Available online at www.sciencedirect.com
SCIENCE DIRECT®
ELSEVIER
Journal of Hydrology 285 (2004) 3–18
Journal of Hydrology
www.elsevier.com/locate/jhydrol

Virtual experiments: a new approach for improving process conceptualization in hillslope hydrology

Markus Weiler*, Jeff McDonnell

Department of Forest Engineering, Oregon State University, Corvallis, OR 97331-5706, USA

Received 23 December 2002; accepted 11 July 2003

Abstract

We present an approach for process conceptualization in hillslope hydrology. We develop and implement a series of virtual experiments, whereby the interaction between water flow pathways, source and mixing at the hillslope scale is examined within a virtual experiment framework. We define these virtual experiments as 'numerical experiments with a model driven by collective field intelligence'. The virtual experiments explore the first-order controls in hillslope hydrology, where the experimentalist and modeler work together to cooperatively develop and analyze the results. Our hillslope model for the virtual experiments (HillVi) in this paper is based on conceptualizing the water balance within the saturated and unsaturated zone in relation to soil physical properties in a spatially explicit manner at the hillslope scale. We argue that a virtual experiment model needs to be able to capture all major controls on subsurface flow processes that the experimentalist might deem important, while at the same time being simple with few 'tunable parameters'. This combination makes the approach, and the dialog between experimentalist and modeler, a useful hypothesis testing tool. HillVi simulates mass flux for different initial conditions under the same flow conditions. We analyze our results in terms of an artificial line source and isotopic hydrograph separation of water and subsurface flow. Our results for this first set of virtual experiments showed how drainable porosity and soil depth variability exert a first order control on flow and transport at the hillslope scale. We found that high drainable porosity soils resulted in a restricted water table rise, resulting in more pronounced channeling of lateral subsurface flow along the soil–bedrock interface. This in turn resulted in a more anastomosing network of tracer movement across the slope. The virtual isotope hydrograph separation showed higher proportions of event water with increasing drainable porosity. When combined with previous experimental findings and conceptualizations, virtual experiments can be an effective way to isolate certain controls and examine their influence over a range of rainfall and antecedent wetness conditions.

Keywords: Subsurface flow; Hillslope hydrology; Virtual experiments; Animation; Hydrograph separation; Residence time; Tracer

1. Introduction

The rate of progress in hillslope hydrology has slowed considerably since the formative work during the International Hydrological Decade (IHD) (Whipkey, 1965; Weyman, 1973; Dunne and Black, 1970a; Dunne and Black, 1970b; and others). Kirkby (1978) and Anderson and Burt (1990) still define the 'benchmarks' in the field, despite new approaches since then, like isotope tracing (Kendall and McDonnell, 1998), new ideas of hyporheic exchange (Bencala, 2000) and

* Corresponding author. Tel.: +1-541-737-8719; fax: +1-541-737-4316.
E-mail address: markus@2hydros.de (M. Weiler).

0022-1694/$ - see front matter © 2003 Elsevier B.V. All rights reserved.
doi:10.1016/S0022-1694(03)00271-3

pressure waves (Torres et al., 1998), new instrumentation like TDR (Grayson and Western, 2001), and electromagnetic induction (Sherlock and McDonnel, 2003). Why is this? We argue that rather than a systematic examination of the first order controls (defined as the main and essential process constraints on water and solute flux) on hillslope hydrology, the field has focused on documentation of the idiosyncrasies of new hillslope environments—examining interesting new effects but without a purposeful hillslope intercomparison to draw out common process behavior (Jones and Swanson, 2001). Even when major hillslope experiments and excavations are undertaken (e.g. McDonnell et al., 1996), the ‘transference value’ to neighboring hillslopes is minimal as a variety of properties change. This suggests that hillslopes, as fundamental landscape features, may be smaller than the size of the representative elementary area (REA) of a catchment. Even if we do understand how the variations of hillslope characteristics scale (Blöschl, 2001), we still have serious difficulty knowing the transfer behavior of effects like bedrock topography, soil properties, and topographic geometries. Some have argued, rather bleakly, that each hillslope is therefore unique (Beven, 2001b). Thus, while there have been hundreds of field experiments in hillslope hydrology since the end of the IHD that have explored, where water goes when it rains, what flow path the water has taken, and how long that water has resided in the hillslope, we appear to be no further towards a common conceptualization of hillslope hydrology.

What is the way forward? The recent American Geophysical Union Chapman Conference on Hillslope Hydrology (http://www.agu.org/meetings/cc01ecall.html) examined current and future prospects in experimental and modeling studies of hillslope hydrology. While many useful individual experimental hillslope investigations have been completed recently (Freer et al., 2002; Montgomery and Gran, 2001; Bishop et al., 1998), we still lack a quantitative framework in which to test and compare first order controls on water and solute mass flux at this scale. Highly complex physically-based Finite-Element-Models have dominated the hillslope modeling literature for the past 20 years (Faeh et al., 1997; Weiler et al., 1998; Calver and Cammeraat, 1993; Sloan and Moore, 1984). Beven and Freer (2001), Seibert (2001), and others have criticized them for their problems of parameter identifiability, uncertainty, and difficulty in objective model testing. We, as a hillslope hydrology community, seem to be in a position, where we often have hillslope models that ‘work’ but for the wrong process reasons (Seibert and McDonnell, 2002).

The modelers alone are not to blame, as a plethora of experimental studies have produced little generalizable potential or definition of appropriate state variables in different environments. Even worse perhaps is that experimentalists have not yet articulated what are the minimal sets of measurements necessary to characterize even a single hillslope! Despite numerous calls for the past two decades (Dunne, 1983; Dunne, 1998), the dialog between experimentalist and modeler still appears out of reach. Little has yet been done to merge experimental and modeling approaches. As Seibert and McDonnell (2002) note, the experimentalist often proposes a perceptual model based on his or her complex and qualitative field observations and experiences, but the modeler usually does not incorporate the experimentalist’s knowledge into the model structure, let alone the model calibration or validation.

This paper attempts to improve conceptualization in subsurface hillslope hydrology by quantifying the interaction between water flow pathways, source, and mixing at the hillslope scale within a virtual experiment framework. Here we define virtual experiments as ‘numerical experiments with a model driven by collective field intelligence’. We argue that these virtual experiments are essentially different to traditional numerical experiments since the intent is to explore first-order controls in hillslope hydrology, where the experimentalist and modeler work together to develop and analyze the results collectively. Thus, we are not primarily fitting or calibrating a model to field experimental results as is typically done in coupled field and modeling hillslope hydrology studies (Binley et al., 1989; Faeh et al., 1997; Bronstert and Plate, 1997; Sloan and Moore, 1984). In addition to the traditional scalar output, visualization is a key interpretive part of the approach. Our work is motivated by frustrations that we have had personally in experiments at various hillslopes, where first order effects often seem difficult to separate from second and third order effects. The work is further motivated by our general philosophy that hillslope models should be simple, with few 'tunable parameters', and might serve ultimately as useful hypothesis testing tools. Here we show how one can test a number of hypotheses within a virtual experimental framework to inform a new organizational structure for hillslope hydrology. Our ideas are motivated by recent works of Alila and Beckers (2001), Seibert and McDonnell (2002), and Troch et al. (2002).

2. A brief review of current hillslope concepts

Our perception of hillslope hydrology and rapid subsurface flowpaths has evolved greatly over the past two decades (for a review of one catchment see McGlynn et al. (2002)). Early studies focused on how rainfall and snowmelt (event water) moved rapidly into the channel during episodes (e.g. Mosley, 1979). With the advent of conservative isotopic tracer studies, there is now consensus that pre-event water stored in the catchment before the episode is the dominant contributor to stormflow in the stream—averaging 75% world-wide (Buttle, 1994). Another consensus is that preferential flow is a ubiquitous phenomenon in natural soils, particularly in steep catchments (Faeh et al., 1997; Weiler and Naef, 2003; Germann, 1990). Today, research focuses largely on mechanisms to explain rapid movement and/or effusion of old water into stream channels. The main process conceptualizations include:

2.1. Transmissivity feedback

In glaciated till-mantled terrain (e.g. Sweden, Canada) or in more temperate or sub-tropical areas, where saprolite is found, the process known as transmissivity feedback (Rodhe, 1987) may dominate the generation of rapid subsurface stormflow. In these instances, vertical recharge into the till or saprolite must first occur before water tables rise into the more transmissive mineral soil. Once the water table rises into this zone, lateral flow begins—and the timing of well response into the mineral soil has been observed by many to coincide with rapid streamflow response (e.g. Kendall et al., 1999; Seibert et al., 2002).

2.2. Lateral flow at the soil bedrock interface

Another commonly observed form of rapid subsurface stormflow production is by way of lateral flow at the soil–bedrock interface, as described by McDonnell (1990). Several recent studies have observed this process in Canada (Peters et al., 1995), Japan (Tani, 1997), USA (Freer et al., 1997), and New Zealand (McDonnell et al., 1998). In steep terrain with relatively thin soil cover, water moves to depth rapidly and perches at the soil–bedrock interface (McIntosh et al., 1999). Since drainable porosity often declines rapidly with depth, the addition of only a small amount of new water (rainfall or snowmelt) is required to produce saturation at the soil–bedrock or soil–impeding layer interface. Rapid lateral flow occurs at the permeability interface through the transient saturated zone. Once rainfall inputs cease, there is a rapid dissipation of positive pore pressures and the system reverts back to a slow drainage of unsaturated soil matrix. Recent work by McDonnell (1997), McDonnell et al. (1996) and Freer et al. (1997) has shown that by mapping the impeding layer surface, one may be able to model the spatial pattern of transient water table development and thus the location of the mobile water flow path (Burns et al., 1998).

2.3. Organic layer interflow

A less widely cited example of rapid subsurface flow production is rapid lateral flow through the litter layer, sometimes called the 'thatched roof effect' (Ward and Robinson, 2000) or pseudo-overland flow (as reported by McDonnell et al., 1991). Recent studies by Brown et al. (1999) and Buttle and Turcotte (1999) have shown using chemical end member mixing and isotopic tracing approaches that this rapid lateral litter layer flow (perched on the mineral soil surface) may be a dominant mechanism in upland forested catchments during summer rainstorms. This is a combination of the high short-term rainfall intensities and water repellency that may develop at these sites during dry periods, especially in burned areas.

2.4. Pressure wave translatory flow

While pressure wave translatory flow was proposed back in the early 1960s by Hewlett and Hibbert (1967) as part of the Variable Source Area concept, recent papers have rejuvenated interest in this area (Rasmussen et al., 2000). Torres et al. (1998) found that 'a pressure head signal advanced through the soil profile on average 15 times greater than the estimated water and wetting front velocities. Hence initial pressure head response appears to be driven by the passage of a pressure wave rather than the advective arrival of new water'. This pressure water was thought to be the cause of the rapid effusion of old stored water from the deeper sandstone groundwater on their hillslopes. While these processes remain poorly understood in many different environments, research to date does show clearly that rapid travel times of fluid pressure head or water content through the unsaturated zone could be interpreted, mistakenly, as preferential or macropore flow (Smith and Hebbert, 1983).

3. On the need for virtual experiments

While general perceptions of lateral delivery mechanisms at the hillslope scale exist, many questions remain that transcend these different mechanisms: (1) how to explain the often observed paradox of high isotope-based pre-event water contribution to the hillslope base in relation to fast 'applied' artificial tracer breakthrough (McDonnell et al., 1998; Anderson et al., 1997) or base cation flushing (Burns et al., 1998; Anderson and Dietrich, 2001); (2) how to derive a mean age of subsurface flow from hillslopes in light of highly skewed age distribution of mobile waters (Vitvar et al., 2002); (3) how to separate first order controls on water fate and transport from other confounding secondary effects (soil depth, vertical porosity and conductivity distributions, slope geometry, etc); (4) how to deal with process equifinality, where multiple 'process responses' may produce similar hillslope hydrographs; and (5) how to explain the very different flow and timing at the hillslope scale based on differences on rainfall amount, intensity and antecedent wetness (Woods and Rowe, 1996; Woods and Sivapalan, 1997)?

Experimentalists lack tools to answer these questions in the field. Modelers are often oblivious to these questions or know to use only flow rate from the base of the slope to test the 'reality' of their simulations. Here, we use virtual experiments (as numerical experiments with a model driven by collective field intelligence) to address these hillslope paradoxes and to clarify, simplify, and classify. These virtual experiments not only provide information on runoff response but also concomitant information on water level response in the hillslope, event water contribution in runoff, residence time distribution, breakthrough curves of artificial tracer experiments, spatial distribution of artificial tracer in hillslope, etc. These additional metrics of hillslope behavior are what hillslope studies since the IHD have shown to be the fundamental descriptors of hillslope response (Bonell, 1998). Thus, virtual experiments can help the experimentalist replicate similar information he or she observes in the field and explore the consequences of different rainfall intensity, duration and frequencies, along with different slope angles, soil properties (hydraulic conductivities, storage) and surface/bedrock topographies. This gives the experimentalist the ability to respond critically to the simulation results and challenge the modeler to develop tools and numerical concepts that are in the line with the experimentalist's view. Visualization and animation of the output provide further ability to see patterns and processes in new and different ways (Blöschl and Grayson, 2001). Finally, virtual experiments provide the possibility to simulate many events that would be prohibitive in the field.

4. The model approach for the virtual experiment

Our hillslope model for the virtual experiments (HillVi) is based on conceptualizing the water balance within the saturated and unsaturated zone in relation to soil physical properties (Seibert and McDonnell, 2002)—this time implemented in a spatially explicit manner at the hillslope scale. Whether using this model or any other, the virtual experiment needs to be able to capture all major controls on subsurface flow processes that the experimentalist might deem important, while at the same time being simple enough and easy to understand to the field-oriented experimentalist. It must also represent the unsaturated and saturated zone explicitly and with tight coupling between them. This is because hillslope studies since the early 1960s have shown that lateral subsurface flow in hillslopes is triggered often by a perched water table within the soil or the soil bedrock interface (Dunne, 1978; Bonell, 1998; McGlynn et al., 2002), converting the unsaturated zone to a saturated zone. We argue that this might be the most common mechanism for delivery of water from slopes into valley bottom and riparian areas. Thus, while not representative of every condition, this most basic 'process starting point' represents a hydrologist's common view of water delivery on hillslopes.

The Dupuit-Forchheimer assumption is a relative simple but useful representation of the main physical process of this perched water table development and lateral flow delivery (Freeze and Cherry, 1979):

$$q(t) = T(t)\beta w \tag{1}$$

where T is the transmissivity, β is the water table slope, and w is the width of flow. We use an explicit grid cell by grid cell approach with a power law transmissivity function to route transient saturated subsurface flow laterally downslope (Wigmosta and Lettenmaier, 1999). In contrast to most of the existing models defining the flow direction a priori by the surface topography, HillVi recalculates the flow direction and thus the partitioning of outflow from each grid cell for each time step based on the local water table gradient. This step is necessary to simulate hillslopes with a spatially variable soil depth as local depressions in the bedrock topography would constrain the outflow in these grid cells to zero (see importance of this variability in recent papers by Woods and Rowe (1996) and Freer et al. (1997)). Mass transport within the saturated zone was implemented by only advective transport from one to another grid cell. However, due to the dispersive nature of the partitioning of outflow from each grid cell, a 'mechanical (convective) dispersion' (Bear, 1972) at the grid cell scale is simulated without defining a dispersion coefficient explicitly. Thus, we neglected molecular diffusion as the high velocities observed at the hillslope during runoff generation are generally dominated by advective transport.

The flow velocity of the particles v is related to the specific discharge or Darcy velocity q (units in LT^{-1} after dividing by area) by:

$$q = n_{\mathrm{eff}} v \tag{2}$$

where n_{eff} has the meaning of an effective porosity with respect to flow (Bear, 1972). The concept of effective porosity is common simplification describing the porosity available for fluid flow and is often assumed to be equal to the porosity n (Bear, 1972). However, especially tracer experiments in soils (e.g. Flühler et al., 1996) showed a lower effective porosity than the total porosity or even a partitioning into a mobile and immobile fraction.

HillVi simulates transport in several domains. Thus, the model can calculate mass movement for different input conditions (line source, constant input concentration, or a pulse input over the whole area) under the same flow simulation. In addition to the simulated flow, transport, and the water balance calculation in the saturated zone, we implemented a simple conceptual interaction between the saturated and unsaturated zones following Seibert and McDonnell (2002) in order to account for the mass exchange under a changing water table. The conceptual description of this interaction is based mainly on the assumption that the proportion of the drainable porosity n_d (analogous to specific yield and effective porosity in a traditional groundwater hydrology sense) to the total porosity n is defined mainly by the soil water characteristic curve.

The mass exchange between the saturated and unsaturated zone under a changing water table is sketched in Fig. 1. The water volume that is available for fluid flow in the saturated zone V_{sat} is given as:

$$V_{\mathrm{sat}} = WAn_{\mathrm{eff}} \tag{3}$$

with W the water table depth and A the area of the grid cell. In addition, the water volume that is available for fluid flow in the unsaturated zone V_{unsat} is given as:

$$V_{\mathrm{unsat}} = (D - W)A(n_{\mathrm{eff}} - n_d) \tag{4}$$

with the total soil depth D. If the water table is falling, mass is transferred (Δm) from the saturated to the unsaturated zone depending on the concentration in the saturated zone ($m_{\mathrm{sat}}/V_{\mathrm{sat}}$) and the change in

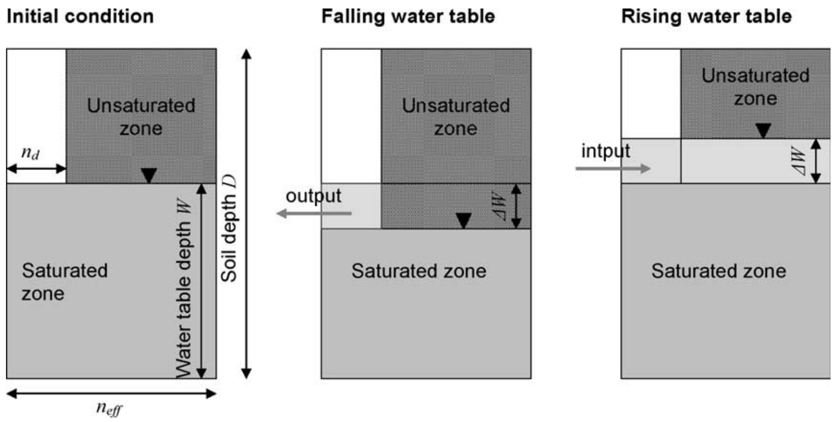

Fig. 1. Conceptualization of the mass exchange between the saturated and unsaturated zone, showing the parameter for the initial condition and the resulting changes for a falling water table (output > input) and a rising water table (output < input).

the water table ΔW.

$$\Delta m = \frac{m_{\text{sat}}}{V_{\text{sat}}} \Delta WA(n_{\text{eff}} - n_{\text{d}}) \quad (5)$$

If the water table is rising, mass is transferred from the unsaturated to the saturated zone depending on the concentration in the unsaturated zone ($m_{\text{unsat}}/V_{\text{unsat}}$) and the change in the water table.

$$\Delta m = \frac{m_{\text{unsat}}}{V_{\text{unsat}}} \Delta WA(n_{\text{eff}} - n_{\text{d}}) \quad (6)$$

The concentrations in the saturated and unsaturated zone are calculated under the assumption of complete mixing in each zone.

In order to simplify the model structure further for event-based virtual experiments in this paper, we did not explicitly account for flow in the unsaturated zone and fluxes and storages between soil and vegetation. We first assume that no rainfall excess overland flow occurs as the virtual experiment is based in a forest environment with highly permeable soils. The flow and storage of the unsaturated zone and the vegetation is accounted for by a parameter that defines the infiltration depth for every unit of rainfall. As long as the calculated infiltration front for each grid cell is above the water table, no rainfall will recharge to the saturated zone and the mass import will be stored in the unsaturated zone. After the time the infiltration front reaches the water table or bedrock, rainfall is equal to recharge. If on the other hand the water table is rising to the soil surface, the resulting saturation overland flow is simulated by removing the excess rainfall and adding the water and mass directly to the outflow without accounting for overland flow routing. This unsaturated zone model is useful only for single events, but can be replaced if one wishes the virtual experiment to account for a more physically based infiltration process.

The model is written in the IDL development environment to enhance the visualization of spatial and temporal results of subsurface flow and transport. IDL has the potential to visualize 4 to 5 dimensional data and generate animation clips of the continuous simulations.

5. Virtual experimental design

Designing a virtual experiment requires much pre-experiment dialog and interaction between the experimentalist and the modeler. The exercise is not one of fitting parameters to an existing experimental output but to explore first order effects of model decisions on 'measured' response. This often follows on from intensive field campaigns, where the experimentalist may have a highly complex yet quantitative view of hillslope runoff generation.

The virtual experiments reported in this paper are based on the design presented below. While these experimental design criteria will no doubt change from experiment to experiment, we present these details as an example that one might conduct and, in so doing, exemplify our approach for this chosen case. Hillslope topography for our virtual hillslope is based on the measured surface and bedrock topography of the Panola hillslope (McDonnell et al., 1996; Freer et al., 2002), where topography and soil depth was surveyed at a resolution of 2 m. The upper and slope parallel boundary was defined as a no flow condition. The water table at the lower boundary was fixed to be equal the bedrock topography, thus reflecting a trench that was installed at the Panola hillslope. A constant water level at the base of the hillslope at a certain depth would reflect direct outflow into a creek or river.

One common problem of transferring the experimental results from one site to another is that either the soil properties change or the geometry is different. The advantage of designing virtual experiments is that, for example, the hillslope geometry can remain constant but the soil properties can change and vice versa. Experimental results often showed that saturated lateral flow in the absence of macropores is influenced mainly by the saturated hydraulic conductivity, the transmissivity profile in the soil, and the drainable porosity. The virtual experiment design in this paper will explore mainly the effect of changing the drainable porosity and thus the changes in the shape of the retention characteristic of the soils. The physical soil properties (water retention curve and saturated hydraulic conductivity) are based on data from the H.J. Andrews Experimental Forest (Ranken, 1974). Both soils have a clay–loamy texture; however, the water retention characteristics are quite different, especially near saturation (Fig. 2). The drainable porosity was determined from the difference between the saturated water content and the water content at a soil water tension of 100 cm (approximately field capacity). The resulting values are $n_d = 0.04$ for the so-called low drainable porosity soil and $n_d = 0.12$ for the high drainable porosity soil. The measured saturated hydraulic conductivities are $K_{sat} = 8.3 \times 10^{-5}$ m s^{-1} and $K_{sat} = 3.3 \times 10^{-4}$ m s^{-1}, respectively. We attribute these differences between the soils to different structural development of the soil, resulting in a larger fraction of macropores and thus a higher saturated hydraulic conductivity for the high drainable porosity soil. Fig. 2 also shows that the rather arbitrary selection of the actual soil water tension attributed to the field capacity is not sensitive to the resulting value of drainable porosity. An effective porosity of 30% and a transmissivity profile with an exponent of 1.3 were kept constant throughout the experiments.

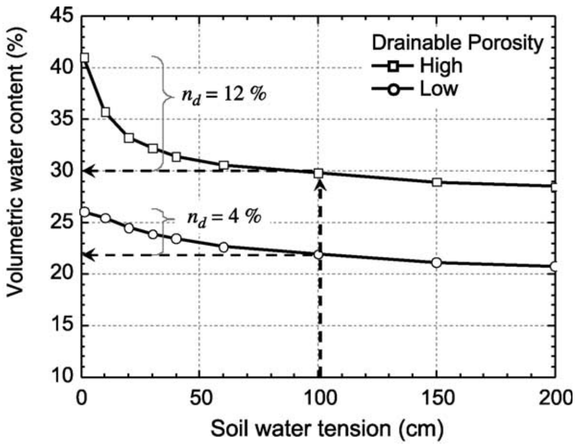

Fig. 2. Water retention characteristic for two soils based on data from the H.J. Andrews Experimental Forest, USA (Ranken, 1974), showing the determination of the drainable porosity based on the concept of field capacity and the water retention curve.

The rainfall event was designed to represent a long duration, low event characteristic of the Pacific Northwest, USA conditions (where the soil properties were taken from) with a total amount of 55 mm and intensities varying between 1.5 and 3.5 mm h^{-1}. We used our simple infiltration module, where infiltration front moves 30 mm for every 1 mm of rainfall. The total simulation time was set to 80 h, with a time step of 15 min.

The input conditions for mass transport were defined in order to be able to 'virtually' describe hydrograph separation and tracer breakthrough from a line source. For calculating event and pre-event water contribution to the subsurface runoff, a constant concentration was added to the rainfall in a separate transport domain. As the concentration in the hillslope was zero at the beginning of the event (reflecting the 100% prevent water concentration), a two component hydrograph separation was calculated knowing exactly the event, prevent and runoff concentration (Genereux and Hooper, 1998). Using a similar approach, the percentage of new water in the soil column can be calculated and thus the spatial distribution of event water in the hillslope can be visualized. Tracer breakthrough from a line source was simulated by adding a constant amount of mass at a simulation time of 3 h in all grid cells located 20 m upslope from the lower boundary (~2/5 of the way up the slope). The mass flux at the slope base within this transport domain was then used to calculate the outflow concentration. The mass in the saturated and unsaturated zone in each grid cell was used to calculate the concentration in the soil column.

6. Virtual experiment results

6.1. Subsurface flow

Analysis and discussion of virtual experiment results is best done by viewing the animated simulations. In this way, the experimentalist and the modeler can interpret the results of the model output. As the continuous simulations can unfortunately not be shown in this paper, we have extracted two ‘snap shots’ from the animated virtual simulations. The two snap shots are then compared for the low and high drainable porosity in order to test the hypothesis of how drainable porosity controls the flow and transport in the hillslope. Each snapshot contains similar elements, showing the temporal and spatial variability of the selected behavior of the hillslope. Subsurface flow behavior is described by the spatial variability of relative flow in the hillslope, the spatial variability of outflow, the temporal variability of the rainfall and total subsurface runoff (Fig. 3). The spatial variability of relative flow in the hillslope is

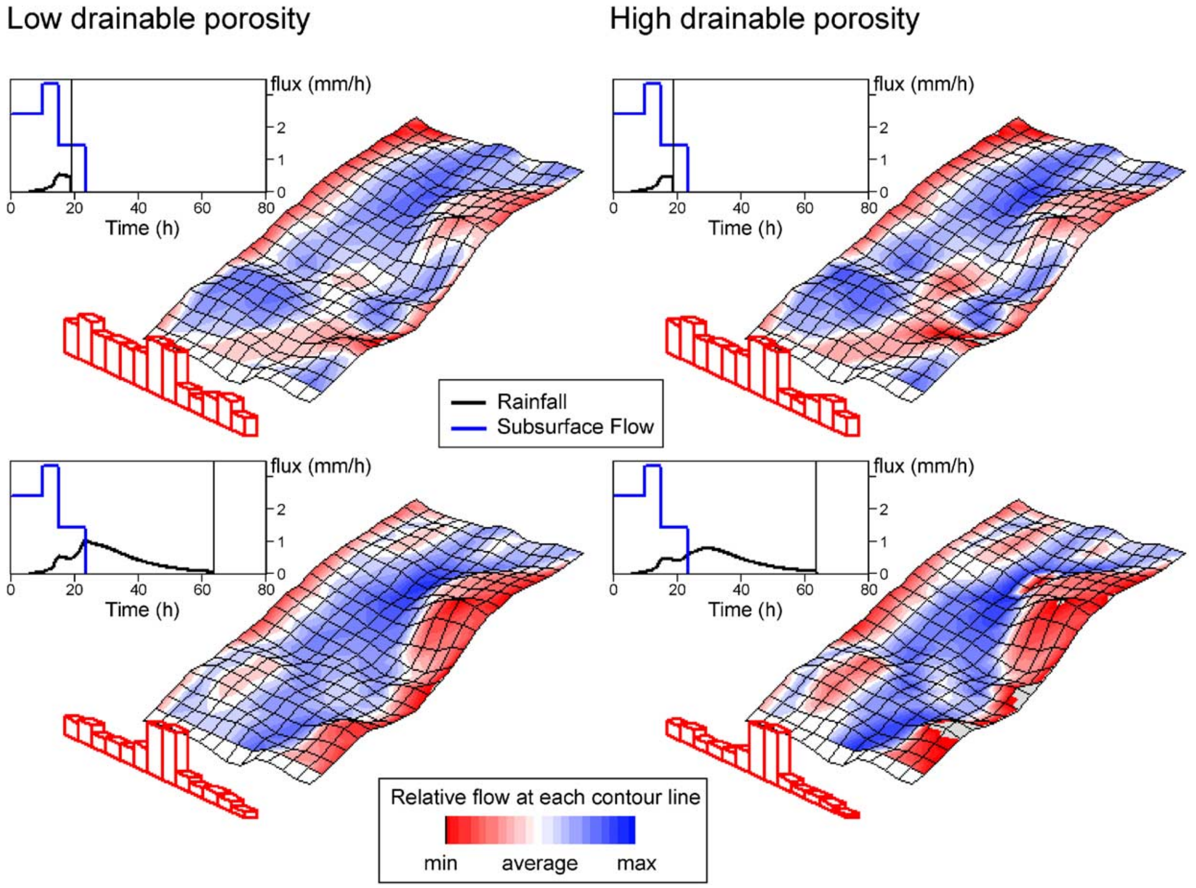

Fig. 3. Two snapshots of the virtual experiments for low and high drainable porosity, showing subsurface flow dynamics described by the spatial variability of relative flow along each contour line in the hillslope (color coded values superimposed on the elevation mesh of bedrock topography), the spatial variability of outflow (bar plot), the temporal variability of the rainfall and total subsurface runoff (flux–time plot, where the thin vertical line is indicating the actual step).

calculated by dividing the actual lateral flow in each grid cell by the average flow along the same distance to the lower boundary. The resulting values are color-coded from low to high relative flow. The color-coded values for each grid are shown on top of the 3-dimensional bedrock topography. The outflow at the lower boundary is represented by a bar plot showing the outflow relative to the average outflow at the simulated time step. The actual simulation time relative to the rainfall input and the subsurface runoff is shown in the upper left corner by an x–y graph representing time and total flux of in- and output together with a moving vertical line depicting the simulation time.

By viewing the flow variability in the hillslope and the variability in outflow at the slope base it appears that flow variability is lower during higher runoff (upper panels in Fig. 3) and higher during lower runoff in the recession (lower panels in Fig. 3). These 'observations' are consistent with field experiments at the Panola site in Georgia (Freer et al., 2002). We interpret the lower runoff variability during the first phase of the hydrograph to be controlled mainly by vertical flow (infiltration) and general wetting up of the soil profile (which does show a less variable behavior). This contrasts with the higher runoff variability during the recession of the hydrograph which is controlled mainly by the lower bedrock topography and thus the accumulation of subsurface flow in hollows (McDonnell et al., 1996). If one compares the results for the high and low drainable porosity, one potential first order control is isolated. The cause and effect relationship between drainable porosity and the selected behavior, like subsurface flow, in the hillslope is now possible to view. Flow variability within the hillslope is higher for the high drainable porosity. Thus, an increase of the drainable porosity within the hillslope results in an increase of subsurface channeling for the same bedrock topography. For both cases, subsurface outflow at the trench begins first in areas with a shallow soil depth, which is also in accordance to observations at the Panola hillslope (Burns et al., 1998; Freer et al., 2002), since the wetting front reaches the bedrock earlier in the shallow soil and thus can generate lateral flow faster as compared to the deeper soil. The shape of the hydrograph is generally quite similar for the two cases of drainable porosity but with a more delayed peak for the high drainable porosity.

6.2. Saturation depth

Examination of temporal and spatial variability of the saturation depth at the hillslope scale is often made by the experimentalist using piezometers or wells. While the Panola hillslope is densely instrumented with piezometers in a 2 m by 2 m grid, most study hillslopes lack this resolution to be able to interpolate the spatial variability of transient saturation on the slope. We can visualize piezometer measurements within the framework of the virtual experiments by showing the depth of saturation of each grid cell within the hillslope (Fig. 4). The depth of saturation is color-coded and overlain on the bedrock topography. In addition, outflow variability is shown by the bar plots (same as for the snapshots for subsurface flow). The graph in the top left corner shows the cumulative soil depth and the cumulative depth of saturation for the actual time step given in Fig. 3.

For the first snapshot near the peak of the hydrograph, the depth of saturation is very different between the low and high drainable porosity case. For the low drainable porosity the depth of saturation is high in the center of the hillslope within some depressions of the bedrock topography but still low, where the soil is shallow (upper left panel in Fig. 4). The resulting cumulative distribution of depth of saturation shows a high variability with a shape similar to the soil depth distribution. For the high drainable porosity, the depth of saturation is generally much lower, which can be seen for the spatial depth of saturation as well as for the distribution of saturation depth. The second snapshot during the recession of the hydrograph shows only a small part of the hillslope with a substantial depth of saturation for both cases of drainable porosity. The low drainable porosity still has a higher average depth of saturation; however, the patterns of saturation depth are more similar compared to the first snapshot time period. The shape of the depth of saturation distribution is different from the shape of the soil depth distribution, as the water table is influenced mainly by the drainage process of the hillslope. This is opposite to the first snapshot, where the water

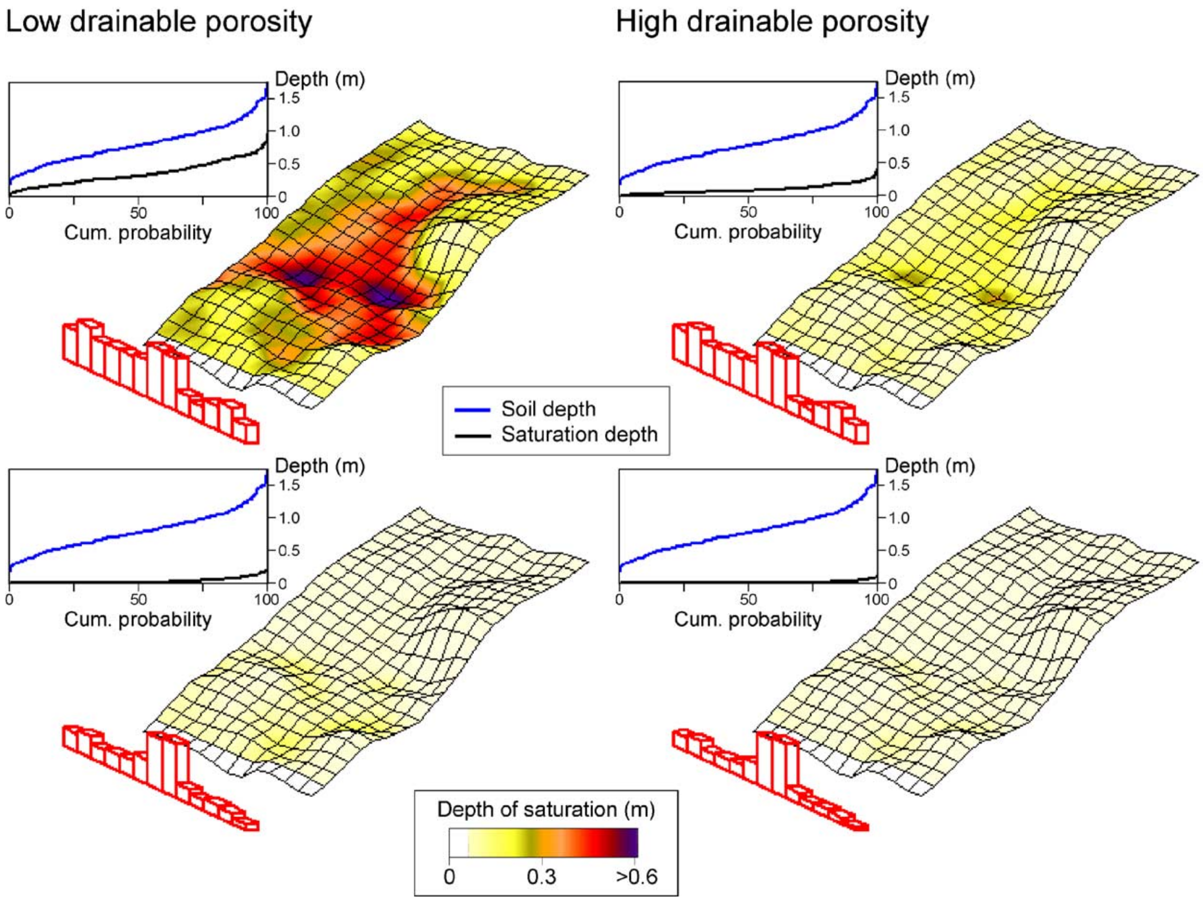

Fig. 4. Two snapshots of the virtual experiments for low and high drainable porosity, showing the dynamics of saturation depth described by the spatial variability of depth of saturation in the hillslope (color coded values superimposed on the elevation mesh of bedrock topography), the spatial variability of outflow (bar plot), and the cumulative soil depth and depth of saturation for the time step (soil depth–probability plot).

table was controlled mainly by rainfall and infiltration. The general difference of the depth of saturation between the two drainable porosity cases is attributed to the influence of the drainable porosity, that controls the rise of the saturated zone in relation to a unit recharge.

6.3. A virtual line source experiment

The temporal and spatial concentration changes in the hillslope for the same rainfall input are visualized in Fig. 5. The spatial variability of the tracer concentration within the soil column (saturated and unsaturated zone) is color coded and shown for each grid cell in relation to the bedrock topography. The concentrations in the soil are shown as concentrations relative to the maximum concentration in the hillslope. The relative variability of tracer concentration in the outflow is visualized by the bar plot. The temporal dynamics and thus the breakthrough curve is depicted by the time-concentration graph in the upper left corner of Fig. 5.

The drainable porosity has a large control on the general behavior of tracer movement in the virtual hillslope. For the high drainable porosity, the tracer plume moves faster than for the low drainable porosity and the breakthrough curve is already completed within the simulation time. As the effective porosity n_{eff} is equal for both cases, only the exchange of mass between the saturated and unsaturated zone can be responsible for the differences. The high drainable porosity soil has a much smaller water

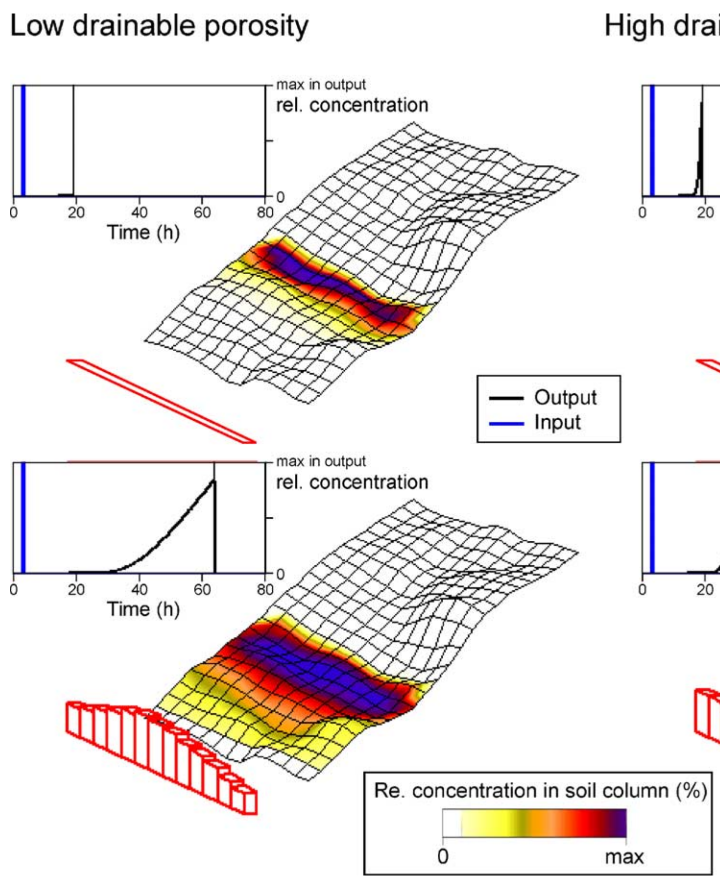

Fig. 5. Two snapshots of the virtual experiments for low and high drainable porosity, showing the dynamics of solute transport described by the spatial variability of the relative tracer concentration within the soil column (color coded values superimposed on the elevation mesh of bedrock topography), the spatial variability of tracer concentration in the outflow (bar plot), and the temporal variability of relative outflow concentration (concentration–time plot, where the thin vertical line is indicating the actual step).

volume in the unsaturated zone compared to the low drainable porosity soil and the water table variations are much higher for the low drainable porosity case. Hence, mass exchange between the saturated and unsaturated zone is much higher for the low drainable porosity soil, resulting in a slower tracer movement and delayed breakthrough.

The subsurface flow variability, itself enhanced by the bedrock topography, affects the spatial movement of the tracer plume. The higher flow variability for the high drainable porosity produces an unequal movement of the tracer and a splitting of the tracer plume, which can be seen especially for the first time snapshot in Fig. 5. This effect is not as pronounced for the low drainable porosity. The concentrations in the runoff at the lower boundary are very similar for the high drainable porosity, due to the muted water table variations and therefore minimal impact on saturated zone concentration. However, the higher water table variations for the low drainable porosity produce a higher concentration change in the saturated zone, resulting in a higher variability of the outflow concentration (lower left panel in Fig. 5).

6.4. Using virtual isotopes for hydrograph separation

Virtual experiments provide an opportunity for isotope tracing experiments and hydrograph separation at the hillslope scale. If the rainfall is simulated with a different constant tracer concentration than the water that is already stored in the unsaturated zone in the hillslope, the event water proportion of the runoff as well as within the soil column in each grid cell in the hillslope can be calculated. The reader is referred to Kendall and McDonnell (1998) for details of this technique. Fig. 6 shows the spatial variability of the event water in the soil column as color-coded values for each grid cell superimposed on the bedrock topography. In addition, the outflow variability of the proportion of event water is shown as bar plots with the upper rectangles representing the 100% value. The outflow hydrograph of total runoff and event water runoff is shown for the appropriate time step in the upper left corner.

Event water percentage in the hillslope and contribution from the hillslope is quite different for the two drainable porosity cases. The hillslope with the high drainable porosity generates a higher proportion of event water than the hillslope with the low drainable porosity. Event water dominates the runoff response especially on the rising limb of the hydrograph (upper right panel of Fig. 6). Event water proportion is lower during the recession compared to the rising limb for both cases. The outflow variability of the event water is very low for the hydrograph recession, but higher for the first snapshot during the rainfall event for both cases. As the event water proportion is controlled mainly by the water that is stored in the hillslope prior to the rainfall event (prevent water), the differences for the low and high drainable porosity cases can be easily explained by the high drainable porosity soil having a much lower

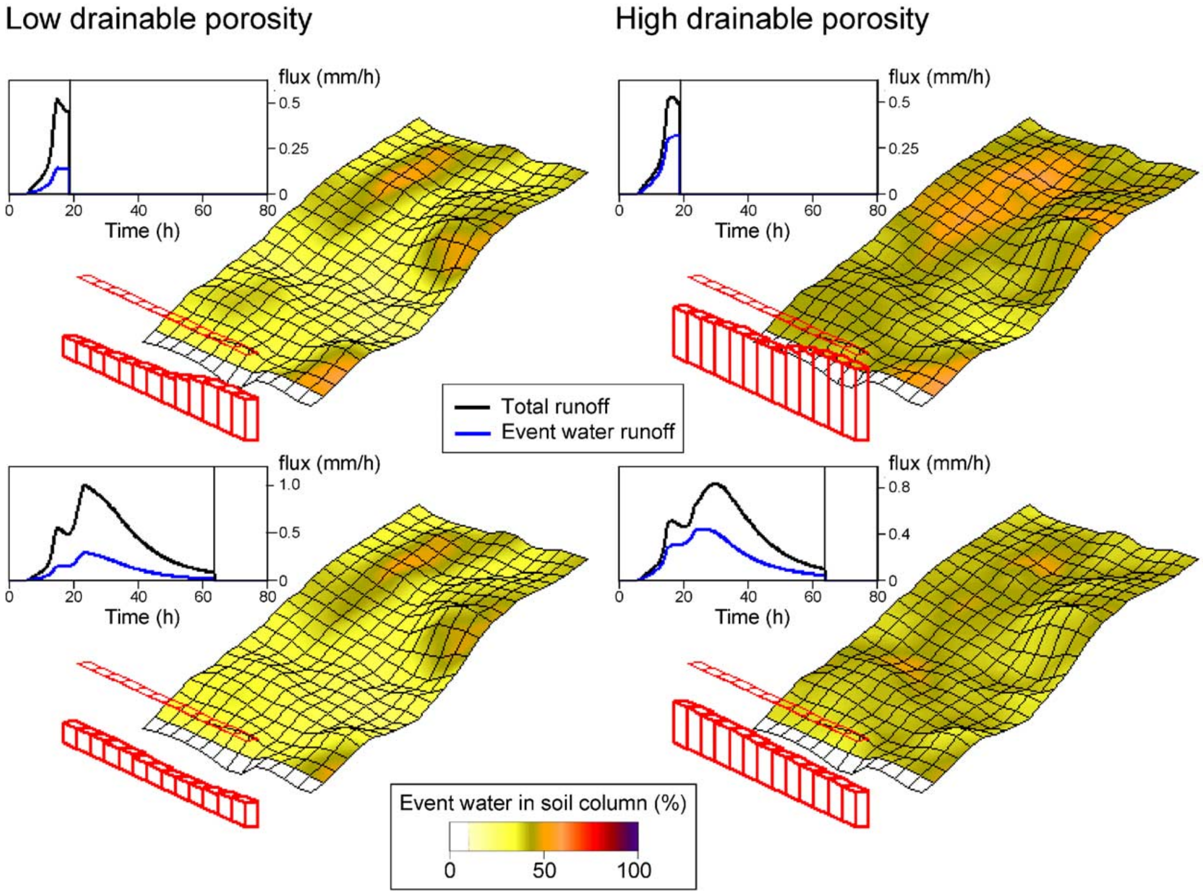

Fig. 6. Two snapshots of the virtual experiments for low and high drainable porosity, showing the dynamics of isotope tracing hydrograph separation described by the spatial variability of the event water percentage within the soil column (color coded values superimposed on the elevation mesh of bedrock topography), the spatial outflow variability of the proportion of event water compared to the 100% value (bar plot), and the outflow hydrograph of total runoff and event water runoff (flow–time plot, where the thin vertical line is indicating the actual step).

pre-event water volume as compared to the low drainable porosity soil. Another striking observation is the spatial variability of event water in the hillslope. Applying hydrograph separation in natural catchments usually assumes a spatially uniform pre-event water contribution (Buttle, 1994). The second snapshot of both drainable porosity cases show a quite variable proportion of event water in the soil column, which is controlled mainly by the bedrock topography and the total water that flushed the grid cell. Since the event water proportion is now spatially variable within the hillslope, the prevent water concentration for the next hypothetical event is then also variable. While some recent literature exists on this topic (Kendall et al., 2001), the virtual experiment offers much potential to examine these internal mixing effects.

7. Discussion

Repeat experiments in hillslope hydrology are rare (McGlynn et al., 2002). While Hooper (2001) has advocated the merits of formal hypothesis testing to the hillslope and catchment hydrology community, direct challenges of hillslope conceptualizations are rarely made. On the few occasions when this has happened, new groups using new approaches have concluded quite different first order controls on water delivery (see the so-called Maimai debate in McGlynn et al. (2002)). Virtual experiments have the potential to test complex perceptions of hillslope behavior by constraining a conceptualization via the multiple outputs of water flux, tracer breakthough, old/new water mixing and age spectra. As seen in the virtual experiments in this paper, the combination of these very diverse 'datasets' constrains the age, origin and pathway of water flux and transport in ways that narrow the possible process possibilities.

Indeed, Beven (2001a) has shown clearly that outflow from the base of a hillslope or outlet of a catchment alone is a weak test of a model or process conceptualization. The virtual experiment can be viewed as an 'equifinality reducing instrument' whereby multiple outputs define how the system works for the right process reasons. We see especially a large potential in combining aggregated spatial information (like, for example, the saturation depth distribution) and to compare it with the few available water table measurements.

The experimentalist is often restricted to point measures of water content, matric potential, and water table depth. It is very difficult then to measure the hillslope dynamics at the scale, where the process occurs—perhaps missing the observation of how the slope as a whole functions. Thus, we force our model or focus our calibration to obey these point measurements. If there is high spatial correlation between individual point measurements of matric potential or water table depth, which is possible in soils with a low drainable porosity, one could relate point measurements of water table fluctuations to the dynamics of soil saturation and perched water table development. However, in many soils, like those described in this paper, this has been a fallacy, as soils with a high drainable porosity tend to generate a poor correlation between individual point measurements and channelize the subsurface flow into a few lateral preferential flow pathways.

The virtual experiments revealed a high variability of the event water in the hillslope after the rainfall event. As this event water variability becomes the pre-event water isotopic signature for the next event, it would be worthwhile for future experiments to explore the effects of this spatial variability on the hydrograph separation approach.

We did not show in detail the potential of the virtual experiments to explore the residence time distribution of water in the hillslope. Since the virtual experiment can be run for months or years, we can simulate a series of natural rainfall events and trace the input to the whole area for one time interval. Thus, the residence time distribution can be directly studied and, the first order controls can again be extracted. We are also able to simulate the natural variations of an input signal like ^{18}O or ^{3}H and see if our simple transfer function models (Maloszewski and Zuber, 1996) are able to capture the residence time distribution, or if new ideas like the power spectrum approach (Kirchner et al., 2000) are more appropriate. We are actively working on this problem now.

Finally, we should ask ourselves how reasonable the results from the virtual experiments are compared to the 'real' experimental finding. Again, as it was not our intention to fit parameters to existing experimental data, but to use this approach to explore first order controls and to show a possible way to an organizational structure for hillslope hydrological studies, we will not conclude with some numbers for optimization measures or graphs showing how well our model performed. Rather, the virtual results can be related qualitatively to other experimental findings. The dynamics of the subsurface flow variability was observed with a similar behavior at the Panola site (Freer et al., 2002) and at the Maimai site (Woods and Rowe, 1996). The channeling of flow pathways and their effects on flushing frequency in soils and concentration of nutrients and other trace elements was observed at the hillslope scale (Burns et al., 1998) and at the pedon scale (Bundt et al., 2001). Brammer (1996) studied in detail the effects of the bedrock topography and the mass exchange between the saturated and unsaturated zone on the movement of an artificial tracer line source. The resulting patterns from the virtual experiments match surprisingly well with his field experimental findings at the Maimai hillslope site.

We are aware that the detailed information from 'multi-year and multi-tracer' field experiments like that performed in Panola will not be always available to guide the virtual experiments. Therefore we propose to use statistical characteristics (average and variance of soil depth, spatial correlation length) and general topographic features (horizontal and vertical curvature) to design an experiment for a virtual hillslope that may be used to explore the characteristics within new environmental settings, where the virtual experiment may go hand-in-hand with new process studies.

8. Conclusion

This paper develops and implements a series of virtual experiments, whereby the interaction between water flow pathways, source, and mixing at the hillslope scale is examined by modeler and experimentalist within a virtual experiment framework. We argue that these virtual experiments are essentially different to traditional numerical experiments since the intent is to explore first-order controls in hillslope hydrology, where the experimentalist and modeler work together to develop and analyze the results collectively. Our results showed that the virtual experimental framework could be a way to examine first order controls on slope dynamics—namely the effect of drainable porosity on flow and transport at the hillslope scale. When combined with previous experimental findings and conceptualizations, virtual experiments can be an effective way to isolate certain controls and examine their influence over a range of rainfall and antecedent wetness conditions. Future work within the virtual model framework could explore depth distributions of drainable porosity and its effect on hillslope runoff and transport as well as effects of hillslope geometry and even soil water–vegetation interactions and their effect on slope runoff non-linearities.

References

Alila, Y., Beckers, J., 2001. Using numerical modeling to address hydrologic forest management issues in British Columbia. Hydrological Processes 15, 3371–3387.

Anderson, M.G., Burt, T.P., 1990. Process Studies in Hillslope Hydrology, Wiley, Chichester.

Anderson, S.P., Dietrich, W.E., 2001. Chemical weathering and runoff chemistry in a steep headwater catchment. Hydrological Processes 15, 1791–1815.

Anderson, S.P., Dietrich, W.E., Montgomery, D.R., Torres, R., Conrad, M.E., Loague, K., 1997. Subsurface flow paths in a steep, unchanneled catchment. Water Resources Research 33 (12), 2637–2653.

Bear, J., 1972. Dynamics of fluids in porous media, Elsevier, New York, 764 pp.

Bencala, K.E., 2000. Hyporheic zone hydrological processes. Hydrological Processes 14, 2797–2798.

Beven, K.J., 2001a. How far can we go in distributed hydrological modelling? Hydrology and Earth System Sciences 5 (1), 1–12.

Beven, K.J., 2001b. On fire and rain (or predicting the effects of change). Hydrological Processes 15, 1397–1399.

Beven, K., Freer, J., 2001. Equifinality, data assimilation, and uncertainty estimation in mechanistic modelling of complex environmental systems using the GLUE methodology. Journal of Hydrology 249, 11–29.

Binley, A., Beven, K., Elgy, J., 1989. Physically Based Model of Heterogeneous Hillslopes: 2. Effective Hydraulic Conductivities. Water Resources Research 25, 1227–1233.

Bishop, K.H., Hauhs, M., Nyberg, L., Seibert, J., Moldan, F., Rodhe, A., Lange, H., Lischeid, G., 1998. The hydrology of the covered catchment. In: Hultberg, H., Skeffington, R. (Eds.), Experimental Reversal of Acid Rain Effects: The Gårdsjön Covered Catchment Experiment, Wiley, New York, pp. 109–136.

Blöschl, G., 2001. Scaling in hydrology. Hydrological Processes 15, 709–711.

Blöschl, G., Grayson, R., 2001. Spatial observations and interpolation. In: Grayson, R., Blöschl, G. (Eds.), Spatial Patterns in Catchment Hydrology; Observations and Modelling, Cambridge University Press, Cambridge, pp. 17–50.

Bonell, M., 1998. Selected challenges in runoff generation research in forests from the hillslope to headwater drainage basin scale. Journal of the American Water Resources Association 34 (4), 765–786.

Brammer, D., 1996. Hillslope hydrology in a small forested catchment, Maimai, New Zealand. MS Thesis. State University of New York College of Environmental Science and Forestry, Syracuse, 153 pp.

Bronstert, A., Plate, E.J., 1997. Modelling of runoff generation and soil moisture dynamics for hillslopes and micro-catchments. Journal of Hydrology 198 (1–4), 177–195.

Brown, V.A., McDonnell, J.J., Burns, D.A., Kendall, C., 1999. The role of event water, a rapid shallow flow component, and catchment size in summer stormflow. Journal of Hydrology 217 (3–4), 171–190.

Bundt, M., Jäggi, M., Blaser, P., Siegwolf, R., Hagedorn, F., 2001. Carbon and nitrogen dynamics in preferential flow paths and matrix of a forest soil. Soil Science Society of America Journal 65, 1529–1538.

Burns, D.A., Hooper, R.P., McDonnell, J.J., Freer, J.E., Kendall, C., Beven, K., 1998. Base cation concentrations in subsurface flow from a forested hillslope: the role of flushing frequency. Water Resources Research 34 (12), 3535–3544.

Buttle, J.M., 1994. Isotope hydrograph separations and rapid delivery of pre-event water from drainage basins. Progress in Physical Geography 18 (1), 16–41.

Buttle, J.M., Turcotte, D.S., 1999. Runoff processes on a forested slope on the Canadian shield. Nordic Hydrology 30, 1–20.

Calver, A., Cammeraat, L.H., 1993. Testing a physically-based runoff model against field observations on a Luxembourg hillslope. Catena 20, 273–288.

Dunne, T., 1978. Field studies of hillslope flow processes. In: Kirkby, M.J., (Ed.), Hillslope Hydrology, Wiley, Chichester, pp. 227–293.

Dunne, T., 1983. Relation of field studies and modeling in the prediction of storm runoff. Journal of Hydrology 65, 25–48.

Dunne, T., 1998. Wolman Lecture: Hydrologic Science…In Landscapes…On a Planet…In the Future, Abel Wolman Distinguished Lecture and Symposium on the Hydrologic Sciences, Water Science and Technology Board, Commission on Geosciences, Environment, and Resources, National Research Council. Hydrologic Sciences—Taking Stock and Looking Ahead. National Academy Press, Washington DC, pp. 10–43.

Dunne, T., Black, R.D., 1970a. An experimental investigation of runoff production in permeable soils. Water Resources Research 6, 478–490.

Dunne, T., Black, R.D., 1970b. Partial area contributions to storm runoff in a small New England watershed. Water Resources Research 6 (5), 1296–1311.

Faeh, A.O., Scherrer, S., Naef, F., 1997. A combined field and numerical approach to investigate flow processes in natural macroporous soils under extreme precipitation. Hydrology and Earth System Sciences 1 (4), 787–800.

Flühler, H., Durner, W., Flury, M., 1996. Lateral solute mixing processes—a key for understanding field-scale transport of water and solutes. Geoderma 70, 165–183.

Freer, J., McDonnell, J., Beven, K.J., Brammer, D., Burns, D., Hooper, R.P., Kendal, C., 1997. Topographic controls on subsurface storm flow at the hillslope scale for two hydrologically distinct small catchments. Hydrological Processes 11 (9), 1347–1352.

Freer, J., McDonnell, J.J., Beven, K., Peters, N.E., Burns, D.A., Hooper, R.P., Aulenbach, B., Kendall, C., 2002. The role of bedrock topography on subsurface storm flow. Water Resources Research 38 (12) 101029/2001WR000872.

Freeze, R.A., Cherry, J.A., 1979. Groundwater, Prentice-Hall, Englewood Cliffs, NJ, 604 pp.

Genereux, D.P., Hooper, R.P., 1998. Oxygen and hydrogen isotopes in rainfall–runoff studies. In: Kendall, C., McDonnell, J.J. (Eds.), Isotope Tracers in Catchment Hydrology, Elsevier, Amsterdam, pp. 319–346.

Germann, P.F., 1990. Macropores and hydrologic hillslope processes. In: Anderson, M.G., Burt, T.P. (Eds.), Process Studies in Hillslope Hydrology, Wiley, New York.

Grayson, R., Western, A., 2001. Terrain and the distribution of soil moisture. Hydrological Processes 15 (13), 2689–2690.

Hewlett, J.D., Hibbert, A.R., 1967. Factors affecting the response of small watersheds to precipitation in humid areas. In: Sopper, W.E., Lull, H.W. (Eds.), Forest Hydrology, Pergamon Press, New York, pp. 275–291.

Hooper, R.P., 2001. Applying the scientific method to small catchment studies: a review of the Panola Mountain experience. Hydrological Processes 15, 2039–2050.

Jones, J.A., Swanson, F.J., 2001. Hydrologic inferences from comparisons among small basin experiments. Hydrological Processes 15, 2363–2366.

Kendall, C., McDonnell, J.J., 1998. Isotope Tracers in Catchment Hydrology, Elsevier, Amsterdam, 839 pp.

Kendall, K.A., Shanley, J.B., McDonnell, J.J., 1999. A hydrometric and geochemical approach to test the transmissivity feedback hypothesis during snowmelt. Journal of Hydrology 219 (3–4), 188–205.

Kendall, C., McDonnell, J.J., Gu, W., 2001. A look inside black box hydrograph separation models: a study at the Hydrohill catchment. Hydrological Processes 15, 1877–1902.

Kirchner, J.W., Feng, X., Neal, C., 2000. Fractal stream chemistry and its implications for contaminant transport in catchments. Nature 403 (6769), 524–527.

Kirkby, M.J., 1978. Hillslope Hydrology, Landscape Systems, Wiley, Chichester, 389 pp.

Maloszewski, P., Zuber, A., 1996. Lumped Parameter Models for the Interpretation of Environmental Tracer Data, Manual on Mathematical Models in Isotope Hydrogeology, International Atomic Energy Agency, Vienna, pp. 9–58.

McDonnell, J.J., 1990. A rationale for old water discharge through macropores in a steep, humid catchment. Water Resources Research 26 (11), 2821–2832.

McDonnell, J.J., 1997. Comment on the changing spatial variability of subsurface flow across a hillslide by Ross Woods and Lindsay Rowe. Journal of Hydrology, New Zealand 36 (1), 97–100.

McDonnell, J.J., Owens, I.F., Stewart, M.K., 1991. A case study of shallow flow paths in a steep zero-order basin. Water Resources Research 27, 679–685.

McDonnell, J.J., Freer, J., Hooper, R., Kendall, C., Burns, D., Beven, K., Peters, J., 1996. New method developed for studying flow in hillslopes. EOS, 77 (47), 465.

McDonnell, J.J., Brammer, D., Kendall, C., Hjerdt, N., Rowe, L., Stewart, M., Woods, R., 1998. Flow pathways on steep forested hillslopes: the tracer, tensiometer and trough approach. In: Tani, M., (Ed.), Environmental Forest Science, Kluwer, Dordrecht, pp. 463–474.

McGlynn, B.L., McDonnell, J.J., Brammer, D.D., 2002. A review of the evolving perceptual model of hillslope flowpaths at the Maimai catchments, New Zealand. Journal of Hydrology 257, 1–26.

McIntosh, J., McDonnell, J.J., Peters, N.E., 1999. Tracer and hydrometric study of preferential flow in large undisturbed soil cores from the Georgia Piedmont, USA. Hydrological Processes 13, 139–155.

Montgomery, D.R., Gran, K.B., 2001. Downstream variations in the width of bedrock channels. Water Resources Research 37 (6), 1841–1846.

Mosley, M.P., 1979. Streamflow generation in a forested watershed. Water Resources Research 15, 795–806.

Peters, D.L., Buttle, J.M., Taylor, C.H., LaZerte, B.D., 1995. Runoff production in a forested, shallow soil, Canadian Shield basin. Water Resources Research 31 (5), 1291–1304.

Ranken, D.W., 1974. Hydrologic properties of soil and subsoil on a steep, forested slope. MS Thesis. Oregon State University, Corvallis, 114 pp.

Rasmussen, T.C., Baldwin, R.H., Dowd, J.F., Williams, A.G., 2000. Tracer vs. pressure wave velocities through unsaturated saprolite. Soil Science Society of America Journal 64, 75–85.

Rodhe, A., 1987. The origin of streamwater traced by Oxygen-18. PhD Dissertation Thesis. Uppsala University, Uppsala, 260 pp.

Seibert, J., 2001. On the need for benchmarks in hydrological modelling. Hydrological Processes 15, 1063–1064.

Seibert, J., McDonnell, J.J., 2002. On the dialog between experimentalist and modeler in catchment hydrology: use of soft data for multicriteria model calibration. Water Resources Research 38 (11) 101029/2001WR000978.

Seibert, J., Bishop, K., Rodhe, A., McDonnell, J.J., 2002. Groundwater dynamics along a hillslope: a test of the steady-state hypothesis. Water Resources Research 39 (1) 101029/2002WR001404.

Sherlock, M.D., McDonnel, J.J., 2003. Spatially distributed groundwater level and soil water content measured using electromagnetic induction. Hydrological Processes 17, 1965–1977.

Sloan, P.G., Moore, I.D., 1984. Modeling subsurface stormflow on steeply sloping forested watersheds. Water Resources Research 20 (12), 1815–1822.

Smith, R.E., Hebbert, R.H.B., 1983. Mathematical simulation of interdependent surface and subsurface hydrologic processes. Water Resources Research 19 (4), 987–1001.

Tani, M., 1997. Runoff generation processes estimated from hydrological observations on a steep forested hillslope with a thin soil layer. Journal of Hydrology 200, 84–109.

Torres, R., Dietrich, W.E., Montgomery, D.R., Anderson, S.P., Loague, K., 1998. Unsaturated zone processes and the hydrologic response of a steep, unchanneled catchment. Water Resources Research 34 (8), 1865–1879.

Troch, P., van Loon, E., Hilberts, A., 2002. Analytical solutions to a hillslope-storage kinematic wave equation for subsurface flow. Advances in Water Resources 25 (6), 637–649.

Vitvar, T., Burns, D.A., Lawrence, G.B., McDonnell, J.J., Wolock, D.M., 2002. Estimation of baseflow residence times in watersheds from the runoff hydrograph recession: method and application in the Neversink watershed, Catskill Mountains, New York. Hydrological Processes 16 (9), 1871–1877.

Ward, R.C., Robinson, M., 2000. Principles of Hydrology, fourth ed., McGraw-Hill, New York, 365 pp.

Weiler, M., Naef, F., 2003. An experimental tracer study of the role of macropores in infiltration in grassland soils. Hydrological Processes 17 (2), 477–493.

Weiler, M., Naef, F., Leibundgut, C., 1998. Study of runoff generation on hillslopes using tracer experiments and physically based numerical model. IAHS Publication 248, 353–360.

Weyman, D.R., 1973. Measurements of the downslope flow of water in a soil. Journal of Hydrology 20, 267–288.

Whipkey, R.Z., 1965. Subsurface storm flow from forested slopes. Bull. Int. Assoc. Sci. Hydrol. 2, 74–85.

Wigmosta, M.S., Lettenmaier, D.P., 1999. A comparison of simplified methods for routing topographically driven subsurface flow. Water Resources Research 35, 255–264.

Woods, R., Rowe, L., 1996. The changing spatial variability of subsurface flow across a hillside. Journal of Hydrology (NZ) 35 (1), 51–86.

Woods, R.A., Sivapalan, M., 1997. A connection between topographically driven runoff generation and channel network structure. Water Resources Research 33 (12), 2939–2950.