were considered: 4, 50, and 100 [μ m], with settling velocities of 0.0000111, 0.0017, and 0.0062 [m s^{-1}] [15], respectively. The corresponding fractions of each particle are 30, 35, and 35 [%] for clay-loam sand, and 10, 10, and 80 [%] for loamy sand.

2.2.3. Climate forcing

Two types of climate conditions are chosen in this study, one corresponding to a humid maritime climate and the other one to a semiarid monsoon climate. For the maritime climate, the meteorological forcing data were obtained from S. Burges' experimental plots at the Urban Horticulture Center located at the north east corner of the University of Washington campus in Seattle, WA, USA. The measurement series from the year 2005, starting on January 1st, were used as the forcing for the first scenario. Precipitation, net radiation, shortwave radiation, relative humidity, air temperature, surface temperature, and wind speed measurements were used as forcing and to calculate potential evapotranspiration for HYDRUS or simulate actual evapotranspiration in tRIBS+VEGGIE. Potential evapotranspiration used as the atmospheric boundary condition for HYDRUS was computed using an energy budget method described in Kampf and Burges [36].

For the second scenario (Table 1), in order to identify several representative precipitation events for the semiarid climate, hourly rainfall data from 1961 to 1990 for the location of Tucson (AZ) were analyzed. Event depths corresponding to the median, 95th, and 99th percentiles for a duration of eight hours (Table 2) were identified. The multiplicative, microcanonical model [50, 54, 61] was used as the stochastic disaggregation method to generate 15-minute rainfall time series. Two characteristic storm events of 8-hour duration were selected from 1,000 disaggregated rainfall ensemble scenarios. The two events can be characterized as a "pulse" and "distributed" storm types, where the former contains $\sim 90\%$ of the storm depth in the first 15-minute interval, while the latter has $\sim 90\%$ of the storm depth delivered during the first hour (Figure 2).

2.2.4. Initial and boundary conditions

In the first scenario, given the long-term duration (1 year) of simulations (Table 1), spatially uniform volumetric soil moisture $0.14 \text{ [m}^3 \text{ m}^{-3}]$ was assigned as the initial condition at each location/depth of the domain. Due to the high conductivity of soil and small scale of the basin, the impact of simplified uniform soil moisture distribution is dissipated in about 15 days from the beginning of simulation (i.e., the spatial variability approaches the range that is characteristic of $\bar{\theta} = 0.14$ obtained in a continuous simulation). Additionally, soil moisture initializations at the event scale for a sub-set of experiments of the first scenario were obtained using the results of a long-term simulation. Their description is tailored to the results discussed in Section 3.3.

The effect of soil moisture initialization is vital for the second scenario of numerical experiments. The initialization conditions for the *semiarid* climate were generated from

Table 2: Event total rainfall depths for a duration of 8 hours corresponding to the median, 95th, and 99th percentiles of the non-exceedance probability for the semiarid climate.

Percentile	Rainfall depth [mm]
Median	1.39
95th	12.95
99th	25.31

Figure 2: Synthetic storm events for the location of Lucky Hills (AZ).

the 11-year simulation described in Ivanov et al. [34], corresponding to a period of 4,150 days that generated 450 (locations) $\times$ 99,600 (hourly) profiles (resolved at 18 depths, from the soil surface to a depth of one meter). Several statistics—the spatial mean, the coefficient of variation, and the mean depth of soil moisture centroid—were computed for each of the spatial distributions. Eight bins uniformly spanning the dry-wet range were used to define eight mean spatial soil moisture contents. For each of the eight means, the conditional distributions of depth of soil moisture centroid were analyzed, where the centroid depths were computed as $\Sigma [\Sigma (\theta_i \times \Delta d_i) \times d_i] / \Sigma (\theta_i \times \Delta d_i)$, where θ_i is the soil moisture, d_i is the depth of soil moisture, and Δd_i is the length of the vertical profile of the soil moisture cell at the i th vertical element. The 1st and 99th percentile of centroid depths were used to select spatial distributions of soil moisture that corresponded to either smaller or higher values of the coefficient of variation, $C_v = \sigma_\theta / \mu_\theta$ (Figure 3, see also Figure S4).

Figure 3: Envelope of centroid depths representing the 1st (the lower envelope) and 99th (the upper envelope) percentiles for eight selected mean soil moisture contents, clayey-loam soil (i.e., the 16 cases of Figure 4). The dashed line represents draining soil-moisture profiles (high $C_v(\theta)$ values), and the solid line represents wetting soil-moisture profiles (low $C_v(\theta)$ values).

Sixteen characteristic soil moisture profiles (Figures 4a - 4b) were selected, i.e., eight pairs of $(\theta, C_v(\theta))$, representing a range of possible soil moisture conditions and variations

across the domain. As Figures S1 and S2 for two soil types show (in Suppl. Materials), the same mean moisture conditions may correspond to contrasting distributions of soil moisture with depth. The above approach was used to determine initializations for both soil types (Figures 4 and S3).

For solving subsurface flow dynamics, the flux (Neumann) boundary condition was specified at the surface (net rainfall) and bottom (zero flux) of the domain, allowing for infiltration, runoff, and exfiltration fluxes. The downslope boundary of the hillslope for subsurface was approximated as a seepage face (atmospheric pressure), where water can leave the domain through the saturated part of the boundary. For surface flow, an open boundary (in the form of free outfall) was assumed at the downstream end [41]. For all other sides of the hillslope, the no-flux (solid slip wall) boundary condition was specified. Since the explicit scheme of the two-dimensional surface flow and sediment transport model of tRIBS+VEGGIE+FEaST is computationally demanding, simulations are performed only for the representative conditions.

3. Simulation results and response analysis

3.1. Comparison of hydrologic simulation results

Because of the differences in representing the process of evapotranspiration in tRIBS+VEGGIE+FEaST and HYDRUS3D, a synthetic scenario with zero evapotranspiration flux was first imposed to compare model simulations of subsurface flow only. The results are illustrated in Figure 5 as the “NO ET” case. The results are shown in terms of a) simulated evapotranspiration using tRIBS+VEGGIE+FEaST potential evapotranspiration calculated using HYDRUS3D; b) the mean of depth-integrated domain soil moisture con-

(a) Wetting fronts (low variability).

(b) Draining fronts (high variability).

Figure 4: Initial soil moisture profiles selected for Lucky Hills, clayey-loam soil. The numbers above each plot are the values of $\bar{\theta}$, which also correspond to the position of the vertical solid line (note that the vertical axis is not of linear scale). The numbers within each plot give the coefficient of variability, $C_v(\bar{\theta})$. At each depth of the model mesh, boxplot represents variation of moisture content over the 450 Voronoi cells.

tent, $\bar{\theta}$, i.e., the total volume of water in soil pores divided by the total volume of soil; c) the coefficient of spatial variation of depth-integrated soil moisture, $C_v(\theta)$, as the spatial standard deviation of depth-averaged soil moisture computed for each mesh location in the X-Y plane divided by the domain mean $\bar{\theta}$; and d) the flow through the hillslope downstream seepage face, Q_{seep} . As seen in the figure, the two models show only marginal differences in the time series of each of these variables for a one-year long simulation. This provides sufficient confidence that possible differences between simulation results of the two models with complete hydrology are likely to be attributed to the differences in modeling evapotranspiration flux.

Subsequently, the one-year long scenario that included evapotranspiration was used. Both models simulated subsurface fluxes and evapotranspiration (Figure 5, the “ET” case). Note that HYDRUS simulation terminated at day 174 because of a failure of the numerical solver. Figure 5a shows that this failure comes at a period with drying conditions and high potential evapotranspiration. The parameters of vegetation stomatal response of tRIBS+VEGGIE+FEaST were marginally calibrated to result in the same cumulative evapotranspiration as obtained by HYDRUS on day 174. This scenario is used as a reference, “control”, simulation throughout the first set of simulations.

3.2. Understanding the differences in spatial variability

The control scenario (Figure 5) illustrates that the two models are in a good agreement, even though moisture sinks associated with evapotranspiration are not simulated in the same fashion. The differences in terms of $C_v(\theta)$ that become more apparent after day 150 are likely to be attributed to HYDRUS having a finer resolution in the horizontal plane,

with the effect of highlighting the convergence of moisture in the swale area (Figure 1).

HYDRUS is useful for demonstrating the three-dimensional soil moisture patterns that develop in the domain and for exploring how these patterns relate to the catchment-averaged soil moisture statistics. To understand spatial patterns of soil moisture evolution in the study domain, soil moisture time series at several locations are analyzed. Three topographically representative mesh locations are selected (Figure 1, white circles): the “upslope” site, located in the headwater region of the domain; the “midslope” site, located near the swale mid-section; and the “downslope” site that corresponds to the location near the seepage face. The time series of soil moisture simulated at three representative depths at all sites are illustrated in Figure 6.

As seen, shallow nodes (2 [cm] from soil surface) respond rapidly to precipitation, and all shallow nodes in the domain have nearly identical moisture responses. This implies that near-surface soil water is essentially de-coupled from the saturated moisture dynamics that take place over a finite region above the soil-bedrock interface, indicating the type of situation that can be described and conveniently captured with simplified two-layer models [e.g., 59]. Middle-depth locations (30 [cm] from soil surface) have nearly identical moisture responses, with the exception of the midslope site, which is located near swale midslope. This midslope site is affected by convergence of subsurface flow. While other midslope nodes are drying, this area receives supplementary moisture both from the upslope headwater area and from the side slopes that drain toward the swale. Deep soil locations (92 [cm] from soil surface) have the most attenuated response to precipitation and transpiration. The deep locations at the upslope site respond gradually to precipitation pulses, whereas deep nodes further downslope stay saturated most of the

time.

Such differences in soil moisture dynamics at different sites and depths are related to vertical and lateral fluxes within the domain. When initialized with uniform soil moisture, HYDRUS simulations begin by drying the soil surface, while deeper soil locations begin to wet, as water is transported following the gravitational gradient. This creates a vertical stratification in moisture that persists throughout the simulation because of the imposed no-flux boundary at the base of the soil. Near the surface, the soil stays relatively dry, with moisture values typically between 0.1-0.2 (Figure 6); precipitation input propagates mostly vertically as a wetting front. At the base of the domain, moisture dynamics are affected both by the recharge from upper soil horizons as well as by the lateral saturated flow above the impermeable lower boundary. Downslope near the seepage face, deep soil locations become saturated within a few days after the start of the simulation and remain saturated most of the time thereafter (Figure 6). The duration of saturation for deep locations decreases further upslope because these areas have an upslope area and therefore less lateral inflow of moisture.

The development and subsequent draining of the saturated zone governs the pulses of seepage face flux out of the domain (Figure 5c). These pulses do not respond rapidly to precipitation but rather develop after a wetting front eventually propagates through the soil to develop and/or recharge the saturated zone at the soil base. Because this process is gradual, the time evolution of flow through the domain combined with the impact of the evapotranspiration process can lead to non-unique states of soil moisture distribution within the domain for a given mean domain soil moisture. For example, both the red circles in Figure 5a, 5b and the vertical lines in Figure 6 mark times at which

477 the mean domain moisture contents, $\bar{\theta}$, are equal, i.e., either 0.19, for the first pair of
 478 circles, or 0.15, for the second pair (Figure 5b). However, the corresponding degrees of
 479 soil moisture spatial variability, e.g., the coefficient of spatial variation of soil moisture, C_v ,
 480 are considerably different (see Figure 5c; the definition of C_v is given in Section 3.1). The
 481 lower degree of variability, co-occurring with the rising stage of the outflow hydrograph
 482 (Figure 5d), indicates a “wetting” domain. The higher soil moisture variability and the
 483 falling hydrograph indicate a “draining” stage in domain moisture dynamics. Figure 7, top
 484 and bottom panels, respectively, illustrates soil moisture distributions that correspond to
 485 these states (the case of $\bar{\theta} = 0.19$). In the “wetting” snapshot (the top panel), the bottom
 486 layer is saturated from prior drainage of the domain. Saturation has partially built up
 487 along the seepage face boundary and along the center of the swale (top left image), and
 488 storm water is propagating vertically through shallower soil layers (top right image). The
 489 upper eleven soil layers (top 0.42 [m]) are all unsaturated, and they each have a uniform
 490 soil moisture spatial distribution that implies negligible unsaturated lateral exchange.

491 In the “draining” snapshot (the bottom panel), only the top three layers of soil (top
 492 0.08 [m]) have spatially uniform soil moisture. The wetting front has already reached the
 493 base of the domain, creating a larger saturated zone: the base layer is fully saturated
 494 in all areas except along the upslope side boundaries (lower left image). Middle layers
 495 have been affected by lateral drainage into the swale (lower right image), creating locally
 496 higher moisture in the swale, particularly near its head. This contributes to a higher
 497 $C_v(\theta)$ overall for the draining case than the wetting case.

3.3. Dry-down experiments

A synthetic case was developed to address the above qualitative reasoning and assess the significance of soil moisture spatial variability on the catchment hydrological response. The four cases of instantaneous soil moisture distribution, corresponding to the two pairs with the same spatially averaged soil moisture ($\bar{\theta} = 0.19, 0.15$), were extracted from the continuous simulation of the control scenario run by both models (Figure 5). These distributions were used to initialize the domain in event-scale experiments that simulated a dry-down resulting from a single input of rain. The selected rain magnitude was 11.8 [mm], corresponding to the median rainfall depth. Precipitation was specified to occur at the beginning of simulation and was distributed in two and a half hours. Figure 8 illustrates the simulation results obtained with both models.

For each particular $\bar{\theta}$, the differences between the pair members (i.e., cases with different initial $C_v(\theta)$) are negligible in terms of mean soil moisture dynamics (Figure 8a). However, the differences in terms of spatial variability, such as the coefficient of spatial variation of depth-integrated soil water $C_v(\theta)$, are appreciable (Figure 8b) and show non-uniqueness with respect to the mean moisture state. For example, the case with the initial $\bar{\theta} = 0.15$ illustrates two pronounced modes of soil moisture variability: the case of “draining” initialization exhibits soil moisture variability continuously decreasing in time; whereas the case of “wetting” initialization causes a strong initial growth of $C_v(\theta)$. The latter observation was addressed by Ivanov et al. [34], who attributed the increase of $C_v(\theta)$ to subsurface saturated exchange triggered by storm water concentrating in the central part of the hillslope in the saturated zone.

Figure 8c illustrates the time series of seepage face flows obtained in these dry-down

experiments. While neither model has been configured to track moisture “age”, conceptually, the results confirm the difference in coupling between storm water and catchment storage depending on the type and degree of soil water spatial variability. For example, when the “wetting” initialization is used, the domain response is substantially delayed by the time period required for rainfall water to propagate to the soil bottom and trigger saturated (or nearly saturated) lateral exchange. The peak flows are observed either after 6–7 days (depending on the model, $\bar{\theta} = 0.19$) or 17 days ($\bar{\theta} = 0.15$) after the beginning of the simulation/rainfall occurrence. For the “draining” initialization, the peak of the flow is reached either immediately after the rainfall event ($\bar{\theta} = 0.15$) or only 3 days later ($\bar{\theta} = 0.19$).

Note that the initially drier state resulted in a substantially longer catchment lag time (17 days). Such an outcome was likely due to both the lower initial hydraulic conductivity of the soil column as well as to evapotranspiration that impedes the propagation of the wetting front. Note further, that in real-world (humid climate) conditions, it would be very likely that another storm would have occurred during the 17-day period prior to the flow peak resulting from the rainfall event on day 1. The end result would be seepage flow as a complex mixture of water from initial storage and subsequent rainfall events, not necessarily being dominated by a single source.

3.4. “Swap-test” experiments

To further address the importance of initial soil moisture distribution on post-storm hydrological dynamics, a “swap-test” set of numerical experiments was designed. In these simulations, the same sets of initial moisture distributions as highlighted in Fig-

ure 5 (red dots) were used; however, these moisture distributions were swapped. Such a design addresses the propagation of error resulting from higher order moments when a correct distribution (i.e., the one obtained in continuous simulation) is substituted with an “equivalent” distribution in terms of the first-order moment. Specifically, for $\bar{\theta} = 0.19$, the three-dimensional distribution of soil water that was obtained on day18 (i.e., $C_v(\theta) = 0.14$, “wetting”) was substituted with the distribution obtained on day 23 (i.e., $C_v(\theta) = 0.25$, “draining”), and vice-versa. Similarly, for $\bar{\theta} = 0.15$, the soil water distribution obtained on day 87 (i.e., $C_v(\theta) = 0.08$, “wetting”) was substituted with the distribution obtained on day 115 (i.e., $C_v(\theta) = 0.26$, “draining”), and vice versa. The simulation results are illustrated in Figures 9 and S5.

The difference between the control simulation and the “swap-test” simulation is almost indistinguishable in terms of mean domain wetness (Figure 9a). Only minor differences are observed, and for both models these differences essentially disappear after 20–30 days after the simulations start. Over the same short-term period, the differences are more appreciable for $C_v(\theta)$ (Figure 9b) and seepage face outflow (Figure 9c). For example, when the “draining” initialization is used ($C_v(\theta) = 0.25$), the peak outflow is $\sim 14\%$ higher than the peak flow simulated in the control scenario; it also occurs 3 days earlier, and the total volume of seepage flux between day 18 and day 174 is $\sim 5.3\%$ smaller than that of the control scenario (tRIBS+VEGGIE results). When the “wetting” initialization is used ($C_v(\theta) = 0.14$), the peak outflow is $\sim 25\%$ lower than the peak flow simulated in the control scenario; it occurs 5 days later and the total flow volume is $\sim 7.7\%$ lower (tRIBS+VEGGIE results).

3.5. *Hydro-geomorphic responses*

While the above simulations result in nearly all rainfall infiltrated into soil, different climate and soil conditions can create situations where a significant fraction of rainfall becomes surface runoff. Therefore, investigating the response of outlet flow and sediment is another facet in assessing the non-unique impacts of antecedent soil moisture variability. In total, 192 simulations were performed resulting from the combination of soil moisture initializations (16), rainfall event percentiles (3), rainfall patterns (2), and soil types (2), as described in Sections 2.2.3 and 2.2.4. Figure 10a illustrates the total runoff depth of an event exiting the domain through the outlet for the 64 cases corresponding to the clayey-loam soil. The median rainfall case generates negligible runoff for this soil type and therefore the results are not shown. All rainfall cases used with loamy-sand soil generate negligible runoff due to the high saturated conductivity (see Table S1 in Suppl. Materials) and are not presented.

Figure 10a shows varying responses in terms of outlet fluxes dependent upon initial soil moisture conditions (all other conditions being identical). Higher runoff is generated for the wetting cases, as compared to the draining cases, due to the soils starting closer to saturation at shallow depths (as can be seen by comparing Figures 4a and 4b). In most cases, a higher mean soil moisture in the subsurface produces greater runoff, except for $\bar{\theta} = \{0.28, 0.31\}$, where the near-surface soil was unsaturated, with water contents lower than those at $\bar{\theta} = 0.25$ (Figures 4a and 4b).

The overall response pattern of sediment loss is mostly similar to that of runoff: greater productions of sediment loss and runoff for the wetting cases than drying cases. However, the generation of sediment loss is not in entire accordance with the runoff response.

Specifically, larger runoff does not always generate higher sediment loss and may result in similar amount of washed sediment. For example, a unique soil loss response is evident for the 99th-percentile rainfall pulse case, while the corresponding generated runoff varied significantly for the sixteen different initial soil moisture conditions. Despite the differences in the runoff volumes between wetting and draining cases, the amount of soil loss is nearly identical (Figure 10b).

Figure 11 shows sixteen time series of discharge and sediment yield corresponding to the set of soil moisture initializations in Figure 4. In each of the subplots, two lines are thicker and use symbols to highlight the cases with the highest peak among the 16 hydrographs (plus symbol) and sedigraphs (circle symbol). For four sets (two percentiles and two rainfall types) of hydrographs, the wetting case with the highest antecedent soil moisture (i.e., the case generating the largest runoff shown as solid magenta lines in all of the subplots) always exhibits the highest peak among hydrographs. However, the flow peak does not always generate the peak in sediment yield; cases producing the peak of sediment yield vary: (a) wetting and mean soil moisture content of 0.157, (b) draining and 0.157, (c) draining and 0.343, and (d) draining and 0.157. These highlighted cases thus illustrate two temporal characteristics of geomorphic response: (1) the peak of sediment yield always occurs only when the preceding rainfall produces the smallest amount of flow (see series with the circle symbol), and (2) the cases with the peak discharge consistently result in considerable erosion, but at the peak time the highest volume of flow does not subsequently guarantee the peak of sediment yield (see series with plus symbol).

Figure 5: Simulation results in terms of (a.) rainfall and the 3-day moving average of simulated evapotranspiration (ET) and potential evapotranspiration (PET), aggregated from hourly to daily temporal resolution; (b.) the mean domain depth-integrated soil moisture content $\bar{\theta}$; (c.) the coefficient of spatial variation of depth-integrated soil moisture $C_v(\bar{\theta})$; and (d.) the seepage flow through the downstream boundary of the domain. Simulations were carried out with the HYDRUS and tRIBS+VEGGIE+FEaST models. Two cases are considered: the “NO ET” case, in which evapotranspiration flux was not simulated; and the “ET” case (also assumed to be the “control” case), in which the simulation of variably saturated flows was combined with the simulation of local moisture sinks to evapotranspiration. The red dots represent times at which soil water spatial distribution was taken as the initialization conditions for synthetic experiments described in Sections 3.3 and 3.4. Note that the numerical solution of the HYDRUS model did not converge after day 170 for the scenario with evapotranspiration.

Figure 6: HYDRUS simulations of volumetric water content (VWC), θ , at observation nodes within the domain along the center trough: the top panel is for the location in the headwater region of the domain (Figure 1); the middle panel corresponds to the location near swale midslope; and the bottom panel corresponds to the location near the seepage face. “Shallow” corresponds to the depth of 0.02 [m], “Mid” is for 0.3 [m] depth, and “Deep” is for 0.92 [m] depth. The vertical lines mark times with same depth-integrated mean soil moisture ($\bar{\theta} = 0.19$ and $\bar{\theta} = 0.15$) but different values of the coefficient of variation C_v . At these times, the soil water spatial distributions were taken as the initialization conditions for synthetic experiments described in sections 3.2 and 3.3.

Figure 7: Snapshots of soil moisture distributions simulated by HYDRUS for the bottom layer (the base of the domain at the soil-bedrock interface), at the depth of 1 [m] (left panels), and middle layer, at the depth of 0.3 [m] (right panels). The top snapshots represent the moisture distributions on day 18, when the domain is wetting, and the mean domain soil moisture $\bar{\theta} = 0.19$, while $C_v = 0.14$. The bottom snapshots represent moisture distributions on day 23, when the domain is draining with $\bar{\theta} = 0.19$ and $C_v = 0.25$.

Figure 8: Illustration of simulation results corresponding to four initialization states paired with respect to the same mean domain moisture content equal to either $\bar{\theta} = 0.19$ or $\bar{\theta} = 0.15$ but differing in terms of their spatial variability. For $\bar{\theta} = 0.19$, “wetting” implies $C_v = 0.14$ and “draining” corresponds to $C_v = 0.25$. For $\bar{\theta} = 0.15$, “wetting” implies $C_v = 0.08$ and “draining” corresponds to $C_v = 0.26$: (a.) the mean domain depth-integrated soil moisture content $\bar{\theta}$; (b.) the coefficient of spatial variation of depth-integrated soil moisture $C_v(\theta)$; and (c.) the seepage flow through the downstream boundary of the domain.

Figure 9: Swap-test simulations in which moisture distributions for $\bar{\theta} = 0.19$ were swapped between day 18 (hour 408) and 23 (hour 528): “ $\bar{\theta} = 0.19$, day 23” implies initialization with the soil moisture distribution from day 23 of the control scenario swapped into a simulation start time of day 18 (upper arrow). “ $\bar{\theta} = 0.19$, day 18” implies initialization with the soil moisture distribution from day 18 of the control scenario swapped into a simulation start time of day 23 (lower arrow). (a.) the mean domain depth-integrated soil moisture content $\bar{\theta}$; (b.) the coefficient of spatial variation of depth-integrated soil moisture $C_v(\theta)$; and (c.) the seepage flow through the downstream boundary of the domain.

(a) Runoff depths for distributed and pulse events.

(b) Sediment loss for distributed and pulse events.

Figure 10: (a) Total runoff depth and (b) sediment loss for two events corresponding to the 95th and 99th percentiles. Distributed (left panels) and pulse (right panels) events were used for the Lucky Hills climate, clayey-loam soil type. The solid line represents the wetting soil moisture cases (low $C_v(\theta)$) of Figure 4a, and the dashed line represents the draining soil moisture cases (high $C_v(\theta)$) of Figure 4b.

Figure 11: Hydrographs (left panels) and sedigraphs (right panels) for distributed rainfall ((a) 95th and (b) 99th percentiles) and pulse rainfall ((c) 95th and (d) 99th percentiles) used for the Lucky Hills climate, clayey-loam soil type. Solid lines represent the wetting initialization cases given in Figure 4a, and dashed lines represent the draining cases given in Figure 4b. Hydrographs and sedigraph series corresponding to the highest peak are highlighted using a thicker line in each subfigure and are also marked with a plus or circle to denote the series with the highest peak discharge (plus) or highest peak sediment yield (circle).

Additional, longer-term simulations were carried out (for 900 hours, see Table 1) to assess the characteristics of event subsurface hydrologic response using the case of more conductive soil—loamy-sand. Figure 12 shows the time scale $\bar{\tau}$, the first moment of the seepage flux hydrograph evaluated over the period of simulation of 900 hours, such that $\bar{\tau} = \Sigma (Q_i \times t_i) / \Sigma Q_i$, where Q_i is the seepage flux and t_i is the simulation time at the time step i . This metric is suitable for characterizing the response time of subsurface flow to initial wetness condition of the domain as well as to the amount and pattern of precipitation forcing. As seen in Figure 12, the response time scale exhibits a peculiar behavior with an inverted relationship between the wetting and draining cases in the regions of $\bar{\theta} < 0.12-0.15$ and $\bar{\theta} > 0.12-0.15$. The draining initialization scenarios result in a slower response scale $\bar{\tau}$ for the drier range of mean wetness conditions in the catchment, with a slow increase over nearly an entire range of $\bar{\theta}$ (excluding the transition from the extremely dry initial state). As the mean soil moisture increases, the seepage flux response is quite similar (Figure S6 in Supplementary Material). The wetting initialization cases show a delayed increase of $\bar{\tau}$ with respect to the mean condition, a peak, and generally higher magnitudes in the wetter range of conditions. As will be argued in Section 4, such a pattern of $\bar{\tau}$ is due to the changing degree of coupling between the storm water and initial saturated storage in the domain.

The clayey-loam soil (not shown) produces negligible seepage flux due to the low saturated hydraulic conductivity (Table S1) that shifts the response to the infiltration excess mechanism (i.e., Figure 10a). The near-surface and deep soil moisture are decoupled within the simulation period of 900 hours, causing negligible seepage flux, except for the cases of initially nearly-saturated soil.

Figure 12: The characteristic time $\bar{\tau}$, the first moment of the seepage face flux hydrograph over the simulation period of 900 hours, as a function of mean antecedent soil moisture and its variability (solid line—wetting case; dashed line—draining case). The distributed rain events were used for the Lucky Hills climate, loamy-sand soil type.

4. Summary and discussion

This study used two representative numerical scenarios to address the effects of non-uniqueness in soil moisture spatial distributions with respect to the mean wetness on both the hydrological and geomorphological responses of a zero-order catchment. The simulations were carried out with two physical models over the short-term, so that the dynamics of near-surface and deep-profile moisture conditions and runoff generation can be elucidated. The dependence of runoff on initial moisture has been explored in other studies, for example in the context of predictability for a flow system with both matrix and macropore flow [86]; geomorphic non-uniqueness has been only partially addressed by [38, 40]. This research demonstrates mechanistically how and why distinct soil moisture patterns arise for the same mean domain soil moisture, and how these distinct patterns propagate into non-unique hydrologic and geomorphic responses. This is done with relatively simple scenarios that focus only on vadose and saturated zone interactions, neglecting the more complex phenomena such as those, for example, arising from preferential flow paths.

4.1. *Near-surface vs. deep-subsurface*

The results of the first set of simulations demonstrate that variable soil moisture distributions can develop for the same mean basin moisture because near-surface soil moisture dynamics can be both coupled and decoupled from the process of subsurface drainage (i.e., in the latter case, flow dynamics deeper in the soil mantle produce soil moisture patterns that are distinctly different from the distribution of near-surface soil water). As a consequence of the non-uniqueness in soil moisture distributions, the responses of seepage outflow to mean basin wetness conditions can also vary, and event response relates less

to near-surface moisture content than to the spatial distribution of deep moisture that determines hillslope seepage outflow.

For example, Figure 7 shows how the statistics of depth-integrated moisture in Figure 5 (e.g., $\bar{\theta}$ and $C_v(\theta)$) relate to the three-dimensional variably-saturated flow dynamics. The two considered cases (the top and bottom panels of Figure 7) are equivalent in terms of the mean water content. Intuitively, however, the hydrological response of the basin to the same hydrometeorological forcing should be different because of the differences in three-dimensional moisture distribution. Qualitative reasoning argues that a rainfall event imposed during a draining state of the domain would lead to faster coupling between the rain (“new”) water and existing storage (“old” water) in the narrow area of higher saturation in the trough (e.g., Figure 7, bottom panel), with the rest of the domain essentially exhibiting decoupling between the near-surface and deep moisture. The outcome is a rapid and short-term contribution of “new” water to the pre-storm outflow occurring through the seepage face. Conversely, in the case when the same rainfall event is imposed during a wetting state (e.g., Figure 7, top panel), initially, seepage face outflow would be dominated by the “old” water. Before coupling with the antecedent (“old”) moisture, storm (“new”) water would have to propagate through the unsaturated zone, reach the water table, and contribute to the lateral saturated exchange. The contribution of “new” water to the seepage flux would therefore be delayed. The drier the initial state of the domain, the longer the delay would be before storm water reaches the downstream outflow boundary. While this reasoning is qualitative and we do not compute water age dynamics in this study, the results offer possible insights and a mechanistic explanation of empirical observations attributing event stormflow to “old” water [e.g., 8, 43, 76]. Further, they

imply that near-surface moisture alone, the commonly measured variable in the field, can be insufficient to characterize the moisture distribution and hydrological response in terms of seepage fluxes and therefore streamflow.

4.2. Hydrologic non-uniqueness

In the simulations of the second scenario that considered eight paired cases of initial moisture, surface runoff generation and fluid movement are governed by near-surface soil moisture conditions, as has been addressed in previous studies [e.g., 9, 57, 58, 62, 77, 85]. The important novelty of results presented here (Figures 10, 11, and 12) is that spatial organization of the soil moisture field (obtained as a result of hydrologic “filtering” imposed by the watershed itself) can lead to vastly different hydro-geomorphic responses, both in terms of flux absolute magnitudes and temporal response.

As already argued in the analysis of results of the first scenario, the key to understanding non-uniqueness of event surface/subsurface hydrologic responses is related to the degree of coupling between the storm water and the distribution of initial storage in the watershed. An illustrative example is the inverted relationship of $\bar{\tau}$ with respect to $\bar{\theta}$ for wetting and draining initializations. Figure 12 shows the higher values of the time scale for the draining cases in the drier range of initial conditions ($\bar{\theta} < 0.12 - 0.15$). For these cases, the near-surface soil moisture does not influence the behavior of seepage flow significantly for both wetting and draining initializations. Instead, the range of areas of saturated conditions near the bedrock determines the response time of seepage flow. In other words, $\bar{\tau}$ is smaller for wetting initializations (which have less saturated areas near the bedrock than draining cases) because lateral seepage flow is very small and does not

last long (please see also Figure S5, Supplementary Materials).

For the wetter range of initial conditions ($\bar{\theta} > 0.12 - 0.15$), the characteristic time $\bar{\tau}$ of the seepage flux hydrograph is generally delayed in the wetting cases, as compared to the draining cases, with convergence of the two for the very wet initial condition. Similar to the above reasoning, the behavior can also be linked to the effective decoupling of soil moisture at the surface and bottom of the soil column for the draining cases: the subsurface flow is dominated by water from the antecedent saturated storage perched on top of the bedrock, and for $\bar{\theta} > 0.12 - 0.15$ the initial saturated flow is sufficiently high. The delayed response in the wetting cases originates from the less-pronounced initially saturated area at the bedrock interface, as compared to the draining cases. This less pronounced interface requires infiltrated water to mix with antecedent near-surface moisture, move down, and contribute to saturated exchange near the bedrock interface before considerable subsurface stormflow is generated. One may also note that for the wettest range of $\bar{\theta}$, the magnitudes of $\bar{\tau}$ become nearly identical for different rainfall intensities (in Figure 12).

4.3. Geomorphic non-uniqueness

A peculiar observation that follows from Figure 10 is that the responses of erosion and sediment transport do not necessarily follow the patterns of surface and subsurface flow responses and a mismatch of hydrographs and sedigraphs peaks can be detected (Figure 11). The physical explanation is that the geomorphic response is additionally mediated by the presence of spatially and temporally varying soil substrate conditions that, in addition to runoff non-uniqueness, can superimpose another set of constraints. More specifically,

soil composed of particles of different sizes can either become mobilized and prone to erosion, or create a relatively firm layer consisting of coarser materials that will hamper erosion. For example, if a preceding soil substrate becomes broken up due to the action of rainfall or overland flow, particles become more mobile because soil contact forces that held particles together are zero; a subsequent runoff event can easily sweep these loose particles, which represents a “flow-limited” condition. If mobile, lighter fractions of soil particles have been already swept away by preceding runoff and what is left behind are heavy coarser particles, subsequent runoff cannot easily generate erosion because coarser material on the surface creates a shielding effect [38], which corresponds to a “source-limited” condition. A more detailed physical reasoning on the effects of rainfall and particle size interaction, and the time scales of morphological variations with respect to the “Shields parameter”, the reader is referred to the study of Kim and Ivanov [38].

In the specific set of results shown in Figure 11, (1) the highest peak of sediment yield preceded by small soil loss rates is because soil was broken up by preceding rainfall detachment (i.e., due to splashing) and particles essentially remained on the surface prior to the subsequent flow event, which entrained and swept away the mobile materials (i.e., the “flow-limited” condition, circle symbols). (2) In contrast, cases continuously producing a large volume of flow cannot be sustained by the sediment source, especially applicable to loose and finer particles (i.e., the “source-limited” condition, series with plus symbols).

Overall, the generation of sediment yield is a strong function of both runoff spatial pattern and soil substrate condition, as opposed to being directly linked to runoff volume and initial soil moisture content. Thus, the non-uniqueness of erosion and sediment transport has a larger number of variability modes than runoff response. For instance, the

sensitivity of geomorphic response to hydrologic dynamics can be dampened for certain conditions, e.g., the 99th percentile, pulse rainfall time series (Figure 10). Note however that this result does not indicate that geomorphic response to the forcing is unique, but is simply caused by a coincidence where variability in sediment transport can produce a similar amount of the total yield for various conditions.

5. Conclusions

This study has illustrated that the hydro-geomorphic response is not only dependent on the overall wetness of a watershed domain (which has been repeatedly shown in previous studies [e.g., 9, 57, 58]), but also on the spatial distribution of soil water. We demonstrate that key to this is the relation between the near-surface and deeper-subsurface soil water, in qualitative terms, their “hydraulic connectivity”.

The non-uniqueness of hydrologic response with respect to the mean wetness has been demonstrated in terms of runoff peak, volume, and time scales characterizing the subsurface stormflow response. The introduced “wetting” and “draining” soil moisture initialization cases facilitate the interpretation of spatial complexity, similar to the “fill-spill” analogy of Tromp-van Meerveld and McDonnell [e.g., 73]. One logical extension proposed is that soil moisture impact on the surface-subsurface responses can be better understood by introducing the notion of the depth-of-moisture centroid that represents the proximity of the center of mass of antecedent soil water to the ground surface.

The geomorphic response exhibits different sensitivities to the non-uniqueness of the hydrologic response that result from the superposition of complexities of erosion and transport processes. We observe sensitivities that vary from strong to very dampened.

The mechanistic interpretation is the dependence of geomorphic response on the composition of soil substrate determining sediment erodibility characteristics. The distribution of soil moisture and erodibility at any given time is a result of continuous evolution in response to the hydro-geomorphic cycles that change soil moisture and lead to the presence or absence of mobile soil particles at the surface. Both factors (soil moisture and soil composition) have to be accounted for to enhance short-term predictability of hydro-geomorphic systems. Admittedly, this may be an improbable task, which will severely hamper deterministic approaches for event-scale sediment and erosion predictions.

Application of the first-principle physics with spatial interactions resulted in complex dynamics that, when combined with mechanistic insights, contributes potential clues to the complex responses observed in field experiments [e.g., 1, 43, 48, 67, 79]. They have causes that emerge directly from the physical laws governing variably saturated flow in soils and erosion and sediment transport on surface. Detailed monitoring programs will continue to be desperately needed to verify and corroborate the emerged characteristics of non-uniqueness of hydro-geomorphic response from numerical simulations. This is particularly true in light of a number of modeling assumptions that need to be relaxed for most natural systems (e.g., bedrock imperviousness, homogeneity and isotropy of soil properties, soil particle size distributions on both intact and deposited layers, etc.).

As a final note, the presented results advance mechanistic interpretation of challenges of the predictability problem in hydro-geomorphology [45], in this case, specifically related to the controlling role of soil moisture *spatial variability*. The overarching message is that operating with the first-moment properties of hydro-geomorphic states is unlikely to improve considerably predictability in a deterministic sense. However, the practical value of

789 models based on first-principles physics will persist because they offer the chance to merge
790 uncertainty and physics. We contend that this represents an opportunity rather than a
791 dead end. Approaches accommodating our inherent “ignorance” about the properties,
792 states, and boundary conditions of hydro-geomorphic systems, yet allowing mechanis-
793 tic space-time interpretations should become a norm and are likely to increase system
794 understanding and predictive capabilities.

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