Connectivity due to preferential flow controls water flow and solute transport at the hillslope scale

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Abstract: Understanding the major controls on water flow and solute transport at the hillslope scale remains a major topic of research despite numerous hillslope experiments at different sites around the world. For example, the influence of lateral preferential flow due to pipes or macropores in the subsurface flow is still unresolved. Experiments show the often paradoxical finding of fast entry of event water and fast mobilization of applied tracers with a concomitant displacement of largely "old" water from the slope base into the stream or riparian zone. Detailed investigations of the actual structures in the soil that promote and conduct lateral water transfer often show relatively short length scales. Thus, many key questions exist, including: How do the preferential flow systems connect across the hillslope during rainfall events? How does this preferential connection affect the mixing of new and old water? We use a new virtual experiment approach to identify the controls of water flow and solute transport at the hillslope scale for two well studied hillslopes in New Zealand and Japan. Previous detailed information on soil pipe and soil matrix properties at these sites provides the necessary dialog for our virtual experiments. Virtual experiments combine the knowledge of experimentalist and modeler to study the influence of pipe density, geometry and pipe length on water flow, nutrient transport, and event and pre-event water contribution. The approach is able to show how different areas within the hillslope connect and disconnect and how these episodic connections might control threshold behavior observed at these and other experimental hillslope sites around the world.

Keywords: *hillslope; conceptual model; pipe flow; hydrograph separation, tracer breakthrough*

1. INTRODUCTION

Subsurface stormflow in steep unchanneled soil-mantled hillslopes is the dominant runoff generation process in many parts of the Pacific Rim. A number of studies have demonstrated specific processes for subsurface stormflow occurrence, including transmissivity feedback, flow through the fractured bedrock, kinematic wave routing and flow through discrete preferential pathways. Perhaps the most common mechansim for rapid subsurface flow on steep, wet hillslopes is lateral preferential flow at the soil-bedrock interface (Mosley, 1979; McDonnell, 1990; Tsuboyama et al., 1994; Weiler et al., 1998; Sidle et al., 2000).

For this study, we define predominantly vertically oriented preferential pathways with lengths comparable to the soil depths as "macropores" and slope parallel preferential flow pathways as "pipes". 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 Britain and in other loess-dominated places of the world (Jones and Connelly, 2002).

Empirical studies of lateral preferential flow found in the literature show how lateral pipeflow controls hillslope response in steep upland forest environments, (Uchida et al., 1999), nutrient flushing, (Buttle et al., 2001), and old water delivery to streams (McDonnell, 1990). Notwithstanding, few process models exist that incorporate these process findings (Jones and Connelly, 2002). Indeed the dialog between experimentalist and modeler in tackling these challenges is extremely limited. The lack of knowledge about lateral preferential flow may be the largest impediment for moving forward in catchment modeling–thus, the effect of soil pipes and other structures on the lateral flow and transport at the hillslope scale is viewed as a major control still awaiting good model-process integration.

While pipeflow is highly complex and highly heterogeneous, experimental work may help to simplify many aspects of the pipeflow model problem. A literature review of pipe morphology formed mainly by root channels reveals some interesting similarities. The goal of our work should be to produce a model structure that avoids calibration–rather we wish to use the process observations to guide a physically-based model approach.

This paper explores how lateral preferential flow systems connect within and across hillslopes and how this in turn controls water flow and solute transport (tracer breakthrough and pre-event water displacement) at the hillslope scale. Two well-studied catchments are used in this study: Maimai in New Zealand and Fudoji in Japan. These two sites offer an unparalleled dataset to enhance the dialog between experimentalist and modeler owing to the multiple data sets that can be brought to bear on the problem, including physical pore pressure data, hillslope trenchflow data, solute chemistry and subsurface flow isotope data.

2. STUDY SITES

2.1. Fudoji

Fudoji is a zero-order watershed and has been a site of ongoing hillslope research since 1997 (Asano et al., 2002; Uchida et al., 2003). Fudoji is located in southeastern Shiga Prefecture, central Japan. The catchment is underlain by weathered and permeable Tanakami granite and covers an area of 0.10 ha. The mean slope in the catchment is 37º and the vegetation consists of dense natural forest, predominately *Chamaecyparis obtusa*. The mean annual precipitation and runoff was 1645 mm and 888 mm, respectively. The soils are predominantly cambisols. The average saturated hydraulic conductivities of the A and B horizons (laboratory measurements of three 100 cm^3 field cores) were 9480 and 235 mm h^{-1}, respectively (Asano et al., 2002). Water retention curves show a high drainable porosity (as defined later) between 0.2 and 0.25.

Two perennial springs contribute to "hillslope discharge" at the base of the experimental hillslope: one from the soil and the other from a crack in the bedrock. For this study focusing on hillslope hydrology, we used the discharge measurements from the soil. Soil pipe outlets with diameters ranging from 3 to 10 cm can be found at the base of the slope. In most of the hillslope area, the soil-bedrock interface is not commonly saturated between events, but most monitored storms do produce transient saturation at the soil-bedrock interface. While pre-event water dominates subsurface flow during most events, new water often forms 20-40% of the total subsurface stormflow during rainfall events (Asano et al., 2002).

2.2. Maimai

The Maimai research catchments are located on the West Coast of the South Island of New Zealand. McGlynn et al. (2002) provided a review of hydrological research at Maimai. Slopes are short (<300 m), steep (average 34º), and have local relief ranging from 100-150 m. Mean annual precipitation 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 (~17 cm) organic humus layer and 250 mm h^{-1} for the mineral soils (~70 cm). 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. Pipe flow was observed regularly along soil horizon planes and along the soil-bedrock interface in this study and in more recent research (Mosley, 1979; McDonnell, 1990; Woods and Rowe, 1996; McGlynn et al., 2002). 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. Data from one rainstorm of this study together with additional results from the studies reviewed by McGlynn et al. (2002) are used in this modeling work.

3. THEORY

3.1. HILL-VI

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 was conceptually implemented into the model HILL-VI. The basic concepts of HILL-VI are described in detail in Weiler and McDonnell (2003). 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 that two storages define the saturated and unsaturated zone for each grid cell, based on DEM and soil depth information. The unsaturated zone is defined by the depth from the soil surface to the water table and time variable water content. The saturated zone is defined by the depth of the water table to the soil-bedrock interface and the porosity n (Figure 1). 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 (Beven, 1982). The transmissivity T is then given by an exponential decline with depth of the saturated hydraulic conductivity:

$$T(z) = \int_{z(t)}^{D} K_s(z)\,dz = \frac{K_0}{f}\left[\exp(-fz) - \exp(-fD)\right] \quad (1)$$

where K_o is the saturated hydraulic conductivity and K_o is the saturated hydraulic conductivity at the soil surface, f is the decay coefficient, z is the depth into the soil profile (positive downward) and D is the total depth of the soil profile.

While these assumptions and model implementations are similar to existing models like DHSVM (Wigmosta et al., 1994) and RHESSys (Tague and Band, 2001), we introduced a new depth function for drainable porosity—a parameter shown to be a key first order process control on transient water table development (Weiler and McDonnell, 2003). The drainable porosity, as defined by the difference in volumetric water content between 0 kPa and 33 kPa soil water potential characterize the water table rise of the saturated zone. Field observations show that the drainable porosity usually declines with depth. Thus, similar to the depth function of hydraulic conductivity a depth function for drainable porosity n_d can be defined as:

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

where n_0 is the drainable porosity at the soil surface and b is a decay coefficient.

We calculate the water balance of the unsaturated zone by the precipitation input, the vertical drainage loss into the saturated zone, and the change in water content. 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. In addition, vertical macropore flow can be simulated by a power law relationship of the relative saturation multiplied by the rainfall input and the exponent β. The water balance of the saturated zone is defined by the drainage input from the unsaturated zone, the lateral inflow and outflow by lateral subsurface flow and the corresponding change of water table depth.

HILL-VI includes a solute transport routine as described by (Weiler and McDonnell, 2003). 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 within HILL-VI. We assume complete mixing and only advective transport in and between the saturated and unsaturated zone and in and between grid cells. Further details can be found in Weiler and McDonnell (2003).

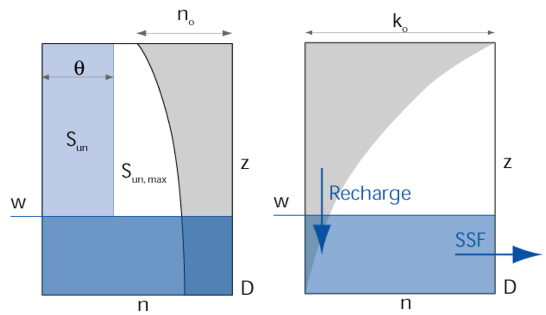

Figure 1. Schematic of the grid cell depth distribution of drainable porosity (left) and saturated hydraulic conductivity (right).

Our approach for adding lateral pipeflow to the HILL-VI structure was to first determine what common features need to be conceptualized as defined by the numerous field investigations of pipeflow in the Pacific Rim (reviewed in Uchida et al. (2001)). These include:

(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

(2) pipe length and connectivity mapping in natural slopes often shows very discontinuous pipe sections, with maximum lengths less than 2 m (Kitahara, 1994)

(3) The location of major pipes within the soil profiles is mostly within a narrow band above the soil-bedrock interface (Uchida et al., 2002)

(4) Initiation of water flow in the pipe flow is assumed to be proportional to the surrounding soil matrix saturated hydraulic conductivity and the square root of hydraulic head (transient water table depth) above the pipe (Sidle et al., 1995).

Based on 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 within and across 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. Based on the selected pipe density, grid cells within the simulated hillslope are randomly chosen and the starting location of each pipe is set. From each starting location, pipes can potentially transmit water only to neighboring cells, thus constraining the pipe length to the grid spacing. 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 (McDonnell, unpub. data; Tsuboyama, pers. com. 2000). The single, final direction of pipe flow is chosen randomly from all possible downslope direction 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 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 initiation within each grid cell is calculated by:

$$q_p(t) = k_p\, A^{0.5} \left(w(t) - z_p\right)^{0.4} \tag{3}$$

where q_p is pipe flow, k_p is the empirical conductivity parameter for pipe flow initiation, A is the grid cell area, w the water table and z_p is the location of the pipe above the same datum. We assume that the outflow of the pipe within the defined end location of each pipe is equal to the pipe flow initiation. Mass 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 is then added to the saturated zone of the cell where the pipe ends.

3.2. Parameterization

The extensive experimental research already completed at the two study sites facilitated parameterization of HILL-VI. Nevertheless, some information was not available and the parameters values were determined by Monte Carlo analysis. Table 1 shows the final parameter values used in the simulations and notes which ones were optimized.

A digital elevation model of the soil surface and the surface of the soil-bedrock interface were derived for both sites with a grid spacing of 3 m using the detailed survey data already available. For the Fudoji watershed, 57 survey points of the soil surface and 18 soil depths measurements were available. For the Maimai hillslope, over 790 survey points and 99 soil depth measurements were available. Mean soil depth at the Fudoji site 74 cm with a standard deviation of 33 cm. The Maimai site showed a quite similar mean depth of 63 cm and a standard deviation of 23 cm.

Table 1. Parameter values used for the simulation at the two sites. (*optimized values)

Parameter	Fudoji	Maimai
n (–)	0.5	0.5
n_0 (–)	0.28	0.13.
b (–)	1.5	3.5
K_o (mm h^{-1})	10.0	5.0
m (–)	0.2	0.15
c (recharge)	18	30.0
β	2.5 *	10.0 *
θ_0	0.20 (θ_r = 0.1) *	0.38 (θ_r = 0.1) *
z_p	0.055 (±0.05) >0.03 above bedrock	-0.87 ±0.05) 0.03 above bedrock
k_p	0.50 *	0.85 *

Table 2. Selected measures of the simulation results with and without pipe flow.

	Fudoji		Maimai	
Pipes	without	with	without	with
Peak flow(mm h^{-1})	0.86	1.1±0.02	0.64	2.0±0.1
Total runoff (mm)	19.5	20.8±0.14	18.5	28.1±0.2
Model Efficiency	0.47	0.68-0.76	0.17	0.70-0.76
Pre-event water (%)	59.2	60.7±0.2	81.1	84.0±0.4
Peak pre-event water (%)	48.8	46.9±0.9	77.3	80.4±0.8
Line source recovery (%)	0.52	1.2±0.13	0.06	1.35±0.53

4. RESULTS

The model was set up with the derived parameter values (Table 1) and simulations were done for each site with and without the imposed pipe system. Definition of the pipe flow system involved three random variables (as described in Sect. 3.1). 20 realizations were simulated and shown as ensembles to view the impact of the pipe flow network geometry on hillslope flow and transport (Figures 1 and 2). We selected one rainfall event for each site for a 60 h period. The observed event characteristics were 70 mm rainfall and 20 mm runoff for the Fudoji site and 55 mm rainfall and 23 mm runoff for the Maimai site (Table 2). In addition to the simulated runoff, hydrograph separation into event and pre-event water components were simulated. We viewed this as another measure of hillslope behavior given the importance of these data at each of the sites as reported by the experimental studies. A line source tracer breakthrough from a "virtual" application of a conservative tracer 30 m upslope of the trench face was also simulated.

The simulated runoff without pipe flow did not describe the measured runoff well at either site. Comparing the runoff where pipe flow was included, the simulated runoff showed much higher model efficiency and more of the observed flashy nature of runoff response at each hillslope. At the Maimai site particularly, pipe flow strongly influenced the runoff response (without pipe flow, peak runoff reduced 30% of observed). The simulated hydrograph separation was very distinct between the two sites (Figures 2 and 3). Fudoji showed a relative low pre-event water contribution of around 60%. In contrast, the total pre-event water contribution at Maimai was around 84%. These results are in agreement with experimental finding of low pre-event water (between 60-75%) at Fudoji (Asano et al., 2002) and comparatively high peak pre-event water percentages at Maimai (between 85-95%) (McDonnell et al., 1991). Simulated hydrograph separation without pipe flow revealed, surprisingly, even lower figures for pre-event water contribution to total flow.

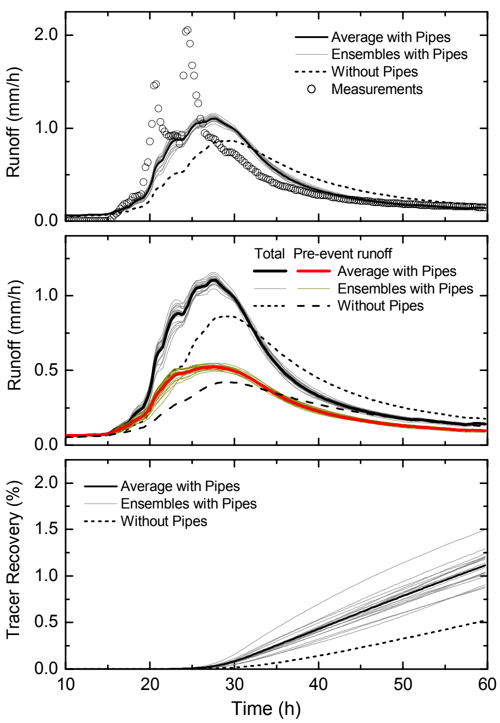

Figure 2. Runoff (top), total and pre-event water runoff (middle), and line source tracer recovery (bottom) for Fudoji.

The results of the simulated tracer breakthrough based on the applied line source into the unsaturated zone are shown as tracer recovery in Figures 2 and 3. The ensembles at both sites, but especially at the Maimai site, revealed a high variability in recovery due to the different pipe network realizations. The recovery rate at both sites for the pipe flow simulations was in the same range. Without pipe flow, tracer recovery was only around 50% at Fudoji and 5% at Maimai. Thus, pipe flow seems to have a much higher influence on tracer movement of a line source than on the bulk percentage of pre-event water. This influence is especially marked for the Maimai site, where the pipes are imbedded in a low permeable soil matrix.

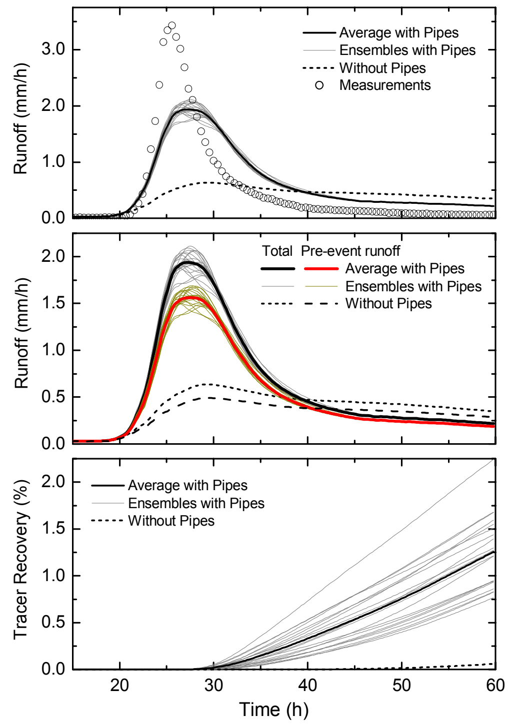

Figure 3. Runoff (top), total and pre-event water runoff (middle), and line source tracer recovery (bottom) for Maimai.

5. DISCUSSION AND CONCLUSION

While subsurface stormflow in steep unchanneled soil-mantled hillslopes is acknowledged to be the dominant runoff generation process in many parts of the world, few studies have explored how preferential flow systems connect across hillslopes and how this in turn controls water flow and solute transport (tracer breakthrough and pre-event water displacement) at the hillslope scale. We modified the HILL-VI model to include a lateral pipeflow routine based on ideas distilled from the pipeflow literature. Two well-studied catchments were used in this study for the model-experimental discussions: Maimai in New Zealand and Fudoji in Japan. These two sites offer multiple data sets that helped constrain the conceptualization and simulation of pipeflow effects on hillslope connectivity.

Our main conclusions are as follows:

1. Hillslopessoils with high drainable porosity are less influenced by lateral pipeflow since water table response to storm rainfall is lower due to high storage potential in the profile.
2. Despite the application of random pipe distributions, event runs of the model developed dendritic patterns of connected pipe flow within the hillslope during the peak of the storm rainfall event.
3. Line source tracer experiments are more influenced by the pipe network geometry than bulk hillslope scale mixing, as shown by the old-new water ratios.

This last point may be a solution to the paradox described by Kirchner (2003) where pre-event water dominates in most published hydrograph separations despite event-based chemical dilution. Drainable porosity, together with transient water table development on the hillslope (as influenced by the presence of pipes) appears to be a first order control on hillslope response.

We acknowledge several shortcomings of this preliminary work, including the fact that we do not present tracer breakthrough field data to support the model observations. Also, the work thus far is for a single event—we still do not know the effects of possible leakage to the underlying bedrock and evapotranspiration between events on event time scale dynamics. Notwithstanding, our work is proceeding to include more events and include additional experimental information from the field. These preliminary findings presented in this paper do show that connectivity due to preferential flow does control water flow and solute transport at the hillslope scale. Other hillslope juxtapositions of drainable porosity and pipe configurations are needed to define the potential extent of this control in catchments generally.

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