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# A virtual experiment on the effects of evaporation and intensity smoothing by canopy interception on subsurface stormflow generation

**R.F. Keim [a,*], H.J. Tromp-van Meerveld [b], J.J. McDonnell [c]**

[a] *School of Renewable Natural Resources, Louisiana State University and LSU Agricultural Center, 227 Renewable Natural Resources Building, Baton Rouge, LA 70803, USA*
[b] *School of Architecture, Civil and Environmental Engineering, Ecole Polytechnique Fédérale de Lausanne, Lausanne 1015, Switzerland*
[c] *Department of Forest Engineering, Oregon State University Corvallis, OR 97331, USA*

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**Summary** The effects of canopy evaporation and intensity smoothing during rain events on hillslope subsurface stormflow are poorly understood. While watershed manipulation experiments have suggested that these processes are important at long timescales, such processes may also be important in storm-timescale responses. Notwithstanding, there are few hillslopes for which both internal subsurface stormflow generation processes and canopy processes are known, so canopy interception effects on subsurface stormflow have not been tested mechanistically. Furthermore, it has not yet been possible to separate the effects of canopy evaporation from intensity smoothing in terms of which component of interception most affects hillslope response. We report a series of virtual experiments (numerical experiments driven by collective field intelligence) using HYDRUS-2D to model flow in a well-studied and characterized research hillslope in Georgia, USA. Previous work has shown that HYDRUS-2D approximates well both measured subsurface stormflow and internal pore pressures at this site. Our virtual experiments compared modeled hillslope response to rainfall and throughfall characteristic of known forest canopy processes in Washington, USA. The experiments generated subsurface stormflow using measured rainfall and throughfall data from three sites within the forest, and using synthetic, simplified throughfall signals containing either evaporation alone or intensity smoothing alone. As expected, results of our virtual experiments driven by field-measured throughfall data showed that evaporative loss delayed the onset of subsurface stormflow, lowered and delayed

* Corresponding author. Fax: +1 225 578 4227.
*E-mail address:* rkeim@lsu.edu (R.F. Keim).


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stormflow peaks, and decreased total flow and the runoff ratio. Virtual experiments based on simplified modeled throughfall (where we separated evaporation from intensity smoothing) showed that canopy evaporation was responsible for most of these effects, while intensity smoothing showed measurable differences only in peak subsurface stormflow rate. Overall, this work has implications for the calibration of watershed models. Not only would ignoring interception miss a major effect of vegetation on subsurface stormflow generation, our work also shows that simply applying some fractional reduction as a scaled input signal (as is customary in watershed modeling studies) may mask important effects on peak flow response in some situations.


## Introduction

Much of the focus on hillslope hydrology is to understand the flowpaths and dynamics of subsurface stormflow (Weiler et al., 2005). While much work has been done to quantify how flow generation depends on soils, geology, and hillslope geometry (e.g., Bonell, 1998; Reggiani et al., 2000; Troch et al., 2003; Buttle et al., 2004), less attention has been given to how vegetation can affect the response of hillslopes to precipitation (Savenije, 2004). Despite the early integration of vegetation into hydrological modeling (Eagleson, 1978), and despite recent interest in ecohydrology (e.g., Rodriguez-Iturbe, 2000; Gurnell et al., 2000; Zalewski, 2000; Kundzewicz, 2002; Bond, 2003; Pringle, 2003; Baird et al., 2004), hillslope hydrology has not fully addressed the effects of vegetation on hillslope hydrological processes. In particular, effects on subsurface stormflow generation are poorly understood.

One important role of vegetation on hillslope hydrology and subsurface stormflow generation is its influence on precipitation inputs to the soil. Evaporation of water intercepted by the canopy can account for up to about half of annual precipitation and 20–60% of total evapotranspiration from forests (e.g., Swift et al., 1975; Shuttleworth, 1988; Calder, 1990; Viville et al., 1993; Dubé et al., 1995; Fujieda et al., 1997). Additionally, transfer of water through the canopy smoothes the intensity of precipitation reaching the soil surface (Trimble and Weitzman, 1954; Keim and Skaugset, 2004). While the link between canopy interception and the volume of runoff over long timescales has been established for some time (e.g., Helvey, 1967), little is understood about how canopy interception might modify the volume and timing of subsurface stormflow during individual storms. In most instances, when forest cover has been reduced or removed in paired watershed studies, peak instantaneous flow rates have increased during small storms (e.g., Thomas and Megahan, 1998; Beschta et al., 2000; Jones and Grant, 2001; DeWalle, 2003), but the timing of peak flows has usually remained unchanged (Harr and McCorison, 1979; Ziemer, 1981; Swank et al., 1988). Unfortunately, paired watershed observations are of limited utility for inferring the role of vegetation in stormflow generation. In part, this is because their black box nature does not illuminate the multiple processes that interact to control watershed response to precipitation, and explanations of storm-scale effects of canopy interception (e.g., Jones, 2000) remain in the realm of hypothesis. Also, paired watershed studies do not always show increased peak flows when vegetation is removed, and the role of vegetation in complex responses is not always clear (Harr, 1979; Thomas and Megahan, 1998). Process-based hillslope studies at smaller spatial and temporal scales (e.g., McDonnell et al., 1996; Kendall et al., 2001; Freer et al., 2002) have generally not considered canopy interception effects.

Designing and implementing an experiment to explicitly measure effects of canopies on hillslope hydrology would be difficult, especially given that evaporation and intensity smoothing occur simultaneously. Therefore, it is not surprising that no studies to date have fully coupled detailed measurements of these processes both separately and in concert. While such an experimental coupling is a laudable goal, it will nonetheless remain difficult to isolate experimentally the effects of these individual processes. This is especially true in the case of explicit coupling of throughfall and subsurface stormflow in complex natural hillslopes where spatial and temporal variability in, for example, rooting, soil properties, and throughfall conspire to generate complexity in flowpaths and hillslope responses.

This research employs the virtual experiment approach (Weiler and McDonnell, 2004) to estimate the effects of canopy interception on subsurface stormflow generation at the hillslope scale and avoid the complexities of field investigations. The virtual experiment, defined as a numerical experiment driven by collective field intelligence, is a learning tool to help identify the most important components of complex processes. Virtual experiments allow investigation of single processes operating within complex natural systems in which the responses of models to varying internal states and boundary conditions provide the basis for inferences about controlling processes and likely results of varying natural conditions. Although field experiments will continue to be important in the development of new understanding and serve as the final tests of conclusions reached by modeling, virtual experiments can be a useful additional tool to develop new understanding. Examples, of this approach include recent work by Bowling et al. (2000), Alila and Beckers (2001), Whitaker et al. (2002), and Keim and Skaugset (2003).

To investigate likely effects of throughfall on subsurface stormflow generation, we calibrated a Richards-equation-based finite element model (HYDRUS-2D of Šimůnek et al., 1999) of a hillslope transect to measured subsurface stormflow at the slope base and to measured internal hillslope pore pressures. We then used field measurements of rainfall and throughfall from a separate field site as input to the hillslope model to estimate the overall effects of interception on subsurface stormflow generation and synthetic throughfall data to estimate the separate effects of canopy evaporation and intensity smoothing.

The virtual experiment approach allows combining data from the two distinctly different field sites. This was necessary because data do not exist at the necessary level of detail at any single field site. The hillslope model was based on the Panola Mountain trenched hillslope. The Panola hillslope has been monitored since 1995 and is instrumented with more than 100 wells and 64 soil moisture measurement stations. Subsurface stormflow has been recorded for more than 150 rainfall events. Previous investigations of subsurface stormflow response at the Panola hillslope research site have examined how soil depth influences the spatial pattern of subsurface stormflow (McDonnell et al., 1996; Freer et al., 2002), the nature of 1-D wetting in the soil profile (McIntosh et al., 1999), solute flushing processes (Burns et al., 1998), threshold response characteristics between rainfall and subsurface stormflow (Tromp-van Meerveld and McDonnell, in press a), and the spatial pattern of transient saturation during storms (Tromp-van Meerveld and McDonnell, in press b). The Panola rainfall and throughfall data were not sufficient for the objectives of this work. Instead, we used data of rainfall and throughfall from Washington, USA, that was originally used for developing models of precipitation transfer through forest canopies at high temporal resolution (Keim and Skaugset, 2004; Keim et al., 2004). By using the virtual experiment approach, we were able to combine the best available data on both rainfall/throughfall and high-resolution hillslope hydrology. The resulting hypothetical setting and juxtaposition allows testing of hypotheses to elucidate coupled canopy-hillslope processes in a general way, without specific meaning for any particular study site.

This work is part of a dialog between experimentalists and modelers whereby visualization of processes internal to well-characterized hillslopes are examined in the context of gross interception and isolated canopy evaporation and canopy smoothing effects on subsurface stormflow. We test the null hypothesis that there is no difference in hillslope response between open rainfall and throughfall. Specifically, we examine how gross interception and separated canopy evaporation and intensity smoothing affect the onset of flow, peak flow response, total runoff amount, runoff ratios, subsurface stormflow recession, and peak pore pressures.

## Methods

### Hillslope model

#### Site description

The Panola Mountain Research Watershed and intensively studied hillslope are located 25 km south of Atlanta, Georgia, USA. The instrumented hillslope is 20 m wide and 48 m long, with an average slope of 13°. The upslope boundary is a small bedrock outcrop, and the lower boundary is a trench dug to bedrock. This trench is divided into 10 2-m wide sections, and discharge from each section (plus five soil pipes) is routed into tipping-bucket gauges and recorded once per minute (McDonnell et al., 1996; Freer et al., 1997, 2002; Burns et al., 1998). Surface topography of the instrumented hillslope is relatively planar, but the bedrock surface topography is highly irregular. The soil on the study hillslope is a light colored sandy loam without clear structuring or layering, except for a $\sim$0.15 m surface organic horizon. There are no observable differences in soil type or texture across the hillslope.

The soils on hillslope positions like the one modeled in this study are of the Ashlar-Wake mapping unit – a multitaxonomic complex composed of mixed, thermic Lithic Udipsamments from the Wake Series and coarse, loamy, mixed thermic Typic Dystrochrepts from the Ashlar series (Zumbuhl, 1998). These soils are hillslope sediments or colluvium from upslope erosional processes. Our specific study hillslope is exclusively the coarse, loamy, mixed thermic Typic Dystrochrepts from the Ashlar series. Soil depth, measured using a 25.4 mm soil corer and small hand auger on a $2 \times 2$ m grid, ranges from 0 to 1.8 m (average 0.63 m) (Freer et al., 2002). Soils are underlain by Panola Granite. A grid of 44 co-located recording tensiometers was installed primarily in the lower half of the hillslope in 1996 (Freer et al., 2002). The average depth of the shallow and deep tensiometers is 0.20 and 0.62 m below the soil surface, respectively.

#### Process knowledge at the Panola hillslope

Panola, like many other well-studied, trenched hillslopes, shows evidence of threshold behavior, in which subsurface stormflow occurs only after perched water tables develop in bedrock depressions at the soil-bedrock interface then connect laterally and feed downslope flow. Recent work by Tromp-van Meerveld and McDonnell (in press a, in press b) has shown that subsurface saturated connectivity and measurable subsurface stormflow at the trench face begins when rainfall exceeds a threshold of about 55 mm. After rainfall, runoff generally ceases within 48 h (Freer et al., 2002; Tromp-van Meerveld, 2004). During winter, when evapotranspiration is low, the period between storms (on average 6 days) is long compared to drainage of the hillslope, such that the hillslope is essentially reset to field capacity before each storm. In this condition, pre-event soil moisture is spatially uniform with depth and across the hillslope (Tromp-van Meerveld and McDonnell, 2005).

#### Model description

HYDRUS-2D (Šimůnek et al., 1999) is a two-dimensional finite element model that simulates movement of water in variably saturated porous media by numerically solving the Richards equation. Especially important for this study is that HYDRUS-2D can model flow regions delineated by irregular boundaries and can accommodate a seepage face boundary through which water leaves the saturated part of the flow domain (the trench face). Tromp-van Meerveld (2004) successfully applied HYDRUS-2D to simulate pore pressures and subsurface stormflow at the Panola hillslope by incorporating measured surface and bedrock topography. In the present study, we modeled saturated and unsaturated flow within the hillslope using HYDRUS-2D. For the two-dimensional modeling, we selected a single longitudinal slice of the hillslope where there were no soil pipes at the outlet and matrix flow dominated discharge. This longitudinal slice intersected the location of nine tensiometers that we used to check internal consistency of the model (Fig. 1). Soil depth in the slice varied from 0.3 to 1.0 m (average 0.71 m).

### Model calibration

We used the measured surface and bedrock topography to generate the finite element mesh. We used a seepage face boundary for the trench face, a zero-flux boundary for the upslope boundary of the hillslope, a free drainage boundary for the lower boundary (deep bedrock), and an atmospheric boundary for the surface (Fig. 1a). The hillslope was divided into five parallel layers (Fig. 1b). The top three layers represented the soil, chosen to represent the observed decline in hydraulic conductivity and drainable porosity with depth. The lower boundary of the third soil layer was the measured bedrock topography. The lower two layers below the soil layers were included to represent seepage into low-permeability materials beneath the soil profile (saprolite and bedrock). All parameters for the soil layers, except saturated hydraulic conductivity, were estimated from the soil moisture release curves from soil cores extracted from the hillslope and measured on a tension table. The saturated hydraulic conductivity of each soil layer and all parameters for the bedrock layers were calibrated by manually minimizing differences between observed and predicted subsurface stormflow and pore pressures (at the location of the nine tensiometers) for the storm of 6–7 March 1996 (which has also been described by McDonnell et al. (1996) and Freer et al. (1997, 2002)). The calibrated saturated conductivity of the three soil layers was additionally constrained to be within one order of magnitude of the measured 0.64 m h$^{-1}$

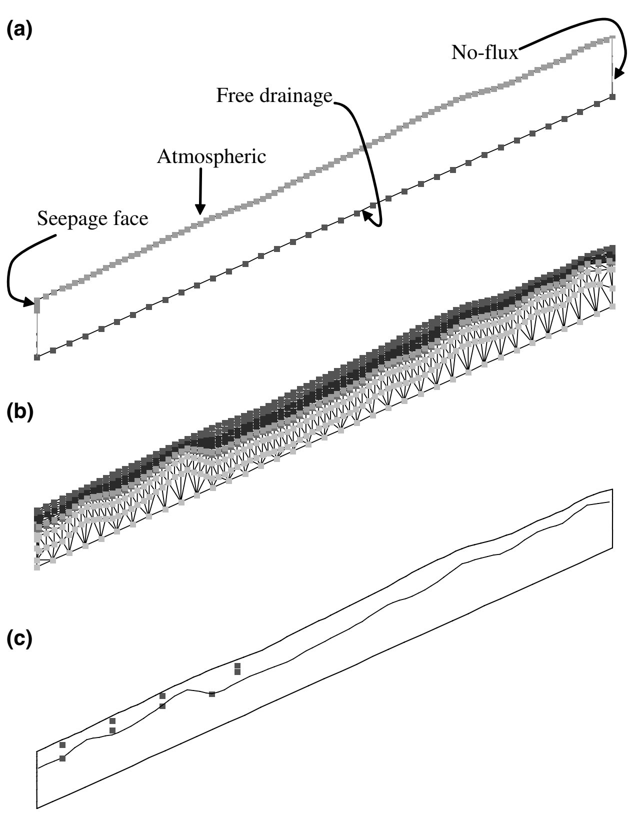


**Figure 1** Schematic representation of the boundary conditions (a), the soil layers (b) and the location of the tensiometers (squares) and soil bedrock interface (line) of the modeled 2-D hillslope transect (c). Vertical exaggeration is 2x.

vertical saturated conductivity in a large intact soil core extracted by McIntosh et al. (1999).

The aim of the calibration was to obtain a reasonable and plausible model for the virtual experiments, not to obtain a model that could reproduce the subsurface stormflow processes at Panola exactly. The calibration storm of 6–7 March 1996 consisted of two component storms: the 49 mm storm on 6 March, which produced little subsurface stormflow (0.4 mm at the 20 m trench), and the 47 mm storm on 7 March, which produced significant subsurface stormflow (24 mm at the 20 mm trench). This storm was chosen because both subsurface stormflow and pore pressure data were available and because we expected that simultaneous calibration to both relatively dry (6 March) and wet (7 March) initial conditions would result in a robust set of model parameters. To set the initial conditions for the hillslope for calibration and simulation, we set the entire modeled slope to an initial pore pressure of −0.5 m and modeled drainage for 24 h before onset of the 6 March storm. As in the field, therefore, the initial state of the hillslope was such that runoff had ceased and pore pressures were nearly at a steady state (Fig. 2).

Rainfall data for the calibration simulation were available as rainfall measured in a tipping-bucket gauge in a nearby opening. However, because the entire hillslope is forested, we used rainfall data modified for estimated canopy interception as input to the hillslope model. The site lacked sufficient instrumentation to model interception physically (e.g., by Rutter et al., 1971); however, Kendall (1993) estimated 1 mm canopy storage and 5% interception loss for winter rainfall at the study site. Therefore, we modified rainfall data to estimate throughfall by applying these values to the 6–7 March storm. We accomplished this by (1) assuming the first 1 mm of rain all stayed in the canopy and (2) choosing a constant assumed rate of evaporation during rainfall (0.56 mm h$^{-1}$) so that total throughfall was 95% of total rainfall. We did not modify the input data for canopy smoothing because the model was run with hourly data for the calibration storm. We modeled precipitation as a spatially uniform input across the hillslope transect in all simulations.

## Throughfall data for model simulations

### Field data

We collected precipitation intensity data in and adjacent to a forest stand in the Cedar Flats Research Natural Area of the Gifford Pinchot National Forest, Cascade Mountains, southwestern Washington, USA. The forest on the site originated about 600 years ago; overstory trees are Douglas-fir (*Pseudotsuga menziesii*), western redcedar (*Thuja plicata*), and western hemlock (*Tsuga heterophylla*) up to 3.1 m diameter and 84 m tall. The canopy is structurally complex because of its age. Senescent large trees have created gaps, and the spatial relationships between canopies of very large trees and younger individuals of shade-tolerant species result in high spatial variability of canopy cover that is reflected in spatial variability of throughfall amount and intensity (Keim et al., 2004; Keim et al., 2005).

The site was equipped with a tipping-bucket raingauge in a large nearby (∼200 m away) opening, and seven under-

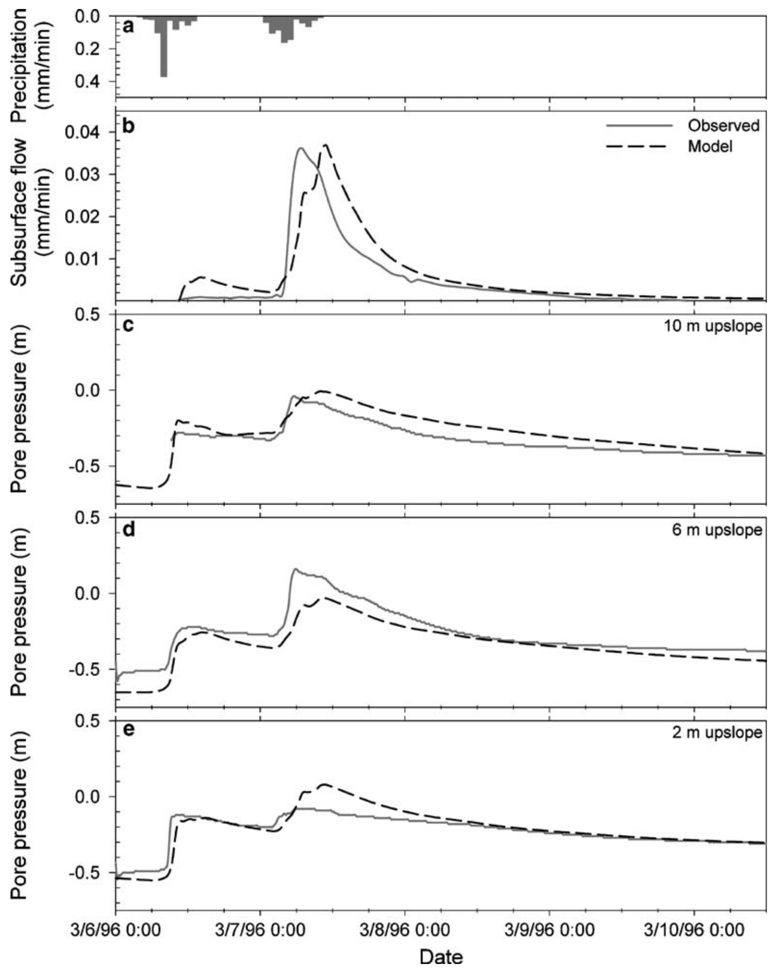


**Figure 2** Precipitation (a), observed and modeled subsurface stormflow (b) and pore pressure at the deep tensiometers located 10 m (c), 6 m (d) and 2 m upslope from the trench face (e). The depth of the tensiometer at 2, 6 and 10 m upslope from the trench face was 0.58, 0.53, and 0.53 m below the soil surface, respectively.

canopy intensity rain gauges. The tipping-bucket rain gauges were calibrated to tip once per 0.254 mm of rainfall. The under-canopy gauges consisted of similar tipping-bucket gauges augmented with two troughs to increase the spatial extent of sampling. The troughs were plastic pipes with a slot cut in each that was 0.02 m wide and 2 m long; each pipe was set at a 22.5° angle above horizontal (slightly less than the funnel in a standard rain gauge) so that travel time to the tipping-bucket was minimized and correct throughfall intensity data were recorded. The area of the two troughs projected to the horizontal exactly doubled the catch area of the gauge, so that each tip recorded 0.127 mm of throughfall.

We modeled hillslope responses to rainfall and throughfall input from a 49 mm rainstorm on 22 August 2001 (comparable in total rainfall to our 6 March storm from the Panola hillslope calibration). We used throughfall data from the three under-canopy gauges with the most contrasting canopy conditions to obtain the greatest range of inputs and modeled hillslope subsurface stormflow responses (Table 1). The model was run with 1-min precipitation and throughfall data.

**Table 1** Matrix of the virtual experiments

| Intensity smoothing | Evaporation | | | |
|---|---|---|---|---|
| | None | Low | Medium | High |
| None | Open | *E = 10%*[a] | *E = 25%* | *E = 50%* |
| Low | *RT = 3 min*[b] | | T4<br>E = 16%<br>RT = 1 min | |
| Medium | *RT = 10 min* | T2<br>E = 2%<br>RT = 6 min | | |
| High | *RT = 25 min* | | T5<br>E = 25%<br>RT = 18 min | |

Experiments based on synthetic throughfall data are italicized.
[a] E, evaporation (% of gross rainfall).
[b] RT, residence time of water in the canopy (a measure of intensity smoothing).

**Synthetic data**

In addition to the measured throughfall data, we generated idealized, synthetic sets of throughfall data to allow independent testing of the effects of evaporative loss and intensity smoothing on hillslope subsurface stormflow response (Table 1). These synthetic data were the result of transform-

**Table 2** Summary of precipitation and observed and modeled subsurface stormflow for the March 6–7 1996 calibration storm

| | Observed | | | Modeled | | |
|---|---|---|---|---|---|---|
| | 6 Mar | 7 Mar | 6–7 Mar | 6 Mar | 7 Mar | 6–7 Mar |
| *Input variables* | | | | | | |
| Total precipitation (mm) | 49 | 47 | 96 | | | |
| Total throughfall (mm) | nd[a] | nd | nd | 45 | 42 | 86 |
| *Output variables* | | | | | | |
| Total subsurface stormflow (mm) | 1.1 | 26.7 | 27.8 | 4.7 | 32.5 | 37.2 |
| Runoff ratio (%) | 2.5 | 63.5 | 32.0 | 10.5 | 77.5 | 42.9 |
| Peak subsurface stormflow rate (mm $h^{-1}$) | 0.06 | 2.17 | 2.17 | 0.34 | 2.21 | 2.21 |
| Peak pore pressure at 10 m upslope (m) | −0.28 | 0.05 | 0.05 | −0.20 | −0.01 | −0.01 |
| Peak pore pressure at 6 m upslope (m) | −0.22 | 0.16 | 0.16 | −0.26 | −0.03 | −0.03 |

[a] nd, no data.

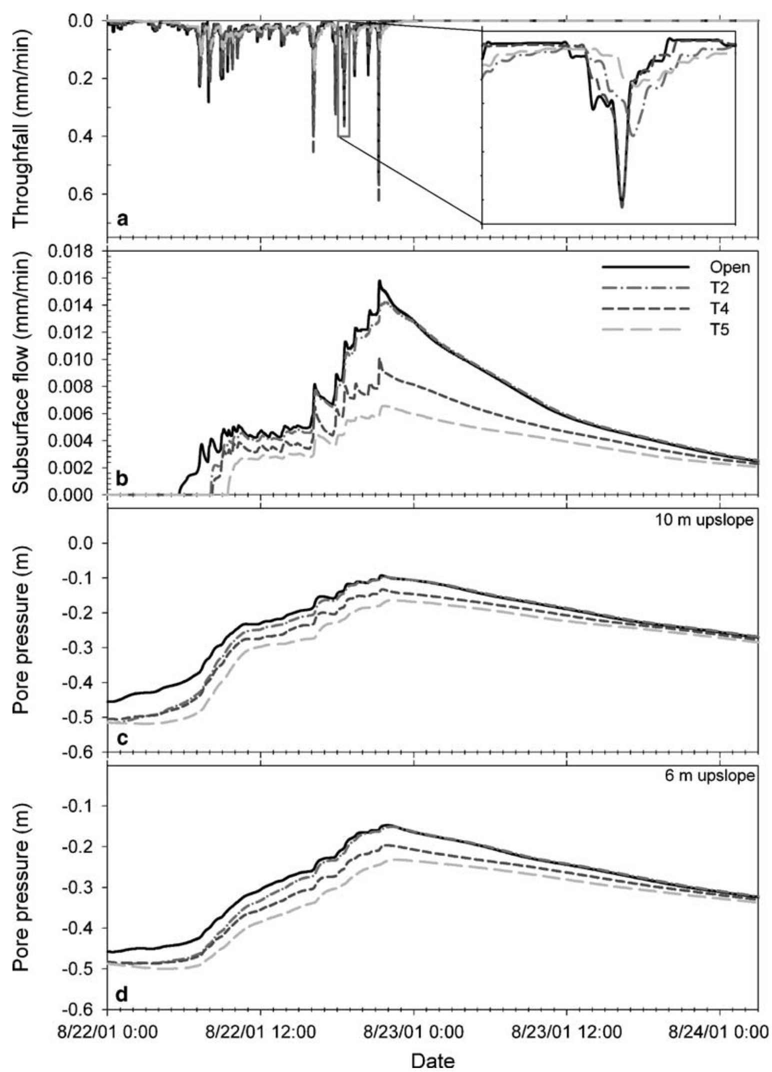


**Figure 3** Observed precipitation and throughfall (a), modeled subsurface stormflow (b) and modeled pore pressure at 10 m (c) and 6 m upslope from the trench (d) for the 2-D transect model for measured precipitation and throughfall inputs. The inset in panel (a) shows more detail for the period 18:00–19:00, 22 Aug 2001.

ing rainfall data according to either a constant-evaporation transformation with no intensity smoothing or an intensity-smoothing transformation with no evaporative loss.

The constant-evaporation transformation was:

$$F(t) = \begin{cases} R(t) - E, & R(t) > E \\ 0, & R(t) \leqslant E \end{cases} \tag{1}$$

where $F(t)$ [mm $h^{-1}$] is the synthetic time series of throughfall, $R(t)$ [mm $h^{-1}$] is the measured time-varying rainfall, and $E$ [mm $h^{-1}$] is a constant effective evaporative loss. We generated synthetic throughfall series $F(t)$ using three values of $E$: 0.15, 0.46, and 1.26 mm $h^{-1}$, which yielded storm-total evaporative losses of 10%, 25%, and 50%, respectively. By comparison, evaporation at the three throughfall observation points was 2%, 16%, and 24% of rainfall (Table 1).

The intensity-smoothing transformation followed the work of Keim and Skaugset (2004), who used a linear system to model smoothing of intensities from rainfall to throughfall:

$$F(t) = \int_0^t R(\tau) g(t - \tau) \mathrm{d}\tau, \tag{2}$$

where $g$ is the smoothing function defining the unit response of $F(t)$ at time shifts $\tau$ after unit input $R(t)$. This method treats the canopy as a watershed using the unit hydrograph approach to streamflow modeling (Dooge, 1959, 1973). Following Keim et al. (2004), we used an exponential form of $g$ to describe smoothing:

$$g(t) = \frac{1}{a} e^{-\frac{t}{a}}, \tag{3}$$

in which $a$ is a parameter equal to the mean residence time of precipitation in the canopy before falling as throughfall. We generated synthetic throughfall series $F(t)$ using three values of $a$: 3, 10, and 25 min. By comparison, mean residence times at the three throughfall observation points during the 22 August 2001 storm were 6, 1, and 18 min (Table 1).

## Results

### Model calibration to observations

The calibrated model reproduced many of the behaviors of subsurface stormflow and internal pore pressures observed during the 6–7 March 1996 storm (Fig. 2). Subsurface flow

**Table 3** Summary of precipitation and modeled subsurface stormflow for the 10 virtual experiments

| Observed throughfall | Open | T2 | T4 | T5 |
|---|---|---|---|---|
| *Input variables* | | | | |
| Total $P$ (mm) | 48.8 | 47.8 | 41.2 | 36.8 |
| Total evaporation (mm) | 0 | 1.0 | 7.6 | 12.0 |
| Evaporation (%) | 0 | 2 | 16 | 25 |
| Mean residence time (min) | 0 | 6 | 1 | 18 |
| *Output variables* | | | | |
| Total $Q$ (mm) | 18.9 | 18.3 | 13.4 | 10.6 |
| Runoff ratio (%) | 39 | 38 | 33 | 29 |
| Peak subsurface flow rate (mm/min) | 0.016 | 0.014 | 0.010 | 0.007 |
| Peak pore pressure 10 m upslope (m) | −0.09 | −0.10 | −0.13 | −0.16 |
| Synthetic: evaporation only | Percent evaporation: | 10% | 25% | 50% |
| *Input variables* | | | | |
| Total $P$ (mm) | | 44.4 | 37.0 | 24.7 |
| Total evaporation (mm) | | 4.4 | 11.8 | 24.1 |
| Mean residence time (min) | | 0 | 0 | 0 |
| *Output variables* | | | | |
| Total $Q$ (mm) | | 15.7 | 10.8 | 5.0 |
| Runoff ratio (%) | | 35 | 29 | 20 |
| Peak subsurface flow rate (mm/min) | | 0.012 | 0.008 | 0.005 |
| Peak pore pressure 10 m upslope (m) | | −0.12 | −0.16 | −0.24 |
| Synthetic: smoothing only | Residence time (min): | 3 | 10 | 25 |
| *Input variables* | | | | |
| Total $P$ (mm) | | 48.8 | 48.8 | 48.8 |
| Total evaporation (mm) | | 0 | 0 | 0 |
| Mean residence time (min) | | 3 | 10 | 25 |
| *Output variables* | | | | |
| Total $Q$ (mm) | | 18.8 | 18.9 | 18.9 |
| Runoff ratio (%) | | 39 | 39 | 39 |
| Peak subsurface flow rate (mm/min) | | 0.015 | 0.015 | 0.015 |
| Peak pore pressure 10 m upslope (m) | | −0.09 | −0.09 | −0.10 |

in response to the 49 mm storm of 6 March was overestimated by 3.6 mm, but modeled subsurface stormflow in response to the 7 March storm was well predicted (Table 2). Modeled peak instantaneous flow matched the observed, but was delayed by 3 h (Fig. 2b) because of a slower modeled rise in the subsurface stormflow rate. The subsurface stormflow recession was well represented by the model. The magnitude and timing of modeled pore pressure changes in the hillslope were similar to the field tensiometer observations (Fig. 2c–d). Overall, although the HYDRUS-2D simulations did not reproduce the behavior of soil water in the experimental hillslope exactly, the calibration resulted in a virtual hillslope that responded plausibly to precipitation and we, therefore, judged it an acceptable test bed for virtual experiments on how canopy interception affects subsurface stormflow.

## Comparing modeled responses to measured rainfall and throughfall

Modeled subsurface stormflow responses to observed throughfall differed from modeled responses to observed rainfall in that the commencement of subsurface stormflow was delayed (Fig. 3b), peak subsurface stormflow was less (Fig. 3b) and peak pore pressure was lower (Fig. 3c–d). The modeled subsurface flow recession was most different for the throughfall inputs that showed the greatest difference in the peak subsurface stormflow rates and peak pore pressures (T4 and T5). Total subsurface stormflow and the runoff ratio were lowest for T5 (Table 3), where throughfall input was most different from the open precipitation input.

In addition to lower modeled peak flows from hillslopes under canopies, subsurface stormflow was generally less responsive to fluctuations in rainfall intensity under canopies. This effect was apparently related to intensity smoothing more than to evaporation; T2, which had the lowest evaporation, showed smoother modeled responses to rainfall than did T4, which had greater evaporation but the lowest residence time (intensity smoothing) (Table 1, Fig. 3b).

Modeled pore pressures at the start of the storm were higher in response to the precipitation in the open. This

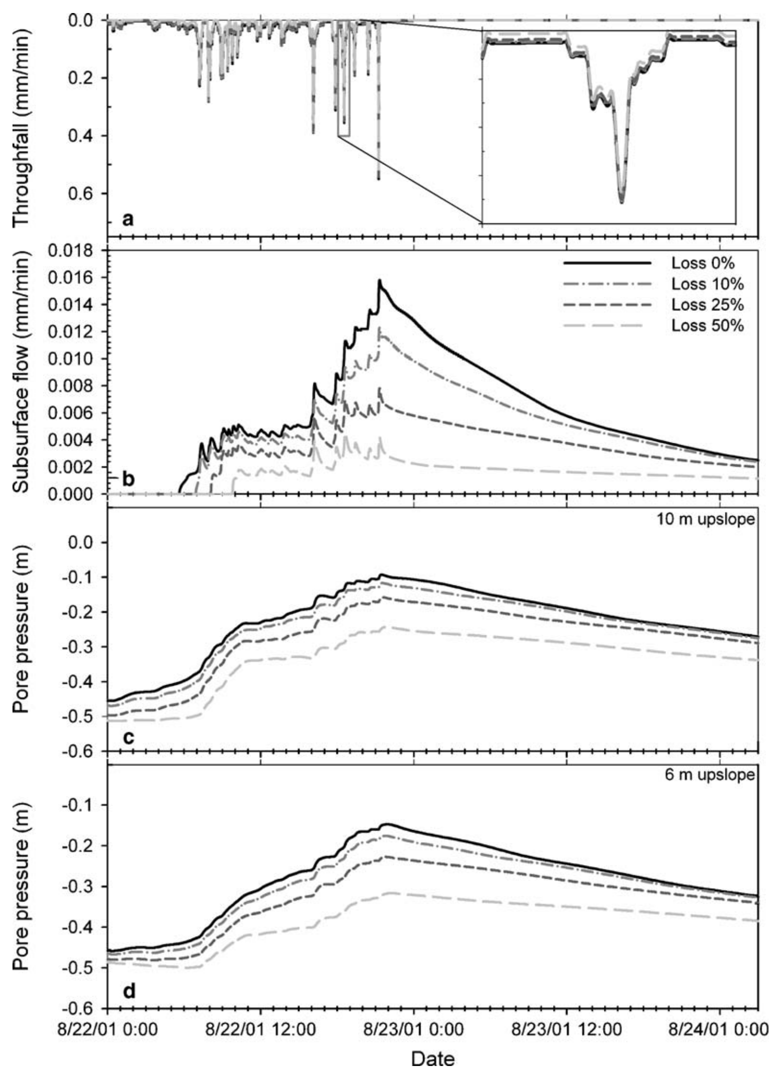


**Figure 4** Calculated throughfall (a), modeled subsurface stormflow (b) and modeled pore pressure at 10 m (c) and 6 m upslope from the trench (d) for the 2-D transect model for 0%, 10%, 25% and 50% evaporation loss. The inset in panel (a) shows more detail for the period 18:00–19:00, 22 Aug 2001.

was because low intensity precipitation on 21 August occurred in the open but not under the forest while canopy storage was being filled, and substantial drip had not yet begun. To determine the influence of this difference in initial conditions before the high intensity precipitation bursts, we also ran models for this storm but set the precipitation to zero prior to 22 August 5:10. This had the expected result that runoff began closer to the same time for the opening vs. canopy model runs, and there were smaller differences in subsurface stormflow and pore pressure between the model runs for the opening and different throughfall inputs. However, the model run with the highest interception losses (T5) still showed the latest start of subsurface stormflow, along with the least and smoothest peak subsurface stormflow and lowest peak pore pressure.

## Modeled subsurface stormflow and pore pressure responses to synthetic throughfall

Evaporation of water in the canopy had a large effect on model predictions of the start of subsurface stormflow, peak subsurface stormflow rate, total subsurface stormflow, the runoff ratio, and peak pore pressure (Fig. 4, Table 3). The start of subsurface stormflow was later and total and peak subsurface stormflow rates were lower as canopy evaporation increased. Modeled peak pore pressures at 0.53 m also decreased with canopy evaporation losses. In contrast, the increases in subsurface stormflow in response to precipitation bursts were similar in magnitude and timing regardless of evaporation rate.

The effect of throughfall smoothing on total subsurface stormflow, pore pressure and the runoff ratio was small (Fig. 5 and Table 3). Modeled peak subsurface stormflow rates decreased slightly as throughfall smoothing increased, and modeled subsurface stormflow responses to bursts were increasingly muted as smoothing increased. Throughfall smoothing led to no substantial differences in pore pressures in the simulations. Overall, intensity smoothing by the canopy had a much smaller effect on subsurface stormflow than did evaporation of precipitation from the canopy.

Comparing the virtual experiment results generated from synthetic 25% evaporation loss (Fig. 4) to those generated from field data collected at T5 (Fig. 3), which also had 25% evaporation loss but included intensity smoothing (Fig. 3, Table 1), shows that both yielded the same runoff ratio for subsurface stormflow. However, subsurface stormflow modeled from the observed throughfall data began later, peaked lower, and was less responsive to intensity changes during rainfall (Fig. 6). Thus, while canopy smoothing alone has no large effect on peak pore pressures or subsurface stormflow response, canopy smoothing in combination with evaporation has a larger effect on subsurface stormflow and peak pore pressure response than evaporation alone.

# Discussion

Comparison of the effects of throughfall differences on subsurface stormflow and peak pore pressures in a virtual hillslope has shown clearly how reductions in net applied water to the slope result in lower subsurface stormflow peaks, delayed peaks, later onset of subsurface stormflow at the trench face, less total flow, and lower runoff ratios and pore pressures. However, isolation of intensity smoothing effects on subsurface stormflow showed measurable differences only in peak subsurface flow rate and not any of the other measured parameter.

Modeled subsurface stormflow responses to varying throughfall in our virtual experiments were not simply scaled by precipitation amount, because the morphological and topographic (surface and subsurface) complexity of the hillslope gave rise to threshold processes that strongly affected runoff generation. Field observations at the Panola experimental hillslope have revealed that infiltrating water ponds in depressions in the bedrock, occurring first at locations of shallow soil (Freer et al., 2002; Tromp-van Meerveld and McDonnell, in press b). Before significant subsurface

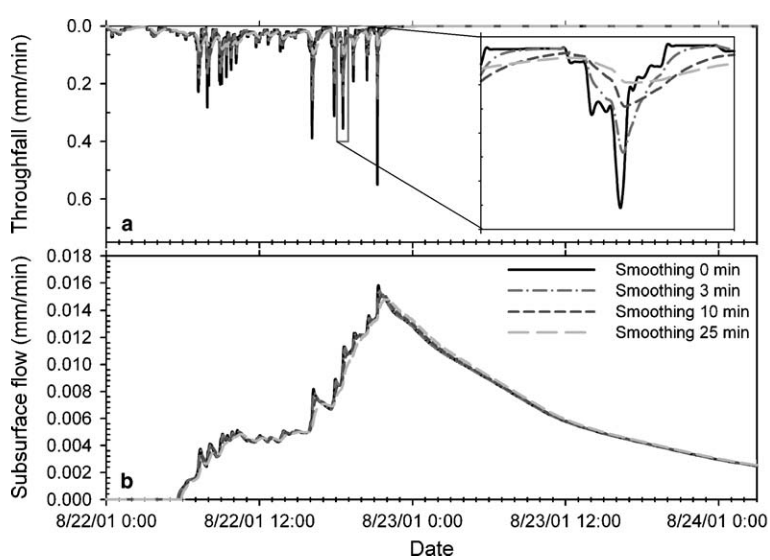


**Figure 5** Calculated throughfall (a) and modeled subsurface stormflow (b) for the 2-D transect model for 0, 3, 10 and 25 min smoothing. The inset in panel (a) shows more detail for the period 18:00–19:00, 22 Aug 2001.

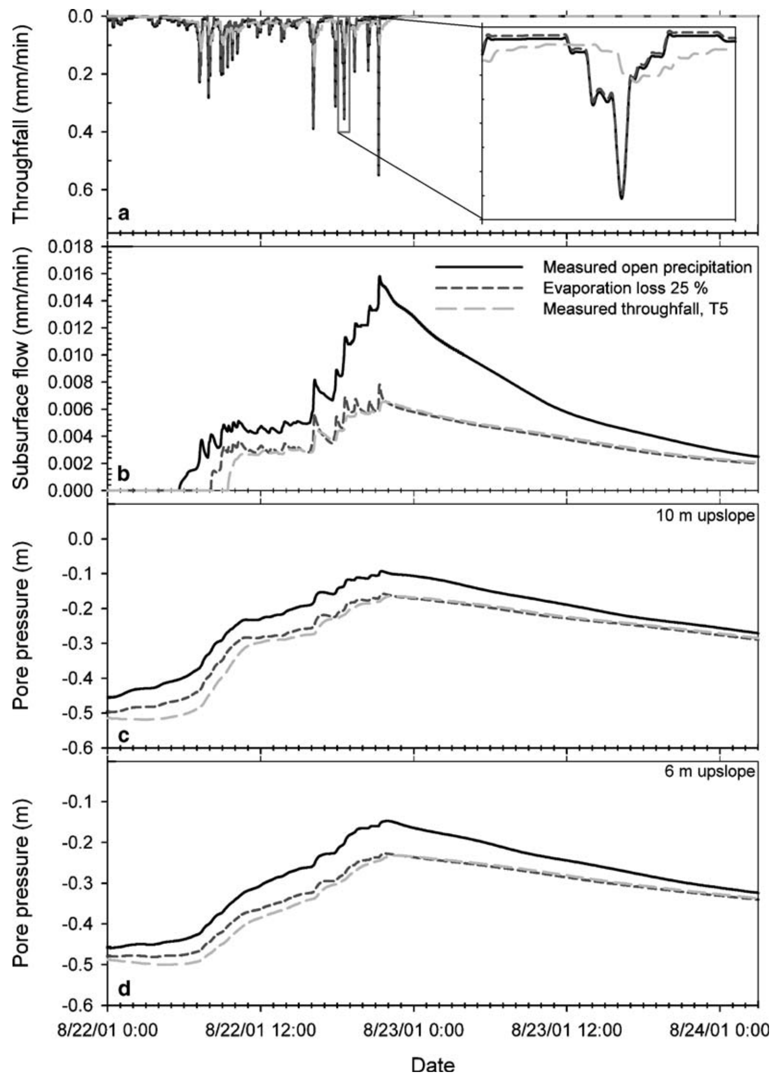


**Figure 6** Calculated throughfall (a), modeled subsurface stormflow (b), and modeled pore pressure at 10 m (c) and 6 m upslope from the trench (d) for the 2-D transect model for measured open precipitation, calculated 25% evaporation but no smoothing, and measured throughfall T5 with 25% evaporation and smoothing. The inset in panel (a) shows more detail for the period 18:00–19:00, 22 Aug 2001.

flow can occur, water levels in these depressions must rise sufficiently to become connected to each other and to the trench face — Tromp-van Meerveld and McDonnell (in press b) termed this the ''fill and spill'' hypothesis: water must ''fill'' storage before it can ''spill'' and generate runoff. Once this connectivity occurs, macropores, especially old root channels, are important in quickly delivering water to the trench face in much of the hillslope (Freer et al., 2002; Tromp-van Meerveld and McDonnell, in press a). The bedrock topography, the importance of connectivity of subsurface saturated areas, and preferential flowpaths are thus responsible for the complex and threshold relationship between precipitation and subsurface stormflow (Tromp-van Meerveld and McDonnell, in press a,b).

Although HYDRUS-2D was able to reproduce threshold behavior by mimicking the spill and fill process (Tromp-van Meerveld, 2004), it was not able to model preferential flow. Despite the fact we purposely chose a section of the hillslope where lateral macropores appear to be least important, we assume the lack of preferential flow in the model is most likely responsible for the slow modeled rise in subsurface stormflow rate compared to observations (Fig. 2). We assume the differences between modeled and observed subsurface stormflow early in the calibration storm were most likely caused by differences in initial conditions.

Un-modeled spatial variabilities in soil and throughfall likely contribute to more rapid hydrological hillslope response in soils with nonlinear soil moisture release curves (e.g., Torres et al., 1998). Thus, we expected the models to, for example, overpredict the time required for initiation of subsurface stormflow at the trench face and underpredict the rapidity of pore pressure response to infiltration. Other likely reasons for errors in model calibration include (3-D) lateral flow from adjacent hillslope transects that included substantial preferential flow (Freer et al., 1997; Burns et al., 1998), errors in estimating soil moisture storage, errors in the estimated soil moisture release curves, and errors in the initial moisture condition at the onset of precipitation. These processes should be investigated to learn more about the complex ways that the hydrological effects of forest canopies interact with hillslope flowpaths, antecedent moisture, and soil structure to affect subsurface stormflow production.

The three modeled throughfall scenarios used in the virtual experiments illustrate the considerable spatial variability that exists in the throughfall process. Several studies have shown that this spatial variation and the resulting infiltration variation contribute to spatial variation in soil moisture patterns in forest soils (e.g., Eschner, 1967; Bouten et al., 1992; Si, 2002; Raat et al., 2002; Zhou et al., 2002; Schume et al., 2003). However, owing to spatial variability of soil thickness and physical properties, it is often not possible to relate patterns in throughfall directly to patterns in soil water content (Raat et al., 2002). Nonetheless, spatial variability of throughfall has important implications for soil hydrology. Consistent and marked spatial differences of water infiltrating into the forest soil (Keim et al., 2005), originating from stemflow or spatial redistribution by the canopy, influence soil moisture and, therefore, the rate of percolation through the unsaturated zone. Consequences may include lateral surface and subsurface flow or rapid recharge to groundwater as infiltration bypasses portions of the soil profile (Weiler and Naef, 2003) and decadal-scale temporal stability of preferential flow paths as observed in a structured forest soil by Hagedorn and Bundt (2002). This research lacked data to fully represent any of these processes in space and time, so we focused instead on likely effects of simple variations in throughfall input (both amount and timing) on subsurface stormflow generation and peak pore pressures.

Our virtual experiments showed that evaporation loss from interception exerted a greater influence on subsurface stormflow than did rainfall intensity smoothing. However, intensity smoothing did reduce flashiness of flow and delayed peaks of both flow and pore pressure. Notwithstanding these results, the relative influence of evaporation and intensity smoothing likely varies by canopy characteristics and hillslope characteristics (e.g., hillslope gradient, hillslope geometry, soil hydraulic conductivity, and soil depth). Interception data for this research were from a canopy of large coniferous trees with high leaf area. Thus, the maximum canopy residence times tested here are likely larger and intensity smoothing effects, therefore, greater than in forests of other regions. The Panola hillslope steepness and soil depth are intermediate between hillslopes typical of plains and mountainous terrain. Subsurface stormflow from flatter hillslopes with deeper soils is likely even less dependent on canopy intensity smoothing, whereas steeper, convergent hillslopes with shallower soils are likely more sensitive to intensity smoothing than is the Panola hillslope.

The results of this work have important implications for the construction and calibration of watershed models. Modelers often use open rainfall gauge data for watershed inputs. Our findings demonstrate clearly that ignoring the transfer of precipitation through vegetation prior to infiltration misses a major effect on subsurface stormflow generation, and that models of forested watersheds that are calibrated to open-field precipitation must unrealistically subsume canopy interception effects in other parameters. While modeling interception evaporation with a temporally constant-evaporation rate would improve results in a watershed model, this approach would still not capture some important effects of the canopy. For example, compared to the simple 25% evaporation loss, modeled runoff from measured throughfall data with the 25% evaporation loss combined with intensity smoothing, began later, peaked lower, and was less responsive to changes in throughfall intensity.

## Conclusions

The effect of evaporation and intensity smoothing by canopy interception on hillslope subsurface stormflow is poorly understood. Until now, it has been impossible to separate the effects of canopy evaporation from intensity smoothing in field experiments in terms of which effect of throughfall affects hillslope subsurface stormflow response the most (if at all). Our virtual experiments using HYDRUS-2D to model subsurface stormflow in a well characterized hillslope showed that throughfall generated lower subsurface stormflow peaks, delayed peaks, later onset of subsurface stormflow, less total flow, and a lower subsurface stormflow runoff ratio compared to open rainfall. By separating evaporation from intensity smoothing, we found that canopy evaporation alone was responsible for most of the effects on subsurface stormflow alteration. Isolation of intensity smoothing effects on subsurface stormflow resulted in measurable differences only in peak subsurface flow rate. Nonetheless, virtual experiments that neglected intensity smoothing failed to reproduce subsurface stormflow behavior in response to measured throughfall. Overall, these results have important implications for calibration of watershed models that ignore the complex effects of canopy interception on infiltration and subsurface stormflow flow generation. Applying a scaled rainfall input to represent interception evaporation is shown by our results to mask important effects on peak subsurface stormflow response and the timing of peak stormflow in some situations.

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