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112 Subsurface Stormflow

Part 10. Rainfall-Runoff Processes

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Introduction

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2. Introduction

Subsurface stormflow is a runoff producing mechanism operating in most upland terrains. Subsurface stormflow occurs when water moves laterally down a hillslope through soil layers or permeable bedrock to contribute to the storm hydrograph in a river. In humid environments and steep terrain with conductive soils, subsurface stormflow may be the main mechanism of storm runoff generation (Anderson and Burt, 1990b; Gutknecht, 1996). In drier climates and in lowlands with gentler topography, subsurface stormflow may occur only under certain extreme conditions (high rainfall and high antecedent soil moisture), when transient water tables form and induce lateral flow to the channel (Wilcox *et al*., 1997).

While an important contributor to the volume of flow in the stream, subsurface stormflow is also responsible for the transport of labile nutrients into surface water bodies (e.g. McGlynn and McDonnell, 2003b). Since the flow path of water in the subsurface often determines the chemistry of waters discharging into the stream and hence the water quality, characterizing this subsurface flow path and the water's age and origin is important (Burns *et al*., 2003). Subsurface stormflow may also enhance positive pore pressure development in steep terrain (Uchida *et al*., 1999; Wu and Sidle, 1995) and may be responsible for landslide initiation (Montgomery *et al*., 1997; Sidle and Tsuboyama, 1992). Thus, subsurface stormflow is of great interest and importance beyond the conventional hydrological literature. This article examines the history of the study of subsurface stormflow processes, reviews theories on the generation of subsurface stormflow, and gives a detailed overview of current research on subsurface flow processes and implications for future research.

Terminology

Subsurface stormflow is also known in the hydrological literature as interflow, lateral flow, subsurface runoff, transient groundwater, or soil water flow. These multiple terms often confuse the process understanding of subsurface stormflow response to rainfall or snowmelt. While some studies have documented subsurface stormflow as unsaturated flow in the unsaturated zone, most studies have shown that subsurface stormflow is a saturated (or near-saturated) water flow phenomenon – due either to the rise of an existing water table into more transmissive soil above (with ensuing lateral flow) or the transient saturation above an impeding layer, soil-bedrock interface or some zone of reduced permeability at depth (argillic horizon, hard-pan, plough layer, etc.).

The literature on subsurface stormflow includes many references to both soil water and groundwater. Inconsistent definitions of these terms have also led to confusion. Here, we define groundwater or the saturated zone as any area in the soil profile with ≥0 kPa matric potential. Soil water, or the unsaturated zone, is the area in the profile with matric potentials of <0 kPa. Conversion from negative to positive potentials may occur very rapidly in the shallow subsurface. Therefore, modifiers such as "transient" groundwater will be used in some instances to indicate parcels of water in the subsurface that change from soil water to groundwater or from being an unsaturated to being a saturated zone, respectively, following our potential-based definition.

Historical Development of Ideas Pertaining to Subsurface Stormflow

As part of early studies on the influence of forest management on watershed hydrology in Switzerland, Engler 1919 was among the first hydrologists to recognize the importance of subsurface stormflow on runoff generation in forested environments. He made observations during numerous rainfall events in two neighboring catchments, conducted infiltration experiments, and performed detailed soil physical measurements of porosity, water content, soil texture, and hydraulic conductivity. He concluded from his experiments that overland flow did not occur even during high intensity rainfall. He observed that water infiltrated into the main root zone and flowed laterally in "uncountable veins" in the soil or at the soil-bedrock interface. Figure 1 shows the original conceptualization of these processes by Engler 1919.

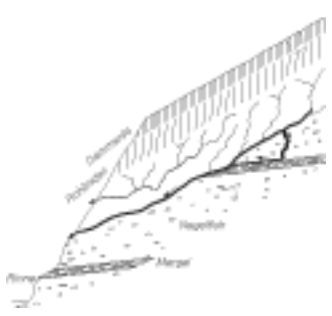


Figure 1. The original perceptual model of subsurface stormflow by Engler 1919. The hatched areas represent the uniform infiltration of water in the humus (*Dammerde*) and the soil (*Rohboden*). Deeper in the profile, water is flowing in "veins" laterally. *Nagelfluh* is the bedrock type for a specific geological setting in Switzerland

Hursh and Brater 1941 were the first to quantify the role of subsurface stormflow *senso stricto* in a watershed. This seminal work showed that the stream hydrograph response to storm rainfall at the forested Coweeta experimental watershed consisted 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.

The rate of progress in the understanding of subsurface stormflow continued to increase through the International Hydrological Decade (IHD (International Hydrological Decade)), with key works by Hewlett and Hibbert 1963 on moisture and energy conditions in a sloping concrete-walled hillslope, Whipkey 1965 on lateral preferential flow, Dunne and Black 1970 on subsurface stormflow around the toe of a hillslope and its interaction with near-stream saturated areas, and Weyman 1973 on saturated wedge development. The most important works during the IHD, however, were those that framed subsurface stormflow within the context of the variable source area concept: Hewlett and Hibbert 1967 in the United States, Cappus 1960 in France, and Tsukamoto 1961 in Japan. Later, Anderson and Burt 1978 provided clear field evidence for how topographic hollows were the key hot spots in the landscape for connected subsurface stormflow to the stream channel.

After comments by Freeze 1972 that subsurface stormflow could not be a dominant runoff process based on Darcy-Richards analytical approaches work by Beasley 1978, Harr 1977 and Mosley 1979 showed clearly that the response time and lateral flux rate of subsurface stormflow on steep forested hillslopes could be fast enough to be a main contributor to channel stormflow in headwater catchments – often via noncapillary pore space or flow in addition to Darcy-Richards like flux in the matrix.

Considerable debate has also surrounded the age and origin of subsurface stormflow. The "Maimai debate" is a classic case study in this regard and, therefore, is used here as an example (see a more extended and detailed review by McGlynn *et al*., 2002). Mosley 1979, 1982 conducted the first comprehensive study of subsurface stormflow at the Maimai catchments in New Zealand. He found a close coincidence in the time of the discharge peak in the stream and the time of the subsurface stormflow peaks, suggesting rapid movement of water vertically in the soil profile and in lateral downslope direction in the form of a saturated wedge. The wedge almost intersected the ground surface at the toe of the slope and tapered off in the upslope direction (Mosley, 1979). Dye tracing experiments of Mosley showed that his excavated "pit faces" (i.e. small trenches roughly 1 m wide and dug down to the soil-bedrock interface) displayed points of concentrated seepage during storm events, usually at the base of the B horizon, at which high rates of outflow were observed. At one site, Mosley 1979 observed that water "gushed" out of two pipes discovered at the base of the B horizon. Mosley's perceptual model considered macropore flow (see details in the next section) 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 (Figure 2a).

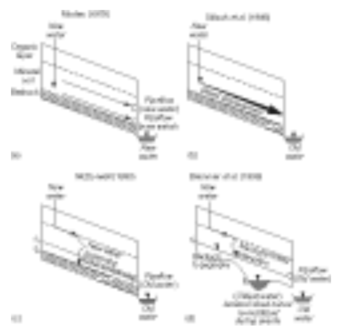


Figure 2. The evolving perceptual model of subsurface stormflow at the Maimai catchment in New Zealand (Reprinted from McGlynn *et al*., 2002. © 2002, with permission from Elsevier)

Pearce *et al*. 1986 and Sklash *et al*. 1986 followed with work at Maimai where they collected samples of rainfall, soil water, and streamflow and analyzed for electrical conductivity (EC (electrical conductivity)), chloride (Cl^{-}), deuterium (δD), and oxygen-18 ($\delta^{18}O$) composition. They found that: (i) most of the mixing of "old" (pre-event) and "new" (event) water occurred in the hillslope; and (ii) subsurface water discharge to the stream was an isotopically uniform mixture of stored water. In other words, the interpretations offered by Pearce *et al*. 1986 and Sklash *et al*. 1986 directly refuted Mosley's 1979 determination that rapid transmission of new water through macropores formed the majority of stream runoff. Sklash *et al*. 1986 postulated the conceptual model that saturated wedges on the lower slopes and groundwater ridges in the valley bottoms developed quickly as infiltrating rain converted the tension-saturated zone into groundwater. This perceptual model negated the need to invoke rapid transmission of new water down slope via macropores in order to explain the stream flow response, since stored water was the main component discharged into the stream channel during events (see Figure 2b).

McDonnell 1989, 1990 and McDonnell *et al*. 1991 combined isotope and chemical tracing with detailed tensiometric recording in an effort to explain the discrepancies between the earlier perceptual models. McDonnell found that: (i) water table longevity at the soil-bedrock interface was very short and showed a close correspondence with hillslope throughflow rate; and (ii) the interconnectedness of pipes in those zones was large enough to account for the rapidity of water table decline and pore water pressure dissipation. As a result, 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 (see Figure 2c). These studies were followed by Woods and Rowe 1996 and Brammer *et al*. 1995 who showed that the topography of the bedrock surface was a key determinant of where subsurface flow was concentrated spatially across the hillslope (see Figure 2d).

The Maimai experiments, along with other field experiments through the early 1990s, achieved a general consensus that: (i) pre-event water stored in the catchment before the rain event is the dominant contributor to stormflow in the stream – averaging 75% worldwide (Buttle, 1994) (*see* 116 (/doi/10.1002/0470848944.hsa120/full)); (ii) vertical (and often also lateral) preferential flow is a ubiquitous phenomenon in natural soils, particularly in steep catchments (Germann, 1990; Tsuboyama *et al*., 1994) (*see* 116 (/doi/10.1002/0470848944.hsa120/full)); and (iii) combining hydrometric, chemical, and isotopic observations in experimental work is necessary to constrain any perceptual model of subsurface stormflow or other runoff producing mechanism (Bonell, 1993; Bonell, 1998).

This scientific debate through the latter half of the twentieth century shows how scientific progress can be made by revisiting and reanalyzing data in one experimental watershed. However, it also highlights the danger in the field of hillslope hydrology of relying upon generalization of findings and processes of only a few selected hillslopes and watersheds. Few studies have compared subsurface flow processes and the dominant flow pathways across many sites (e.g. Beasley, 1978; Scherrer, 1997). The synthesis work of Dunne 1978 still stands as the most complete tabulation and intercomparison of subsurface stormflow data to date.

Flow Regimes of Subsurface Stormflow

Subsurface stormflow describes all runoff generation processes in the hillslope close to the soil surface that result in a stream channel hydrograph response during a precipitation or snowmelt event. This response may be coupled directly to flow in preferential pathways like macropores and layers or areas with high permeability. However, rapid subsurface stormflow response may also result from a fast hydraulic response of connected saturated areas in a hillslope in response to infiltrating precipitation (Burt and Butcher, 1985). The main flow regimes at the hillslope scale may be subdivided into homogeneous matrix flow and preferential flow.

Homogeneous Matrix Flow

Lateral matrix flow can be a viable subsurface stormflow process if water is already stored in the soil within connected saturated and close-to-saturated areas. These areas may respond quickly to an increase in hydraulic gradient and cross-sectional area due to infiltrating water. This process may occur in slopes where a high-permeable soil layer with high infiltration capacity is situated (parallel) above a low-permeable soil layer (e.g. bedrock, argillic horizon, etc). Since the water in storage on a typical hillslope is large relative to the rainfall depth, this matrix regime often results in a large contribution of pre-event water to the stream as only a small amount of event water is necessary to increase the hydraulic gradient and cross-sectional area in the slope and create a connected transient groundwater body. This flow process is described in the literature often in terms of translatory flow (Burt, 1989), transmissivity feedback (Rodhe, 1987), or lateral flow at the soil-bedrock interface (Tani, 1997).

Preferential Flow

Preferential lateral flow occurs either in distinctive structures in the soil where water flows only under gravity (macropores) or in areas with a higher permeability than the surrounding soil matrix. Macropores in the soil or fractures in the bedrock that are oriented predominantly slope parallel may transport water efficiently and rapidly from the hillslope to the stream (Beven and Germann, 1982). Laterally oriented macropore flow may dominate in many forest environments where macropores are generated by plant roots and burrowing animals. Macropores that are enlarged by erosion and connected over several meters are often termed *soil pipes* (Anderson and Burt, 1990a; Jones, 1971). If a connected network is developed because of internal erosion and eluviation and connection of macropores, piping can provide effective drainage augmentation to hillslopes. However, disconnected macropores that connect hydraulically during storms can also result in an effective drainage of the hillslopes (Weiler *et al*., 2003). If the underlying bedrock is more permeable, water can infiltrate into the bedrock and then percolate vertically into bedrock fissures and cracks – negating the macropore enhancement of subsurface stormflow on the timescale of a rainfall event. The preferential flow regime is described in the literature often in terms of lateral preferential flow (Tsuboyama *et al*., 1994) and pipeflow (Uchida *et al*., 1999), or lateral preferential flow at the soil-bedrock interface (McDonnell, 1990).

High permeability layers are areas in the slope with a coarse texture and large pore and void space. These are often found in talus slopes, landslide debris, peri-glacial solifluction deposits, or unconsolidated moraine material. Erosion of fine sediments by turbulent flow in areas with already coarse soil material increases the hydraulic conductivity and makes these areas particularly conductive. This flow regime can be best envisioned by an extension of the surface streams into the hillslope where many "small" subsurface streams connect preferentially the hillslopes with the streams (Sidle *et al*., 2000). This flow process is described in the literature in terms of flow in "high-permeable layers" (Scherrer, 1997).

Current Research in Subsurface Flow Processes

Subsurface Flow Initiation

Various experimental observations have shown that the water table development near the base of the soil profile can be very rapid following the onset of precipitation. This response is often a key control on the initiation of lateral flow due to the combination of four factors: (i) an increase of the hydraulic gradient; (ii) an increase of the cross-sectional area of flow; (iii) the rise of the water table into more permeable upper soil layers; and (iv) a connection of transient groundwater bodies across the hillslope. There is still debate among scientists about the controls on rapid water table response, but vertical preferential flow during infiltration and a decline of the drainable porosity with soil depth are thought to be the main causes for rapid subsurface flow initiation capable of generating subsurface stormflow.

Vertical Preferential Flow

Vertical preferential flow in natural soils is an almost ubiquitous phenomenon (Flury *et al*., 1994; Weiler and Naef, 2003). Different types of heterogeneities in soils result in different preferential flow processes. For example, macropore flow is common in soils with animal burrows, roots, and earthworm channels; heterogeneous matrix flow (a type of preferential flow) in soil with spatially heterogeneous soil properties and wettability. How water is transported rapidly in the soil bypassing the unsaturated zone depends on the initiation processes of preferential flow and the interaction of the preferential flow pathways (e.g. macropores) with the surrounding soil matrix. How the two processes determine vertical water delivery into the lower soil profile can be distinguished experimentally. Figure 3 shows results from two sites where a dye (Brilliant Blue FCF) was added to 70 mm of simulated rainfall that stained the flow pathways upon infiltration into the soil. Later, the soil profiles were excavated and pictures were taken of the vertical soil sections. The stained areas can be then classified with image analysis techniques (Weiler and Flühler, 2004). Figure 3(a) shows that infiltration is governed by a high interaction between macropores and the soil matrix and that macropore flow is initiated close to the soil surface. These processes will result in a slow response in the lower profile. In the example in Figure 3(b), macropore flow is also initiated close to the soil profile but interaction between water flow in the macropores and the surrounding soil matrix is very low, resulting in narrowly stained features. Here, water flow in macropores is often turbulent and mostly driven by gravity. Hence, water delivery to the lower soil profile is very rapid. While macropores may comprise only a small part of the total soil porosity (e.g. 0.35–0.77%, e.g. Weiler and Naef 2003), they account for almost all the water flow at or near saturation within the profile. The resulting water movement in the soils is very heterogeneous and certain areas within the soil may be completely bypassed. These processes often defy the Darcy-Richards formulations that rely exclusively on capillary-driven fluid flow (*see* 66 (/doi/10.1002/0470848944.hsa070/full) for full treatment of classical flow in porous media theory).

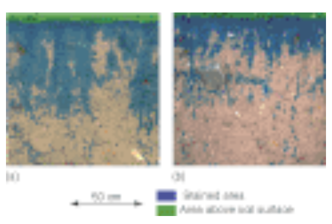


Figure 3. Dye patterns from two different sites: Rietholzbach experimental watershed, Switzerland, (a) and Heitersberg near Zurich, Switzerland (b). Note the spatial heterogeneity of dyed water and preferred nature of water flow vertically within the soil profile

Another observation sometimes made in soils when water content is close to field capacity is that only a small fraction of water input into the soil results in a strong hydrodynamic response in the lower soil profile. Torres *et al*. 1998 found that this hydrodynamic response can be on an average 15 times greater than the estimated water and wetting front velocities. While these processes remain poorly understood, research to date shows clearly that rapid travel times of fluid pressure head or water content through the unsaturated zone could be mistaken as preferential or macropore flow (Rasmussen *et al*., 2000; Smith and Hebbert, 1983).

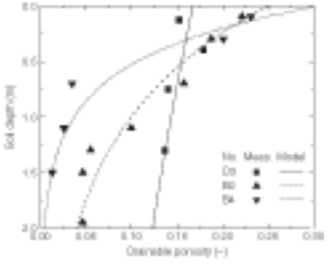
Drainable Porosity

The conversion of unsaturated to saturated conditions or the rise of an existing groundwater table at depth depends on the drainable porosity (or specific yield) of the soil. Drainable porosity is commonly defined as the difference in volumetric water content between 0 kPa and 33 kPa soil water potential (approximately field capacity) (Weiler and McDonnell, 2004). This represents the porosity of the soil draining within a short duration – characteristic of water table rise in the saturated zone. Drainable porosity commonly decreases with depth because of an increase in the bulk density, a decrease in macroporosity, and a change in soil texture and pore structure.

The decrease in drainable porosity with depth can be observed in different soils by comparing measurements of the water content at saturation and field capacity at several depths within the soil profile. Figure 4 illustrates the change of drainable porosity with depth in three different soils. The values show a high-drainable porosity in the topsoil and a slower (site no. D3) or a more rapid (site no. B4) decrease in drainable porosity with depth. Since the drainable porosity cannot be zero within the soil profile, an exponential model can be used to describe the observed change with depth:

$$n_d(z) = n_0 \exp\left(-\frac{z}{m}\right) \tag{1}$$

Figure 4. Measured and fitted drainable porosity with depths of three different soils (No. of sites see Table 1)



where n_0 is the drainable porosity at the soil surface, m is a decay coefficient, and z is the depth into the soil profile (positive downward). An exponential model can be fitted to the measured values to produce a depth distribution of drainable porosity. Table 1 gives the values of the two fitting parameters n_0 and m for 13 different soils. The fitted values of the parameter m show a pronounced decline of the drainable porosity with depth for almost all sites, except for site no. A4. At this site, the texture changes to a more sandy soil at depth, resulting in an increase of the drainable porosity with depth. For all other sites, the parameter m ranges from 0.294 (rapid decrease) to 6.496 (slow decrease). The drainable porosity at the soil surface n_0 does not influence the parameter m; however, the grassland soils generally show a lower n_0 than the forest soils. A low drainable porosity in the lower soil profile indicates that only a small amount of water is needed to raise the water table and, thus, to increase the potential for lateral subsurface flow.

Table 1. Values of the Two Parameters n_0 and m of the Exponential Depth Function for Drainable Porosity and the Goodness-of-Fit Information (Efficiency) for Soil Data of 13 Different Sites

No.	Site	n_0	m	Efficiency
Data by Weiler 2001, Grassland soils				
A1	Rietholzbach	0.083	0.294	0.915
A2	Heitersberg	0.037	2.353	0.867
A3	Koblenz	0.058	0.339	0.953
A4	Niederweningen	0.016	–0.457	0.817
Data by Ranken 1974, Forest soils				
B1	Pit 1, upslope	0.268	0.881	0.929
B2	Pit 2, upslope	0.271	1.206	0.946
B3	Pit 5, midslope	0.237	1.717	0.957
B4	Pit 12, downslope	0.356	0.762	0.912
Data by Yee 1975, Forest soils				
C1	Klickitat soil, Pit 2, Granophyric gabbro, 70% slope	0.250	3.152	0.878
C2	Bahannon soil, pit A, Tyee sandstone, 65% slope	0.131	2.756	0.949
Data by Rothacher *et al*. 1967, Forest soils				
D1	Reddish-brown Lateritics over weathered breccia	0.347	1.052	0.984
D2	Yellow–brown Lateritics over rotten rock	0.253	1.146	0.856
D3	Regosol over breccia	0.168	6.496	0.377

For example, a sensitivity study implementing the exponential model for drainable porosity decline in Hillvi (/doi/10.1002/0470848944.hsa246/full) (http://faculty.forestry.ubc.ca/weiler/hillvi.html (http://faculty.forestry.ubc.ca/weiler/hillvi.html)), a subsurface hillslope flow model, reveals that for a 10% steep hillslope using m-values of 0.3, 0.5, and 1.5, the response time is 14.4, 3.8, and 1.4 times faster, respectively, than for the same hillslope with a constant drainable porosity. The response time in this example is calculated as the time subsurface flow increases to half of the simulated constant rainfall intensity. This example highlights the importance of the drainable porosity decline in the soil profile for the vertical response of the groundwater level in the hillslope and, hence, the lateral response of subsurface flow.

Topographic Control on Lateral Subsurface Flow

There is now wide consensus that in areas with steep slopes, thin soils, and matrix hydraulic conductivities above the maximum rainfall intensity, water moves vertically to depth (as matrix or preferential flow), perches at the soil-bedrock or an impeding layer at depth, and then moves laterally along the lowest depths of the profile (Freer *et al*., 2002; McDonnell, 1990; Peters *et al*., 1995; Sidle *et al*., 2000; Tani, 1997; Tsukamoto *et al*., 1982; Uchida *et al*., 2002). Hence, the bedrock topography may control the direction and accumulation of flow more directly than the topography of the surface. Hillslope experiments using piezometer and tensiometer data as well as flow volumes recorded at a trench face suggest that rapid saturated subsurface flow occurs as narrow "ribbons" of saturated flow along the bedrock topographic surface (Hutchinson and Moore, 2000; Woods and Rowe, 1996). The bedrock surface is the main "pathway" for mobile transient saturated flow during events (Freer *et al*., 2002; McDonnell *et al*., 1996).

Isotope composition and major ion concentration of subsurface flow have been used to test the physical mechanisms revealed by the pore water pressure and topographic analysis. ^{18}O analysis of collected trench flow often showed little evidence of "new water" breakthrough during storm events (Burns *et al*., 1998). The Panola hillslope at Panola, Georgia, USA, is a useful example of hillslope flow behavior. For example, the Panola trench SO_4 chemistry showed no concentration-discharge relationship (Burns *et al*., 1998). The only variability in concentrations of subsurface flow at the trench face was between neighboring troughs, suggesting waters of slightly different ages were mobilized within the hillslope. Burns *et al*. 1998 found that base cation concentrations in hillslope subsurface stormflow were generally related to flushing frequency, where parts of the trench with the highest bedrock surface drainage area had consistently lower mean base cation concentrations than other trench face positions with lower bedrock surface drainage areas.

The relative contributions of different parts of the Panola hillslope change with total precipitation, antecedent wetness, total flow, and season (Tromp-van Meerveld, 2004). This suggests that the bedrock topography might not be the only dominant control on subsurface flow as was suggested by the analysis of only a few storms by McDonnell 1997 and McDonnell *et al*. 1996 and Freer *et al*. 1997, 2002. Together with bedrock contributing area, soil depth may control subsurface stormflow dynamics during storms, especially for smaller events (Buttle *et al*., 2004). At the Panola hillslope, subsurface storm flow volume of smaller events (less than the 55 mm rainfall threshold) or events with dry antecedent conditions are controlled largely by soil depth variations in space (Tromp-van Meerveld, 2004), while the bedrock topography is the primary control during medium to large storms. The shift in dominant areas of flow appears to be related to increasing antecedent soil moisture, storm size, and total flow (Freer *et al*., 2002).

Despite the fact that bedrock topography and soil depth variability are important controls on subsurface stormflow, few models have conceptualized and implemented this into simulation models. Recently, Weiler and McDonnell 2004 incorporated soil depth variability into a subsurface flow model and simulated subsurface stormflow for the Panola hillslope. They could show that soil depth variations not only have a large influence on the spatial variation of subsurface flow but control largely the total subsurface flow volume produced. Figure 5 shows observed and modeled subsurface flow for simulations using Hillvi (/doi/10.1002/0470848944.hsa246/full) (http://faculty.forestry.ubc.ca/weiler/hillvi.html (http://faculty.forestry.ubc.ca/weiler/hillvi.html)), a spatially explicit saturated and unsaturated water balance model for the actual measured soil depth variations at the Panola and Maimai experimental hillslopes and for an assumed constant average soil depth at those sites. All other model parameters were kept constant. For the Panola hillslope in particular, the differences in subsurface stormflow response are considerable – changes are also notable for the Maimai hillslope. These differences seem to be related mainly to the total variance of soil depth and the measured spatial correlation length of soil depth.

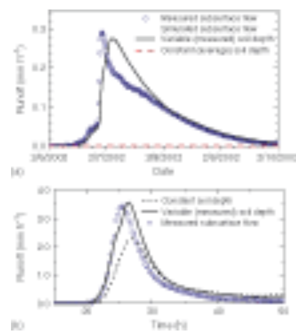


Figure 5. Simulated and measured subsurface flow response for the experimental hillslope at Panola (USA) – (a) and Maimai (NZ) – (b) assuming a constant soil depth and the measured variable soil depth

The preferential nature of flow along the soil-bedrock interface may also partly account for the often observed discrepancy between measurements of saturated hydraulic conductivity at the plot scale and the calculated hydraulic conductivity assuming lateral homogeneous matrix flow from measurements of lateral subsurface flow at the hillslope scale (Brooks *et al*., 2004; Uchida *et al*., 2002; Weiler *et al*., 1998). As the results in Figure 5 show, subsurface flow can increase by orders of magnitude because of the variations in soil depth without changing the saturated hydraulic conductivity.

Subsurface Flow in Soil Pipes/Macropores

In many environments, subsurface stormflow is dominated by lateral macropore flow, ranging from subarctic wetland (Woo and DiCenzo, 1989) and boreal forest (Roberge and Plamondon, 1987) to tropical rain forest (Elsenbeer and Lack, 1996) and semiarid land (Newman *et al*., 1998). These macropores are commonly referred to as *soil pipes* and concentrated subsurface flow in these natural soil pipes is called as *pipeflow* (e.g. Jones, 1971; Jones, 1981). Empirical studies of lateral pipeflow found in the literature show that lateral pipeflow controls hillslope response (e.g. Uchida *et al*., 1999), nutrient flushing (Buttle *et al*., 2001), and old water delivery to streams (Freer *et al*., 2002; McDonnell, 1990) and to riparian zones (McGlynn and McDonnell, 2003a).

Recently, soil pipe morphology has been studied by means of dye tracer, fiberscope, and ground penetrating radar. Results show that: (i) soil pipes in gentle moorland environments are complex and long (>50 m) (e.g. Holden and Burt, 2002; Jones, 1987), while soil pipe connectedness and length mapping in forested steep hillslopes show that soil pipes are highly discontinuous with maximum lengths of usually only a couple of meters (e.g. Michihata *et al*., 2001; Noguchi *et al*., 1999); and (ii) the position of soil pipes in gentle moorland environments varies from being shallow to deep within the peat layer. The pipe can be entirely within the peat, at the peat-substrate interface, or entirely within the substrate (e.g. Holden and Burt, 2002). The position of the soil pipes in steep forested hillslopes is often at or near the soil-bedrock interface (e.g. McDonnell, 1990; Terajima, 2002).

Runoff characteristics of pipeflow have been examined at hillslopes in Japan, the United Kingdom, North America, and Peru. These studies suggest that the maximum discharge of pipeflow is determined mainly by soil pipe diameter. Many studies have also shown that a precipitation threshold exists when pipeflow starts to dominate subsurface flow. The observed precipitation threshold depends strongly on the antecedent soil water content (Noguchi *et al*., 2001; Tromp-van Meerveld, 2004; Uchida *et al*., 1999). Figure 6 shows the linear relationship between pipeflow and total subsurface stormflow that was observed in several forested hillslopes around the world (Uchida *et al*., 2005).

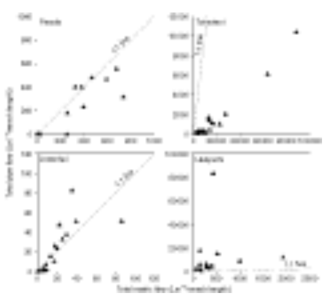


Figure 6. Relationship between total stormflow and total pipe flow of each storm at (a) Panola, Georgia, USA, (b) Toinotani, Kyoto, Japan, (c) Jozankei, Hokkaido, Japan and (d) Hakyuchi, Tokyo, Japan. Data for Jozankei and Hakyuchi compiled from Kitahara *et al*. 1994 and Ohta *et al*. 1981, respectively

Only recently have some models considered the role of pipeflow on runoff generation (Faeh *et al*., 1997; Jones and Connelly, 2002; Kosugi *et al*., 2004; Weiler *et al*., 2003). Nevertheless, these studies have not yet fully incorporated field perceptions into their numerical models. Although they are now viewed as a major subsurface flow control, the effect of soil pipes and other structures on lateral flow and transport at the hillslope scale still awaits good model-process integration.

Thresholds and Nonlinearities

Since the IHD, hydrologists have believed that subsurface stormflow rate changes linearly with changes in rainfall magnitude (Hewlett and Hibbert, 1967). Only recently has the nonlinearity of the subsurface stormflow process been fully realized (Buttle *et al*., 2004; McDonnell, 2003). Many recent studies have examined the relation between the amount of precipitation and the volume of subsurface storm flow – documenting thresholdlike, nonlinear relationships between the two. In this section, we describe briefly how subsurface storm flow volume changes with rainfall amount and discuss what processes control the change in the hydrological response of subsurface storm flow.

Field Evidence of Threshold and Nonlinear Responses

Despite the lack of any formal recognition conceptually, data in the hillslope hydrology literature suggest that precipitation thresholds for subsurface stormflow generation may be a widespread phenomena. Revisiting data from many trenched hillslope studies (e.g. Mosley, 1979; Whipkey, 1965) suggests that the precipitation threshold for subsurface stormflow commencement lies commonly between 15 and 35 mm. More recent studies show that the precipitation threshold depends on the antecedent moisture conditions in the hillslope (Guebert and Gardner, 2001; Noguchi *et al*., 2001; Peters *et al*., 1995; Uchida *et al*., 1999). Tani 1997, for example, has pointed out that after the threshold was reached, there was an almost 1:1 relation between precipitation above the threshold and subsurface flow. A simplified summary of the relation between storm total precipitation and the threshold for subsurface flow is given in Figure 7 for four sites. These relations suggest that while hillslopes vary tremendously from place to place, the subsurface stormflow initiation threshold and the slope of the line thereafter may be an emergent property of the hillslope hydrological system.

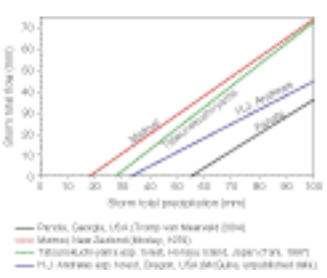


Figure 7. Schematic representation of the thresholdlike relationship between total storm precipitation and storm total flow under wet antecedent conditions. The lines represent the best fit lines through maximum storm total subsurface stormflow data points

The recent analysis of 147 storms at the Panola trenched hillslope in the United States is one of the first to explicitly examine and document the relation between rainfall and subsurface stormflow over a large range of storm sizes, seasons, and antecedent wetness conditions (Tromp-van Meerveld and McDonnell, 2004). They found that macropore and matrix flow had very similar thresholds (around 55 mm) for significant (>1 mm) subsurface flow at the trench. The value of the threshold precipitation was related secondarily to antecedent soil moisture. Only 8 of the 147 storms (an average of 4.3 storms per year) exceeded the precipitation threshold to produce significant lateral subsurface stormflow in the Panola watershed. For these storms, the total amount of subsurface flow was between 30% and 80% of the total precipitation after the precipitation threshold.

Searching for First-Order Control of Thresholdlike Responses

Notwithstanding the observation of thresholdlike subsurface stormflow, hydrologists still lack a clear understanding of the processes responsible. Recent studies have shown that there are at least two common controls for a threshold response: (i) interconnection of lateral preferential flow paths; and (ii) development and extension of transient saturated areas. Tani 1997 observed a large subsurface runoff response after smaller areas of transient saturated zones in the hillslope became connected. Using data from another catchment in Japan, Sidle *et al*. 2000 postulated that with increasing wetness, subsurface flow expands over greater slope distances, such that during very wet conditions the hillslopes become linked to the channel system to enhance subsurface flow significantly.

Recently, Spence and Woo 2003 and Tromp-van Meerveld and McDonnell (2004) proposed a fill and spill mechanism to account for threshold behavior at the hillslope scale. Tromp-van Meerveld and McDonnell (2004) used a spatially distributed grid of piezometers to obtain spatial and temporal information of transient saturation at the Panola trenched hillslope. Both piezometer data and a two-dimensional finite element model indicated that for storms smaller than the observed precipitation threshold, subsurface saturation occurred first in areas where the soil was shallowest and then expanded to areas with deeper soils, but did not flow further downslope because it was blocked downslope by a bedrock ridge. During larger storms, transient saturation filled up local depressions in the bedrock microtopography before water spilled laterally over the bedrock ridges between the depressions and flowed further downslope through the bedrock lows (Figure 8). Only during storms larger than the precipitation threshold did these transient saturated areas become connected to the trench face, resulting in significant (>1 mm) subsurface flow. Figure 8 illustrates how soil depth and bedrock microtopographic relief (i.e. the bedrock lows) together control the connection of subsurface saturated areas to the trench face causing the observed threshold for subsurface flow (see also Figure 7).

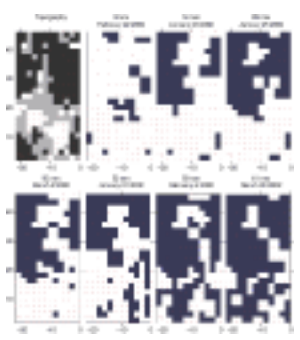


Figure 8. The location of areas of shallow soil depth (<0.75 m) (in black) and the areas with high (>4) bedrock topographic index (grey) (a) and the observed spatial distribution of subsurface saturation at the soil-bedrock interface across the Panola hillslope with increasing precipitation (b—h). The shaded area represents the area where transient subsurface saturation was observed; the unshaded area indicates the area where no subsurface saturation was observed. The diamonds represent the locations of the piezometers. Linear triangulation was used to interpolate between the measurement points

Outlook and Conclusion

Subsurface stormflow is the main mechanism of runoff generation in many catchments around the world and is significant for the initiation of landslides, for the flushing of labile nutrients and hence the water quality changes, and the hydrograph response in streams. Most subsurface stormflow studies published to date have focused on only a few storms at any given site. Up until recently, it has been difficult to derive general hydrologic principles from single research studies within intensively studied small basins (Jones and Swanson, 2001). Comparing different well-instrumented sites allows extraction of the commonalities or major differences between sites and helps define the first-order controls on subsurface stormflow generation and flow pathways.

A question for future research on subsurface stormflow is how we should compare different sites and what tools we should use to define first-order controls. First, we need to continue with our experimental investigations to understand better the internal processes. New experimental technologies like ground-penetrating radar (Huisman *et al*., 2003) or electromagnetic induction (Sherlock and McDonnell, 2003) will be important. Detailed measurement of flow combined with isotope and chemical tracing in time and space will be essential to constrain our conceptual and numerical models of subsurface stormflow.

Site intercomparison will require an organizational framework to summarize and evaluate the comparative analysis. Scherrer and Naef 2003 proposed a decision tree to define the dominant hydrological flow processes on a variety of grassland sites in Switzerland. They showed, rather convincingly, that the decision tree approach can be a powerful tool to clarify the first-order controls on dominant runoff processes. A decision tree may be a useful organizational framework for subsurface stormflow science to summarize and organize comparative analyses and to provide a structure for defining the hierarchy of process controls that are necessary for model development.

Process controls and new conceptualization for model development will also be a prerequisite for simulating and predicting the links between biogeochemical-hydrological aspects of subsurface stormflow. Recently, attempts have begun to derive the primary controls of transport processes and nutrient flushing in hillslopes dominated by subsurface stormflow (e.g. Stieglitz *et al*., 2003; Weiler and McDonnell, 2005). Nevertheless, the observed variability for soil properties and structural features (e.g. macropores and pipes) still need to be implemented and conceptualized. When introducing variability (as shown, for example, in Figure 5), we may not be able to predict deterministically where and when water is flowing at a specific location, but we may be able to describe the distribution of subsurface flow variability and its effects on water age, origin, and timing.

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