Effect of spatial heterogeneity of runoff generation mechanisms on the scaling behavior of event runoff responses in a natural river basin

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This paper presents a theoretical investigation of the effects of spatial heterogeneity of runoff generation on the scaling behavior of runoff timing responses. A previous modeling study on the Illinois River Basin in Oklahoma had revealed a systematic spatial trend in the relative dominance of different runoff generation mechanisms, attributable to corresponding systematic trends in landscape properties. Considering the differences in the timing of hillslope responses between the different runoff mechanisms, this paper explores their impacts on the catchment-scale runoff routing responses, including how they change with spatial scale. For this purpose we utilize a distributed, physically based hydrological model, with a fully hydraulic stream network routing component. The model is used to generate instantaneous response functions (IRF) for nested catchments of a range of sizes along the river network and quantitative measures of their shape, e.g., peak and time to peak. In order to separate the effects of soil heterogeneity from those due to basin geomorphology, the model simulations are carried out for three hypothetical cases that make assumptions regarding landscape properties (uniform, a systematic trend, and heterogeneity plus the trend), repeating these simulations under wet and dry antecedent conditions. The simulations produced expected and also surprising results. The power law relationship between the peak of the IRF and drainage area is shown to be flatter under wet conditions than under dry conditions, even though the (faster) saturation excess mechanism is more dominant under wet conditions. This result appears to be caused by partial area runoff generation: under wet conditions, the fraction of saturation area is about 30%, while under dry conditions it is less than 10% for the same input of rainfall. This means travel times associated with overland flow (which mostly contributes to the peak and time to peak) are, in fact, longer during wet conditions than during dry conditions. The power law relationship between peak and drainage area also exhibits a scaling break at around 1000 km², which can be shown to be related to the peculiar geomorphology of the catchment.


1. Introduction

This paper reports on a diagnostic study of event runoff responses on the 2500 km² Illinois River Basin in Oklahoma, carried out as part of the Distributed Model Intercomparison Project Phase 2 (DMIP2), which is coordinated by NOAA [Smith et al., 2009]. Having completed a previous study of runoff generation mechanisms on this catchment [Tian et al., 2011; Li et al., 2011], in this paper we focus more on the timing response, as represented by event instantaneous response functions (IRFs) at the catchment scale. In particular, we seek to understand the process controls on the IRFs, including how they vary with scale (size of subcatchment) and location (position on the river network in relation to the catchment outlet), and the manifestation of the spatial heterogeneity of runoff generation processes on this scaling behavior.

The motivation for this project is twofold. The primary motivation was the results of a previous distributed modeling study [Li et al., 2011] on this catchment, which had shown that its runoff generation response exhibited considerable heterogeneity, while being underlain by a systematic variation or trend. Li et al. [2011] showed that the heterogeneity of soil hydraulic conductivity and topographic slope, along with a systematic spatial trend (both generally increasing in the downstream direction, from headwaters to the outlet), contributed to a corresponding spatial trend in the relative fractions of saturation excess overland flow (Dunne runoff) and subsurface stormflow. Runoff generation in headwater catchments toward the eastern part of the catchment was dominated by saturation excess mechanism, whereas runoff generation in subcatchments...
closer to the outlet was dominated by subsurface stormflow, with a systematic transition in their relative dominance along a gradient from the headwaters to the outlet. Previously, Vivoni et al. [2007] had also found similar behavior in the nearby Baron Fork catchment at Eldon, Oklahoma, using another distributed hydrological model. The immediate question that arises is, how does the spatial heterogeneity of runoff generation mechanisms that result from the heterogeneity of landscape properties manifest in the runoff routing (timing) response of the catchment? In many practical applications there is a tendency to treat runoff generation (i.e., volume) and runoff routing (i.e., timing) responses as independent processes and to decouple their conceptualizations. This may be adequate in large catchments where the network residence time is the dominant component of overall catchment residence time. However, in small and intermediate catchment sizes the hillslope response time can be a dominant component, and therefore, heterogeneity of hillslope responses can still impact the runoff routing response at the catchment scale. In this paper, we wish to explore the ramifications of the heterogeneity of runoff generation behavior on the runoff routing response, using the same distributed model as used before [Li et al., 2011] but highlighting more the processes associated with runoff timing.

Ever since the introduction of the geomorphologic instantaneous unit hydrograph (GIUH) by Rodriguez-Iturbe and Valdes [1979], there has been a lot of research activity on the physical controls of the GIUH. Considering the GIUH as the probability density function (pdf) of travel times and characterizing the shape of the GIUH in terms of dispersion (of travel times), a number of subsequent studies have explored the relative contributions of geomorphology, channel hydraulics, hydraulic geometry, and hillslope timing to the total dispersion [Rinaldo et al., 1991; Snell and Sivapalan, 1994; Robinson et al., 1995; Saco and Kumar, 2002; Botter and Rinaldo, 2003]. These studies also addressed the question of how the shape of the unit hydrograph (or, generally, the instantaneous response function) changed with changing scale (catchment size). The general consensus is that the relative contribution of hillslope dispersion decreases with increasing catchment size, whereas geomorphologic dispersion increases. Hydrodynamic dispersion is unimportant at any scale, whereas kinematic dispersion remains important at all scales. In the Illinois River Basin, the subject of this paper, we have a unique situation where there is a systematic change to the hillslope timing response with downstream distance, i.e., increasing contributions of (relatively slower) subsurface stormflow relative to the (faster) overland flow response. How does this phenomenon impact not only the shape of the IRFs but also how the shape changes with increasing scale? This question is the second motivation for this work.

Guided by these motivations, the objective of this paper is to explore how the spatial heterogeneity and systematic trends in landscape properties (topographic slope and saturated hydraulic conductivity) impact the shape of the IRFs associated with individual events [Robinson et al., 1995; Saco and Kumar, 2002]. In this case, we want to explore how the relative roles of the geomorphologic, hillslope, and kinematic dispersion components of total dispersion change with both catchment size and the level of antecedent wetness of the catchment and the role of the spatial heterogeneity of runoff generation mechanisms. A major advantage of this study over many previous ones is that we will be using a comprehensive distributed hydrological model (distributed at the representative elementary watershed (REW) or subcatchment scale) that simulates both runoff generation processes and runoff routing processes in detail, including hillslope routing, and we can include the effects of the spatial heterogeneity of soil hydraulic conductivity and topographic slope on subsurface stormflow within hillslopes, the hydraulics of overland flow on hillslopes, and the hydraulics of flows in the channel network in a realistic and physically defensible manner. By including a more advanced treatment of channel network routing, this model is also able to explicitly capture the effects of kinematic dispersion [Saco and Kumar, 2002], which is otherwise difficult to include in lumped routing models.

The remainder of this paper is organized as follows. We begin in section 2 with a brief description of the THREW model, the distributed, physically based model that will be used throughout the study, including some details of its channel network routing component. We then introduce the study area and use data analysis and results of previous modeling studies [Li et al., 2011] to illustrate the nature of spatial heterogeneity exhibited by the catchment and how it impacts runoff generation mechanisms. We then present the details of numerical experiments that we carried out with the use of this model to elucidate the effects of spatial heterogeneity of runoff generation mechanisms and how they impact on the shape of the IRFs, including how they change with catchment size. The results in both cases are interpreted in terms of what has been learned through previous theoretical studies, including those related to the GIUH and its extensions. We conclude with a brief summary of the main results, their implications for design practice, and a discussion of avenues for further research.

2. Background

2.1. Distributed, Physically Based Hydrological Model

This is fundamentally a model diagnostic study, using a model already calibrated to a specific catchment to explore some general questions about runoff event responses. Here we present some details of the model being used, namely, the Tsinghua Hydrological Model, or THREW. THREW is a spatially distributed, physically based hydrological model based on the REW approach pioneered by Reggiani et al. [1998, 2001]. Within the REW approach, the representative elementary watersheds serve as fundamental building blocks, which are linked together by the river channel network. Each REW is further divided into a number of subregions according to their hydrological function and known (or assumed) organizational structure. The number of subregions can vary according to the major landscape conditions. For example, a snow-covered zone is important for cold regions but could be neglected where snowfall, accumulation, and melt are not significant [Tian et al., 2006]. The subregions included in this work are a saturated zone, an unsaturated zone, a vegetated zone, a bare soil zone, a substream network, and the main channel reach.

THREW is quite comprehensive in that it includes all key processes contributing to the water balance of
catchments. It incorporates surface runoff (overland flow) by both infiltration excess (Horton runoff) and saturation excess (Dunne runoff) mechanisms and also includes subsurface stormflow. Groundwater discharge from the regional confined aquifer is not explicitly included in the version used in this paper but could be included if necessary and, indeed, has been included in previous studies. The model, including its previous versions, has been applied successfully in many catchments in Australia, Europe, China, and the United States [Lee et al., 2007; Tian et al., 2006, 2011; Mou et al., 2008; Li et al., 2011]. For more details about THREW, see Lee et al. [2007], Tian et al. [2006, 2011], and Mou et al. [2008].

The input data for THREW are organized at the REW level, i.e., those data with different spatial resolutions should be processed to match the scale of REW by either upsampling or downsampling methods. More detailed information about the application of THREW at the Illinois River Basin near Tahlequah, especially the parameter calibration procedure, is given by Tian et al. [2011] and Li et al. [2011] and will not be repeated here.

Since this modeling study focuses on runoff timing, details of the runoff routing on the hillslopes, with respect to both overland flow and subsurface flow and associated hillslope routing, and the transport processes within the channel network are more directly relevant. The details of these have been presented before in different contexts [Reggiani et al., 2001; Lee et al., 2006]. For completeness, a summary of the routing algorithms is presented in Appendix A. Broadly speaking, overland flow on hillslopes is modeled through the use of a kinematic wave approximation, applicable to the saturated areas only. Subsurface stormflow timing is modeled through the use of lumped reservoir routing at the REW scale, with a closure relation (i.e., storage-discharge relationship) that is derived through spatial integration of Darcian flow in heterogeneous catchments [Lee et al., 2006]. Runoff routing through the river network is modeled through the use of coupled mass and momentum (force) balance equations written directly at the scale of the river network [Reggiani et al., 2001] in the form of coupled difference equations. For the purposes of this application, we use a diffusive wave approximation of the momentum balance equation. The network model assumes a rectangular channel cross section, with the channel width assumed to increase in the downstream direction according to a power function relationship with upstream drainage area. Channel roughness is chosen on the basis of known bed material and is then calibrated to match observed flood hydrographs.

2.2. Study Area

The analyses of runoff event responses are carried out in a specific catchment, namely, the Illinois River Basin, but the questions explored are general. The study catchment is located at the border of Oklahoma and Arkansas and has a drainage area of about 2500 km². The climate is of the temperate continental type, with mean annual precipitation of about 1140 mm and mean annual free water (potential) evaporation of about 1060 mm. The dominant land cover is deciduous broadleaf forest, according to the 2001 National Land Cover Data (NLCD) based on the International Geosphere Biosphere Program (IGBP) classification system [Eidenshink and Faindein, 1994]. The basin is occupied by porous limestone overlain by cherty soils, which leads to high infiltration capacities and efficient subsurface drainage [Peters and Easton, 1996; Smith et al., 2004]. The shape of the basin is somewhat unique, with an upper half that is leaf-like, similar to most basins, up to a drainage area of about 1000 km², but the remainder of the area forms what can literally be described as a “bottleneck,” a part of the catchment that is long and narrow from about 1000 km² up to the catchment outlet at a drainage area of 2500 km². This catchment shape is reflected in a complex channel network structure, with a two-part mainstream length versus area relationship. One of the questions that will be explored will be what its impact is on the scaling of IRF peaks and times to peak.

2.3. Spatial Variability of Runoff Generation

Li et al. [2011] have previously applied the THREW model to the Illinois River Basin for the period of 1 October 1996 to 30 September 2006. For the application of the THREW model, the whole Illinois River Basin has been subdivided into 83 REWs, as shown in Figure 1. Most of the data used in this work were provided by the Office of Hydrologic Development, National Weather Service, NOAA, as a contribution of DMIP2 [Smith et al., 2009]. Among them, hourly streamflow data were provided at multiple locations, for example, Savoy, Watts, and Tahlequah, as shown in Figure 1. The upstream contributing areas of these three locations are about 434, 1680, and 2454 km², respectively, and they are nested, i.e., Savoy is upstream of Watts and Watts is upstream of Tahlequah. The model has been calibrated against the observed streamflow at Tahlequah, with a Nash-Sutcliffe coefficient of 0.73. It was then validated against the observed streamflow at Savoy, resulting in a Nash-Sutcliffe coefficient of 0.64, and at Watts, with a Nash-Sutcliffe coefficient 0.80. Note all these coefficient values were for the period 1 October 1996 to 30 September 2005. For more details about these data, the parameter calibration procedures used, and more detailed results on the spatial variability of runoff generation mechanisms, see Li et al. [2011].

The modeling study by Li et al. [2011] revealed that there is a systematic spatial trend in the runoff partitioning within the Illinois River Basin, as shown in Figure 2, and this is the motivation for the present study. Figure 2 (left) shows the fractions (expressed as percentages of the total annual volume of runoff generation) of subsurface stormflow and saturation excess (Dunne) overland flow, which are estimated from model simulations and then averaged over the annual time scale. Because of the porous limestone and relatively permeable soils within this basin, infiltration excess runoff (Horton runoff) occurs only rarely and can thus be deemed negligible [Tian et al., 2011; Li et al., 2011]. Figure 2 (left) clearly shows that the fraction of subsurface stormflow (out of the total runoff volume annually) increases from upstream to downstream (east to west, from as low as 11% to as high as 92%), and the volume fraction of Dunne overland flow decreases in a corresponding manner (from east to west, from as high as 89% to as low as 7%). These changes in the downstream direction are fairly systematic but do exhibit some variability around the systematic trend.

By way of explanation, Li et al. [2011] also showed that the spatial trend of runoff partitioning predicted by the model is due to the corresponding spatial trends in the
Figure 1. Illinois River Basin near Tahlequah, Oklahoma, and the delineation into 83 representative elementary watersheds (REWs). Note the locations of three gauging stations and the outlines of the first-order catchment.

Figure 2. Control of spatial variability of landscape properties on the spatial pattern of runoff generation mechanisms in Illinois River Basin. (left) Percentage of subsurface stormflow and saturation excess overland flow out of the total volume of annual runoff generation within each REW. (right) Mean values of saturated hydraulic conductivity and topographic slope averaged across each REW.
landscape properties that control runoff generation processes and, in particular, saturated hydraulic conductivity (derived from the State Soil Geographic (STATSGO) Database) and topographic slope (derived from digital elevation models (DEMs)). The occurrence and magnitudes of both Dunne overland flow and subsurface stormflow are, of course, strongly linked since they are both connected to the position of the water table. When considering the relative fraction of each runoff component within the total volume of runoff generation in a year, one should also recognize that a “competition” exists between Dunne overland flow and subsurface stormflow. With high hydraulic conductivity or steep topographic slope or both, more soil water flows out of the ground surface as subsurface runoff, leading to smaller soil water and groundwater storages, leading to a smaller saturated area fraction and hence a smaller fraction of Dunne overland flow generation. With the same climate and soil properties, the fraction of saturated area in a steep catchment would be less than that in a flat catchment, whereas the fraction of subsurface flow would be higher. The same applies to catchments with more permeable soils as opposed to less permeable soils, everything else being equal. Therefore, in general, one would expect subsurface runoff to be more dominant with increasing hydraulic conductivity or topographic slope or both. Figure 2 (right) presents estimated values of REW-scale averages of saturated hydraulic conductivity and the REW-scale averaged land surface slope estimated for all 83 REWs. On the basis of the arguments presented above, one can easily associate the spatial trends in the two runoff generation mechanisms to these drainage properties of the landscape, as shown in Figure 2 (right).

[15] Note that the results reproduced above from Li et al. [2011] are model-based predictions and inferences, and there is no independent confirmation that the reported spatial variability of runoff generation mechanisms is actually true. In spite of this, the results are quite remarkable and raise the questions as to how such spatial variability manifests itself in event-scale runoff responses and how they scale with catchment size and how they impact on regional flood frequency. This is motivation for the present study, which will assume the previous results from Li et al. [2011] and will use the previously calibrated model to investigate the runoff timing responses at catchment scale.

[16] Figure 3 presents the fraction of subsurface stormflow within the total runoff volume as predicted by the model, not only at the annual scale (as in Figure 3 (top)) but also at the monthly scale for two selected months, March and September, as a function of locations of the 83 REWs (location is expressed in terms of cumulative drainage area in the downstream direction). The month of March is representative of the wet season, and September is representative of the dry season. On average, the percentage of subsurface stormflow is smaller in the wet month and larger in the dry month. This can be explained in terms of the position of water table depth. Usually, in a basin, a 1 year period can be divided into two segments, the wet season (including winter and spring) and the dry season (including summer and fall). In the wet season, the basin experiences wetter soil moisture and a higher water table level that is closer to the surface; in the dry season, the basin experiences drier soil moisture and a low water table level that could be detached from the land surface. When the water table position is high, saturation excess can be large, and hence the volume fraction of subsurface stormflow would be smaller. When the water table position is low, saturation area fraction is smaller, meaning most of the runoff will occur via subsurface stormflow. In Figure 3 there is considerable scatter: the large scatter at small scales is because these subcatchments are distributed all around the catchment and reflect the full heterogeneity of landscape properties. Note also the steady increase of subsurface stormflow contribution in REWs located beyond 1000 km²; these are located within the bottleneck region, which exhibits an increase of both hydraulic conductivity and topographic slope with downstream position.

[17] As a matter of interest, the effects of the spatial (downstream) and temporal (seasonal) variations of the dominant runoff generation mechanisms can also be seen in the shapes of other streamflow signatures. Figure 4 presents the regime curves (scaled mean monthly variation of streamflow, i.e., mean monthly flows divided by the mean annual flow) and the flow duration curves (hourly flows normalized by the mean hourly flows), all of which are estimated from observed streamflows at Savoy, Watts, and Tahlequah for the period September 1996 to October 2006. The regime curves, which are a standard way to describe intra-annual variability of runoff responses, all exhibit obvious seasonality, i.e., high mean monthly discharge in the wet season and low mean monthly discharge in the dry season. Li et al. [2011] suggested that this seasonality in the streamflow is in phase with the seasonality of the two runoff components. Surface runoff contributes significantly to the high flows and dominates in the wet season, and subsurface runoff contributes mostly to low flows and dominates in the dry season. It is interesting to note that the observed regime curve at Savoy is higher during the wet season than at the downstream locations (Watts and Tahlequah) but reverses itself and is lower in the dry season. This suggests that there is an increase in the dominance of slow subsurface runoff from upstream to downstream and a corresponding decrease of surface runoff. The hourly flow duration curves (FDCs) at the three locations also show similar spatial trends, as seen in Figure 4 (bottom). Over the low-flow range the normalized FDC at Savoy is significantly lower than the FDCs at Watts and Tahlequah, while over the high-flow range it is quite the opposite. This suggests a higher contribution of subsurface stormflow at Tahlequah, whereas the opposite is the case at Savoy. These observations are consistent with model predictions within Illinois River Basin (not presented here for reasons of brevity) in terms of the spatiotemporal trends in the runoff generation mechanisms.

3. Results of Numerical Experiments

[18] The goal of the study is to try and understand how the spatial variability of runoff generation processes might manifest in event-scale runoff (timing) responses of subcatchments of various sizes within the Illinois River Basin. The methodology we have adopted is to use the previously calibrated THREW model to carry out several numerical experiments, now under hypothetical conditions, under scenarios involving different assumptions about the spatial heterogeneity of the landscape properties. All parameters
adopted here are the same as obtained by Li et al. [2011], including calibrated values, except for the Manning’s roughness factors, which are changed to 0.13 for overland flow and 0.08 for channel flow on the basis of land cover data and channel measurements [Smith et al., 2009]. In this way, we want to address the following questions: How does the spatial heterogeneity of runoff generation mechanisms propagate through the river network? How does it manifest in the shape of the instantaneous response functions at the catchment outlet and at various points within the catchment? How does the shape of the IRF, as characterized by the peak of the IRF and the time to peak, scale with the size of the catchment (i.e., scaling behavior)?

3.1. Event-Scale Responses: Runoff Generation

[19] Because of the interest in the IRFs and how their scaling behavior is affected by spatial heterogeneity of runoff generation mechanisms, all numerical experiments will be carried out for a hypothetical rainfall pulse of 1 h duration and 30 mm/h intensity, which is uniformly applied throughout the Illinois River Basin. Considering the size of the catchment and the main runoff generation mechanisms, the choice of a 1 h storm is deemed close enough to an instantaneous pulse.

[20] Because of the focus on spatial heterogeneity of runoff mechanisms (arising from the spatial heterogeneity of landscape properties), the event simulations are repeated.

Figure 3. Spatial heterogeneity of the fraction of subsurface stormflow volume (out of total runoff volume) at annual and monthly scales within the Illinois River Basin. All presented values are averaged from the model predictions for each REW for the period 1 October 1996 to 30 September 2006. Drainage area is used to denote the relative location of the REWs within the whole Illinois River Basin.
for three scenarios, cases I–III, depending on the extent of spatial heterogeneity assumed, as will be explained below. In particular, motivated by the spatial trends shown in Figure 2, in this study we focus on the spatial distributions of topographic slope and hydraulic conductivity. Figure 5 (triangles) presents the magnitudes of estimated topographic slope and saturated hydraulic conductivity, averaged over each of the 83 REWs, as functions of cumulative upstream drainage area, which is used as a surrogate for relative location along the river network. (Note that the REW at the outlet has a cumulative drainage area of 2454.3 km$^2$, but its properties are averaged over its local drainage area, which is 39.3 km$^2$.) In spite of the scatter in the plot, there is a tendency for both properties to increase with increasing drainage area (or distance downstream), especially after a threshold area of about 1000 km$^2$. This is especially the case for saturated hydraulic conductivity.

[21] On the basis of these trends, the three scenarios simulated are as follows: Case I assumes almost complete spatial homogeneity. Saturated hydraulic conductivity and topographic slope are averaged throughout the entire Illinois River Basin and are applied uniformly to each REW (denoted by the red dashed lines in Figure 5). Case II assumes that the hydraulic conductivity and topographic slope vary systematically (deterministically) from upstream to downstream (i.e., the random heterogeneity is removed from the observed natural heterogeneity). As shown in Figure 5, the increasing trends of hydraulic conductivity and topographic slope in the downstream direction are approximated (fitted) by an approximate exponential function (blue diamonds in Figure 5). All other soil properties are assumed to be uniform across all REWs. Case III is the default case; all landscape properties (e.g., saturated hydraulic conductivity, porosity, and the Brooks-Corey exponent) take on their original values as estimated from the STATSGO Database, and topographic slope values are directly derived from DEMs.

[22] In all three cases, however, the stream network properties, such as channel slope, length, and width, are allowed to remain variable and are maintained the same as those originally derived from the DEMs and from topographic maps and previous study reports since they do not impact the hillslope runoff generation processes. In summary, compared to case I (completely uniform), in case II we are trying to explore the effect of a systematic increasing trend of
hydraulic conductivity and topographic slope with increasing (upstream) drainage area, and in case III we are trying to explore the effects of both the apparent increasing trends and the natural variability (scatter) around the trends.

In addition to the rainfall intensity and duration and the adopted landscape properties (topographic slope and saturated hydraulic conductivity), runoff responses can be significantly affected by antecedent soil moisture and groundwater level. Note that the antecedent conditions are essentially a legacy of many past events (sometimes even more than a year) and arise from the history of interactions between the climatic forcing during these previous events and the landscape characteristics that constitute the catchment. In this work, the event simulations were initialized by using antecedent conditions (i.e., for water storages for all subregions in all REWs) that were generated by running the THREW model for the particular choice of landscape properties for the period 1 October 1996 to 30 September 2006 with the use of observed climatic data and the chosen landscape properties (as the case may be). From the model predictions, in each case, we then derive the mean monthly subregion water storages and use them as the corresponding antecedent conditions for the event simulations, as reported later, under wet conditions (using typical March values of antecedent conditions) and under dry conditions (using typical September values).

We first look at the impact of heterogeneity of landscape properties (systematic variation as well as random scatter) on the volume fractions of saturation excess overland flow and subsurface stormflow generated as a result of the applied rainfall pulse (i.e., 30 mm/h for 1 h). For each of the three cases, the simulation period is 30 days (720 h), which is deemed sufficiently long to capture the tail of subsurface runoff recession.

The accumulated volumes of runoff generation by the Dunne mechanism and subsurface stormflow resulting from the rainfall pulse are estimated from the model outputs for each of the 83 REWs. Figure 6 shows the fraction of total event runoff volume that is contributed to by subsurface stormflow under two different antecedent conditions: wet (March) and dry (September). Note here that the magnitudes of the fraction of subsurface runoff shown in Figure 6 are averages estimated over the entire upstream drainage areas for the particular location on the river network.

Figure 5. Spatial variability of landscape properties across the Illinois River Basin: (top) topographic slope and (bottom) saturated hydraulic conductivity. The points are mean values of hydraulic conductivity and topographic slope, averaged across each REW. In case I, landscape properties are assumed to be spatially uniform; in case II, landscape properties vary systematically with upstream drainage area; in case III, landscape properties are actual measured values.
The results show that for all three cases, subsurface runoff is more dominant under the dry condition than under the wet condition (as expected, considering the results previously presented in Figure 3 for conditions roughly equivalent to case 3). For example, the volume fraction of subsurface stormflow for the uniform case (case I) is about 30% under wet conditions and increases to about 60% under dry conditions. The change from wet to dry antecedent conditions results in a lowering of the water table and a reduction in soil moisture in the unsaturated zone. While subsurface stormflow and saturation excess overland flow are both dependent on water table depth, subsurface stormflow is somewhat more linearly related to water table position (a consequence of the routing model used), and in addition, it continues even after the precipitation ceases. On the other hand, saturation excess runoff is nonlinearly dependent on the water table position, governed by saturation thresholds, and therefore, the reduction of saturation excess runoff can be much stronger when changing from wet to dry conditions. This partially explains the contrasts between wet and dry conditions seen in Figure 6.

This may also be the reason for the differences between cases I, II, and III seen in Figure 6. It is not surprising that the percentage of subsurface runoff is constant at all catchment scales for case I because of the assumed uniformity of the landscape properties. For case II, the fraction of subsurface runoff volume is increasing slowly with catchment size when the drainage area is less than 1000 km$^2$, and the increase is steeper after 1000 km$^2$. This is due to the slow increase of saturated hydraulic conductivity and topographic slope up to about 1000 km$^2$ and a rapid increase afterward. The same trend holds, on average, for case III, although there is considerable scatter due to the organization of the tributaries and REWs and the heterogeneity of landscape properties. Comparing cases II and III, one can see that the inclusion of full heterogeneity leads to a decrease of subsurface runoff fraction, on average (and correspondingly increase of saturation excess runoff fraction). This must be due to the interaction of the spatial heterogeneity and process nonlinearity associated with saturation excess overland flow. In the present case the nonlinearity is in the relationship between water table depth and fraction...
of saturated areas, and the heterogeneity is associated with antecedent soil moisture and water table position. This is consistent with many previous studies that have explored the effects of spatial heterogeneity of soil hydraulic properties on surface runoff by both infiltration excess and saturation excess mechanisms [see Smith and Hebbert, 1983; Sivapalan et al., 1987].

3.2. Event-Scale Responses: Runoff Timing and IRFs

Here we utilize the concept of the instantaneous response function to describe the runoff timing responses at the catchment scale in each of the three cases. The IRF can be defined as the travel time distribution of a unit volume of runoff generated from a specified rainfall pulse [Robinson et al., 1995; Saco and Kumar, 2002] and is only applicable to the specific event of interest. The procedure to construct the IRFs from model simulations was as follows: (1) record the hourly discharges (720 h hydrographs) directly resulting from the chosen rainfall pulse, i.e., 1 h rainfall pulse at intensity of 30 mm/h; (2) subtract the (initial) base flow from the hourly discharges and thus obtain the direct runoff hydrograph; (3) calculate the total volumes of streamflow under these direct runoff hydrographs; (4) normalize the direct runoff hydrographs by the corresponding estimated total runoff volume. This way, the total area under each IRF will be equal to 1 (unity), as it should.

The IRF (similar to the unit hydrograph) reflects two types of transformations effected by the catchment (considered as a combination of hillslopes and the channel network): advection and dispersion. Advection contributes to an average time delay between the time of runoff generation and the time of arrival at the catchment outlet, which can be attributed, to first order, to the mean travel distance and mean flow velocity. Dispersion refers to the variance of travel time, which may include four distinct components: geomorphologic, kinematic, hydrodynamic, and hillslope. Geomorphologic dispersion arises mainly from the multiplicity of flow paths in the catchment and the associated variance of their travel distances. Kinematic dispersion arises from the spatial variability (systematic and random) of travel velocities within the channel network. Hydrodynamic dispersion arises from variability of flow velocities across the channel cross section, caused mostly by boundary friction effects, and is often deemed negligible compared to geomorphologic and kinematic dispersions [Rinaldo et al., 1991; Snell and Sivapalan, 1994; Saco and Kumar, 2002; Botter and Rinaldo, 2003]. Hillslope dispersion refers to a combination of all of the above effects but operating at the hillslope scale. Since the THREW model uses a hydraulics-based network routing model and explicitly models the routing of flows within the hillslopes (i.e., REWs), it is able to capture most of these runoff transformations to a reasonable level of approximation.

The IRFs derived in this way from model simulations for cases I, II, and III are presented in Figures 7 and 8 for wet and dry antecedent conditions, respectively. In each case we present the results for four different nested catchments having (upstream) drainage areas of 21.7, 433.9, 1680.4, and 2454.3 km². Note that the latter three scales correspond to the streamflow gauging stations at Savoy, Watts, and Tahlequah, respectively, as shown in Figure 1. The catchment of size 21.7 km² is a first-order catchment (again, see Figure 1 for its location, where its boundary is highlighted with a thick line).

The results presented in Figures 7 and 8 share several common features, which can be highlighted regardless of the nature of heterogeneity adopted. Under both wet and dry conditions, the shapes of the IRFs undergo consistent transformations with increasing catchment size: (1) total dispersion associated with the IRF increases, (2) the peak of the IRF decreases and the time to peak increases, and (3) there is also a corresponding shift in the skewness, with the IRF changing over from a left-skewed hydrograph (i.e., positive skew) to a right-skewed one (i.e., negative skew). These results are consistent with what has been described in the literature, i.e., the dominant contribution to total dispersion at small scales being hillslope dispersion (which tends to be left skewed) and at large scales being geomorphologic dispersion [Robinson et al., 1995; Botter and Rinaldo, 2003], which is related to the catchment’s area function (which tends to be right skewed). Thus, the left skewness of the IRF is a reflection of the dominance of hillslope dispersion, whereas the right skewness is a reflection of the dominance of geomorphologic dispersion.

There are also differences between the IRFs under wet conditions (Figure 7) and under dry conditions (Figure 8). The best way to characterize these differences is to compare the peaks of the IRF and the times to peak for these two conditions. On average, one can see that changing from wet to dry conditions leads to a reduction of the peak value and an increase in the time to peak, especially for the larger catchments. The exception is the response for the smallest catchment with an area of 21.7 km² (which will be discussed separately). The change from wet to dry conditions leads to an increase in the contribution of (slower) subsurface flow (see Figure 3). The discharge resulting from the imposed rainfall pulse is also smaller during the dry conditions, which means channel network velocities will be smaller (see later discussion for details). Thus, both hillslope and network travel times are longer, contributing to a higher degree of filtering and a stronger attenuation of the peaks and longer delays, or times to peak. Figure 8 also shows that the transition to right-skewed IRFs is especially stronger under dry conditions and for large catchments, perhaps because of the effect of stronger subsurface stormflow contributions.

However, there is another difference between the results presented in Figures 7 and 8, and this has to do with the rate at which the peak of the IRF decreases with increasing drainage area. These results are presented separately in Figure 9 on the basis of estimates along the river network corresponding to the locations of the 83 REWs, repeated for both wet (March) and dry (September) conditions. Three features can be noted in these results. First, all three cases almost coincide (which is not surprising based on what we saw in Figures 7 and 8). Second, the results presented on log-log paper exhibit a two-part power law relationship, with a scaling break at about 1000 km² (note this is roughly where the bottleneck occurs). Third, and most interestingly, the power law relationship is relatively flatter under wet conditions and steeper under dry conditions. This is more obvious for the 10–1000 km² range. Table 1 lists the scaling exponents of these power law regressions for the two scaling ranges. Under wet conditions, the power law exponent
is about $-0.2$ over the 10–1000 km$^2$ range and about $-1.06$ in the 1000–2500 km$^2$ range. Under dry conditions, these values change to $-0.37$ and $-1.32$, respectively. We note in passing that we have repeated the simulations for 10 and 60 mm/h rainfall pulses, and the last two features persist. How can we explain the last two features on the basis of our understanding of the runoff generation and routing processes, as included in the model?

[34] We will first look at the possible reasons for the break in the scaling exponent in Figure 9. This cannot be due to differences in the heterogeneity of dominant runoff generation mechanisms. The break happens in all of these cases. Channel routing is therefore the most likely reason for the break. Channel routing is controlled by channel velocity on the one hand and distribution of travel distances, i.e., network geomorphology, on the other.

[35] We first examine the model predicted channel velocities across the entire network. The THREW model simulates channel velocities across the entire network as a function of time for the entire simulation period. For the rectangular pulse that is imposed here, the velocity variations mimic the variation of flows themselves, exhibiting a rising limb and a falling limb, with velocities systematically also changing in the downstream direction as the flood wave propagates down the river network and experiences changes in the hydraulic geometry. To provide guidance toward the interpretation of the shape of the IRFs, we present only the maximum channel velocities achieved in the stream reaches associated with each REW during the events. These results are presented separately for the wet and dry conditions in Figure 10. They indicate relatively constant peak velocities across the network, yet peak velocities under wet conditions (due to higher discharges) are about 100% larger than those under dry conditions. This means that residence time in the channel network under dry conditions is almost twice as that under wet conditions.

[36] Clearly, the results presented in Figure 10 did not show a major change in the channel velocities past the 1000 km$^2$ threshold area. We can therefore guess that network geomorphology may be the most likely explanation for the breaks in slope seen in Figure 9 and for the exponents presented in Table 1. Indeed, the explanation may lie in the classic Hack’s [1957] law, which describes a geometric relationship between a catchment’s mainstream channel length $L$ and its drainage area $A$, as given by

\[
L = aA^\delta;
\]  

(1)

Figure 7. Instantaneous response functions (IRFs) resulting from a rectangular pulse of rainfall with an intensity of 30 mm/h and a duration of 1 h under a wet antecedent (March) condition. In case I, landscape properties are assumed to be spatially uniform; in case II, landscape properties vary systematically with upstream drainage area; in case III, landscape properties are actual measured values.

Figure 7. Instantaneous response functions (IRFs) resulting from a rectangular pulse of rainfall with an intensity of 30 mm/h and a duration of 1 h under a wet antecedent (March) condition. In case I, landscape properties are assumed to be spatially uniform; in case II, landscape properties vary systematically with upstream drainage area; in case III, landscape properties are actual measured values.
where $\alpha$ and $\beta$ are the coefficient and scaling exponent, respectively. Sivapalan et al. [2002] noted that estimates of the scaling exponent in many catchments hovered around 0.6 on average. In the Illinois River Basin, the equivalent power law relationship between $L$ and $A$ is seen to exhibit a break in slope, as shown in Figure 11, with the scaling exponent being about −0.6 over the range 10–1000 km$^2$ and about −1.6 over the range 1000–2500 km$^2$. This break can be easily attributed to the unique shape of the basin, containing a narrow bottleneck part dominating the catchment beyond a drainage area of about 1000 km$^2$, along with the meandering of the river channel within it. We can therefore infer that the break in the power law relationship between peak of the IRF and drainage area, shown in Figure 9, must be caused by the basin’s particular shape, as reflected in Figure 11.

[37] However, the explanation for the differences in the power law exponents for the wet and dry conditions is not so straightforward; in fact, it is somewhat counterintuitive, which makes it really interesting and potentially insightful. Having presented the shape of the Hack’s law relation in this catchment, in the ideal case of a fast catchment (fast hillslope response and fast channel network response), the power law exponents for the $q_p$ (peak of IRF) versus $A$ relationship must approach −0.6 and −1.6 (representing the shape of the catchment’s width or area function). On the other hand, when the hillslope response time increases, let’s say if it dominates network residence time (e.g., under very high channel velocities), then the catchment IRF will mainly reflect the hillslope IRF; hence, there will be little attenuation of the catchment IRF with increasing drainage area, and the power law exponents must approach 0.

**Table 1. Exponents of the Power Law Relationships Between the Drainage Area and the Peak of Catchment Instantaneous Response Function**

<table>
<thead>
<tr>
<th>Condition</th>
<th>Exponent 1</th>
<th>Exponent 2</th>
</tr>
</thead>
<tbody>
<tr>
<td>Wet condition</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Case I</td>
<td>−0.20</td>
<td>−1.06</td>
</tr>
<tr>
<td>Case II</td>
<td>−0.18</td>
<td>−0.95</td>
</tr>
<tr>
<td>Case III</td>
<td>−0.20</td>
<td>−0.86</td>
</tr>
<tr>
<td>Dry condition</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Case I</td>
<td>−0.37</td>
<td>−1.32</td>
</tr>
<tr>
<td>Case II</td>
<td>−0.37</td>
<td>−1.02</td>
</tr>
<tr>
<td>Case III</td>
<td>−0.34</td>
<td>−0.85</td>
</tr>
</tbody>
</table>
In other words, increasing dominance of hillslope residence time must lead to a flatter line, while decreasing hillslope residence time must lead to power law exponents that approach the Hack’s law exponents.

However, the results presented in Figure 9 contradict this initial, intuitive expectation. We have already seen that under wet conditions the (faster) saturation excess overland flow dominates the hillslope response. On the other hand, under dry conditions, the (slower) subsurface stormflow was shown to dominate. However, what we see is that the case where the dominant hillslope response is generally slower (i.e., dry conditions) is the one that is exhibiting the steeper power law relationship. Why is this? We believe that the answer lies in the fact that the magnitudes of the peaks and times to peak of the IRFs are governed by the residence time distribution of the faster component of the hillslope response, which is saturation excess overland flow. Subsurface stormflow, even when it is dominant volumetrically, is too slow to have a dominant control on the peak and time to peak of the IRFs. The key to the explanation is the nature of hillslope IRFs, including particularly the overland flow response. These are discussed in detail next.

In Figure 9 we present characteristic hillslope IRFs predicted by the model for different cases involving different combinations of soil hydraulic properties and antecedent wetness: (1) high versus low clay content (which impacts the soil hydraulic and retention properties) and (2) wet versus dry antecedent conditions. When clay content is high, one would expect subsurface stormflow to be less dominant and Dunne overland flow to be dominant. In addition, under wet conditions, one would expect the saturation area fraction to be larger than under dry conditions. Consider now the results presented in Figure 12 for the dry conditions. In this case, the IRF under high clay content shows a larger time delay than under low clay content, which can only be due to the time delay associated with overland flow routing across saturated areas. This is repeated under wet conditions, although to a slightly lesser degree. Consider now the results for soils with high clay content, under wet and dry conditions, and also for soils with low clay content. In all cases, it appears that the hillslope IRFs appear more dispersed whenever the saturation area fraction is larger. This is true even when subsurface stormflow dominates the runoff volume under dry conditions, i.e., the hillslope IRF from the REW with low clay content.
content appears more dispersed than that from the REW with high clay content. This is an interesting result and is reinforced by the results presented in Figure 13. This is also an apparent, visually misleading result in that the subsurface stormflow impacts mainly on the recession behavior, manifesting in a long tail of the recession limb, which is overwhelmed by the overland flow response at short time scales that are easily visible.

[40] Figure 13 presents simulated saturated area fractions for all 83 REWs under the three cases (cases I, II, and III) under wet and dry conditions. The results show that under wet conditions, saturation excess runoff is dominant in terms of volume and the saturation area fraction up to about 900 km$^2$ is of the order of 30%, meaning that the width of the overland flow region could be quite large. In cases II and III (which allow for heterogeneity) it then declines rapidly with increasing drainage area, a result of the increase in topographic slope and saturated hydraulic conductivity. This means that the residence time of (fast) overland flow can be quite significant because of the relatively long travel distances (across the saturation area), especially for catchments smaller than 900 km$^2$. On the other hand, under dry conditions, the saturated area fraction is less than 8% (at scales smaller than 900 km$^2$), meaning smaller overland flow lengths and tighter residence time distributions of the faster flow pathways. The consequence of this is that the supposedly faster catchment (i.e., the wet one) actually behaves like a slow one in terms of controlling the peak, while the supposedly slower catchment (i.e., the dry one) instead behaves like a fast one. Taken together, however, these results and those presented in Figure 12 give a plausible explanation for the magnitudes of the power law exponents seen in the results of Figure 9. Note that we repeated these simulations for rainfall pulses of 10 and 60 mm/h, and while the results changed substantially in detail, the general conclusions we have made here remain valid.

[41] From the catchment IRFs, we also extracted the magnitudes of the times to peak (h). Figure 14 shows the corresponding scaling behavior of the time to peak for both wet (March) and dry (September) antecedent conditions. First, one could see that in all cases the time to peak increases with drainage area, and this increase is faster over the range of 1000–2500 km$^2$ than over the range of 10–1000 km$^2$. As listed in Table 2, under wet conditions the power law exponent is about 0.2 over the 10–1000 km$^2$ range and about 1.1 over the 1000–2500 km$^2$ range. Under dry conditions, these values change to 0.3 and 1.4, respectively. Once again, the change in steepness of this scaling behavior roughly mirrors the mainstream length versus area relationship presented in Figure 11. For catchments larger than about 100 km$^2$, the time to peak is longer under dry conditions than under wet conditions. This is perfectly understandable, considering that channel velocities are about 50% smaller (see Figure 10), hillslope contributions are more dominated by subsurface flow (see Figure 3), and saturation areas are smaller (Figure 13), meaning that overland flow delay is minimal. Therefore, time to peak is controlled primarily by network routing. However, for catchments smaller than about 100 km$^2$, the time to peak appears to be actually longer under the wet condition than for the dry condition, despite the fact that surface runoff...
Figure 12. Hillslope IRF for typical first-order REWs, showing the effect of soil hydraulic properties and antecedent condition on the shape of hillslope IRF. With respect to soil properties, Dunne runoff generation is dominant in REWs with high clay content, as manifested by the higher peak. With respect to antecedent condition, a dry condition implies lower water table and smaller saturation area and less Dunne runoff generation, which contributes to a faster decrease of hillslope IRF compared to the wet condition.

Figure 13. Scaling behavior of the fraction of saturated area on hillslopes within each of the REWs, resulting from a rectangular pulse of rainfall with an intensity of 30 mm/h and a duration of 1 h. Values of the saturation area fraction are averages over the entire simulation period (720 h). The horizontal axis represents the total upstream drainage area of each REW, which is used to denote the relative location of each REW. In case I, landscape properties are assumed to be spatially uniform; in case II, landscape properties vary systematically with upstream drainage area; in case III, landscape properties are actual measured values.
dominates volumetrically and channel velocities are faster. This confirms our supposition that at small scales the timing of hillslope overland flow controls the scaling of the IRF peaks. As discussed, this is due to the larger saturation area fractions that occur under wet conditions. Overall, these results seem to indicate that in this case, the time to peak is governed by the residence time in the river network at scales larger than 100 km$^2$, while at smaller scales it is governed by the residence time on the hillslopes associated with overland flow. These results strongly indicate that partial area runoff generation is, indeed, an important consideration for runoff routing purposes in situations where saturation excess overland flow is a significant component of runoff generation.

One can now return to Figures 7 and 8 and discuss some obvious differences between the IRFs derived for cases I, II, and III caused by the effects of heterogeneity. As highlighted in Table 3, increasing heterogeneity gives rise to a sharper reduction of the fraction of subsurface runoff in the first-order catchment, especially under dry conditions. This gives rise to relatively smaller values of hillslope dispersion, which is reflected in the shapes of the IRFs presented in Figures 7 and 8 (especially the latter, under dry conditions). The differences between the IRFs for all cases, however, tend to be reduced or eliminated with increasing catchment size.

We can now propose an explanation for the anomalous result associated with the smallest 21.7 km$^2$ catchment. In this catchment, we can see it is the wet condition.

Table 2. Exponents of the Power Law Relationships Between the Drainage Area and the Time to Peak

<table>
<thead>
<tr>
<th>Condition</th>
<th>Exponent 1</th>
<th>Exponent 2</th>
</tr>
</thead>
<tbody>
<tr>
<td>Wet condition</td>
<td>0.19</td>
<td>1.10</td>
</tr>
<tr>
<td>Case I</td>
<td>0.20</td>
<td>1.15</td>
</tr>
<tr>
<td>Case II</td>
<td>0.21</td>
<td>1.16</td>
</tr>
<tr>
<td>Dry condition</td>
<td>0.32</td>
<td>1.20</td>
</tr>
<tr>
<td>Case I</td>
<td>0.32</td>
<td>1.39</td>
</tr>
<tr>
<td>Case II</td>
<td>0.35</td>
<td>1.41</td>
</tr>
</tbody>
</table>

Figure 14. Scaling behavior of the time to peak of the instantaneous response function resulting from a rectangular pulse of rainfall with an intensity of 30 mm/h and a duration of 1 h. The horizontal axis represents the total upstream drainage area of each REW, which is used to denote the relative location of each REW. In case I, landscape properties are assumed to be spatially uniform; in case II, landscape properties vary systematically with upstream drainage area; in case III, landscape properties are actual measured values.
that causes the wider (more dispersed) IRF compared to the narrower one shown for dry conditions. The explanation is now clearly evident (see Figure 12). Although subsurface stormflow dominates, the saturation area fraction is smaller under dry conditions, and hence, it is the overland flow part of the hillslope IRF that controls the peak of the catchment IRF and the width (variance) of the residence time distribution. One can see that with respect to the anomalous results (1) in the shape of the IRF curves for the 21.7 km$^2$ catchment presented in Figures 7 and 8 and (2) in the steepness of the power law relationship between peak of the IRF and drainage area presented in Figure 9, the same consistent explanation applies, namely, partial area runoff generation and the impact on the residence time distribution of the fast hillslope responses.

### 4. Summary and Conclusions

[44] In this paper we have investigated the effects of the spatial heterogeneity of runoff generation processes on event runoff (timing) responses, using a calibrated model implemented in the Illinois River Basin in Oklahoma. This study has been motivated by previous distributed model predictions [Li et al., 2011] that there is a significant spatial trend in the dominant runoff generation mechanisms, with headwater basins being dominated by saturation excess overland flow, whereas subcatchments close to the catchment outlet are dominated by subsurface flow. This spatial heterogeneity and associated trend were connected to the corresponding spatial heterogeneity and systematic spatial trends in the underlying landscape properties, in particular, saturated hydraulic conductivity and topographic slope, with both increasing (on average) in the downstream direction toward the catchment outlet.

[45] In order to investigate the effects of spatial heterogeneity we carried out numerical experiments where we considered three different cases, starting with a catchment with uniform landscape (hillslope) properties, followed by a catchment where the soil hydraulic conductivity and topographic slope were assumed to vary systematically in the downstream direction as a function of upstream drainage area. The final case included the full heterogeneity of these two properties (identical to estimated values for the actual catchment). The numerical experiments were repeated for both wet and dry antecedent conditions. In each case we simulated the response of the catchment to a single rectangular pulse of rainfall with an intensity of 30 mm/h and a duration of 1 h. We estimated from model simulations the rates of runoff generation by both saturation excess overland flow and subsurface stormflow (and their relative fractions) and instantaneous response functions (IRFs) that captured the nature of the runoff timing response.

[46] Three major conclusions can be reached on the basis of the results presented in this paper. First, increasing (ancendent) wetness of a catchment, on average, leads to a higher fraction of the total runoff being contributed to by saturation excess overland flow. This has to do with the fact that the rate of increase of saturation excess runoff with the rise of the water table is quite nonlinear. The effects of spatial heterogeneity of runoff generation on the runoff timing appear to be relatively small, especially at large scales, compared to the effects of network geomorphology and antecedent conditions.

[47] The estimation of event-scale runoff routing responses has confirmed results of many past studies about process controls on the shape of the IRFs and how the shape changes with the size of the catchment. The IRFs of small catchments are governed to a large extent by the hillslope IRF, whereas at large scales they are controlled by network geomorphology (or geomorphological dispersion). The change of shape of the catchment IRFs with increasing drainage area can be quantified by their peaks and times to peak and plotting them against drainage area on a log-log plot. In both cases the results revealed a two-part power law relationship, with a break in the neighborhood of 1000 km$^2$. This scaling break is easily explained by means of the mainstream length versus drainage area relationship (i.e., Hack’s law), which also exhibits such a break in scaling in the study catchment.

[48] However, the power law relationship between the peak of the IRF and drainage area turned out to be actually flatter under wet conditions than under dry conditions, which is counterintuitive. We attributed this result to the dynamics of contributing areas that produce saturation excess overland flow. A catchment otherwise dominated by saturation excess overland flow (in volumetric terms) under wet conditions actually behaves like a slow catchment, while a catchment dominated by subsurface stormflow under dry conditions actually behaves like a fast catchment simply because of differences in the extent of saturation areas, as far as determining the peak and time to peak is concerned. The heterogeneity of runoff generation mechanisms did assist in the transfer of control from hillslope dominance to network dominance because the saturated area fraction actually decreased with increasing drainage area. These insights and conclusions have been possible only because in this case we had used a distributed model that is able to incorporate these features

---

**Table 3. Percentage of Subsurface Runoff in the Total Volume of Runoff Generation Resulting From a Pulse of Rainfall**

<table>
<thead>
<tr>
<th>Condition</th>
<th>A1 (21.7 km$^2$)</th>
<th>A2 (433.9 km$^2$)</th>
<th>A3 (1680.4 km$^2$)</th>
<th>A4 (2454.3 km$^2$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Wet</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Case I</td>
<td>Case II</td>
<td>Case III</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Case II</td>
<td>Case III</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Dry</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

*The rainfall pulse has a duration of 1 h and an intensity of 30 mm/h. The value of the percentage of subsurface runoff is averaged through the corresponding upstream drainage area.*
explicitly, including particularly partial area runoff generation and network routing.

This is admittedly a site-specific modeling study, and some of the heterogeneity observed here may not exist in other catchments. In fact, a neighboring catchment, the Blue River Basin in Oklahoma, experiences much more random heterogeneity of landscape properties than in the Illinois River Basin [Li et al., 2011]. Nevertheless, there are many more examples of basins where runoff is generated by a combination of overland flow and subsurface stormflow, and the insights gained from this theoretical study may still be useful or transferable to other catchments, if only to serve as hypotheses for more detailed field studies. The present study could present a useful model to interpret results of detailed field studies, such as tracer studies, that are aimed at measuring and interpreting distributions of residence times and how they scale with catchment size in different climatic and geologic settings [McGlynn et al., 2003; McGuire et al., 2005; Hrachowitz et al., 2009].

Finally, the timing response of runoff has an immediate connection to the estimation of flood peaks during storm events of different durations, to their extension to flood frequency, and then to how the flood frequency curves scale with catchment size. There have been several research efforts in the past that have tried to understand the climatic and landscape controls on the scaling behavior of flood peaks and the flood frequency curve. These studies have revealed that the scaling behavior of flood peaks and the flood frequency curves can be attributed to the interaction, in the time domain, between precipitation time scales and characteristic time scales of catchment response, especially time scales associated with processes that govern the flood peaks [Robinson and Sivapalan, 1997; Blöschl and Sivapalan, 1997; Menabde and Sivapalan, 2001; Sivapalan et al., 2005]. The results presented in this paper will have important ramifications for the interpretation of observed flood frequency scaling behavior through such phenomenological studies, including efforts to improve the scientific basis of flood frequency regionalization [Gupta et al., 1994; Blöschl and Sivapalan, 1997]. This is an area where immediate extension of the work presented in this paper is highly relevant.

Figure A1. Channel routing framework in the THREW model.

Appendix A: Routing Algorithms in the THREW Model

Runoff generated by both surface runoff (overland flow) and subsurface runoff is included in the THREW model. The model uses lumped algorithms at REW scale to handle the routing of these flows over or through the hillslopes. Once surface runoff (by the Horton and/or Dunne mechanism) is generated, it is routed as overland flow across the t zone (subnetwork zone, the fraction of a REW occupied by ephemeral rills and gullies). The averaged (REW scale) overland flow velocity is given by a REW-scale Manning’s equation:

\[
v' = \frac{1}{n'_m} \left[ \frac{y'}{2/3} \sin(\gamma') \right]^{1/2},
\]

where \( n'_m \) is the effective Manning’s roughness at REW scale, \( y' \) is water depth within the t zone, and \( \gamma' \) is topographic slope. Subsurface stormflow timing is modeled through the use of lumped reservoir routing [Lee et al., 2007], applicable at the REW scale:

\[
q_{sub} = \alpha K_s \frac{y_s}{Z} \left( \frac{y_s}{Z} \right)^{\beta},
\]

where \( q_{sub} \) is the discharge from subsurface water storage to the channel (m/s), \( K_s \) is the averaged hydraulic conductivity (m/s), \( S \) is topographic slope, \( y_s \) is the average depth from water table to the bedrock (m), \( Z \) is total soil depth (m), and \( \alpha \) and \( \beta \) are coefficients accounting for the effects of the topography and spatial heterogeneity within a REW.

The overland flow from the t zone from each REW will be contributing to the channel network, which will then be routed downstream. The network routing model involves the coupling of the general conservation laws for both mass and momentum [Reggiani et al., 2001]. Generally, a main channel segment, or reach, has two upstream reaches and one downstream reach. For example, in Figure A1, reach 3 has reaches 12 and 13 as its upstream reaches and reach 1 as its downstream reach.
The network-scale water balance equation is given by

\[ \frac{dS_i}{dt} = Q_h + \sum Q_n - Q_{out}, \quad (A3) \]

where \( S_i \) is water storage at local reach \( i \) (equal to \( A_iL_i \), where \( A_i \) is the cross-sectional area of local reach and \( L_i \) is the reach length), \( Q_h \) is the flow that enters the channel network from the hillslope, and \( Q_n \) is the inflow from upstream nodes (equal to \( v_iA_i \), where \( v_i \) is the velocity at upstream reach \( k \) and \( A_i \) is the cross-sectional area of upstream \( k \)). \( Q_{out} \) is the outflow at reach \( i \) (equal to \( v_iA_i \), where \( v_i \) is the velocity at local reach \( i \)). The channel cross-sectional area \( A_i \) at any time step is estimated by dividing the water storage \( S_i \) by the channel reach length \( L_i \), while the velocity \( v_i \) is calculated using the momentum balance equation (equation (A4)).

In this paper, for simplicity, we simplify the full network-scale momentum balance equation by adopting the kinematic wave approximation. In this case, the averaged flow velocity in each reach is then given by

\[ v = \frac{1}{n_i}' \left[ \frac{R_i^{1/3}}{P_i} \right] \left \{ m_i \sin(\gamma_i) \pm \sum \left \{ \frac{1}{4} y_i (m_i + m_j) \cos(\theta_i) \right \} \right \} \frac{1}{2} y_i m_i \quad (A4) \]

where \( n_i \) is the effective Manning’s roughness of the current reach, \( R_i \) is the averaged hydraulic radius, \( P_i \) is the averaged wetted perimeter, \( L_i \) is the length of the current reach, \( m_i \) is the cross-sectional area of the current reach, \( \gamma_i \) is the average channel slope, \( y_i \) is the averaged water depth in the channel, \( \gamma_i \) is the topographic slope, \( m_i \) is the cross-sectional area of the neighboring upstream or downstream reach, and \( \theta_i \) is the angle between the current reach and its neighboring reach. The sign before \( \sum \) is positive for upstream reaches and negative for downstream reaches. The term \( \frac{1}{2} y_i m_i \) exists if and only if the current reach is at the outlet of the whole basin.

The wetted perimeter and cross-sectional area are both functions of channel geometry and discharge rate. In this version of the THREW model, the channel cross section is assumed to be rectangular, and the width of each reach is given by the following hydraulic geometry relationship:

\[ w_i = w_0 \left( \frac{A_i}{A_0} \right)^{0.5}, \quad (A5) \]

where \( w_0 \) is a known channel width corresponding to a drainage area \( A_0 \) and \( A_0 \) is the drainage area corresponding to the channel width to be estimated, \( w_0 \).

Acknowledgments. We are grateful to the DMIP2 team at the National Weather Service, NOAA, for providing most of the data used in this study and for providing financial support to attend DMIP2 meetings. The research was partly funded by financial support provided by the University of Illinois through the AESIS project (Barbara Minser, PI) and by the National Science Foundation (NSF ERFI-0835982, Ximing Cai, PI). This support is gratefully acknowledged. We dedicate this paper to Steve Burgess for his significant and diverse contributions to hydrology over an illustrious 40 year career.

References


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