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Conceptualizing lateral preferential flow and flow networks and simulating the effects on gauged and ungauged hillslopes

Markus Weiler[1] and J. J. McDonnell[2]

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[1] One of the greatest challenges in the field of hillslope hydrology is conceptualizing and parameterizing the effects of lateral preferential flow. Our current physically based and conceptual models often ignore such behavior. However, for addressing issues of land use change, water quality, and other predictions where flow amount and components of flow are imperative, dominant runoff processes like preferential subsurface flow need to be accounted for in the model structure. This paper provides a new approach to formalize the qualitative yet complex explanation of preferential flow into a numerical model structure. We base our examples on field studies of the well-studied Maimai watershed (New Zealand). We then use the model as a learning tool for improved clarity into the old water paradox and reasons for the seemingly contradictory findings of lateral preferential flow of old water where applied line sources of tracer appear very quickly in the stream following application. We evaluate the model with multiple criteria, including ability to capture flow, hydrograph composition, and tracer breakthrough. We generate output ensembles with different pipe network geometries for model calibration and validation analysis. Surprisingly, the range of runoff response among the ensembles is narrow, indicating insensitivity to specific pipe placement. Our new model structure shows that high transport velocities for artificial line source tracers can be reconciled with the dominance of preevent water during runoff events even when lateral pipe flow dominates response. The work suggests overall that preferential flow can be parameterized within a process-based model structure via the structured dialog between experimentalist and modeler.

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1. Introduction

[2] Hillslope hydrology is still poorly understood despite numerous hillslope trenching campaigns (J. J. McDonnell et al., Slope Intercomparison Experiment: Forging a new hillslope hydrology, submitted to *Hydrological Processes*, 2006, hereinafter referred to as McDonnell et al., submitted manuscript, 2006), some dating back almost one hundred years [*Engler*, 1919]. *Hursh and Brater* [1941] were among the first studies to quantify the role of subsurface stormflow. Their seminal work showed that the stream hydrograph response to storm rainfall at the forested Coweeta experimental watershed was composed of two main components: channel precipitation and subsurface stormflow. Later, *Hoover and Hursh* [1943] showed that soil depth, topography, and hydrologic characteristics associated with different elevations influenced peak discharge. While the rate of progress in understanding subsurface stormflow increased substantially through many field campaigns during the International Hydrological Decade (IHD) [e.g., *Whipkey*, 1965; *Dunne and Black*, 1970; *Weyman*, 1973], we have entered the new IAHS Decade on Prediction in Ungauged Basins with little ability to make nontrivial predictions of subsurface stormflow behavior on slopes that have not yet been trenched and gauged. Even when we do have a detailed trenched hillslope, extrapolation to a neighboring site in the same catchment is often impossible.

[3] Progress is being made in developing new theory [*Troch et al.*, 2002] and new field diagnostics [*Scherrer and Naef*, 2003] of hillslope hydrology. However, these approaches often ignore what many hillslope investigations since (and including) *Engler* [1919] have observed: lateral preferential flow domination of stormflow response. These preferential lateral flow networks have been described anecdotally in studies around the world, from semiarid hillslopes [*Newman et al.*, 1998] to subtropical sites [*Freer et al.*, 2002], from steep forested Pacific Rim sites [*Tani*, 1997] to grassland sites in the Swiss Alps [*Weiler et al.*, 1998]. Recent intercomparison studies have shown that lateral preferential flow is often highly threshold dependent, with a certain local rainfall amount threshold necessary to activate lateral preferential flow [*Uchida et al.*, 2005]. One particularly vexing issue in the context of this network-like hillslope response [*Tsuboyama et al.*, 1994] to storm rainfall (and snowmelt) is the often paradoxical accompanying

[1]Department of Forest Resources Management and Department of Geography, University of British Columbia, Vancouver, British Columbia, Canada.

[2]Department of Forest Engineering, Oregon State University, Corvallis, Oregon, USA.

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finding that most of the water emanating from these preferential flow networks at the slope base is water that was stored in the soil profile prior to the rain event [*McDonnell*, 1990]. While the chemistry of this water is often variable (sometimes showing flushing of soluble products in the soil, sometimes not), this finding of old (preevent) water domination of lateral preferential flow is widespread in humid regions [*Sklash et al.*, 1986; *Uchida et al.*, 2006].

[4] The grand challenges for the field of hillslope hydrology have been summarized recently by McDonnell et al. (submitted manuscript, 2006) stemming from the first Slope Intercomparison Experiment (SLICE, http://sinus.unibe.ch/boden/slice/). These challenges include: intercomparison and classification of hillslope behaviors; distinguishing and resolving hillslope pressure wave response, quantifying the local effects of bedrock permeability of hillslope discharge, developing new theory for hillslope network behavior and developing new measurements strategies at the hillslope scale. One of the greatest challenges to the field currently is conceptualizing and parameterizing the effects of lateral preferential flow on gauged and ungauged hillslopes. Our current physically based and conceptually based models often ignore such behavior. However, for addressing issues of land use change, water quality and other predictions where flow amount and components of flow (even time and geographical source components of flow), dominant runoff processes like preferential subsurface flow need to be accounted for in the model structure.

[5] So how might we describe lateral preferential flow in our models? *Faeh et al.* [1997] used a layer of higher conductivity and a kinematic wave approximation to implement preferential flow in their numerical hillslope model QSOIL. *Bronstert* [1999] used a similar approach in his HILLFLOW model. *Beckers and Alila* [2004] recently implemented a preferential flow routine into the distributed hydrology soil vegetation model (DHSVM). Their approach subdivides the soil into two storage components where flow is driven by Darcy's law assuming different flow velocities. They implemented a threshold parameter for when the preferential flow storage component is filled. There are also many attempts to model vertical macropore flow [*Beven and Clarke*, 1986; *Germann and Beven*, 1985; *Weiler*, 2005], which may serve as an additional guideline to describe lateral preferential flow. While a useful start, most approaches did not incorporate knowledge about commonly observed features of preferential flow pathways and new experimental findings of water flow in preferential pathways, and they typically did not model solute transport to further validate their models. To do this requires original thinking in terms of how to embed site specifics with model structure generality. *Beven* [2000] discussed this in the context of the uniqueness of field measurements as a limitation on model representations. *Phillips* [2003] outlined a philosophical approach that may be a possible way forward for conceptualizing and parameterizing the effects of lateral preferential flow on hillslope hydrology. This involves bridging the qualitative (idiographic) approach with the quantitative (nomothetic) approach that seeks explanation based on the application of laws and relationships that are valid everywhere and always. While *Phillips* [2003] advocates that particularities of place and time can be treated as boundary conditions in this way, our current models of hillslope and catchment hydrology to date ignore the particularities of place and time. We argue that this one-size-fits-all model structure approach exacerbates the equifinality problem as outlined by *Beven and Freer* [2001]. Our experimental evidence from field studies in trenched watersheds, especially in the context of lateral preferential flow, suggests that preferential flow processes vary considerably from site to site. In this paper we embrace the peculiarities of site as a necessary part to explain how such a site might respond to precipitation. This paper is an attempt to combine the quantitative (idiographic) with the qualitative (nomothetic) approach (the basic laws of flow in porous media with observations of Pacific Rim hillslope conditions where lateral preferential flow often controls hillslope response) into a model structure that can be used as a learning tool for further understanding of lateral preferential flow. Our approach attempts to conceptualize the effects of preferential flow of old water in humid catchments, particularly in the Pacific Rim, by bringing lateral preferential flow into a formal model structure. Doing so forces the alliance between traditional flow models in porous media with the often qualitative and complex field descriptions of preferential flow behavior. This also forces a dialog between experimentalist and modeler by necessitating the simplest possible description of preferential flow by the field scientist and the simplest and most parsimonious description of preferential flow in the numerical code by the modeler. The goal of our work is to produce a model structure that minimizes calibration. We use the process observations to guide a physically based model approach with the objective of combining flow and transport to form a tool for better understanding and resolving the old water paradox. The specific objectives are (1) formalize the highly qualitative yet highly complex explanation of preferential flow into a model structure and (2) use the model as a learning tool for improved clarity into the old water paradox and reasons for the seemingly contradictory findings of lateral preferential flow of old water, at high runoff response ratios but where applied line sources of tracer appear very quickly in the stream following application.

2. Maimai Catchment

[6] A watershed exemplar of the old water paradox in steep, wet, preferential flow dominated terrain is the Maimai catchment in New Zealand (see review by *McGlynn et al.* [2002]). Here studies have debated over the years the precise mechanisms for water delivery to the channel because of the seemingly conflicting results from different study approaches. *Mosley* [1979, 1982] found a close coincidence in the time of the discharge peak in the stream and the time of the subsurface stormflow peaks from a series of small trenches on the steep, wet hillsides in the Maimai M8 catchment, implying rapid movement of rainwater vertically in the soil profile and in lateral downslope via connected soil pipes. Mosley's perceptual model considered macropore flow to be a "short-circuiting" process by which water could move through the soil at rates up to 300 times greater than the measured mineral soil saturated hydraulic conductivity and contribute to coincident hillslope and catchment hydrographs. *Pearce et al.* [1986] and *Sklash*

et al. [1986] followed with work in the same catchment by collecting samples of rainfall, soil water and streamflow and analyzed for electrical conductivity, chloride, deuterium, and ^{18}O composition. Using this tracer-based methodology, they found no evidence to support the macropore short-circuiting perceptual model and rather, posed an alternative perceptual model of subsurface water discharge to the stream as an isotopically uniform mixture of stored water. While no causal mechanism for this old water delivery was determined in situ, *Sklash et al.* [1986] invoked groundwater ridging theory [*Gillham*, 1984] to explain how such large old water signatures might appear in the channel so quickly and in such large amounts. *McDonnell* [1989, 1990] and *McDonnell et al.* [1991a] combined isotope and chemical tracing with detailed soil matric potential measurements in an effort to explain the discrepancies between the earlier perceptual models. *McDonnell* [1990] proposed a new conceptual model where, as infiltrating new water moved to depth, water perched at the soil-bedrock interface and "backed up" into the matrix, where it mixed with a much larger volume of stored, old matrix soil water. This water table was dissipated by the moderately well-connected system of pipes at the mineral soil-bedrock interface. Follow-on work by *Brammer* [1996] and *McDonnell et al.* [1996] used line source tracer applications on a Maimai hillslope to trace preferential flow more directly than the bulk subsurface stormflow mixture described by old water. Surprisingly, bromide applied 30 m upslope of a large trench [*Woods and Rowe*, 1996] appeared within 6 hr of application (during a rainfall event). The lateral preferential flow of old water (>90% based on isotopic mixing analysis) was laced with bromide that made its way vertically from the line source and then into the narrow ribbons of highly mobile flow at the soil bedrock interface. Over the succeeding 3 months of rainfall events, >80% of the bromide tracer was recovered, each time with a combination of pulses appearing at the trench face at the peak of events along with the slower diffusive wave moving downslope through the matrix, ultimately reaching the slope base after 90 days. The perceptual model of *McDonnell* [1990] has been used to explain behavior of hillslopes in other hydroclimate settings in Georgia [*Freer et al.*, 2002], Ontario [*Peters et al.*, 1995], and Japan [*Tani*, 1997] where transmissive soils overlie largely impermeable bedrock.

[7] Mean annual precipitation in the M8 watershed at Maimai averages 2600 mm, and produces approximately 1550 mm of runoff. A moderately weathered, nearly impermeable early Pleistocene conglomerate underlies silt-loamy Blackball Hill soils. Study profiles showed an infiltration rate of 6100 mm h^{-1} for the thick ($\sim$17 cm) organic humus layer and 250 mm h^{-1} for the mineral soils. Water retention curves show a low drainable porosity between 0.08 and 0.12. *Mosley* [1979] found that soil profiles at vertical pit faces in the Maimai M8 catchment revealed extensive lateral and vertical preferential flow pathways which formed along cracks and holes in the soil and along live and dead root channels. Preferential flow was observed regularly along soil horizon planes and along the soil-bedrock interface in this study and in more recent research. In the Maimai M8 catchment, *Woods and Rowe* [1996] excavated a 60 m long trench face at the base of a planar hillslope in the Maimai M8 catchment. They measured subsurface flow with an array of troughs. Rainfall and subsurface flow data from this study (10 min time step, 25.01.1993 to 14.05.1993), from a study afterward by *Brammer* [1996] that measured rainfall and subsurface flow, water table response in the hillslope as well as performed a bromide tracer experiment (24 March 1995 to 10 May 1995), together with additional results from studies reviewed by *McGlynn et al.* [2002] are used in this paper.

3. Theory and Methods

[8] Our methodology represents the dialog between experimentalist and modeler where Maimai serves as the "place" for these discussions. Key questions for conceptualizing and parameterizing the effects of preferential flow in a functional sense are: Where would one place the soil pipes in the model elements? What size would they be? How continuous would they be? How variable would their characteristics be and how might they vary in space on the slope? How would water mix between the soil pipe and the matrix? Here, we build upon our recent work that explores the dialog between experimentalist and modeler [*Seibert and McDonnell*, 2002] and notions of virtual experiments [*Weiler and McDonnell*, 2004, 2005] where the modeler and experimentalist work together to better understand a natural system. In this discussion we use the term macropore to describe predominantly vertically oriented preferential pathways with lengths comparable to the soil depths [*Weiler and Naef*, 2003] and pipes as slope parallel preferential flow pathways. These pipes can either be formed by soil fauna (mole and mouse burrows) or more frequently in forest soils by dead root channels (sometimes eroded). In this study we do not consider the continuous, large pipe networks that were frequently observed in peat and loess watersheds [*Jones and Connelly*, 2002].

3.1. Model: Hill-vi

[9] We use a physically based hillslope model Hill-vi as the foundation for discussion between experimentalist and modeler. Field observations of soil pipe density, geometry and pipe length were conceptually implemented into the model Hill-vi. The basic concepts of Hill-vi are described by *Weiler and McDonnell* [2004]. Here we review only the basics of the model structure as the foundation for this new pipe flow analysis. The model is based on the concept of two storages that define the saturated and unsaturated zone for each hillslope grid cell, which is based on DEM and soil depth information. The unsaturated zone is defined by the depth from the soil surface to the water table and its time variable water content. The saturated zone is defined by the depth of the water table above the soil-bedrock interface and the porosity n. Lateral subsurface flow is calculated using the Dupuit-Forchheimer assumption and is allowed to occur only within the saturated zone. Routing is based on the grid cell by grid cell approach [*Wigmosta and Lettenmaier*, 1999]. The local hydraulic conductivity in the soil profile is described by a depth function [*Ambroise et al.*, 1996]. The transmissivity T is then given by a parabolic decline with depth of the saturated hydraulic conductivity:

$$T(z) = \int_z^D K_s(z)dz = \frac{K_0 D}{m}\left(1 - \frac{z}{D}\right)^m \quad (1)$$

where K_o is the saturated hydraulic conductivity at the soil surface, m is the power law exponent, z is the depth into the soil profile (positive downward) and D is the total depth of the soil profile.

[10] While these assumptions and model implementations are similar to existing models like DHSVM [*Wigmosta et al.*, 1994] and RHESSys [*Tague and Band*, 2001], *Weiler and McDonnell* [2005] introduced a depth function for drainable porosity n_d taking into consideration that the drainable porosity usually declines with depth:

$$n_d(z) = n_0 \exp\left(-\frac{z}{b}\right) \qquad (2)$$

where n_0 is the drainable porosity at the soil surface and b is a depth scale. Hill-vi calculates the water balance of the unsaturated zone by the precipitation input, the vertical drainage loss into the saturated zone, and the actual evapotranspiration. Actual evaporation is calculated based on the relative water content in the unsaturated zone and the potential evaporation [*Seibert et al.*, 2003]. Drainage from the unsaturated zone to the saturated zone is controlled by a power law relation of relative saturation within the unsaturated zone and the saturated hydraulic conductivity at water table depth z and a power law exponent c [*Weiler and McDonnell*, 2005]. The water balance of the saturated zone is defined by the drainage input from the unsaturated zone, the lateral inflow from upslope (in terms of water table) cells, outflow to downslope cells by lateral subsurface flow and the corresponding change of water table height. The effect of the vegetation cover on precipitation is described by applying a slightly modified version of the throughfall equation developed for coastal forest of the Pacific Rim in the USA by *Rothacher* [1963].

[11] Hill-vi includes a solute transport routine as described by *Weiler and McDonnell* [2004, 2005]. This is an important added constraint for evaluating model output reasonability and another model performance validation tool. Key observations from the experimentalist like new/old water ratios, line source breakthrough, or residence time calculations can be reproduced with Hill-vi. We assume complete mixing in and only advective transport between the saturated and unsaturated zone and in and between grid cells (numerical dispersion cannot be prevented). The effective porosity for solute transport is assumed to be 80% of the total porosity. Further details can be found in the work of *Weiler and McDonnell* [2004].

3.2. Implementation of Pipe Flow Concepts Into Hill-vi

[12] Our approach for adding lateral pipe flow to the Hill-vi structure was to first determine what common features need to be conceptualized as defined by the numerous field investigations of pipe flow in the Pacific Rim (reviewed by *Uchida et al.* [2001]). These include the following: (1) Measured pipe diameter is often within a narrow range [*Uchida et al.*, 2001] and pipe diameter does not usually restrict flow rate in the pipes [*Weiler*, 2005]. (2) Pipe length and connectivity mapping in natural slopes often shows very discontinuous pipe sections, with maximum lengths less than 2–5 m [*Anderson and Weiler*, 2005; *Kitahara*, 1993]. (3) The location of major pipes within the soil profiles is mostly within a narrow band above the soil-bedrock interface or above a soil layer interface [*Uchida et al.*, 2002]. (4) Water flow in the soil pipes is proportional to the hydraulic head (transient water table depth) above the pipe and a constant that is related to hydraulic conductivity of the soil matrix, internal pipe roughness and tortuosity, hydraulic gradient, and pipe dimensions [*Sidle et al.*, 1995].

[13] On the basis of this distillation of the experimental evidence from the literature, lateral pipe flow simulation within Hill-vi is approached with the assumption that pipe geometry and distribution in the hillslope is defined by the pipe density (fraction of grid cells where a pipe starts) and the mean and standard deviation of the height of the pipes above the bedrock. On the basis of the selected pipe density, grid cells within the simulated hillslope are randomly chosen and then the height of the pipe above bedrock within each grid cell is set randomly based on the defined normal distribution. From each starting location, one pipe can potentially transmit water only to neighboring cells, thus constraining the pipe length to the grid spacing (2 m for the Maimai hillslope). This is in keeping with pipe mapping results in New Zealand and Japan, where pipes are rarely observed to be continuous for more than some meters [*Tsuboyama et al.*, 1994]. The single, final direction of pipe flow is chosen randomly from all possible downslope directions based on the bedrock topographic surface. The pipe height within the starting cell location is again randomly chosen from a Gaussian probability distribution defined by the mean and standard deviation of the height of the pipes above the local bedrock. This approach implements numerically, our current "best generalized" understanding of pipe geometry and uses randomly defined parameters for all the details of what we do not know well or are impossible to measure. Pipe flow within each grid cell is calculated by

$$q_p(t) = k_p A^{0.5} \left(w(t) - z_p\right)^{\alpha} \qquad (3)$$

where q_p is pipe flow, k_p is the empirical conductivity parameter for pipe flow initiation (includes hydraulic conductivity of the soil matrix, internal pipe roughness and pipe tortuosity and hydraulic gradient), A is the grid cell area, w the water table height, z_p is the location of the pipe above the same datum and α the slope of the log linear regression between hydraulic head and pipe flow [*Sidle et al.*, 1995]. We assume that the outflow of the pipe within the defined end location of each pipe is equal to the pipe flow. Solute transport by pipe flow is based on the concentration of the solute in the saturated zone at the starting cell location and the simulated pipe flow. The transported mass by pipe flow for each time step is then added to the saturated zone of the cell where the pipe ends.

3.3. Parameterization

[14] A digital elevation model of the soil surface and the surface of the soil-bedrock interface were derived with a grid spacing of 2 m using the detailed survey data already available [*Woods and Rowe*, 1996]. For the Maimai hillslope, over 790 survey points and 99 soil depth measurements were available. Mean soil depth of the simulated hillslope is 0.74 m with a standard deviation of 0.29 m. The simulated part of the hillslope is 52 m long and 40 m wide (trench section 1 to 20). Initial conditions for each simulation were determined by matching the simulated and mea-

Table 1. Parameter Values in Hill-vi for All Simulations

Symbol	Parameter	Values	Source (Approximate Value)
N	porosity	0.45	difference between saturated and residual water content using water retention curve (0.4–0.45)
n_0	drainable porosity at the soil surface	0.11	in situ measurement of matric potential (tensiometer) and water content (TDR) in various depths (0.10–0.12)
b	depth scale	3.5	see above (3.0–5.0)
K_o, m h^{-1}	saturated hydraulic conductivity at the soil surface	6.0	soil core measurements in the topsoil (~6.0)
m	power law exponent	2.5	see above (2.5–3.0)
c	power law exponent (drainage)	35.0[a]	estimated from water retention curves (10.0–40.0)
E_{pot}, mm h^{-1}	potential evapotranspiration	0.24	estimation for Maimai during summer [*Seibert and McDonnell*, 2002]
z_p, m	location of the pipe above bedrock	0.05 ± 0.03[a]	pipes are mainly located within 10 cm above soil-bedrock interface. We assumed a standard deviation of 3 cm representing the natural variability.
k_p	empirical conductivity parameter	0.45[a]	fitting parameter for pipe flow initiation flow (see equation (3))
ρ, m m^{-2}	pipe density	1.0	measurements of pipe occurrence and length in the hillslope during a staining and excavation experiment
α	slope between hydraulic head and pipe flow	0.4	laboratory experiments determined a range from 0.320 to 0.424 [*Sidle et al.*, 1995].

[a]Optimized parameter within the range of experimental evidence.

sured subsurface flow prior to the rainfall event by changing uniformly the water content in the unsaturated zone.

[15] The extensive experimental research already completed at the study sites facilitated parameterization of Hill-vi. Nevertheless, some parameters could not be determined or only estimated within a possible range. We used Monte Carlo analysis to optimize these parameters (see more details in Table 1) for our focal event on 25.01.1993 using the Nash-Sutcliffe efficiency as optimization criteria. First parameters were optimized for the model without pipe flow and then with pipe flow without changing the initially calibrated parameters. The initial soil moisture content was calibrated for the focal event, but not changed thereafter. The initial water table height was set to 0.02 m above the bedrock for all events and all cells. Table 1 shows the final parameter values used in the simulations along with the data sources from the experimental studies.

4. Results

4.1. Subsurface Flow Response With and Without Pipe Flow

[16] We used the rainfall-runoff event on 25 January 1993 to calibrate Hill-vi including the pipe flow routine. This event produced 23.2 mm of runoff at the hillslope trench after 55.8 mm rainfall. The maximum rainfall intensity of a nearby tipping bucket rain gauge was 10 mm h^{-1} and generated a peak subsurface flow of 3.4 mm h^{-1}. This corresponded to specific discharge of 944 l s^{-1} km^2 and shows the extreme flashy nature of this hillslope, with very high peak discharge production. Hill-vi was not able to reproduce the measured discharge without activating pipe flow routine using a realistic range of parameters for peak runoff reproduction (Table 2 and Figure 1). The overall model performance was calculated with the Nash-Sutcliffe efficiency and the root mean square error (RMSE) (Table 2). Since the definition of the pipe flow system involved three random variables (as described in Section 3.2), 20 realizations were simulated with the same parameters but with different pipe network geometries in each realization. These ensembles were analyzed statistically (mean and coefficient of variation) and are shown as a minimum and maximum range in the Figure 1 to view the impact of the pipe flow network geometry on hillslope flow and transport.

[17] The simulated runoff from the model that included pipe flow in the model structure adequately captured the measured hydrograph response in terms of peak flow, total runoff, high efficiency and low RMSE (Table 2 and Figure 1). For this event, pipe flow contributed a significant amount of 45% to peak flow and 49% to total runoff. Hillslope runoff is plotted on a logarithmic scale in Figure 1b and shows that the simulation with pipe flow was able to capture the extreme runoff response stretching over three orders of magnitude and the following hydrograph recession.

[18] The model was validated on two other events (7 April 1993 and 24 March 1995) and on a continuous time series of several weeks (24 March 1995 to 10 May 1995). This validation procedure ensured that the model was able to simulate the hillslope response for different rainfall event characteristics and for different runoff responses. For the event on 7 April 1993, peak flow was delayed slightly and underestimated. Notwithstanding, peak flow simulations were still nearly twice as high as the simulations using the model without pipe flow. The model

Table 2. Model Calibration Event on 25 January 1993[a]

Criteria	Observations	Simulation Without Pipes	Simulation With Pipes	
			x	CV, %
RMSE, mm h^{-1}	-	0.39	0.11	4.29
Efficiency	-	0.732	0.979	0.18
Efficiency (*log*(q))	-	0.929	0.964	0.22
Peak flow, mm h^{-1}				
Total	3.43	1.98	3.09	0.97
Pipe	-	0.0	1.40	7.01
Runoff, mm				
Total	23.18	22.65	23.55	0.19
Pipe	-	0.0	11.58	7.31
Preevent water, %				
Peak flow	-	80.6	82.2	0.32
Total flow	-	83.1	85.1	0.32

[a]Duration 60 hours and 55.8 mm rainfall.

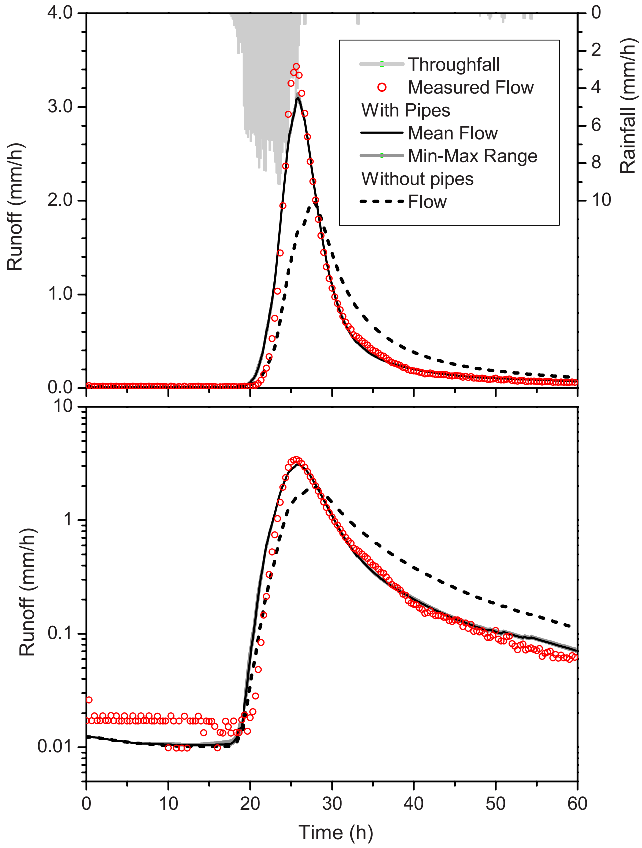

Figure 1. Comparison of measured and simulated subsurface flow for the calibrated storm event on 25 January 1993.

performance for the simulations with pipe flow was adequate (Table 3) and the overall response was well captured (Figure 2a).

[19] The simulation of the second event (24 March 1995) with 43.2 mm of rainfall produced a relatively small runoff response compared to the calibration event shown in Figure 2b. The performance results are listed in Table 4. The simulations without pipe flow again significantly under predicted peak flow and resulted in low performance measures. The simulations with pipe flow captured the runoff response in terms of overall model performance and peak response. Analysis of the simulated continuous time series (24 March 1995 to 10 May 1995) including pipe flow showed that the model performed well (RMSE = 0.12 mm h^{-1} and E = 0.77) in particular compared to the simulation without pipe flow (RMSE = 0.19 mm h^{-1} and E = 0.49). A graphical comparison (not shown due to space restrictions) also confirmed that the low flow in between events and the response to smaller events was well captured.

[20] Another interesting aspect of the model calibration and validation analysis is shown by the ensembles generated with the different pipe network geometries. In Figures 1 and 2, the range of runoff response among the ensembles was narrow, indicating insensitivity to specific pipe placement. The coefficient of variation for the model performance measures and for the total runoff and peak flow in Tables 2–4 was always very low (<5% for the model performance). The variation for the peak pipe flow and pipe flow amount was around 10%. This variation was compensated by higher matrix flow resulting in a very similar total runoff response among the realizations. These results demonstrate that runoff response within the Maimai hillslope was not dependent on any single simulated pipe network geometry.

4.2. Preevent Water Contribution With and Without Pipe Flow

[21] We applied a virtual tracer to the model rainfall with a constant concentration (i.e., the event water concentration) using the approach of *Weiler and McDonnell* [2004]. We set the concentration of the water in the saturated and unsaturated zone to zero (preevent water concentration) and simulated solute transport within the hillslope and analyzed the simulated concentration in the runoff using the standard two component hydrograph mixing analysis [*Kendall and McDonnell*, 1998]. For each simulated storm event, we calculated the preevent water contribution for the peak and total flow for the simulations with and without pipes (Tables 2–4). The simulations showed that the peak flow as well as the total flow contribution of old water was very similar between the simulation with and without pipes. For most of the realizations, preevent water contribution was in fact slightly higher for the simulations with pipe flow. These results are surprising as common (hydrological) sense would predict that more new or event water (and less preevent water) would be detected in lateral hillslope runoff where water was flowing through pipes. A more detailed analysis of the contribution of preevent water is shown for the 25 January 1993 hydrograph (Figure 3). The simulations with and without pipes were distinctively different; however, the relative preevent water contribution was quite similar for the two different simulations. The variations of the preevent water hydrograph among the realizations for the different pipe networks was very low, which was also noted in the low CV values in Tables 2–4 (in general <0.5%).

4.3. Artificial Tracer Experiment

[22] We used the results of a Maimai bromide tracer experiment on the same study hillslope of *Brammer* [1996] (24 March 1995 to 10 May 1995, coinciding with the analyzed storms in the this paper) to learn about and compare the observed and simulated subsurface tracer transport in our Hill-vi model with and without soil pipes. We used Hill-vi to replicated the *Brammer* [1996] experiment (reported by *McDonnell et al.* [1996]) at Maimai

Table 3. Model Validation Event on 7 April 1993[a]

Criteria	Observations	Simulation Without Pipes	Simulation With Pipes x	Simulation With Pipes CV, %
RMSE, mm h^{-1}	-	0.26	0.13	3.61
Efficiency	-	0.500	0.873	1.06
Efficiency (log(q))	-	0.840	0.904	0.62
Peak flow, mm h^{-1}				
Total	2.01	0.822	1.51	1.76
Pipe	-	0.0	0.79	10.34
Runoff, mm				
Total	18.1	18.18	19.08	0.21
Pipe	-	0.0	9.73	11.19
Preevent water, %				
Peak flow	-	80.1	79.6	0.43
Total flow	-	83.5	84.4	0.34

[a]Duration 100 hours and 62.6 mm rainfall.

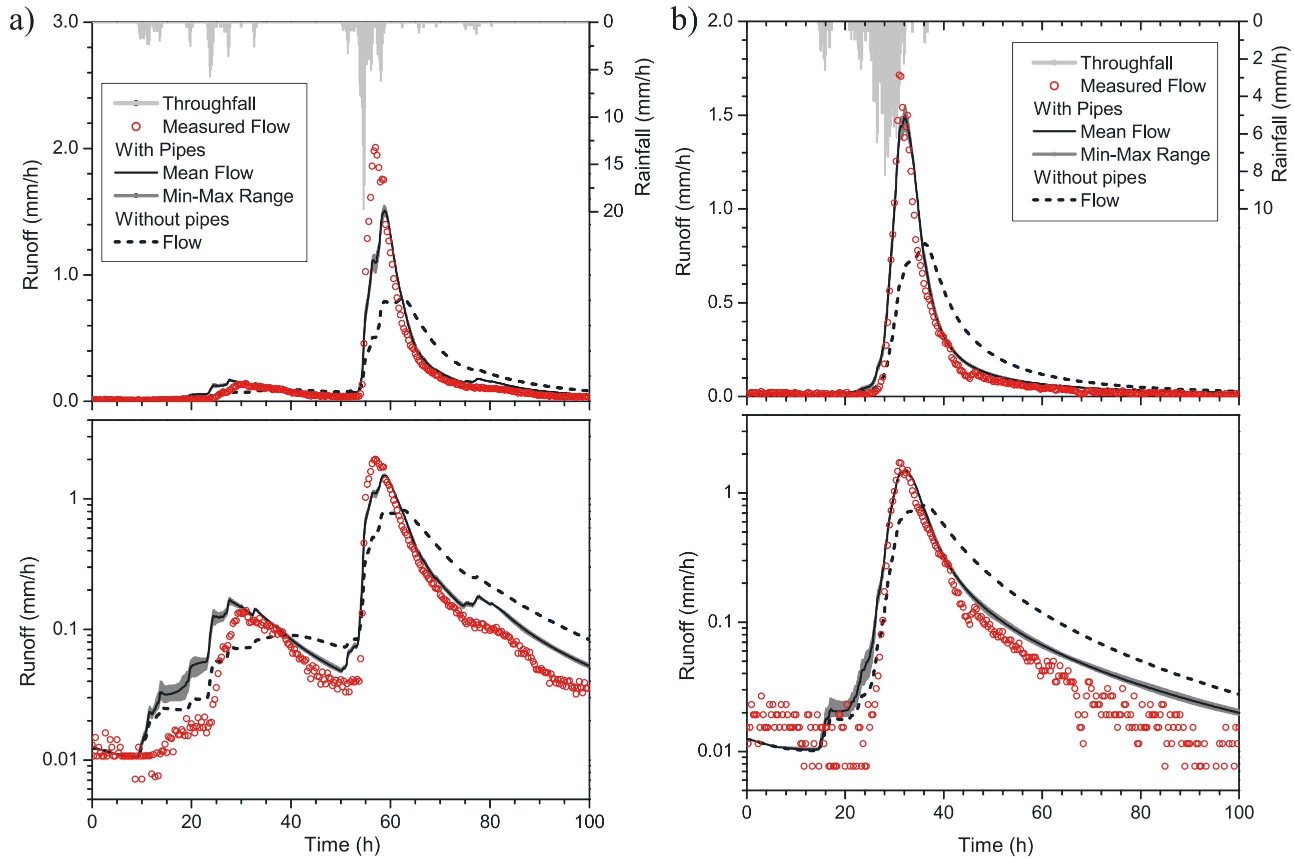

Figure 2. Model validation of measured and simulated subsurface flow for (a) the storm event on 7 April 1993 and (b) the storm event on 24 March 1995.

where he applied the bromide tracer 30 m upslope from the trench along a line and measured the bromide concentration with an ion-selective electrode for 6 weeks at the individual trench sections at irregular time intervals. We used the same application time and location and simulated tracer transport with Hill-vi for the observation period without and with pipes. The event on 24 March 1995 (see also Table 4 and Figure 2b) was the first storm after the tracer application.

[23] Simulations with pipes for the line tracer experiment showed a maximum tracer velocity (velocity of first appearance of tracer at the trench) of 0.12 cm/s and a peak tracer velocity of 0.05 cm/s. The time was calculated between the time when precipitation started and the first tracer appearance, defined as 5% of peak concentration. For the simulations without pipes, the maximum velocity was 0.06 cm/s and the peak velocity was 0.018 cm/s. These differences can also be seen in Figure 4. The measured concentrations of subsurface flow in the individual trench sections were

Table 4. Model Validation Event on 24 March 1995[a]

Criteria	Observations	Simulation Without Pipes	Simulation With Pipes	
			x	CV, %
RMSE, mm h^{-1}	-	0.19	0.06	4.04
Efficiency	-	0.596	0.960	0.34
Efficiency (*log*(q))	-	0.767	0.883	0.40
Peak flow, mm h^{-1}				
Total	1.82	0.81	1.49	2.27
Pipe	-	0.0	0.79	10.12
Runoff, mm				
Total	13.2	15.4	15.65	0.18
Pipe	-	0.0	7.96	9.78
Preevent water, %				
Peak flow	-	84.6	86.2	3.54
Total flow	-	86.2	87.9	0.38

[a]Duration 100 hours and 43.2 mm rainfall.

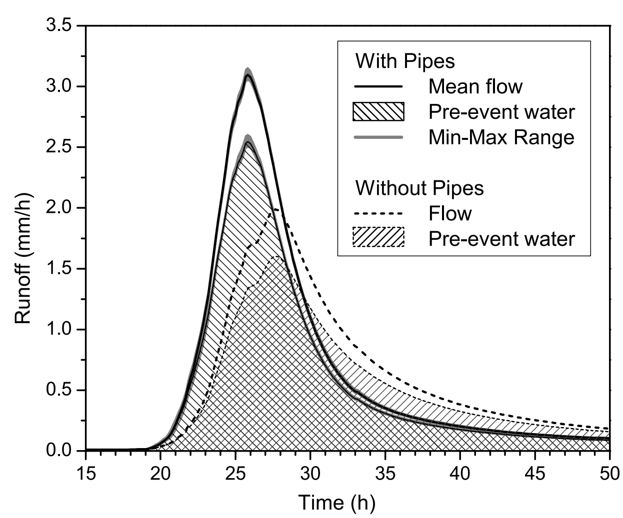

Figure 3. Simulated contribution of preevent water with and without pipes for the hydrograph of the event on 25 January 1993.

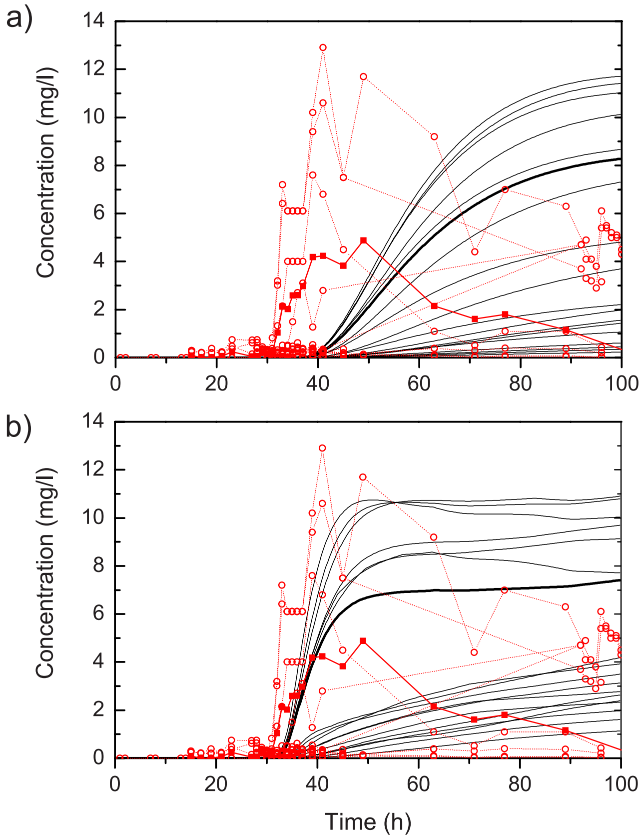

Figure 4. Measured (circles) and simulated (lines) bromide concentration at the base of the hillslope in each trench section for (a) simulation without pipes and (b) simulations with pipes. The thicker line shows the average tracer concentration simulated for the subsurface flow at the base of the hillslope. Time is given in hours after tracer application.

different from the simulated concentrations without pipes. Simulated flow velocities were too slow and concentrations among the sections were too similar. The simulated concentrations for the hillslopes with pipes reflected the observed fast response after 30 hours. The variability of concentrations among the "trench" sections was larger than for the simulations without pipes and more similar to the observed variability described by *Brammer* [1996]. For the first event, the introduction of the pipe network into the hillslope model increased the tracer velocity by a factor of two to four. However, both models were not able to simulate the decline in concentrations during the recession of the flow hydrograph.

[24] The tracer recovery for the whole period (24 March 1995 to 10 May 1995) is shown in Figure 5. The recovery after 40 days was very similar for the simulations with and without pipes. However, there were significant differences within the first few days (see enlarged inset in Figure 5). The variability among the different pipe network ensembles was large relative to the variability among the ensembles for water flow and preevent water contribution. The pipe network configuration had a much higher influence on tracer movement of a line source than on the proportion of preevent water or water flux. *McDonnell et al.* [1996] calculated a total tracer recovery of 82% for his experiment. This calculation is admittedly highly uncertain due to the intermittent measurements of bromide concentrations; however, the value is close to the simulated value of nearly 70%.

5. Discussion

[25] One of the greatest challenges to the field of hillslope hydrology is parameterizing the effects of lateral preferential flow. While our current physically based and conceptually based models ignore such behavior, incorporating it is essential if we are to move from models that mimic hydrology to ones that capture dominant flow pathways. Throughout the history of hillslope hydrological field investigations and subsequent model exercises there has been creative tension between a search for fundamental laws and generalities that are independent of place with description of the unique and complex peculiarities of a given site. The dialog between *Freeze* [1972] and *Hewlett* [1974] is a classic example of this, where Freeze questioned how subsurface stormflow could explain observed runoff dynamics in the stream channel during rainfall events based on model estimates using Darcy's law. Meanwhile Hewlett, with his knowledge of place (in his case the Coweeta Experimental Forest with the presence of macropores), could not accept the purely Darcian view applied to a homogeneous hillslope as an explanation for his observed site behavior. This paper has resumed the spirit of discussions between Freeze-Hewlett now some 30+ years on. Our experiences with both the experimental work and modeling work at Maimai suggest a need for balancing the quantitative with the qualitative approaches. Experimentalists at Maimai (and elsewhere) have almost always invoked preferential flow as a fundamental building block to their perceptual model(s) to describe the observations of flow and transport. However, many modelers at Maimai (and elsewhere) have assumed that including preferential flow into their modeling framework is either not important or impossible to parameterize. Applying Hill-vi to the Maimai hillslope and incorporating a pipe flow routine helped us to understand how randomly placed, individual pipe sections conspire with subsurface topography to produce network-like behavior at the hillslope scale. We also observed that pipe flow and pipe network presence in the model does not contradict other findings with respect to tracer and isotope mixing observations. Our work embraces the *Phillips* [2003] philosophy where we embed site specificity within a general, but malleable model structure. We suggest that this may be a new way forward for simulating the effects of lateral preferential flow on hillslope hydrology.

5.1. Predicting Subsurface Stormflow From an "Ungauged" Hillslope

[26] The classic challenge in preferential flow conceptualization and modeling is to make a nontrivial prediction of flow and transport on a "neighboring hillslope," that is one next to the heavily gauged experimental slope where the pipe flow network is described and modeled. Such a test has hitherto not been accomplished in the hillslope hydrology literature. To test the ability of the parameterized Hill-vi model to predict the response for another hillslope in the Maimai watershed, we used a second hillslope that was set up and measured by *Woods and Rowe* [1996],

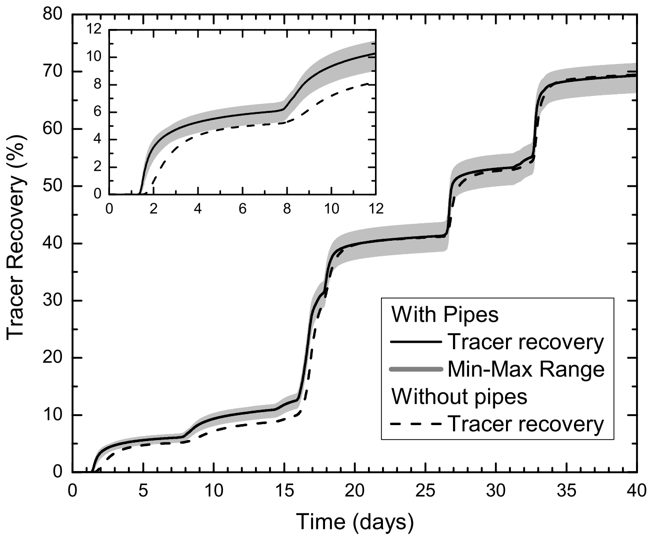

Figure 5. Tracer recovery for the line tracer experiment for the simulations with and without pipes.

corresponding to their trench section 21 to 30. This hillslope is much smaller than the one used in the preceding analyses (18 m wide and 28 m long), the soil is shallower (average depth = 0.67 m), more homogeneous (standard deviation of soil depth = 0.06 m) and the hillslope is more planar. We implemented the different topography and soil depth into Hill-vi and simulated two events without any recalibration. Figure 6 shows the results for two storm events (25 January 1993 and 7 April 1993). Compared to the same storm events at the larger hillslope, the smaller slope shows a smaller peak flow for both events per unit area that is well reproduced by the model. While our model tends to predict an earlier response and a slower recession, these tests show the power of a model that is based on combining quantitative with qualitative approaches when defining a model to simulate preferential flow processes in steep, forested hillslopes. While the exact configuration of the pipe flow network is unknown on this slope, the approach outlined in this paper provides a reasonable nontrivial approximation of flow and transport.

5.2. Resolving the Old Water Paradox

[27] What have we learned from the model vis-à-vis the old water paradox? The overall values of preevent water contribution simulated for the three different storms (84–88%) were in agreement with experimental findings of high preevent water contribution in Maimai. *Pearce et al.* [1986] found for several storms approximately 75–85% preevent water contributions in streamflow for the neighboring M6 and the M8 catchment of the simulated hillslope. For hillslope positions similar to the one simulated in our study, *Sklash et al.* [1986] found approximately 90% preevent water in collected subsurface stormflow. *McDonnell et al.* [1991b] estimated for two throughflow pits in the M8 catchment a preevent water contribution of 84% and 75% for a 45 mm rainfall event. *McGlynn et al.* [2004] presented hydrograph separations for a 70 mm rainfall event at different subcatchments in the Maimai watershed. For the central part of the studied hillslope, they estimated 93% preevent water contribution and for the small neighboring M15 catchment, they estimated a preevent water contribution of 65%. Unfortunately, no isotope data were collected and analyzed for the time series that were available for this study. However, the published results for the same hillslope and catchment or from catchments close by showed that the simulated separation is in close agreement with the experimental results used to evaluate the model.

[28] A restatement of the old water paradox (as per *Kirchner* [2003]) at the Maimai hillslope would be that while lateral subsurface stormflow responds promptly to rainfall inputs, fluctuations in passive tracers are strongly damped. This indicates that stormflow in these catchments is mostly old water. One could argue that the old water paradox is not a paradox at all but simply an expression of the different “velocities” of flow and transport [*Beven*, 1989; *Bishop et al.*, 2004]. In addition, pressure velocities may be very high relative to particle and Darcy velocities, but need specific conditions to occur [*Rasmussen et al.*, 2000].

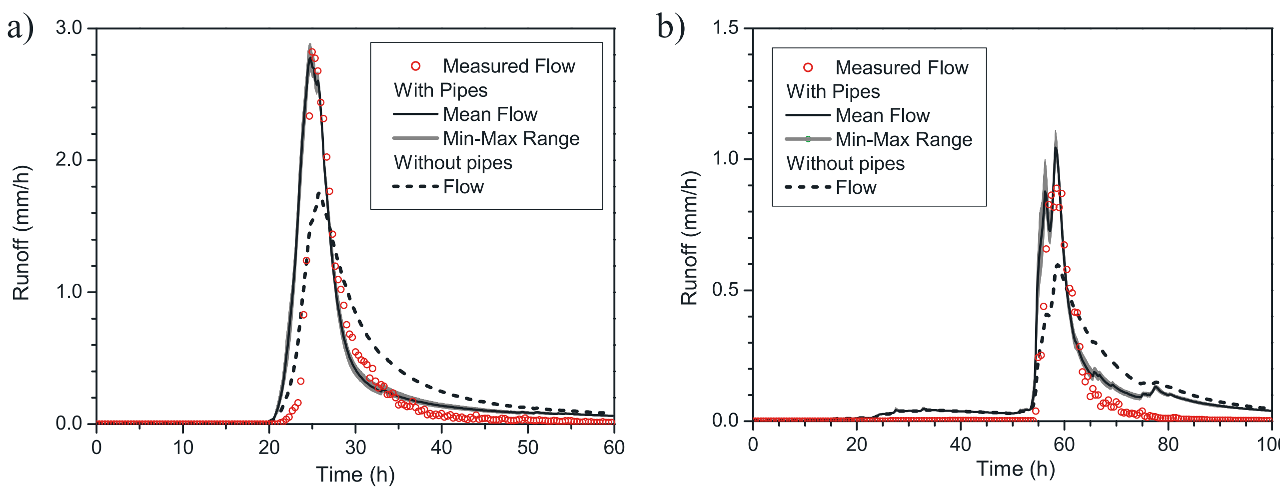

Figure 6. Uncalibrated model application to a neighboring hillslope in the Maimai watershed for (a) the storm event on 25 January 1993 and (b) the storm event on 7 April 1993.

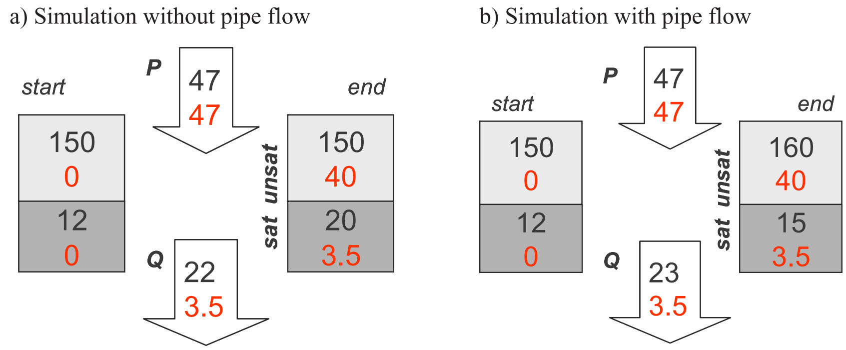

Figure 7. Comparison of sources and mixing of event and preevent water in the Maimai hillslope for the storm event on 25 January 1993. The diagram compares the total amount of water (mm) in the different stores (black values, top in box) at the beginning and the end of the storm with the values of event water (red values, bottom in box) for rainfall, runoff, and the unsaturated and saturated zone.

[29] In addition to our observed rapid hydrograph response and its domination of preevent water, artificial tracer experiments of subsurface flow at Maimai have shown very high water velocities. For example dye tracer experiments by *Mosley* [1979] in the Maimai catchment showed for several pits for a distance of 1–4 m, maximum transport velocities (first appearance of tracer) of 0.17 to 0.81 cm/s for a low intensity storm event and 1.2–2.1 cm/s for a high intensity storm event. When we compared the simulations with pipes for the hillslope with the natural events of 1995 with Mosley's results, the simulated velocities are only between 5% and 70% of the velocities Mosley observed. However, considering the total travel distance of the tracer being about 10 times the distances Mosley performed his experiments, the simulated velocities appear to be reasonably close to the observed values. In addition, since Mosley did his experiments for a very short travel distance, he probably observed the tracer velocity within one continuous pipe. The water in the simulation was routed through several pipe elements (as would be in nature) to reach the base of the hillslope.

[30] So is there a paradox concerning rapid old water effusion at the Maimai hillslope (and for that matter at the many other experimental hillslopes in the hydrological literature)? In Figure 7 we compare the sources and mixing of event and preevent water in the hillslope based on the simulation with Hill-vi for the storm event on 25 January 1993. Most important for the contribution of preevent water is the amount of water that is stored in the hillslope prior to the event that is available for mixing with the incoming event water. In the case of the Maimai hillslope, 150 mm of water that is available for transport (effective porosity) was stored in the unsaturated zone and 12 mm in the saturated zone. The rainfall changes the amount of event water in the unsaturated zone at the end of the simulation to 40 mm, which is a new water proportion of approximately 25–27%. In the unsaturated zone, the mixing ratio throughout the hillslope is strongly dependent on the actual soil depth (water storage) and maximum water table response since vertical flow and transport processes dominate. The highest proportion of new water (up to 50%) can be observed in the shallow soil close to the upper hillslope boundary. In the saturated zone, the average amount of event water at the end of the simulation is 3.5 mm for the simulation with and without pipe flow, which represents a new water proportion of 17.5% in the simulation without and 23% in the simulation with pipe flow. The spatial variability of the proportion of event water in the hillslope is strongly dependent on the flow pathways. Areas with relatively high total lateral flow (convergent areas and areas in the lower part of the slope) show a lower proportion of event water than the other areas since lateral flow processes dominate the mixing behavior.

[31] However, why do we still observe (and are able to simulate) high transport velocities for the artificial line source tracer despite the dominance of preevent water during the runoff event? If we place the results of the tracer recovery from the line tracer experiment (Figure 5) into the old water perspective at Maimai following the first event (days 0 to 5 in Figure 5), only 5.5% of the applied tracer was recovered at the base of the hillslope for the simulation with pipe flow. So, despite the fact that the tracer is transported rapidly through the pipe system in the hillslope, resulting in a rapid breakthrough and large velocities, a large proportion of the tracer remains in the soil. Part of the remaining tracer is then mobilized during each succeeding rainfall event. This behavior is in complete agreement with the observation and simulation of a high preevent water contribution for each rainfall-runoff event. The artificial tracer experiment provides results for a longer time interval than the typical event-based hydrograph separation as well as information on individual fast movement of tracer in the hillslope during a rainfall event. We observed a similar behavior when analyzing the tracer experiments at another hillslope in the H.J. Andrews research forest in Oregon, USA [*McGuire et al.*, 2007]. Both of these data sets show that high transport velocities for the artificial line source tracer can be reconciled with the dominance of preevent water during runoff events.

5.3. From Individual Soil Pipes to Hillslope-Scale Preferential Flow Networks

[32] Runoff response at the Maimai hillslope is not dependent on any single simulated pipe network geometry. This means that the unknown location of individual pipes in the hillslopes only slightly affects the runoff response. Water appears to be able to find alternative routes through the hillslope; we would speculate that in any hillslope that is favorable to the development of pipes, a potential pipe network exists that is a set of all potential pathways with

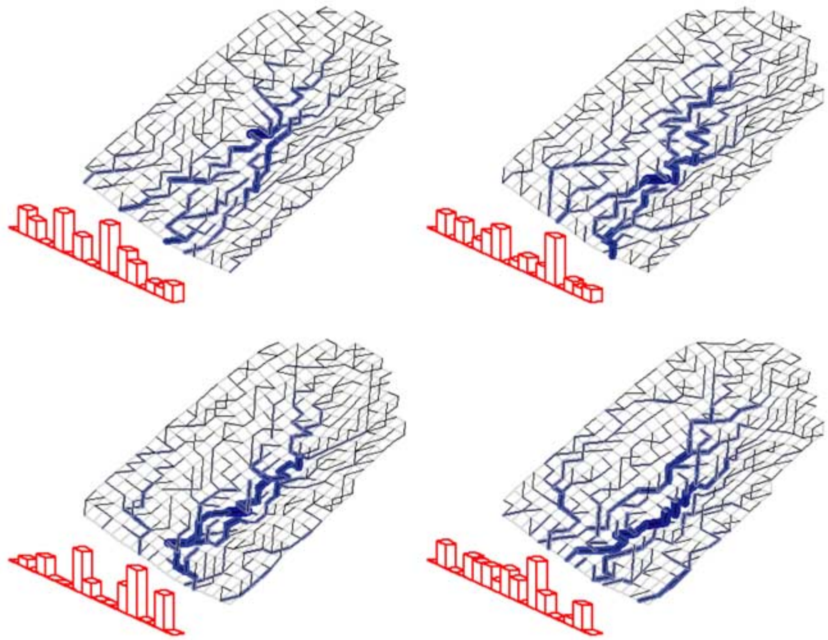

Figure 8. Four realization of the potential pipe network (black lines) and the actual pipe network (blue lines) for the storm event on 25 January 1993 at a time of 27 hours (1 hour before peak flow). The thickness of the actual blue pipes reflects the relative amount of water flowing in an individual pipe. The red bar graph at the base of the hillslope shows the relative pipe outflow from each grid cell.

low resistance. During a rainfall event, the actual pipe network is a subset of links that are wetted and flowing. This behavior is explored in Figure 8, where the model assumes a dense potential pipe network that is different for each realization (in its direction as well as in its vertical location). While Figure 8 shows only four realizations at one time step, it illustrates and compares the different actual pipe networks in relation to the different potential networks. For example, only a minor difference in the random configuration of the potential pipe network can result in a more dentritic actual pipe network (lower left) or in a more slope-parallel pipe network (upper left and lower right). The resulting outflow variation at the base of the hillslope (red bars) is different for each realization, however, as shown in the result section, the total flow and pipe flow is very similar for each realization. These relatively stable pipe flow simulations hold great promise for dealing with pipe flow and other preferential flow in hillslope modeling, since our ability to map deterministically the potential and actual pipe network in a hillslope is very limited. Currently, only destructive methods including dye tracer staining [*Anderson and Weiler*, 2005] can be used to explore a small proportion of the cumulative actual pipe network in a hillslope. Therefore a stochastic approach as we have chosen seems to be a sound way to include a potential pipe network into a hillslope model.

[33] Since the actual pipe network depends on the potential network, we also investigated the effect of the average pipe density in the hillslope for the storm event on 15 January 1993. The average pipe density is defined as the ratio of total pipe length and total area of the hillslope. The results in Figure 9 show that increasing the pipe density is related to an increase in total flow (peak flow) as well to an increase in the proportion of pipe flow. Increasing pipe density is nearly linearly related to an increase in peak flow; however, the proportion of pipe flow increases is more irregular due to the number of pipes that directly drain into the hillslope trench. Figure 9 also shows that for smaller pipe densities, the relationship between flow and pipe density becomes more hysteretic where pipe flow is higher during the rise of the hydrograph than during the recession. Overall, these network analyses suggest that the potential pipe network does not need to be known explicitly since the actual network will develop based on the path of lowest

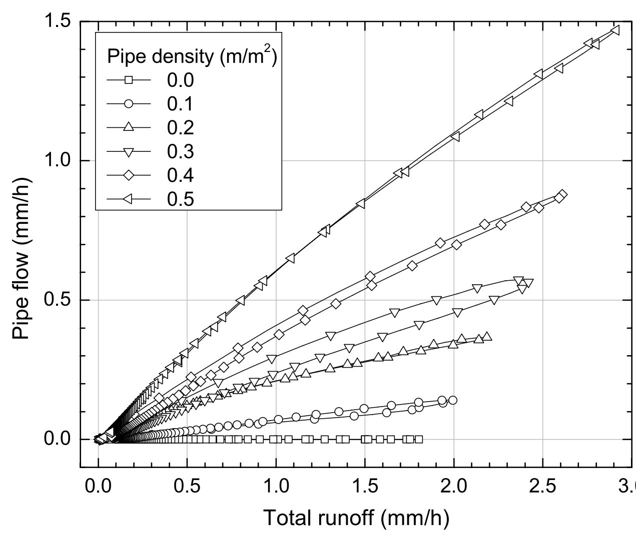

Figure 9. Relation between simulated total runoff and pipe flow for the storm event on 25 January 1993. The effect of pipe density is shown for a selection of six different densities.

resistance and consequently the topological layout of the pipe network is not important.

6. Conclusion

[34] Subsurface stormflow in steep unchanneled soil-mantled hillslopes is acknowledged to be a dominant runoff generation process in many parts of the world. While many studies have described anecdotally lateral preferential flow via soil pipes, few studies have quantified such behavior or explored explicitly how preferential flow systems connect across hillslopes and how this in turn controls water flow and solute transport (tracer breakthrough and preevent water displacement) at the hillslope scale. We modified the Hill-vi model to include a lateral pipe flow routine based on ideas distilled from the literature and our own observations. We acknowledge that these findings are only a first step toward generalizability and there is still a cost of including additional parameters that may themselves need calibration. Nevertheless, our approach represents the dialog between experimentalist and modeler for addressing questions of where to place the soil pipes in the model elements, what size they should be, how continuous they should be and how variable their characteristics should be across the slope. Our main conclusions are as follows:

[35] 1. Despite the application of randomness to generate the pipe network, model simulations showed the development of an actual network based on the path of lowest resistance that was in detail dependent on the pipe distribution, but the runoff response of the hillslope was not dependent on any single simulated pipe network geometry. Overall, the analyses suggest that the potential pipe network (topology) does not need to be known explicitly to simulate lateral preferential subsurface flow at the hillslope scale.

[36] 2. The pipe network configuration had a much higher influence on tracer movement of a line source than on the proportion of preevent water. However, simulated high transport velocities for the artificial line source tracer can be reconciled with the dominance of preevent water during runoff events. Although some tracer is transported rapidly through the pipe system in the hillslope, a large proportion of the tracer remains in the soil. The artificial tracer experiment provides results for a longer time interval than the typical event-based hydrograph separation as well as information on individual fast movement of tracer in the hillslope during a rainfall event. This point offers a solution to the so-called old water paradox [*Bishop et al.*, 2004; *Kirchner*, 2003] where preevent water dominates in most published hydrograph separations despite event-based chemical dilution.

[37] 3. Applying the model to predicting runoff at another ‘‘ungauged’’ hillslope in the Maimai watershed was promising. The chosen way of conceptualizing and parameterizing lateral preferential flow using a combination of deterministic and random elements seems to provide a good starting point, but further evaluation of this approach in different watersheds and for other hillslopes is necessary to broaden the validity of this approach.

[38] **Acknowledgments.** We thank Taro Uchida for useful discussions of his pipe flow review and Ross Woods and Lindsay Rowe for use of their flow data. We also thank the many students and post docs who have worked at Maimai and/or with these Maimai data: Dean Brammer, Brian McGlynn, Jan Seibert, Kellie Vache, Jim Freer, Taro Uchida, and Mario Martina. Finally, we thank Keith Beven for his challenge to investigate how to conceptualize and parameterize preferential flow effects.

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J. J. McDonnell, Department of Forest Engineering, Oregon State University, Corvallis, OR 97331, USA.

M. Weiler, Department of Forest Resources Management, University of British Columbia, 2037-2424 Main Mall, Vancouver, BC, Canada V6T 1Z4. (markus.weiler@ubc.ca)