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The importance of surface controls on overland flow connectivity in semi-arid environments: results from a numerical experimental approach

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Abstract:

In semi-arid environments, the characteristics of the land surface determine how rainfall is transformed into surface runoff and influences how this runoff moves from the hillslopes into river channels. Whether or not water reaches the river channel is determined by the hydrological connectivity. This paper uses a numerical experiment-based approach to systematically assess the effects of slope length, gradient, flow path convergence, infiltration rates and vegetation patterns on the generation and connectivity of runoff. The experiments were performed with the Connectivity of Runoff Model, 2D version distributed, physically based, hydrological model. The experiments presented are set within a semi-arid environment, characteristic of south-eastern Spain, which is subject to low frequency high rainfall intensity storm events. As a result, the dominant hydrological processes are infiltration excess runoff generation and surface flow dynamics. The results from the modelling experiments demonstrate that three surface factors are important in determining the form of the discharge hydrograph: the slope length, the slope gradient and the infiltration characteristics at the hillslope-channel connection. These factors are all related to the time required for generated runoff to reach an efficient flow channel, because once in this channel, the transmission losses significantly decrease. Because these factors are distributed across the landscape, they have a fundamental role in controlling the landscape hydrological response to storm events. Copyright © 2013 John Wiley & Sons, Ltd.

KEY WORDS runoff; hydrological connectivity; CRUM; surface; vegetation; slope; semi-arid; overland flow

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INTRODUCTION

In semi-arid hydrological systems, the amount of river flow, and potential flood hazard, resulting from rainfall storm events is determined by both the generation of runoff at a point in the landscape and the hydrological connectivity of that generated runoff with the river channel network. The generation of significant river flow in semi-arid environments has been shown by previous work to be related to slope length (Wainwright and Parsons, 2002), gradient (Kirkby *et al*., 2002), surface flow path convergence (Reaney *et al*., 2007), infiltration rates (Puigdefabregas *et al*., 1999), terrace structures (Bellin *et al*., 2009; Lesschen *et al*., 2009; Meerkerk *et al*., 2009) and the spatial configuration of vegetation (Cammeraat and Imeson, 1999; Ludwig *et al*., 2005). This paper presents novel results exploring the interactions between the spatial factors controlling runoff generation and the hydrological connectivity of runoff at the storm event and hillslope scale in a semi-arid environment.

The concept of hydrological connectivity is defined as 'the passage of water from one part of the landscape to another and is expected to generate some catchment runoff response' (Bracken and Croke, 2007). At a point in time, there is either a transfer of water from one point to another or not, and hence, the connection can be seen as a Boolean relationship. As the timescale increases, there is a distribution of connected and disconnected periods, and hence, it is possible to determine a probability of the pair of points being connected. As the spatial scale increases, the landscape will have a distribution of connection probabilities between the many possible pairs of points, and it is the nature of the landscape that will control the form of this distribution (Reaney *et al*., 2011). The different factors that control the hydrological connectivity can be separated into 'structural' and 'functional' (Turnbull *et al*., 2008). The structural factors are those that remain static during the time period of interest, such as topography and vegetation patterns. Functional factors relate to the dynamic processes that occur during the storm event to create the active connections, such as the fluid dynamics of overland flow or the interaction with the storm event to create the flow path lengths. Over longer time periods, structural factors may shape the functional controls, such as the response of vegetation patterns to runoff connectivity creating banded vegetation (Puigdefabregas, 2005).

The key controls on the generation of runoff and overland flow connectivity can be grouped into those

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factors that affect the infiltration rate and those that affect the transmission of water across the hillslope (Bracken and Croke, 2007 for a more lengthy discussion). The infiltration rate affects the time during the storm event at which runoff is generated, the time when depression storage overflows and the transmission losses of the overland flow. The combined result of these effects means that the infiltration rate has a significant control over the discharge from the slope. The spatial configuration of vegetated and bare areas on a hillslope has been shown to affect the amount of runoff generated (Bromley *et al*., 1997; Ludwig *et al*., 2005) due to the vegetated and bare areas having significantly different infiltration characteristics. The factors that affect the transmission include slope length, gradient and flow convergence. Slope length is a universal parameter of every point in the landscape that affects both the distance to the channel and the contributing area above a runoff generating point. The distance to the channel is important because it controls the time required to reach the outflow and hence determines the time available for runoff to be lost as either transmission losses or to cross a disconnecting point on the hillslope. The upslope contributing area determines the amount of run on and affects the soil moisture store, which in turn affects the infiltration rate and the amount of runoff that will be generated. Slope gradient affects many factors related to the generation and transmission of overland flow (Figure 1). Figure 1 shows that with increasing slope gradient, the amount of water that can be held in depression storage decreases, the soil depth decreases and overland flow velocity increases. The surface depression storage and the overland flow velocity are very sensitive over the range 0° to 5°, and the depression storage is extremely sensitive over the range 0° to 2°. The soil depth is linearly related to the surface gradient and hence is not sensitive over a particular range. Slope length, gradient and soil depth have historically been managed in semi-arid environmental to increase local infiltration and water storage through the construction of agricultural terraces. The abandonment of these structures in the late 20th century has caused significant alterations to the hydrological behaviour of these landscapes (Bellin *et al*., 2009; Lesschen *et al*., 2009; Meerkerk *et al*., 2009).

Across semi-arid catchments, there are strong differences in the flow path convergence. These differences affect the distance to an area of concentrated flow, which previous work has shown is a key factor in promoting hydrological connectivity (Reaney *et al*., 2007). Field observations reported by Bull *et al*. (2000) suggested that areas with shorter distance to the flow concentration had greater runoff connectivity. The convergence of the surface flow paths will affect the amount of discharge in two ways. Firstly, the concentration of water into a smaller surface area results in more water being infiltrated in these areas. This infiltration in turn leads to a decrease in the infiltration capacity and hence a decrease in the transmission losses. This decrease therefore results in more efficient flow and greater discharge. The second factor relates to the friction imposed on the flow. The friction decreases with increasing flow depth, resulting in faster flows and greater discharge (Smith *et al*., 2010, 2011). These key controls on runoff generation and connectivity are universal factors found across landscapes and hence are an important macro-scale control on the runoff characteristics of catchment systems.

Limited research exists that explores the interrelationship between structural and functional connectivity (Bracken and Croke, 2007; Turnbull *et al*., 2008). In this paper, we consider the dynamic interactions between the structural and the functional connectivity at the time scale of a storm event and the spatial scale of a hillslope. We explore four factors in detail, two structural and two functional. The structural connectivity factors considered are: i) the effects of the vegetation patterns with different infiltration rates and; ii) the geomorphic structure of the hillslope, in terms of the slope length and cross-slope form. The functional factors considered are: iii) the infiltration rates in response to changing soil moisture; and iv) the overland flow fluid dynamics as expressed through changes in flow rate and efficiency with changing flow

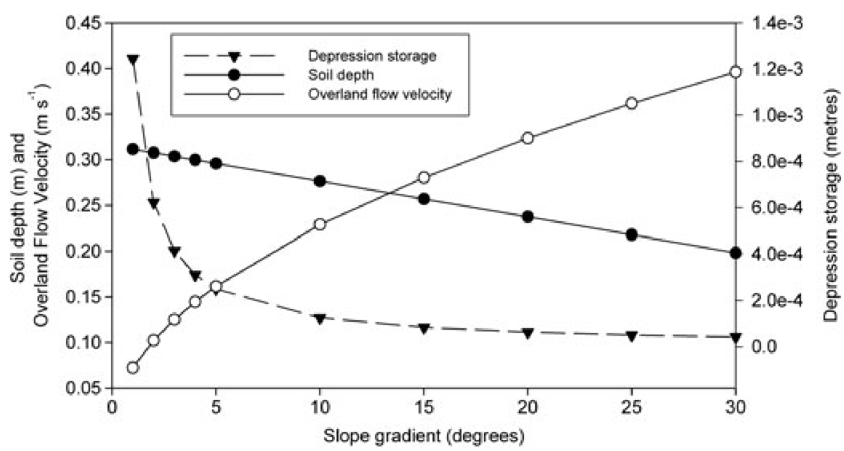

Figure 1. Effects of slope gradient on depression storage, soil depth and overland flow velocity. The effect on the surface depression storage is calculated from the relationship presented in Kirkby *et al*. (2002). The effect on the overland flow velocity is calculated using the Darcy – Weisbach equation. The effect on soil depth is taken from a regression model based on field measurements (Reaney, 2003)

depth. Each of these controlling factors varies across the landscape and over time, and hence, it is difficult to undertake controlled, systematic experiments to determine the dominant controls using field-based studies. Therefore, to enable the required level of experimental control, a numerical experiment based approach has been adopted utilizing a hydrological simulation model, Connectivity of Runoff Model, 2D version (CRUM-2D) (Reaney *et al.*, 2007). This approach enables the separation of the different effects that each parameter has on the generation and the transmission of runoff from the hillslope.

METHODS

Each of the four surface factors (slope length, gradient, flow path configuration and vegetation cover pattern) was investigated over a range of values determined from field data and literature-based sources. These values were then utilized within a numerical experimental approach to assess the sensitivity of the runoff generation and overland flow connectivity processes to that surface control. The effect of the surface control on both the runoff coefficient and the peak discharge rate has been investigated. The simulation model CRUM-2D was selected to represent the hydrological environment. The physical characteristics of the environment were simplified to explore the full range of possible impacts on developing and sustaining hydrological connections that would not be possible to measure in a field experimental situation. Simplification also increased the signal strength of each surface factor enabling a clearer understanding of the controls to be developed.

Simulation context: study area

The location used to provide the simulation context of this research is the Rambla de Nogalte in south-east Spain (37.586°N 1.928°W). This location was selected because it is representative of many semi-arid environments around the world. The basin is located on the border of the Provinces of Almeria and Murcia and receives on average 234 mm of rain annually [Lorca rain gauge (37.671°N, 1.699°W), average for 1958 – 1987]. The main ephemeral channel is a broad wandering gravel bed river with maximum channel widths of 300 m and has a catchment area of 171 km^2. The relief of the catchment is relatively high with maximum and minimum heights of 1300 m and 545 m, respectively. The downstream sections are composed of wide flat alluvial channels, but there are also areas of steep gorge-like bedrock channels, especially in the southern tributaries. The Rambla de Nogalte catchment is composed predominantly of metamorphic rocks, notably phyllites with some greywackes, but there are also extensive areas of clayey conglomerates. Soils are thin throughout the catchment due to the steep slopes and sparse vegetation. Land use within the catchment is predominantly tree crops, with a mix of almonds and olives. There are also extensive areas of matorral, especially on steep slopes adjacent to the ephemeral channels. Other land uses include land ploughed ready for tree planting, small areas of vines and scrub (less dense naturally occurring vegetation similar to matorral including rosemary (L. *Rosmarinus*), thyme (L. *Thymus*) and anthyllis (L. *Anthyllis cytisoides*).

For this research, a base surface was created to reflect characteristics of the hillslopes found both in the Rambla de Nogalte catchment and across south-eastern Spain (Bracken *et al.*, 2008 for more details). Field research was carried out to measure and characterize infiltration rates, surface roughness, soil depth, vegetation patterns and slope characteristics (Reaney, 2003). These field measurements were utilized to generate a base-case typical hillslope that represented the mid-point in the distribution for each characteristic. The details of the base hillslope and storm, range of parameter values and the details of the source are given in Table I.

The connectivity of runoff model (CRUM-2D)

The model simulations in this paper were performed using the CRUM-2D (Reaney, 2003; Reaney *et al.*, 2007). CRUM-2D is a physically based, dynamic, distributed model, which represents the processes occurring during a short-duration storm event in a semi-arid environment. The vertical component of the model represents interception, surface storage, infiltration, soil moisture storage and recharge. Interception is modelled as a non-leaking store that covers a fraction of the ground surface. The depth of the surface depression storage is related to roughness and slope gradient by the relationship presented in Kirkby *et al.* (2002). The infiltration rate is determined using the simplified Green and Ampt model of Kirkby (1975, 1985), which relates the infiltration rate to the soil moisture:

$$i_t = a + \frac{b}{\theta} \tag{1}$$

Where a and b are coefficients and θ is the soil moisture. The surface flow velocity is determined using the Darcy – Weisbach equation:

$$v = \sqrt{\frac{8gDs}{ff}} \tag{2}$$

Where v is the flow velocity ($m\,s^{-1}$), ff is the friction factor (fixed), g is the gravity constant, D is the flow depth and s is the sine of the slope of the energy gradient. The model is spatially distributed on a grid because the even spacing of model cells does not make any prior assumptions about the nature of the hydrological connectivity. To ensure that the connectivity patterns can be fully represented, we routed the water across the surface using the multiple flow path algorithm FD8 (Quinn *et al.*, 1991). The model simulations were performed with a time step of 1 s and have been shown to accurately represent the surface flow dynamics in semi-arid environments (Reaney, 2003; Reaney *et al.*, 2007). Full details of the model can be found in Reaney *et al.* (2007).

Table I. Model parameters derived from field measurements and base simulation results. This configuration represents the mid-point in the distribution for each factor

Parameter	Value	Justification
Slope size	50 m × 50 m	Related to slope lengths found within the Rambla de Nogalte catchment from field measurements and DEM analysis (50 m resolution). The range of hillslope lengths calculated was between 5 – 400 m for the study region.
Cell size	1 m × 1 m	Balances spatial resolution with computational cost
Slope gradient	6°	Related to characteristic slope gradients found within the Rambla de Nogalte catchment from field measurements and DEM analysis (50 m resolution)
Surface roughness	Independent random elevations based on an exponential distribution with a mean of 1 mm	Characteristic value determined from the results of field measurements using a pin metre at a resolution of 1 cm in the cross section (Reaney, 2003).
Soil depth	0.288 m	Soil depth was measured by excavating the rainfall simulation plots after the experiments had been completed and related to slope gradient (Reaney, 2003).
Initial soil moisture	45%	Represents the soil moisture conditions part way through a storm event before the main high intensity rainfall pulse from field data. Field measured after rainfall simulation experiments (Reaney, 2003)
Vegetation	None	No vegetation in the base case
Infiltration parameters (mm hr^{-1})	Bare: $a = 11$; $b = 9$ Vegetated: a = 46; b = 4 Range: $a = 10 - 75$ and $b = 1 - 12$	The infiltration parameters for the Simplified Green and Ampt model (Kirkby, 1975, 1985) were determined using a sprinkler-based rainfall simulator, based on the design of Cerdá *et al.* (1997) on vegetated (anthyllis) and bare areas (Reaney, 2003).
Friction factor, *ff*	1.25	Related to roughness values measured in the field, presented in Reaney (2003). Constant with depth.
Storm length	5 min	Typical high-intensity rainfall pulse duration for the region (Bracken *et al.*, 2008).
Storm intensity	75 mm hr^{-1}	Typical rainfall rates for the region and used in the rainfall simulation experiments to determine the infiltration parameters. Of the measured rainfall, 0.6% occurs at this intensity or above (Bracken *et al.*, 2008).
Base runoff coefficient	4.87%.	The total discharge was 0.76 m^3
Base peak discharge	0.008 $m^3 s^{-1}$	This is the peak discharge for a single 1 s timestep

Typical geomorphic configuration

From a combination of fieldwork and digital elevation model (DEM) analysis, the hillslope lengths and gradients were calculated. Hillslope lengths were defined as the maximum overland flow distance to a defined river channel from the hydrological divide at the top of the slope. The range of hillslope lengths calculated was between 5 – 400 m for the study region. The calculated slope gradients ranged from 1° to 30°.

The data on hillslope length and gradient were then used to generate a series of experimental virtual surfaces, which were created with a range of distances to the nearest channel using a sine wave and a separate slope component. These slopes have a constant slope gradient in the *y*-axis, and the sine wave was added to the *z*-axis. As the wavelength was decreased, so too was the amplitude, thus creating surfaces with a similar slope gradient. Sample surfaces are shown in Figure 2. This range of surfaces was selected to explore the range of hillslope characteristics and the resulting impact on hydrological connectivity.

Storm event

The rainfall storm event used in the simulation is based on a high intensity rainfall pulse event typical of south-eastern Spain. The storm event lasted for 5 min at an intensity of 75 mm hr^{-1}. Of the measured rainfall in the Rambla de Nogalte catchment, 0.6% occurs at this intensity or above. This rainfall pulse was based on an observed storm event in the catchment, which generated a flood event. The return interval for this event was calculated as between 6 and 7 years by Bull *et al.* (2000). Only the main rainfall pulse was investigated because this high intensity rainfall is thought to drive the runoff generation and the creation of connected runoff to the river channel.

Typical infiltration rates

The parameters of the simplified Green and Ampt equation (Kirkby, 1975) for bare and vegetated areas were investigated over the range derived from a set of rainfall simulation experiments (Reaney, 2003). This gave the range of values $a = 10 - 75$ at intervals of five and $b = 1 - 12$ at intervals of one, thus giving 168 parameter combinations. This therefore gave a minimum steady state infiltration rate of 16.7 mm hr^{-1} and a maximum rate of 155 mm hr^{-1}.

Generation of vegetation patterns

Two approaches were utilized for the investigation of vegetation patterns. The first considers the fragmentation of

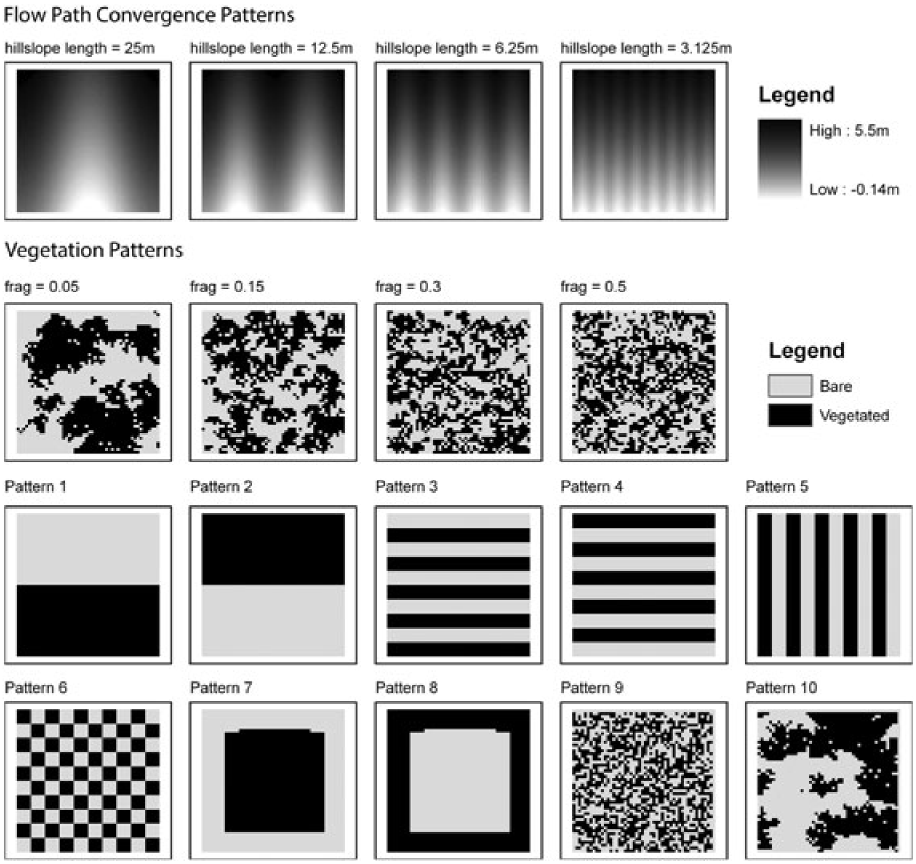

Figure 2. Different slope configurations. Each slope has a grid resolution of 50 × 50 with a cell spacing of 1 m. a, Different distances to the nearest channel; b, Sample fragmented vegetation pattern realizations (black represents vegetation) all show 50% vegetation cover; c, Different defined vegetation patterns (black represents vegetation), all show 50% vegetation cover. In all cases, the hillslope flows from the top to the bottom

vegetation across the hillslope and the second considers systematic patterns.

The generation of fragmented vegetation patterns utilizes the algorithm presented by Tischendorf and Fahrig (2000). This algorithm generates vegetation patterns with variable amounts of fragmentation. A cell was chosen at random on the surface, and the probability of it being defined as vegetated is a function of the surrounding vegetation and the value of the *frag* parameter. A uniform random number was chosen and if it was less than the *frag* value, then the cell became vegetated. However, if the cell had a vegetated neighbour, the cell became vegetated. This algorithm created clustering in the vegetation pattern. The algorithm was used to generate 1000 different realizations of the stochastic patterns for analysis. By the visual comparison of the generated patterns and field measured patterns, it was found that *frag* values of 0.05 – 0.3 gave realistic results comparable with those observed in the Rambla de Nogalte catchment. Example patterns are shown in Figure 2.

The systematic patterns that were investigated all have 50% vegetation cover (Figure 2). The patterns are organized into four sets. Patterns 1 and 2 are reversed versions of each other, one with all of the vegetation in the lower part of the slope and the other with it in the upper part. Patterns 3 to 6 consider the effects of vegetation banding or tiger strips (Thiery *et al*., 1995) on the discharge. The width of the bands is 5 m. The difference between Patterns 3 and 4 is whether the cells next to the river channel are vegetated or not. Pattern 5 is rotated through 90° to give bands running in the same direction as the flow paths. Pattern 6 is a combination of Patterns 3 and 5 to give a chess board effect. Pattern 7 has a contained area of vegetation, whereas Pattern 8 is the inverted image of Pattern 7. Patterns 9 and 10 consider the effects of clustering. Pattern 9 is a near random distribution, whereas pattern 10 has a large amount of clustering.

A third set of vegetation-based experiments was created to quantify the importance of changes in the land cover in relation to the distance from the slope outflow. This objective was achieved through creating a set of 48 vegetation maps with a 2-m band of vegetation that was placed at an increasing distance from the slope outflow with an increment of 1 m per map.

RESULTS

Slope length

The discharge hydrographs for slopes lengths between 5 and 50 m are shown in Figure 3a. The simulations with slope lengths greater than 20 m show the same rising hydrograph limb and the same peak discharge. The slopes with shorter lengths have shorter travel times to

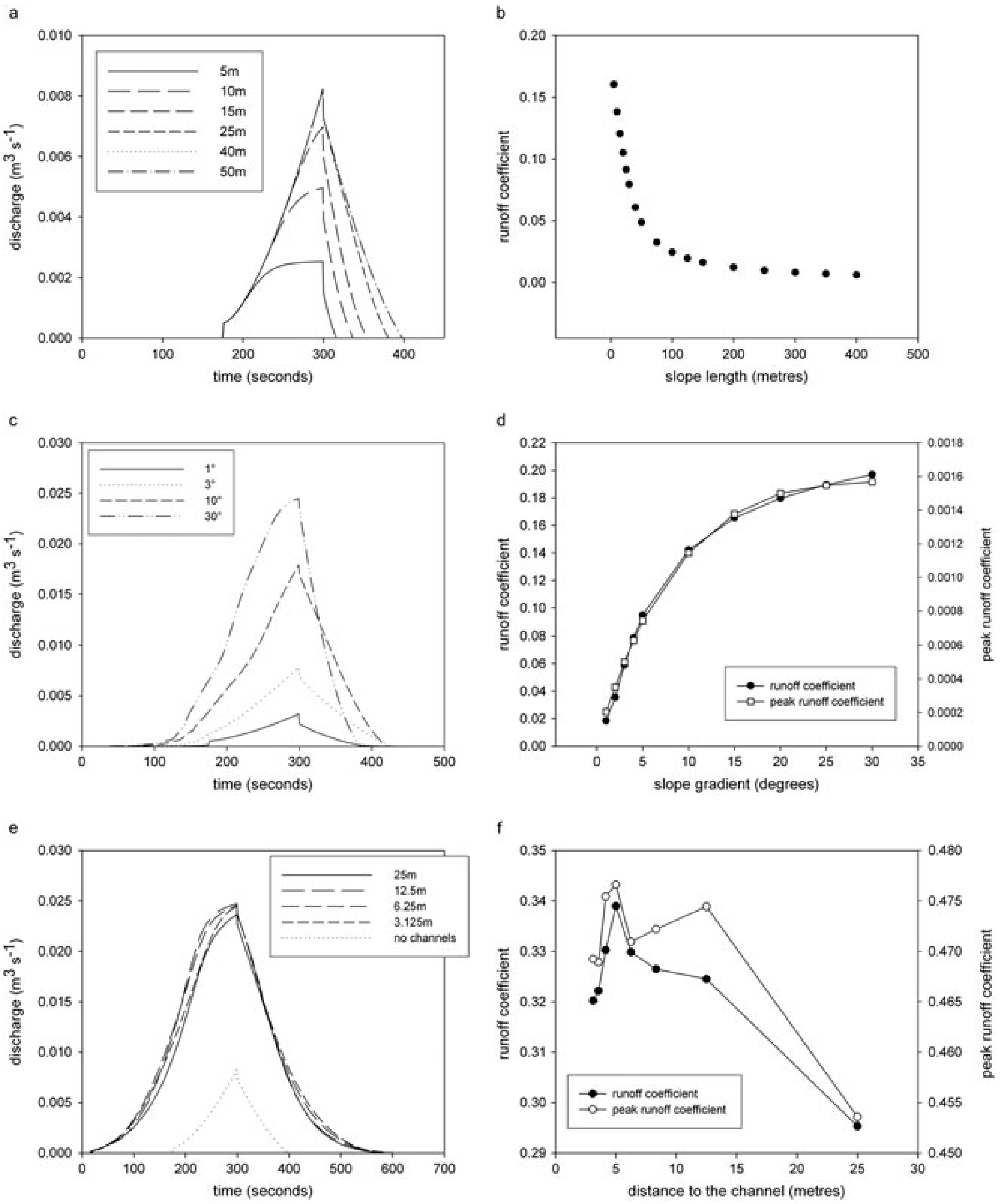

Figure 3. a, Effects of slope length on the simulated discharge hydrograph form; b, Effects of slope length on the runoff coefficient; c, Effects of slope gradient on the hydrograph form; d, on total and peak discharge; e, Effects of the distance to the channel on discharge; f, Effects of the distance to the channel on total and peak discharge. In all cases, the rainfall starts at 0 s and ends at time 300 s at an intensity of 75 m hr^{-1}

the slope base. Therefore, the runoff generated on a larger percentage of the slope is able to leave the slope as discharge. As the slope length increases, the slope of the receding limb decreases, resulting in a longer time after the end of the rainfall until the cessation of runoff. The relationship between the slope length and the runoff coefficient for slope lengths from 5 m to 400 m are shown in Figure 3b. The relationship shows an inverse power law decrease in the runoff coefficient with increasing slope length. This relationship will be affected by the slope gradient, the infiltration characteristics and the storm characteristics. The decrease in the runoff coefficient with increasing slope is caused by transmission losses as the runoff flows down the slope. With an overland flow velocity of 0.1 m s^{-1}, it will take over 16 min for runoff generated 100 m from the base to reach the slope outflow. If flow only occurs for 100 s after the end of the storm, the maximum distance it can travel is 10 m. The introduction of concentrated flow will reduce the transmission losses and is investigated below.

Slope gradient

The effect of slope gradient on the hydrograph form is shown in Figure 3c, and the effect on the total and peak discharge is shown in Figure 3d. As the slope gradient increases the steepness of the rising limb of the hydrograph, the peak and total discharge increase. These increases are greatest over the range 1° to 5°.

Flow path convergence

The discharge hydrographs from the different flow path convergence surfaces are shown in Figure 3e. The introduction of flow convergence onto a surface through the presence of channels results in a significant increase in both total and peak discharge when compared with a surface without flow convergence. The introduction of channels creates slopes on the surface that have gradients greater than 6° of the base case slope. This change decreases the surface depression storage and hence the amount of infiltration excess required to initiate runoff. This accounts for the shorter time to runoff. The flow concentration increases the flow depth and hence the flow velocity. The maximum velocity on the planar surface is $0.067\,\mathrm{m\,s^{-1}}$, whereas on the surface with a single channel, the maximum velocity increases to $0.3\,\mathrm{m\,s^{-1}}$. The runoff coefficient reaches its maximum value for a surface with a distance to the channel of 5 m in the *x*-axis (Figure 3f) with distances greater and less than 5 m giving a lower runoff coefficient.

Effect of infiltration rate

The effects of different infiltration characteristics on the runoff coefficient and the peak runoff coefficient are shown in Figure 4a and 4b, respectively. Over much of the parameter space of the infiltration equation, there is no runoff generated for this storm. The change in the storm runoff coefficient increases linearly above a threshold infiltrate rate.

Combined effects of slope length and slope gradient

Because there is a high degree of interaction and interdependencies between parameters, each should not be considered in isolation. Both the slope length and the slope gradient showed a significant effect on the total and peak discharge. Whereas the slope length has a negative relationship with the runoff coefficient, the slope gradient has a positive relationship. The interactions between these factors are shown in Figure 4, parts c and d.

Figure 4c shows an increase in the runoff coefficient with increasing slope gradient and decreasing slope

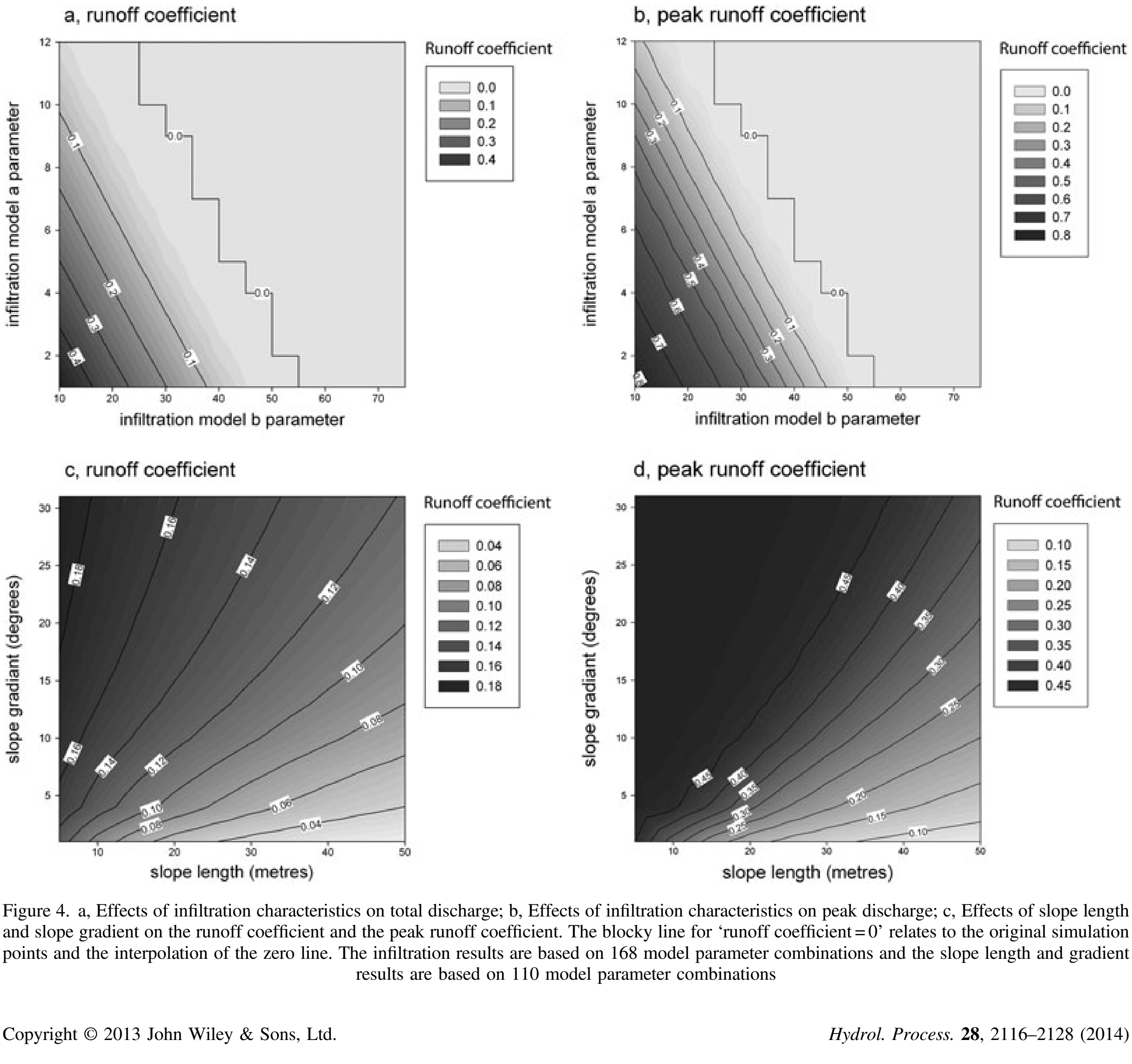

Figure 4. a, Effects of infiltration characteristics on total discharge; b, Effects of infiltration characteristics on peak discharge; c, Effects of slope length and slope gradient on the runoff coefficient and the peak runoff coefficient. The blocky line for 'runoff coefficient = 0' relates to the original simulation points and the interpolation of the zero line. The infiltration results are based on 168 model parameter combinations and the slope length and gradient results are based on 110 model parameter combinations

length. The runoff coefficient is sensitive to changes in the slope gradient below 5°; however, this sensitivity decreases with increasing slope length. As the slope gradient increases, the runoff coefficient increases; therefore, a longer, high gradient slope will behave like a shorter, lower gradient slope. Figure 4d demonstrates that the peak runoff coefficient responds to slope length and gradient in a similar way to the total runoff coefficient. However, the peak runoff coefficient is only affected by increases in slope gradient up to a threshold gradient. Above this point, the peak runoff coefficient does not tend to increase with increasing slope gradient

Effects of vegetation patterns

The runoff coefficient from each surface is plotted against the *frag* value for 1000 vegetation pattern realizations (Figure 5). The greatest variability and both the minimum and the maximum runoff coefficient occur at the smallest value of *frag*. This relates to the vegetation patterns with the greatest amount of clustering. Therefore, all of the vegetation may occupy the upper slope, thus connecting the runoff source areas directly to the slope outflow. It is also possible that the vegetation may occupy the lower slope, thus absorbing the runoff generated upslope. As the patterns become more fragmented, the variability of the runoff coefficient decreases ,and the standard deviation decreases. However, the mean remains constant at 0.4%. At *frag* values above 0.2, there is little change in the distribution of runoff coefficient values.

The introduction of areas with different infiltration rates results in significant changes in both total and peak discharge (Table II). The discharge hydrographs from Patterns 3 through 6 and Patterns 9 and 10 are shown in Figure 6.

There are strong differences in the discharge hydrographs produced by Patterns 1 and 2. Pattern 1, with the vegetation on the lower part of the slope, did not produce any runoff. The vegetated area is able to infiltrate all of the runoff generated in the bare area. Pattern 2, with the bare area at the base of the slope, gave a runoff coefficient of 4.6%; over ten times greater than for the average infiltration conditions. Because no runoff sink is located downslope of the bare area, the runoff source is strongly connected to the outflow.

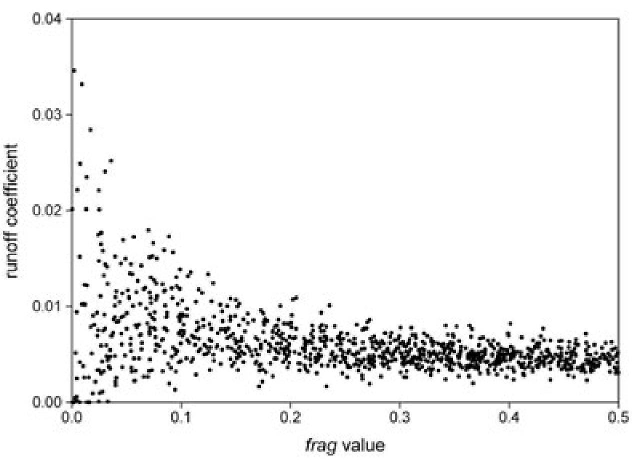

Figure 5. Results from 1000 different realizations of the vegetation pattern generator showing the effect of different levels of vegetation clustering on the slope runoff coefficient

Table II. Presents total and peak discharge values for the different vegetation patterns shown in Figure 2

Pattern in Figure 2	Runoff coefficient (%)	Peak runoff coefficient (%)	Total Q % of no pattern	Peak Q % of no pattern
No pattern (average infil)	0.4	3.7	100	100
1	0.0	0.0	0.0	0.0
2	4.6	15.8	1098.1	429.1
3	0.0	0.0	0.0	0.0
4	1.6	4.8	385.2	131.9
5	1.2	4.5	281.9	122.9
6	0.7	2.4	178.8	64.9
7	2.8	9.1	660.0	247.7
8	0.0	0.0	0.0	0.0
9	0.4	1.5	100.4	40.7
10	1.5	5.7	368.7	154.6

The hydrographs for Patterns 3 to 6 differ in all sections of their discharge hydrographs. Pattern 6 did not produce any runoff, because the 5-m-wide strip of vegetation was large enough to absorb and disconnect all of the runoff generated upslope. Pattern 4 has the highest runoff coefficient at 1.6% (385%) and the highest peak runoff coefficient at 4.8% (131%). All of the runoff generated on the 5 m bare strip is directly connected to the outflow and hence the greater amount of runoff, which is able to leave the slope. From Figure 6, it can be seen that Pattern 4 generates a hydrograph with a steep rising limb. Patterns 5 and 6 both have vegetation cover of 50% at the slope out flow. Pattern 5 gave a runoff coefficient of 1.2% (282%), and Pattern 6 gave a runoff coefficient of 0.7% (179%). The early stages of both hydrographs are the same but then the rising limb of Pattern 5 continues to rise, whereas the rising limb of Pattern 6 declines. This is associated with the large amount of bare area, which is directly connected to the outflow on Pattern 5. This set of results shows that the discharge is highly sensitive to the vegetation cover at the slope base. Pattern 3 has no bare areas at the slope base and has the minimum discharge. Pattern 4 has 100% of bare areas at the slope base and has the maximum discharge. Both Patterns 5 and 6 have 50% bare area at the slope base. On Pattern 6, the bare area only extents for 5 m upslope, whereas for Pattern 5, it extends to the top of the slope, and hence, Pattern 5 has the higher discharge.

Pattern 7 has a total runoff coefficient of 2.8% (660%), whereas Pattern 8 did not produce any discharge. This can also be related to the land cover at the base of the slope. Pattern 7 has a bare area at the base of the slope and hence much greater discharge. Patterns 9 and 10 consider the effect of clustering on the discharge. Pattern 9 has a total runoff coefficient of 0.4% (100%) and a peak of 1.5% (41%). The small scale random pattern of vegetation gives a total discharge almost identical to the averaged conditions but a significantly lower peak discharge value.

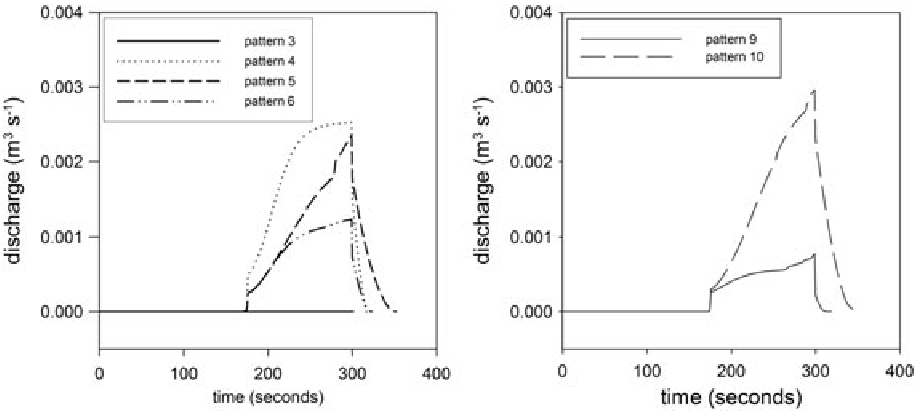

Figure 6. Discharge hydrographs from different vegetation patterns

However, the peak is significantly reduced. Pattern 10 has a higher total runoff coefficient of 1.5% (369%). This may be related to the large bare area connected to the slope outflow. These experiments show that the infiltration characteristics at the slope outflow are very important in determining the discharge from the whole slope. A narrow buffer strip is able to absorb all of the runoff generated upslope. The width required to do this will depend upon the upslope area and the characteristics of the storm event.

There is a non-linear response in the total discharge to the position of the 2-m-wide vegetation strip on the slope (Figure 7). As the strip moves further from the outflow there is an increase in the total discharge. This is related to the increase in the bare area directly connected to the slope outflow. The change in behaviour as the strip moves more than 12 m relates to the size of the source area. When the strip is located in the lower 12 m of the slope, it is located within the source area. Once it moves out of the source area, its impact is decreased, and the discharge increases with the increasing effective slope length.

DISCUSSION

From the modelling results presented, three main features have been revealed as important in controlling the discharge: 1) the distribution of flow path lengths; 2) the slope gradient; and 3) the spatial organization of the hillslope.

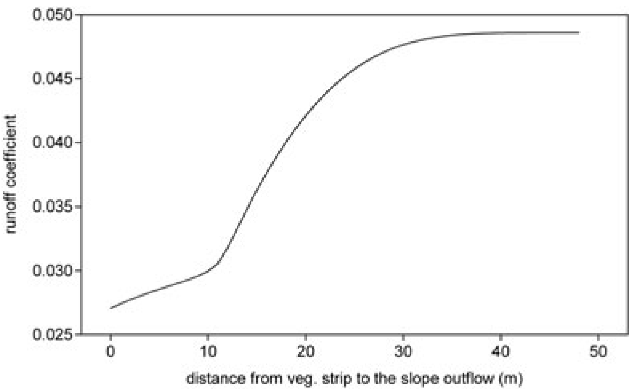

Figure 7. Effects of a strip of vegetation at varying distances from the slope base. Results are from 48 model simulations with the 2 m strip moving at 1 m intervals on the slope

The distribution of flow path lengths

The importance of slope lengths is shown in a number of the modelling results presented. This structural connectivity control affects the total amount of discharge produced from the slope. As the slope length increases, the distance that water must flow over the infiltrating surface increases and hence the greater the time available for transmission losses, which leads to lower runoff coefficients. These results are in agreement with the field results presented by Lal (1997); Gabriels (1999) and van de Giesen *et al*. (2000) and the modelling results of Yair and Lavee (1985) and some of the results of Wainwright and Parsons (2002). Wainwright and Parsons (2002) found that increasing slope length caused a decrease in the runoff coefficient with measured storm events. However, they found that with constant rainfall rates (time averaged from the measured events) the runoff coefficient did not decrease with increasing slope length. The results presented in this paper utilized constant intensity rainfall and did find a decrease the runoff coefficient with increasing slope length. However, the rainfall intensity used in this study enabled the generation of runoff early in the storm event, whereas the lower intensities used by Wainwright and Parsons (2002) may not have caused the generation of runoff until later in the storm event. Wainwright and Parsons (2002) also utilized a 1D slope model rather than the 2D slope used in this paper. The 2D formulation allows for flow concentrations and hence efficiencies of flow not represented in a 1D model.

The results from the simulations with flow path convergence suggest that the reduction in discharge on surfaces with larger distances to the channel can be related to the increase in the flow path length. This increase is caused by the water flowing across the slope to the channel before flowing down slope. Therefore, there is a greater amount of transmission losses and hence lower discharge. These results can be related to the slope length experiments because the introduction of channels onto the slope manipulates the distribution of slope lengths. Hence, this research suggests that there is an optimal spacing for channels for this storm. If the channels are spaced closer together than this value, then there is not enough runoff generated to create efficient channel flow. If the channels are further apart, then the flow is too shallow to create efficient channel flow and more water is lost on the slope as transmission losses. The peak runoff coefficient values show two peaks, one at a distance of 5 m and the second at a distance of 12.5 m (Figure 3f). The 12.5 m peak may relate to there being sufficient catchment area to produce efficient channel flow, whereas the 5 m peak may relate to the shorter travel times across the side slopes. Therefore, it is not the broad scale slope length that is the key control on the amount of water reaching the channel but the distance to a flow concentration. As was shown in some of the results presented by Lal (1997), on longer slopes where rills were able to form, the runoff coefficient increased. This increase in runoff coefficient is due to the decrease in the distance to a flow concentration. Therefore, to fully appreciate the impact that slope lengths have on catchment hydrology, measurements need to be taken at an appropriate scale to capture the small-scale flow concentrations. The slope modelling is able to suggest that the decrease in runoff is due to the transmission losses. This effect is also shown in the investigation of the infiltration rates leading to the curvilinear relationship between the infiltration rate and the total discharge. The distribution of slope lengths within a catchment will have a significant effect on the catchment discharge. There is therefore a tight coupling between the structural connectivity driver of the topography and how this shapes the functional connectivity in the form of the flow hydraulics.

The distribution of slope lengths will be affected by both topography and land management. The topography will affect the drainage density, which will be strongly related to the distribution of slope lengths. Catchments with a high drainage density have a shorter average slope length, and hence, these results predict that they will have higher runoff coefficients. Land management affects the distribution of slope lengths through the introduction of artificial breaks on the slope in the form of roads, terraces, irrigation channels and check dams. Each of these landscape features potentially disconnects the catchment area upslope from the lower slope. In the case of irrigation channels and roads, the generated runoff is efficiently transported to another point in the catchment where it may be stored or released as a concentrated point source. The slope lengths can be extracted from a DEM and a map of the efficient channels. This mapping would enable the estimation of the likely catchment-scale runoff coefficient and the identification of the areas of the catchment that would give the highest slope discharges.

Slope gradient

The slope gradient is a structural connectivity factor that affects a number of parameters, which have a strong influence on the generation and transmission of overland flow. On steeper slopes, the soil depth is shallower (Figure 1), and hence, there is a greater reduction in the infiltration capacity with the same amount of rainfall. The decrease in depression storage means that less infiltration excess is required to initiate overland flow. The increase in the flow velocity results in the runoff reaching the base of the slope in less time, hence the steeper rising hydrograph limbs. On the higher slope gradient discharge hydrographs, water is able to travel from the top of the slope to the outlet within the time period, and hence, the effective contributing area is not able to expand at the same rate as in the earlier part of the storm. This effect leads to a convex form of the hydrograph before the end of the storm event. There is a far greater change in overland flow velocity with slope gradient than in the infiltration rates with decreasing soil depth (Figure 1), and hence, the key driver of the different runoff responses will be the changes in slope gradient.

The threshold slope gradient for runoff leaving the hillslope is dependent upon the slope length with longer slopes reaching the threshold at a higher slope gradient. The location of this threshold can be related to the runoff flow velocity across the slope. On the shorter slopes, the runoff velocity required for runoff generated at the top of the slope to travel to the outflow within the storm duration is less and hence can be achieved at a lower slope gradient. It is possible to calculate the time required for runoff to reach the outflow using the Darcy – Weisbach equation and the time required to fill the depression store. The size of the depression store varies with slope gradient and the time required to fill these stores is given by the difference between the rainfall intensity and the infiltration rate. The maximum travel distance was taken as the diagonal of the surface, i.e. from top left to bottom right. Using the rainfall intensity as above (75 mm hr^{-1}) and infiltration model parameters (41 mm hr^{-1} at 30% soil moisture), the times required to reach the out flow have been calculated (Figure 8). For the standard 300 s storm used in this paper, all of the area to the right of the 300 s isoline would not produce runoff from the whole slope. Therefore, slopes with these combinations of length and gradient would not be able to produce their maximum peak discharge, and hence, it reduces the contributing area and thus the size of the generated flood event.

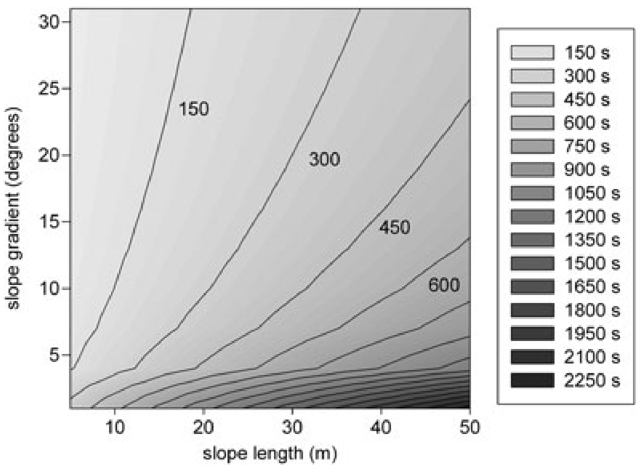

Figure 8. Effects of slope length and gradient on the maximum time in seconds required to reach the outflow as predicted with the Darcy – Weisbach equation

The slope gradient affects the amount of energy available for overland flow, increasing the overland flow velocity and hence affecting the functional connectivity. An increase in the slope gradient decreases the depression storage and the soil depth. From the investigation of the effects of slope gradient on the slope discharge, it was found that both the peak and total discharge are positively related to the slope gradient. The slope gradient effect interacts with the slope length effects. An increase in the slope gradient has been shown to increase the runoff coefficient. Thus, the correct combination of a longer steep slope will have the same runoff coefficient as a shorter slope at a lower gradient, as shown in Figure 8. This relationship is in line with the results found by AbuAwwad and Shatanawi (1997) for basin-scale runoff in Jordan. The reduction in infiltration, as shown by the greater runoff, is also comparable with the results reported by Chaplot and Le Bissonnais (2000) where an increase in runoff was found for increased slope gradient under natural and simulated rainfall. However, the results are in contrast with those reported by Poesen (1984) and Bryan and Poesen (1989) where an increase in slope gradient resulted in a decrease in runoff. This was reported as being due to the erosion of the surface crust exposing the material below, which had a greater infiltration capacity. This process is not included in the model representation and hence it is not possible for the model to be used to investigate this behaviour.

The depth of the surface depression store is very sensitive to slope gradients less than 5° because the surface depression store has a strong influence on the generation and transmission of runoff. From a 50 m DEM of the Rambla de Nogalte catchment, 56% of slopes are less than 5° and 67% are less than 6°. Therefore, there will be considerable variation in discharge among these similar slopes. The model considers the slope of the surface at the spatial scale of 1 m. As the spatial scale changes, the measurement of slope also changes. This is important when up-scaling the results for slope gradient to the landscape scale.

Slope gradients may be altered through land management practices such as terracing and ploughing. The building of terraces reduces the surface slope to effectively zero over most of the hillslope. The regular ploughing of the slopes to remove non-crop vegetation moves large amounts of material down slope and in-fills the valley bottom. This reduces the slope gradient and increases the size of the valley bottom, thus reducing the slope length. The construction of check dams alters the valley floor slope gradient. These landscape management practices therefore have the potential to significantly alter the landscape hydrological dynamics. This effect has been seen in the changing hydrological response of catchments following the abandonment of agricultural structures such as terraces (Bellin *et al.*, 2009; Lesschen *et al.*, 2009; Meerkerk *et al.*, 2009).

The spatial organization of the hillslope

The spatial organization of the slope affects the structural connectivity and can be considered in terms of the surface roughness / topography or in terms of the spatial arrangement of infiltration characteristics. The importance of vegetation patterns in the functioning of semi-arid environments has long been recognized (Cerdá, 1997; Puigdefabregas *et al.*, 1999; Ludwig *et al.*, 2005). The results presented in this paper show the impact of the vegetation pattern on the amount of runoff leaving the slope. The model results show that the response of hillslopes with a fragmented vegetation cover is non-linear and complex, supporting the findings of Cerdá (1997) and that the spatial pattern of vegetated and bare areas, on a slope with the same vegetation cover, can have a significant impact on the amount of runoff reaching a channel at the base of the slope. This supports the comments of Morgan (1995), Fitzjohn *et al.* (1998) and Cammeraat and Imeson (1999) that the spatial configuration of the vegetation patterns needs to be considered when upscaling from the hillslope to landscape scale.

The contribution of an individual model cell to the hillslope discharge is related to its location on the slope. Cells located nearest the channel have the greatest influence while cells located at the top of the slope have the least influence. Therefore, a change in land use close to the channel will have a greater influence than a corresponding change on the upper slope sections. Results demonstrated that the cells located at the base of the slope control the timing of the onset of slope discharge. However, it is the characteristics of all of the cells on the slope that determine if the contributing area will extend to the top of the slope. The investigations of the effects of the vegetation patterns on discharge found that it was the characteristics of the 5 – 10 m strip at the base of the slope, which was the essential factor in determining the slope discharge. However, the response of the hillslope cells nearest to the channel is not independent of the hydrological behaviour of the rest of the slope. The cells located directly above the connected cells may still play a significant role in the hillslope's storm response. The runoff that is generated at these points may be simulated to infiltrate lower down the slope (Reaney, 2008). This water is then stored in the soil and affects both the infiltration rate and the amount of available storage. This feedback effect will then generate more runoff and hence increase the slope discharge. Therefore, the disconnected areas still play a role in the generation of connected flow further down slope.

General discussion

The results presented in this paper for semi-arid hillslopes complement results from temperate hydrological environments. Hopp and McDonnell (2009) used a similar virtual experimental approach to investigate the relative importance of slope angle, soil depth, storm size and bedrock permeability on the discharge from a hillslope (Panola, Georgia). They found that the primary factor driving the hydrological response was the bed rock permeability or the hydraulic conductivity contract between the bedrock and the soil and the second most important factor was found to be slope gradient due to how it affected the ability of bedrock depressions to sore and overflow water (fill and spill). These key factors are also relevant for semi-arid hillslopes, but with the hydrological processes occurring at the surface rather than at the soil-bedrock interface. Recent temperate hydrology studies have highlighted the importance of the soil-bedrock interface for the generation of connected hydrological flows (e.g. Tromp-van Meerveld and McDonnell, 2006a,2006b; Jencso and McGlynn, 2011), and these similarities suggest that the hydrological connectivity dynamics between temperate and semi-arid system may be more similar than previously thought.

The results can be combined into a framework for conceptualizing semi-arid hillslope hydrological systems on the basis of sources and sinks of runoff with the spatial configuration of these processes being the key to determining the system response. The spatial configuration will determine the strength of the hydrological connectivity. Within semi-arid hydrological systems, dominated by infiltration excess runoff generation, whether a point is a source or a sink for runoff water will depend upon its infiltration characteristics in relation to the rainfall intensity and its spatial location on the hillslope. The infiltration rate will determine if rainfall is partitioned into surface flow or infiltrated into the soil. The infiltration rate is determined by the characteristics of the soil and is strongly influenced by the presence of vegetation, with infiltration rates being significantly greater under the plants opposed to the interplant areas. The spatial location of a point on the hillslope determines the potential for additional water from run on from upslope. This additional water alters soil moisture and the local infiltration rate and hence can increase the amount of water that can be partitioned into runoff generation. These factors are then set within a larger spatial structure of the landscape, which influence how the generated runoff moves to the catchment outflow and how geological and soil types are distributed, each affecting the propensity for runoff generation. These factors are further influenced by agricultural structures, such as terraces, and water harvesting structures, such as check dams. When moving between these different scales, care must be taken in make predictions because of the complex interactions between processes (Cammeraat, 2002).

CONCLUSIONS

Understanding the interaction between structural and functional aspects of hydrological connectivity is crucial to furthering our understanding of catchment hydrology and developing new, holistic ways of managing river environments. The results from the modelling experiments show that three surface factors are important in determining the form of the discharge hydrograph: slope length, gradient and the infiltration characteristics at the base of the slope. The CRUM-2D model predicts that an increase in the slope length will lead to a decrease in the runoff coefficient for the slope because of transmission losses following the cessation of rainfall. An increase in the slope gradient produces an increase in both the predicted total and peak discharge. This response is due to the reduction in the soil depth and the surface depression storage and the increase in flow velocity with increasing slope gradient. This increase in velocity decreases the time available for transmission losses and hence leads to the increase in the discharge. The introduction of two contrasting infiltration regimes on a slope can alter the predicted total and peak discharge from a slope. The hillslope discharge was found to be most sensitive to the infiltration rates at the base of the slope. A small change in this area will lead to a large change in the discharge, whereas a large change on the upper sections of the slope is predicted to give a small change in discharge.

The results presented in this paper demonstrate that there are likely to be significant differences in the spatial pattern of the hydrological response to a storm event at the landscape level. Some of the factors that control this response are investigated in this paper and include topographic (slope length, gradient, flow path convergence), soils (infiltration rates) and vegetation patterns. These relationships can be upscaled to the landscape through a series of data processing steps. Information on the topographic factors can be extracted in a systematic way across large spatial areas using detailed digital terrain information. Spatial mapping of infiltration rates is difficult due to the natural variability but inferences can be made between the rates and geology and land cover. Detailed information about the vegetation patterns may be gained from airborne remote sensing. Using this dataset and relationships presented in this paper, a risk map for the probable floodwater source areas for a large catchment could be calculated. Although this approach would not take into account the non-linear scaling effects when moving from the hillslope to the landscape scale, the map would allow the rapid targeting of flood risk mitigation measures in a cost effective way.

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REFERENCES

AbuAwwad AM, Shatanawi MR. 1997. Water harvesting and infiltration in and areas affected by surface crust: examples from Jordan. *Journal of Arid Environments* **37**: 44–452.

Bellin N, van Wesemael B, Meerkerk A, Vanacker V, Barbera GG. 2009. Abandonment of soil and water conservation structures in Mediterranean ecosystems: a case study from south east Spain. *Catena* **76**: 114–121.

Bracken LJ, Cox NJ, Shannon J. 2008. The relationship between rainfall inputs and flood generation in south-east Spain. *Hydrological Processes* **22**: 683–696.

Bracken LJ, Croke J. 2007. The concept of hydrological connectivity and its contribution to understanding runoff-dominated geomorphic systems. *Hydrological Processes* **21**: 1749–1763.

Bromley J, Brouwer J, Barker AP, Gaze SR, Valentin C. 1997. The role of surface water redistribution in an area of patterned vegetation in a semi-arid environment, south-west Niger. *Journal of Hydrology* **198**:1–29.

Bryan R, Poesen J. 1989. Laboratory experiments on the influence of slope length on runoff, percolation and rill development. *Earth Surface Processes and Landforms* **14**: 231.

Bull LJ, Kirkby MJ, Shannon J, Hooke JM. 2000. The impact of rainstorms on floods in ephemeral channels in southeast Spain. *Catena* **38**(3): 191–209.

Cammeraat LH. 2002. A review of two strongly contrasting geomorphological systems within the context of scale. *Earth Surface Processes and Landforms* **27**: 1201–1222.

Cammeraat LH, Imeson AC. 1999. The evolution and significance of soil-vegetation patterns following land abandonment and fire in Spain. *Catena* **37**: 107–127.

Cerdá A. 1997. The effect of patchy distribution of *Stipa tenacissima* L. on runoff and erosion. *Journal of Arid Environments* **36**: 37–51.

Cerdá A, Ibanez S, Calvo A. 1997. Design and operation of a small and portable rainfall simulator for rugged terrain. *Soil Technology* **11**: 163–170.

Chaplot V, Le Bissonnais Y. 2000. Field measurements of interrill erosion under different slopes and plot sizes. *Earth Surface Processes and Landforms* **25**: 145–153.

Fitzjohn C, Ternan JL, Williams AG. 1998. Soil moisture variability in a semi-arid gully catchment: implications for runoff and erosion control. *Catena* **32**: 55–80.

Gabriels D. 1999. The effect of slope length on the amount and size distribution of eroded silt loam soils: short slope laboratory experiments on interrill erosion. *Geomorphology* **28**: 169–172.

Hopp L, McDonnell JJ. 2009. Connectivity at the hillslope scale: identifying interactions between storm size, bedrock permeability, slope angle and soil depth. *Journal of Hydrology* **376**: 378–391. DOI: 10.1016/j.jhydrol.2009.07.047

Jencso KG, McGlynn BL. 2011. Hierarchical controls on runoff generation: Topographically driven hydrologic connectivity, geology, and vegetation. *Water Resources Research* **47**: W11527. DOI: 10.1029/2011WR010666

Kirkby M. 1975. Hydrograph modelling strategies. In *Processes in Human and Physical Geography*, Peel R, Chisholm M, Haggett P (eds). Heinemann: London; 69–90.

Kirkby M. 1985. Hillslope hydrology. In *Hydrological Forecasting*, Anderson MG, Burt TP (eds). John Wiley and sons: Chichester; 3–37.

Kirkby M, Bull L, Reaney S. 2002. The influence of land use, soils and topography on the delivery of hillslope runoff to channels in SE Spain. *Earth Surface Process and Landforms* **27**: 1459–1473.

Lal R. 1997. Soil degradative effects of slope length and tillage methods on alfisols in Western Nigeria. III. Soil physical properties. *Land Degradation & Development* **8**: 325–342.

Lesschen JP, Schoorl JM, Cammeraat LH. 2009. Modelling runoff and erosion for a semi-arid catchment using a multi-scale approach based on hydrological connectivity. *Geomorphology* **109**: 174–183.

Ludwig JA, Wilcox BP, Breshears DD, Tongway DJ, Imeson AC. 2005. Vegetation patches and runoff–erosion as interacting ecohydrological processes in semiarid landscapes. *Ecology* **86**: 288–297.

Meerkerk AL, van Wesemael B, Bellin N. 2009. Application of connectivity theory to model the impact of terrace failure on runoff in semi-arid catchments. *Hydrological Processes* **23**: 2792–2803.

Morgan RPC. 1995. *Soil erosion and conservation*. Longman: Essex; 198.

Poesen J. 1984. The influence of slope angle on infiltration rate and Hortonian overland flow volume. *Zeitschrift für Geomorphologie Supplementband* **49**: 117–131.

Puigdefabregas J. 2005. The role of vegetation patterns in structuring runoff and sediment fluxes in drylands. *Earth Surface Processes and Landforms* **30**: 133–147.

Puigdefabregas J, Sole A, Gutierrez L, del Barrio G, Boer M. 1999. Scales and processes of water and sediment redistribution in drylands: results from the Rambla Honda field site in Southeast Spain. *Earth-Science Review* **48**: 39–70.

Quinn P, Beven K, Chevallier P, Planchon O. 1991. The prediction of hillslope flow paths for distributed hydrological modelling using digital terrain models. *Hydrological Processes* **5**: 59–79.

Reaney SM. 2003. Modelling runoff generation and connectivity for semi-arid hillslopes and small catchments. PhD Thesis, University of Leeds, http://etheses.whiterose.ac.uk/720/.

Reaney SM. 2008. The use of agent based modelling techniques in hydrology: determining the spatial and temporal origin of channel flow in semi-arid catchments. *Earth Surface Processes and Landforms* **33**(2): 317–327.

Reaney SM, Bracken LJ, Kirkby MJ. 2007. The use of the connectivity of runoff model (CRUM) to investigate the influence of storm characteristics on runoff generation and connectivity in semi-arid areas. *Hydrological Processes* **21**: 894–906.

Reaney SM, Lane SN, Heathwaite AL, Dugdale LJ. 2011. Risk-based modelling of diffuse land use impacts from rural landscapes upon salmonid fry abundance. *Ecological Modelling* **222**(4): 1016–1029. DOI: 10.1016/J.ECOLMODEL.2010.08.022

Smith M, Bracken LJ, Cox NJ. 2010. Toward a dynamic representation of hydrological connectivity at the hillslope scale in semi-arid areas. *Water Resources Research* **46**: W12540. DOI: 10.1029/2009WR008496

Smith M, Cox NJ, Bracken LJ. 2011. Modelling depth distributions of overland flows. *Geomorphology* **125**: 402–413.

Thiery JM, D'Herbes J-H, Valentin C. 1995. A model simulating the genesis of banded vegetation patterns in Niger. *Journal of Ecology* **83**: 497–507.

Tischendorf L, Fahrig L. 2000. How should we measure landscape connectivity? *Landscape Ecology* **15**: 633–641.

Tromp-van Meerveld HJ, McDonnell JJ. 2006a. Threshold relations in subsurface stormflow: 1. A 147-storm analysis of the Panola hillslope trench. *Water Resources Research.* DOI: 10.1029/2004WR003778

Tromp-van Meerveld HJ, McDonnell JJ. 2006b. Threshold relations in subsurface stormflow: 2. The fill and spill hypothesis: an explanation for observed threshold behavior in subsurface stormflow. *Water Resources Research.* DOI: 10.1029/2004WR003800

Turnbull L, Wainwright J, Brazier RE. 2008. A conceptual framework for understanding semi-arid land degradation: ecohydrological interactions across multiple-space and time scales. *Ecohydrology* **1**: 23–34.

Van de Giesen NC, Stomph TJ, de Ridder N. 2000. Scale effects of Hortonian overland flow and rainfall–runoff dynamics in a West African catena landscape. *Hydrological Processes* **14**: 165–175.

Wainwright J, Parsons AJ. 2002. The effect of temporal-variations in rainfall on scale dependency in runoff coefficients. *Water Resources Research* **38**: art-1271.

Yair A, Lavee H. 1985. Runoff generation in and semi-arid zones. In *Hydrological Forecasting*, Anderson MG, Burt TP (eds). John Wiley & Sons Ltd.: London, UK; 183–220.

