Article

Inferring Sediment Transport Capacity from Soil Microtopography Changes on a Laboratory Hillslope

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Abstract: In hillslope erosion modeling, the Transport Capacity (T_c) concept describes an upper limit to the flux of sediment transportable by a flow of given hydraulic characteristics. This widely used concept in process-based erosion modeling faces challenges due to scarcity of experimental data to strengthen its validity. In this paper, we test a methodology that infers the exceedance of transport capacity by concentrated flow from changes to soil surface microtopography sustained during rainfall-runoff events. Digital Elevation Models (DEMs) corresponding to pre- and post-rainfall events were used to compute elevation change maps and estimate spatially-varying flow hydraulics ω taken as the product of flow accumulation and local slope. These spatial data were used to calculate a probability of erosion PE at regular flow hydraulics intervals. The exceedance of T_c was inferred from the crossing of the PE = 0.5 line. The proposed methodology was applied to experimental data collected to study the impact of soil subsurface hydrology on soil erosion and sediment transport processes. Sustained net deposition occurred under drainage condition while PE for seepage conditions mostly stayed in the net erosion regime. Results from this study suggest pulsating erosion patterns along concentrated flow networks with intermittent increases in PE to local maxima followed by declines to local minima. These short-range erosion patterns could not be explained by current T_c -based erosion models. Nevertheless, T_c -based erosion models adequately capture observed decline in local PE maxima as ω increased. Applying the proposed approach suggests a dependence of T_c on subsurface hydrology with net deposition more likely under drainage conditions compared to seepage conditions.

Keywords: transport capacity; soil erosion; deposition; microtopography

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1. Introduction

The sediment transport capacity (T_c) is a term used in geomorphology to refer to different processes depending on the domain of the application such as fluvial, hillslope, aeolian and costal geomorphology [1]. In hillslope erosion, the T_c concept describes an upper limit to the flux of sediment transportable by a flow of given hydraulic characteristics. This concept was introduced to the hillslope erosion research community as early as Ellison [2] and later popularized by Meyer and Wischmeier [3]. The T_c concept has been used to propose a foundational principle in hillslope soil erosion modeling that suggests a coupling between detachment rate D_r , detachment capacity D_c , sediment transport rate T_r and transport capacity T_c [4]. The T_c concept is currently used in many physically-based soil erosion models such as the Water Erosion Prediction Project (WEPP), the Kinematic Runoff and Erosion Model (KINEROS), the European Soil Erosion Assessment (EUROSEM) and the Rangeland Hydrology and Erosion Model (RHEM) to determine when sediment

transport processes switch from net erosion to deposition. Geomorphic processes such as alluvial fan formation and dynamics are also modeled using the T_c concept [5].

Many predictive models of T_c have been proposed mostly from flume experiments on non-cohesive beds [6]. In these predictive models, T_c is a function of a measure of flow hydraulic forces (e.g., shear stress, stream power, etc.) and intrinsic characteristics of the transported material (e.g., sediment specific gravity, particle size, critical shear stress, etc.). The effects of various hydraulic factors on T_c have been explored over the years to adjust T_c to specific extrinsic conditions, e.g., [7–11]. Wainwright, Parsons [12] cautioned that current applications of the T_c concept to hillslope erosion are based on extrapolations from the fluvial models that were untested in the shallow and rapidly changing flow conditions encountered on hillslopes. Furthermore, Wainwright, Parsons [1] challenged the existence of a constant flow transport capacity that can be calculated from clear water hydraulics because flow hydraulic characteristics are constantly changing with the addition of new sediment.

The coupled detachment- T_c model proposed by Foster and Meyer [4] was challenged by many studies, particularly when attempting to measure T_c in varying surface and shallow subsurface hydrologic conditions. Using data from a dual-box experiment, Huang, Wells [13] concluded that a decoupling of detachment and transport was more consistent with observed erosion/deposition patterns, and this conclusion was further supported by other studies, e.g., [14,15]. Challenges to the T_c concept warrant more data-driven research to clarify the conceptual definition of T_c . This need for more data contrasts with the historical lack of experimental methodologies to quantify the key indicator of T_c exceedance, deposition.

The mutually exclusive or simultaneous nature of erosion and deposition processes has been the object of much debate. Attempts have been made in the field of estuarine sediment dynamics to clarify this question, resulting in laboratory evidence supporting the exclusivity of the processes [16–19], while field measurements support the simultaneous occurrence hypothesis [20–22]. In hillslope soil erosion and sediment transport modeling, the Foster and Meyer [4] coupling of erosion and deposition processes (coupled model hereafter) makes no explicit description of the mutual exclusivity of both physical processes, but mathematically integrates them into their net resultants (net erosion or net deposition), which are considered to be mutually exclusive at every point along a hillslope position for a given flow condition. The mathematical formulation of the Hairsine and Rose [23] model, by contrast, suggests that both erosion and deposition processes occur simultaneously so will be referred to as the simultaneous model hereafter.

Compared to studies in which detachment processes are of interest, those specifically focused on deposition are more complex and often require additional controls on boundary conditions, e.g., [13,24], or measurement of soil surface elevation with mechanical relief-meters, e.g., [25,26], laser scanner, e.g., [27–29] or image-based methods, e.g., [30–32]. Recent advances in the fields of computer vision and photogrammetry have led to the development of Structure from Motion (SfM) photogrammetry, a simple and low-cost alternative to traditional image-based 3D reconstruction methods. SfM has been successfully used in various geoscience and hydrology applications and is currently viewed as the technology that popularized image-based 3D reconstruction for these disciplines, e.g., [33–36]. SfM makes possible the routine collection of surface-change information during experimental erosion and hydrology studies, thus presenting a unique opportunity to address many erosion- and particularly deposition-related questions.

While soil surface elevation measurement techniques are becoming increasingly accessible, no accepted framework exists to utilize elevation change data to determine transport capacity. The objectives of this paper were to (1) propose a methodology to quantify T_c from surface change information collected during rainfall simulation experiment, and (2) apply this new methodology to a study evaluating the effect of shallow subsurface hydrology on T_c .

2. Theory: Transport Capacity and Soil Surface Elevation Change

As implemented in coupled hillslope erosion models, the transport capacity concept describes an upper limit to detachment rate, beyond which no net detachment occurs and deposition dominates. This T_c deficit approach is currently used in many major hillslope soil erosion models (e.g., WEPP, RHEM, EUROSEM) to predict space- and time-varying detachment in eroding rills using a variant of the following framework:

$$Dr \propto Dc \left(1 - \frac{G}{T_c} \right), \quad (1)$$

where Dr is the detachment rate with units of $M L^{-2} T^{-1}$, Dc is the detachment capacity (function of flow hydraulics, $M L^{-2} T^{-1}$) and G is the sediment load in the same unit as T_c (often $M L^{-1} T^{-1}$) [37]. Equation (1) implies that when the eroding flow reaches transport capacity, $G = T_c$, then $Dr = 0$ and any additional sediment added to the flowing runoff would be deposited.

Equation (1) often predicts shifts in dominant detachment and transport processes as flow hydraulics change along the hillslope gradient (example deposition at toe slopes of concave hillslopes). The implications of the coupled model as expressed in Equation (1) is that if the sediment transport capacity designates the maximum sediment load that a flow of a given hydraulic can carry, then flow paths on the hillslope where flow hydraulics are below transport capacity would experience net erosion, those where transport capacity has been exceeded would undergo net deposition and those at transport capacity would experience no net change. In other words, for an area along the hillslope flow path where T_c has been exceeded, the probability PE that erosion would occur in this area is greater than the probability PD that deposition would occur.

At any given point i on a hillslope flow path where flow conditions are characterized by hydraulic parameter Γ (e.g., shear stress or stream power), sediment transport status is determined by one of several possibilities:

$$\begin{aligned} \Gamma > \Gamma_{cr} \wedge G < T_c, & \text{Erosion,} \\ \Gamma < \Gamma_{cr} \wedge G < T_c, & \text{No net change,} \\ \Gamma > \Gamma_{cr} \wedge G > T_c, & \text{Deposition,} \\ \Gamma < \Gamma_{cr} \wedge G > T_c, & \text{Deposition,} \\ G = T_c, & \text{No net change,} \end{aligned} \quad (2)$$

where Γ_{cr} is the critical value of the hydraulic parameter that should be exceeded before erosion starts, and $\wedge$ is the logical and operator. The sediment transport status at the point in consideration is then a function of the spatial distribution of Γ_{cr} and the sediment flux G , which is controlled by the erodibility of all the upslope points draining to the point, thus controlling the sediment flux G .

For hillslope positions where $\Gamma > \Gamma_{cr}$, the set of possibilities is reduced to:

$$\begin{aligned} G < T_c, & \text{Erosion,} \\ G > T_c, & \text{Deposition,} \\ G = T_c, & \text{No net change.} \end{aligned} \quad (3)$$

One can define two complementary probabilities PE and PD as:

$$\begin{aligned} PE &= P[G \leq T_c] = P\left[\frac{dZ}{dt} \leq 0\right], \\ PD &= P[G > T_c] = P\left[\frac{dZ}{dt} > 0\right], \end{aligned} \quad (4)$$

where Z is the elevation at point i , t is time and $PE + PD = 1$.

One of the implications of the coupled erosion-deposition concept as formulated in Equation (1) is that once the flow reaches transport capacity, any additional sediment detached (as the result of the detachment capacity, D_c) is deposited. From this assertion, one can posit that at transport capacity, $PE = PD = 0.5$, i.e., soil particles in the areas of flow path where transport capacity is reached have as much chance of being eroded as deposited.

Assuming steady state conditions and ΔZ , the integration of elevation change over the duration of an erosive event is,

$$\begin{aligned} PE &= P\left[\frac{dZ}{dt} \leq 0\right] = P[\Delta Z \leq 0], \\ PD &= P\left[\frac{dZ}{dt} > 0\right] = P[\Delta Z > 0]. \end{aligned} \quad (5)$$

By observing elevation change on the soil surface after an erosive event, one can determine areas of erosion and deposition where $\Delta Z \leq 0$ and $\Delta Z > 0$, respectively. Since PE and PD are complementary, calculating one gives information on the other.

3. Materials and Methods

3.1. Experimental Infrastructure

The data presented in this paper were collected from rainfall simulation experiments aimed at understanding the effect of subsurface hydrology on soil erosion and channel network development. A detailed description of the experimental setup can be found in [14], but a concise description is presented here for reference. The soil used in these experiments was an unmapped silt loam (73% silt, 17% clay and 10% sand). The soil was packed in a soil box to form a laboratory hillslope 9.75 m $\times$ 3.66 m in dimensions (Figure 1).

The hillslope was set on a constant 5% slope, with side slopes set at 5% allowing flow convergence to a center line along the length of the box. Soil in the box was 0.3 m-thick and underlain by a layer of highly porous landscape fabric and a sand layer in which water table control tubes were buried. An impermeable rubber membrane was placed underneath the sand layer as an isolation between the hydrologically-controlled soil layer and an underlying pea gravel layer. The pea gravel layer provided structural support to the soil and was set on the 5% slope to which the soil layer conformed.

The soil box was equipped with a system of water table control made of a 0.15 m diameter pipe running the 9.75 m length of the soil box at a 5% slope. Inside this water table control pipe, vertical plates created a series of 0.2 m long cascading compartments. Each of these compartments was connected to a water table control tube within the sand layer. At the upslope end of the soil box, a perforated water reservoir delivered subsurface interflow to the soil layer to control the upslope subsurface boundary condition. In addition, the surface boundary condition at the upslope end of the box was controlled with two tubes providing run-on flow at a controlled rate. A 0.25 m tall baffle plate located 2 m from the downslope end of the box separated the upslope study area from a downslope buffer zone.

The rainfall simulator used in this study was based on a trough design. Each trough was equipped with four Veejet nozzles (Spraying Systems Co. Wheaton, IL, USA) fed at a constant pressure of 41.4 kPa. A total of 11 troughs were aligned along the length of the soil box at a 0.92 m spacing.

Soil erosion and deposition were tracked during each rainfall simulation using SfM reconstruction. A digital single lens reflex camera Canon EOS Digital Rebel XT (Canon, Tokyo, Japan) was used to take photographs of the soil for SfM three-dimensional (3D) reconstruction. The camera was mounted on a rail 3.5 m above the soil surface and traveled along 2 tracks spaced 0.5 m on either side of the centerline of the box. On average, it took 35 pictures per track to cover the length of the plot with an overlap of 88% between pictures. The 3D reconstruction process used 28 ground control points (GCPs) that were measured with a Nikon NPR 352 total station (Trimble Navigation Ltd., Dayton, OH) with a maximum distance measurement accuracy of ± 2.02 mm and positional precision of 0.3 mm.

for the experimental conditions. Pictures were processed into 3D point clouds using the SfM software Agisoft PhotoScan v 1.3 [38] and transformed into Digital Elevation Models (DEM) with ArcGIS v 10.3 [39].

Figure 1. Schematics showing the different hydraulic control mechanisms of the laboratory hillslope used (a) and picture of the hillslope in use (b).

3.2. Experimental Setup and Data Description

Data used in this paper were collected with the goal to investigate the effect of subsurface hydrology on soil erosion and channel network development. A total of 8 rainfall simulation runs were performed, corresponding to two replicates of high (H) and low (L) rainfall intensities on a soil under free drainage conditions (LD, HD) and seepage conditions (LS, HS). Rainfall intensities were 17 mm/h for L and 34 mm/h for H events. During each rainfall, run-on was applied at a rate equal to the rainfall intensity over the 35.69 m² area of the soil box. These rates were 1.7×10^{-4} m³/s for L and 3.4×10^{-4} m³/s

for H. Seepage condition was achieved by setting the water table at the soil surface with the water table control pipe in Figure 1. Rainfall duration was 1 h for L events and 30 min for H events, such that the same nominal cumulative volume of $Q = 1.45 \text{ m}^3$ was applied to the soil surface.

During each event, rainfall was applied intermittently, and the soil surface photographed for the 3D reconstruction between each rainfall application. For the H event, the soil surface was digitized at 0, 5, 10, 15, 20 and 30 min, whereas for the L event, soil surface digitization occurred at 0, 10, 20, 30, 40 and 60 min. Data from the last event in each treatment (20–30 min for H and 40–60 min for L) were discarded from this analysis since these events were longer than the preceding four events in a series. With this digitization schedule, soil surfaces at each time increment for an L event received the same amount of rainfall and runoff as the corresponding time increment for an H event. The first four installments of each event received 0.24 m^3 of water, while the last received 0.48 m^3 . A series of 3-mm-resolution DEMs corresponding to each time increment were resampled at 300 mm and used for the T_c estimation.

3.3. Transport Capacity from Elevation Change Information

In this study, DEMs of eroding soil surfaces were available for each time increment of H and L events. During each event, the incremental elevation difference maps ΔDEM_t were calculated by subtracting each DEM_t at time t from the DEM_{t-1} at time $t-1$ in the same event sequence.

The ΔDEM_t maps are considered here as expressions of the erosion and deposition processes that occurred during the timeframe between t and $t-1$. These maps were used to calculate the probability of net erosion PE using Equation (5). A Cullen and Frey graph (skewness vs. kurtosis plot) was used to determine which probability distribution best described the distribution of the ΔZ values calculated across all plots and runs. A logistic model was found to best fit the ΔZ values and was, therefore, used to filter elevation change values and compute probabilities and quantiles in the ΔDEM_t :

$$PE = P_{\text{logistic}}(\Delta Z \leq 0). \quad (6)$$

With this probabilistic approach, T_c is attained at hillslope locations where $PE = 0.5$, i.e., there is as much chance for erosion as there is for deposition. Areas where $PE > 0.5$ are in net erosion regime whereas those where $PE < 0.5$ are in deposition regime. Because each rainfall simulation event caused a net soil erosion, PE was expected to be >0.5 over the entire soil box but may be less than 0.5 in areas of the soil box where deposition dominates.

Because T_c is a flow-hydraulics-dependent entity, its value is expected to vary on the hillslope as flow becomes more and more concentrated. The likelihood of T_c exceedance is, therefore, a spatially varying function. For each run, the spatially-varying PE values were determined for incremental changes in flow hydraulic conditions. To characterize flow hydraulics, flow accumulation maps were determined from post-event DEMs. Flow accumulation (Fac) was used in this paper as a surrogate for spatially-varying runoff discharge, assuming uniform hydraulic conductivity. The flow accumulation is the accumulated weight of all cells flowing into each downslope cell and is calculated from the elevation map. In this paper, a weight of 1 was assigned per cell and hence, the accumulated flow at any given cell is the number of cells draining through that cell. Local slopes were determined from elevation maps resampled at 0.3 m resolution. From flow accumulation and local slope maps, surface flow hydraulics emulating stream power ω were calculated as:

$$\omega = Fac \times Slope. \quad (7)$$

This approximation of flow hydraulics with variable ω is a simplification of the stream power index (the product of the contributing area and slope), which is used to measure the erosive power of water flowing through a watershed, e.g., [40]. The distribution of ω on each map was divided into 10 equal-interval flow hydraulics groups covering the range

of ω values. In each group, PE was determined using Equation (6) and T_c is exceeded for a given flow hydraulic group when $PE < 0.5$. Raster data of ΔZ , Fac and Slope were all sampled at 0.3 m to limit the effect of DEM uncertainty on the analysis.

4. Results

4.1. Flow Hydraulics and Elevation Change

The average planimetric and elevational accuracies of 3D point clouds (evaluated using the total station as reference) were, respectively, 2.4 mm and 0.6 mm. These accuracy values are within the distance measurement accuracy of the total station instrument used to survey GCPs. The precision of these point clouds, measured as the variability of SfM-estimated GCP coordinates within an event sequence, was 0.4 mm in the XY direction and 0.3 mm in the Z direction. The 0.3 m analysis resolution adopted for the DEMs in this paper was, therefore, appropriate to limit the effect of precision uncertainty in the spatial analysis.

Figure 2 shows examples of maps produced from the pre- and post-event DEMs. Additional maps for all runs are provided as supplemental material. Changes in elevation (Figure 2a) show mostly low magnitude elevation changes across the soil surface but with spatial clustering of erosion and deposition processes. At this coarse resolution, patterns of erosion and deposition along flow concentration pathways can be identified as linear features formed by clustering pixels of comparable elevation change magnitude. Flow accumulation (Figure 2b) was dictated by the global topography of the soil surface that allowed flow convergence towards the centerline of the soil box. Towards the downstream end of the plot, the area below the baffle plate was kept undisturbed and acted as a buffer zone, leading to shallower slopes in this area (Figure 2c).

All average elevation changes were negative, consistent with the net soil loss experienced during each rainfall event. Average elevation changes were -0.5 mm for the LD, -0.2 for HD, -0.3 for LS and -0.5 for HS, suggesting no discernable effect of soil subsurface hydrology or rainfall-runoff intensity on the average soil surface deflation.

Values of the stream power proxy ω derived from Figure 2b,c are plotted against observed elevation changes in Figure 3. For all replicates and treatments, a fan-shaped distribution was observed between ω and ΔZ . The range of variation of ΔZ was consistently wide for low ω values and rapidly narrowed as ω increased. The effect of subsurface hydrologic condition was observed on Figure 3. The range of ΔZ variation was wider under seepage conditions than it was under drainage conditions. Standard deviations of ΔZ ($\sigma\Delta Z$) were 2.6 mm for LD and 2.4 mm for HD, but increased to 3.9 mm and 3.8 mm for LS and HS, respectively (Table 1).

Table 1. Average and standard deviation ($\sigma\Delta Z$) of elevation change ΔZ after 0.43 m^3 of runoff on the soil box under drainage and low flow (LD), drainage and high flow (HD), seepage and low flow (LS) and seepage and high flow (HS).

Treatment	Replicate	$\sigma\Delta Z$ (mm)	Average ΔZ (mm)
LD	1	2.2	−0.1
	2	3.0	−1.0
HD	1	2.5	−0.3
	2	2.4	−0.2
LS	1	4.5	−0.2
	2	3.3	−0.4
HS	1	4.0	−0.3
	2	3.6	−0.6

Figure 2. Examples of maps showing elevation change obtained by subtracting pre- from post-event elevation models (a), surface flow accumulation (b) and slope at 0.3 m resolution (c).

Figure 3. Cont.

Figure 3. Change in elevation (ΔZ) as a function of stream power approximation (Flow accumulation $\times$ Slope) for (a) the low intensity event under drainage condition (LD), (b) the high intensity event under drainage condition (HD), (c) the low intensity event under seepage condition (LS) and (d) the high intensity event under seepage condition (HS).

4.2. Effect of Soil Subsurface Hydrology on the Probability of Erosion

Figure 4 graphs probabilities of erosion calculated with Equation (6) as a function of the stream power proxy ω . One can note from this figure that the probability of erosion followed a sinusoidal pattern with decreasing peaks and valleys as ω increased. At the upstream boundary of the soil box (low ω), all treatments had approximately the same probability of erosion. Average PE for the upstream boundary was 0.57 ± 0.02 , suggesting erosion-dominant conditions. The behavior of PE at the initial stage of flow concentration differed based on soil subsurface hydrology. Under seepage treatments (LS and HS), the probability of erosion initially increased to a maximal value, then started declining to follow a sinusoidal pattern, but only dropped below 0.5 for the LS treatment. Under drainage conditions (LD and HD), however, PE mostly decreased from the initial 0.57 probability position and dropped below the 0.5 mark before sharply increasing at high ω .

Figure 4. Probability of elevation loss (erosion) as a function of the stream power proxy Flow accumulation $\times$ Slope (Omega).

For LS and HS, PE values stayed mostly above the 0.5 line but briefly decreased below the 0.5 line at $\omega = 388$ and $\omega = 501$ under the LS treatment. Under drainage conditions, PE curves intercepted the 0.5 threshold values at lower ω values, after which deposition was sustained before a return to erosion-dominated conditions at high ω values. The LD treatment intercepted the $PE = 0.5$ threshold at $\omega = 335$ on the decreasing segment and at $\omega = 499$ on the increasing limb of the PE - ω curve. For this treatment, deposition was dominant between $\omega = 335$ and $\omega = 499$. The HD PE - ω curve crossed the 0.5 threshold at $\omega = 305$, after which deposition was dominant, and returned to the erosion-dominated condition past $\omega = 479$. At very high ω values, all PE curves dramatically increased likely due to the edge effect and flow through the buffer zone, which became concentrated through narrow paths within the depositional delta formed by successive rainfalls.

5. Discussion

5.1. Implications to the Coupled Detachment- T_c Concept

In current process-based hillslope erosion models using the T_c concept, T_c is modeled in channels as a function of flow hydraulics and soil intrinsic properties while detachment rate is a function of the T_c deficit (difference between T_c and sediment load). The consequence

of this coupling between T_c and detachment is that on a plain slope of sufficiently high erodibility drained by a one-dimensional channel receiving runoff and sediment from sheet and splash (or interrill) areas, erosion rate gradually decreases with distance if the channel slope is fixed. Overall, in our study, a general decrease in the probability of erosion (after a brief initial increase under seepage condition) was observed as the proxy to stream power ω increased. This general decrease in PE with ω is partially consistent with the coupled detachment- T_c model. Nevertheless, the sinusoidal pattern observed across all treatments for the $PE-\omega$ function is not consistent with current coupled detachment- T_c formulations and highlights the need for more laboratory and field data to better gain insight into transport mechanisms of sediments. Perhaps the coupled detachment- T_c approach proposed by [4] adequately describes rates of net erosion over broad distances but does not inform on the actual dynamics of sediment transport over the hillslope.

It is also possible that the changes of direction of the $PE-\omega$ curves at local maxima of the sinusoidal patterns occur when multiple concentrated flow pathways merge, forming larger channel systems. In an experimental laboratory study, Ref. [41] showed that increase in sediment load at channel confluence often requires an increase in shear stress via flow constriction and increased velocity for the excess sediment to be transported. When these hydraulic adjustments do not occur, channel depth decreases as sediment load increases at the confluence [41]. The sinusoid $PE-\omega$ pattern observed is also consistent with a series of cascading headcuts, whereby plunge pools formed at the base of headcuts leave a vanishing depositional trail in the downstream direction. A natural hillslope is often drained by a dendritic and complex network of channels merging and bifurcating depending on spatially-variable soil properties and microtopography. Erosion processes on natural slopes are also diverse, including splash and sheet detachment, headcut formation and migration, bank erosion, etc. One would, therefore, expect the behavior $PE-\omega$ curves in Figure 4 to be common on natural hillslope. More research is needed to fully understand the dynamic of sediment transport on actively eroding hillslope for the next generation of hillslope erosion models.

While not the primary focus of this study, we found that detachment and deposition processes were both influenced by subsurface hydrology. Drainage conditions promoted reduced erosion probability and increased deposition opportunities. The dependence of soil erosion on subsurface hydrology has been shown by many, e.g., [14,42–45], but little is known on the effect of these transient near surface conditions on deposition. Nouwakpo and Huang [14] noted that visually-observed deposition patterns from the same experiment presented in this paper suggest higher rate of deposition under drainage conditions. By estimating the probability of erosion as a function of surface flow hydraulics, these earlier observations were confirmed.

Some of the many prediction equations for T_c may already provide a framework to partly account for the effect of subsurface hydrology on T_c . The equation of [46] proposes, for example, an equation that relates T_c to the difference between flow shear stress and critical shear stress. Since critical shear stress has been shown to be larger under drainage condition [43,47], one can expect T_c to be systematically reduced under drainage conditions. Nevertheless, many questions remain on the effect of subsurface hydrology on transport processes and an increased combination of traditional erosion measurement techniques with microtopographic change data will help advance understanding of these processes.

5.2. Limitations of the Proposed Approach and Sources of Uncertainty

The sediment transport capacity concept is currently embedded in many hillslope erosion models but is supported by limited experimental data [1]. The methodology developed in this paper uses formulations of the coupled detachment- T_c to infer a probability-based estimation of T_c as a hydraulic condition whereby a point along the hillslope flow path has as much chance to be eroded as it does to be deposited upon. As formulated in this paper, the probabilities of erosion and deposition PE and PD are complementary (i.e., $PE + PD = 1$) in the experimental conditions where runoff discharge along flow concentration pathways

was high enough to exceed critical values for incipient particle movement (i.e., $\Gamma > \Gamma_{cr}$). As such, this paper assumes that under the experimental conditions, the probability of erosion is non-zero along flow concentration pathways. The proposed methodology is, therefore, adequate in experimental conditions where flow hydraulics is high enough to eliminate situations where $\Gamma < \Gamma_{cr}$. In these low flow situations, probabilities PE and PD may be undefined and the proposed methodology unsuitable.

Sources of uncertainty in this study include the assumption that steady state conditions prevailed throughout the rainfall simulation. In fact, immediately following rainfall stoppage, eroded soil particles would be forced to deposit due to lack of flow for transport along the channel network and thus, may have influenced elevation changes measured in this paper. While the magnitude of these end-of-rain deposition processes was not quantified, they were not expected to consistently impact patterns of sediment transport in all 8 rainfall simulation experiments. Furthermore, the parameter ω calculated as the product of flow accumulation (drainage area) and local slope is assumed to be proportional to the spatially varying stream power function. In the coupled detachment- T_c concept formulated in Equation (1), net detachment is a function of the sediment flux G and not directly related the flow hydraulic stresses. Mapping sediment flux across the eroded surface was, however, not directly available in our study. A proxy to the direct mapping of sediment flux could be achieved by accumulating ΔZ values along flow paths but at the expense of a greater bias between the ΔZ -weighed ω and PE .

6. Conclusions

In this paper, a methodology was proposed to determine hydraulic conditions expressed as a proxy ω of the stream power along hillslope flow paths where transport capacity T_c is reached by defining erosion and deposition probability functions PE and PD and assuming that T_c is reached where $PE = PD = 0.5$. Elevation change data obtained during a study of the effect of subsurface hydrology on soil erosion processes served as an example to demonstrate the proposed methodology. Elevation change ΔZ followed a logistic distribution from which PE and PD were calculated. Overall a general decrease in the probability of erosion was observed as hydraulic levels increased, a result predicted by the coupled detachment- T_c concept. Nevertheless, sinusoidal features in the detailed PE - ω function is inconsistent with the coupled detachment- T_c concept and suggest pulsating erosion patterns along concentrated flow networks with intermittent increases in probability of erosion to local maxima followed by declines to local minima. Additionally, this study confirms earlier work supporting the need to develop a satisfactory framework that accounts for transient field conditions such as subsurface hydrology in the estimation of soil erosion parameters. This study highlights the value of experimental studies combining traditional soil erosion measurement techniques with more advanced three-dimensional reconstruction techniques to characterize changes to soil surface microtopography.

Supplementary Materials: The following are available online at <https://www.mdpi.com/article/10.3390/w13070929/s1>, Figure S1: Flow accumulation (Flow acc.), elevation change (Delta Z) and slope maps obtained after incremental rainfall/runoff events 1Q, 2Q, 3Q, 4Q and 6Q from the 9.75 m $\times$ 3.66 m plots. $Q = 0.24 \text{ m}^3$.

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