Original Research

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A) Forest soil F2



B) Grassland soil G1



Upscaling flow and transport processes from hillslope to catchment scale requires connecting hillsope soil structure quantification to its impact on those processes at the landscape scale. We address this by integrating classical pedological soil-landscape pattern description with the process knowledge gained from soil physics and hydrology.

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Preferential Flow Effects on Infiltration and Runoff in Grassland and Forest Soils

Due to the heterogeneity of soil hydraulic properties, there are many ways in which natural soils respond to rainfall, making upscaling of flow processes from plot to catchment scale difficult. The objectives of this study were to qualitatively characterize the flow pathways on forest and grassland hillslopes and to quantitatively define the relevant parameters controlling surface runoff generation by using infiltration and dye tracer experiments supplemented with measurements of the saturated hydraulic conductivity $K_{sat'}$ the structural porosity $n_{\rm SP'}$ and the pore-size distribution PSD. While infiltration excess overland flow dominates in grassland, forest soil structure characterized by relatively high values of K_{sat} and n_{sp} enhances the infiltrability of the soil and consequently prevents or at least reduces surface runoff. The dye patterns suggest that macropores are more efficient in forest than in grassland soil. The low efficiency of grassland soil macropores in transporting all water vertically downward can be explained by (i) the fine and dense few topsoil layers caused by the land use that limit water flux into the underlying macropores and (ii) their restricted number, their tortuosity, and the restricted interaction between macropores and the matrix below the topsoil layer. The larger root water uptake of forest soil as compared to grassland soil can be viewed as an additional factor enhancing its storage capacity and, consequently, may reduce the generation of surface runoff. It remains unclear, however, what effect the low interaction between macropores and soil matrix in the upper part of the subsoil has on surface runoff in grassland soil; this should be investigated in future studies.

Abbreviations: BB, Brilliant Blue; DC, dye coverage; F [soil], forest; G [soil] grassland; OC, organic carbon; PSD, pore-size distribution; RC, runoff coefficient; SPW, stained pathway width; TDR, time domain reflectometry.

Preferential flow is well recognized as a potentially important infiltration mechanism in soils because it may increase the leaching potential of soil, resulting in the accelerated transport of nutrients, pesticides, and pathogens (e.g., Jarvis, 2007; Köhne et al., 2009a,b). Preferential flow of infiltrating water in macropores may bypass most of the soil porous matrix (e.g., Beven and Germann, 1982) and limit the storage, filter, and buffer functions of soils (e.g., Clothier et al., 2008). Macropore flow is a subset of preferential flow that occurs in continuous root channels, earthworm burrows, fissures, or cracks in structured soil (e.g., Gerke, 2006; Hendrickx and Flury, 2001). Its initiation during infiltration depends on the initial matrix water content, intensity and amount of rainfall, matrix conductivity, and soil surface contributing area (e.g., Jarvis, 2007; Köhne et al., 2009a,b).

Apart from physical and geometric characteristics of the macropore domain, such as size, continuity, and surface area, the presence of relatively low permeability of linings and coatings of macropore walls (Gerke and Köhne, 2002) that restrict the local (i.e., mostly lateral) mass exchange is also an important factor controlling water flow in macropores (e.g., Jarvis, 2007). The degree of preferential flow has been shown to increase with increasing rainfall intensity (e.g., Gjettermann et al., 1997; Alaoui et al., 2003) and to be stronger under ponding than under sprinkling irrigation conditions (e.g., Hamdi et al., 1994). The initiation of preferential flow does not require the full saturation of the entire soil profile (Jarvis, 2007). It can be initiated if the soil surface layer approaches near-saturated conditions.

Although their fraction of the soil pore volume is restricted, macropores are reported to transport most of the water and solutes under specific soil conditions. For example, Alaoui and Helbling (2006) reported that for an estimated macropore volume of only 0.23 to 2% of the total soil volume, macropores transported 74 to 100% of the total water vertically downward. In other studies, Lin et al. (1996) reported that just 10% of the soil porosity (macro- and mesopores) contributed 89% of total water flow. As already demonstrated in

an untilled soil by Ehlers (1975), the maximum infiltrability of conducting channels exceeded 60 mm h^{-1} , although the volume of these channels totaled only 0.2% of that of the soil.

Dye tracer experiments are a valuable technique for characterizing preferential flow. In the dye tracer method, infiltration patterns are evaluated semiquantitatively based on dye coverage distributions (Flury and Flühler, 1994), by counting the number of stained pathways contributing to preferential flow (Gjettermann et al., 1997), or by defining classes of stained pathway widths (SPWs) and analyzing vertical and horizontal dye pathways soil cross-sections (Weiler and Flühler, 2004; Bachmair et al., 2009). Despite the progress in image analysis techniques that facilitate quantitative examination of the link between texture and morphology of soil horizons (e.g., Kulli et al., 2003; Alaoui and Helbling, 2006), no attempt was made to link qualitative analyses of dye tracer experiments with the hydrodynamic aspects in terms of water movement. In addition, each dye tracer experiment is related to specific soil structural properties and initial moisture conditions, which further complicate any comparison of experimental results (Forrer et al., 2000).

Although vertical preferential transport has been investigated intensively over the last two decades, less attention was paid to investigate the relationships between vertical preferential flow and runoff processes occurring at hillslope scale. Scherrer (1996) performed sprinkling experiments on grassland and arable land in regions of Switzerland with soils of different geological origins. He reported a high variability in runoff generation that depends on the local conditions, mainly soil properties and geology. Using irrigation experiments, Weiler et al. (1998) reported that runoff generation on grassland was mainly controlled by a highly permeable A horizon; the network of vertical macropore system in deeper layers controlled bypass flow. In contrast to the grassland, the same authors (Weiler et al., 1998) found a stronger subsurface flow response for forest soils that had a bimodal PSD. Markart et al. (1997) reported favorable conditions of the forest regarding fast runoff mitigation. As reported in Price et al. (2010), in a comparison with soils impacted by human land use, soils underlying forest trees generally feature low bulk density and high saturated hydraulic conductivity, total porosity, and macroporosity, as a result of ample litter cover, organic inputs, root growth and decay, and abundant burrowing fauna (Lee and Foster, 1991). In contrast, soils exposed to human impact are often stripped of organic-rich upper horizons and compacted by heavy machines or by cow trampling, increasing bulk density and reducing infiltration rates (Celik, 2005; Li and Shao, 2006). Due to their heterogeneity, natural soils respond to rainfall in many ways (Scherrer et al., 2007). Despite the progress in developing new approaches (Troch et al., 2002) and field investigations (Scherrer and Naef, 2003) on hillslope hydrology, there are only few attempts to include preferential flow in the model's structure (Weiler and McDonnell, 2007).

For grassland sites in Switzerland, Scherrer et al. (2007) suggested investigating the complex interaction between soil vegetation, field slope, soil clay content, and initial soil moisture before establishing schemes for upscaling the flow processes to the catchment scale. In addition, hydrodynamic aspects in terms of storage capacity have to be compared between grassland and forest soil since no clear tendency could be concluded from the existing literature on this topic (e.g., Badoux et al., 2006). This study is motivated by the need to develop methods that support predictive modeling of the impact of macropore flow on the overall flow processes occurring at the landscape scale under different field slopes and two different vegetation types.

The objectives of this experimental investigation were to qualitatively characterize the flow processes on grassland and forest hillslopes and to quantitatively define the relevant parameters controlling surface runoff generation. Mainly the vertically oriented soil water flow pathways are examined using infiltration and dye tracer experiments supplemented with soil textural and structural data (i.e., saturated hydraulic conductivity $K_{\rm sat}$, structural porosity $n_{\rm SP}$, and PSD).

Materials and Methods Overview and Context

The experiments were performed during summer 2008 (April-October) at the experimental area called Innerrüteni/Hälfis in Kandergrund (Fig. 1) located 1110 m above sea level on the eastern side of the Kander Valley, which spreads out in a north-south direction in the Bernese Oberland in Switzerland (Furrer et al., 1993). On the western side of the investigated area, the Kander River is located at 793 m above sea level and can have streamflow values as high as 7.7 m³ s⁻¹ (i.e., mean value for the period 1961– 980) (Schädler and Weingartner, 1992). On the eastern part of the investigated area, the glacier deposit of the Kandergrund is divided into two main compartments (Furrer et al., 1993): a Moraine till deposit on the western side and slope debris on the eastern side (Table 1, Fig. 1). The forest in which soil hillslopes were selected consists of fir-beech forest. The mean annual temperature is 5.9°C, and the total annual precipitation is 1274 mm. The region of Kandergrund has an area of 3209 ha. The cropland occupies 1571 ha, forest occupies 843 ha, and the unproductive land has a surface of 794 ha. The land use is extensive in some areas because of the restricted fertile ground and its inaccessibility. Therefore, grassland is subject to alternate use: it is grazed and used as pasture. In the experimental area, 45 d before the irrigation experiments, grassland was grazed using a tractor (Hürlimann Prestige 88). Twenty days later (15 d before our irrigations), the grassland was used for a couple of days as pasture for the cows.

On 22 Aug. 2005, the region was subject to flood of the Kander River, which recorded a streamflow of 85 m³ s⁻¹ for a daily precipitation of 71 mm measured on 21 August, followed by a second



Fig. 1. Location of the catchment under consideration (Innerrüteni, Kandergrund, Bern, Switzerland). On some sites (within the circles) only surface runoff was measured under controlled irrigations to complete the data of the surface runoff versus slope. (A) Map of Switzerland with the location of the hillslopes; (B) precise location of the forest (F1, F2, and F3) and grassland hillslopes (G1, G2, and G3)

day (22 August), with total precipitation of 48 mm causing considerable damage (SwissMeteo, 2006). In Switzerland, the flood of August 2005 caused total property damages of three billion Swiss Francs (Bezzola and Hegg, 2007). Continuing precipitations over large areas in addition to high soil saturation led to exceptionally large discharges and high lake levels. Flooding, erosion, overbank sedimentation, landslides, and debris flow deposition were the dominant damage-causing processes. When considering the entire event and a longer time period, this event is rare, but not exceptional. Therefore, such flood events of similar magnitude and extend have to be expected in the future (Bezzola and Hegg, 2007).

Site Description

The soil is described as a Cambisol (Soil Survey Staff, 2003) in grassland (G) soil and in the forest (F) soil (Table 1). Its texture consists of clay loam to a depth of 0.60 m in grassland and silt loam in forest soil to 0.75 m depth. Its organic carbon (OC) content varies from 9 to 9.6 g 100 g⁻¹ in grassland and from 8.3 to 12.5 g 100 g⁻¹ in forest soil. The soil pH values range from pH 5 to 8 in topsoil and subsoil, respectively, in both soils (Table 1). The slopes of sites F and G varied between 17.2 and 31.3° and between 10.6 and 18.0°, respectively. At both locations, the soil is relatively

Table 1. Basic properties of soils on grassland (G1–G3) and forest (F1–F3) hillslopes (Kandergrund, Switzerland).†

		Particle si	ze distributio				
Plot	Depth	Sand (>60 µm)	Silt (2–60 µm)	Clay (<2 µm)	Texture	Organic carbon	pН
	m					%	
F1	0.10-0.15	20.0	62.9	17.1	Silt loam	12.53	5.5
	0.25-0.30	16.4	61.6	22.0	Silt loam	n.a.	5.5
	0.50-0.55	14.2	51.9	33.9	Silty clay loam	10.03	8.0
	0.70-0.75	20.5	46.7	32.8	Clay loam	8.27	8.0
F2	0.10-0.15	23.2	61.1	15.7	Silt loam	6.57	6.0
	0.25-0.30	30.5	56.4	13.1	Silt loam	n.a.	7.0
	0.50-0.55	25.0	51.6	23.4	Silt loam	7.14	8.0
	0.70-0.75	23.0	44.2	32.8	Clay loam	7.66	8.0
F3	0.10-0.15	17.5	54.4	28.1	Silty clay loam	12.73	5.5
	0.25-0.30	13.0	46.2	40.8	Silty clay	n.a.	8.0
	0.50-0.55	21.1	36.9	42.0	Clay	10.46	8.0
	0.70-0.75	17.3	43.4	39.3	Silty clay loam	9.93	8.0
G1	0.05-0.10	28.2	45.1	26.7	Clay loam	n.a.	5.5
/ G2	0.15-0.20	27.9	41.4	30.7	Clay loam	8.90	5.5
02	0.25-0.30	28.9	41.8	29.4	Clay loam	n.a.	5.0
G3	0.35-0.40	38.2	36.2	25.6	Clay loam	9.57	7.5
	0.45-0.50	36.5	39.7	23.8	Loam	n.a.	n.a.
	0.55-0.60	35.0	32.4	32.6	Clay loam	n.a.	n.a.
	0.05-0.25 0.25-0.35	24.1 25.2	42.1 41.2	33.8 33.6	Clay loam Clay loam	n.a. 7.97	5.5 5.5
	0.35-0.45	25.6	46.9	27.5	Clay loam	9.61	6.0
	0.45-0.55	22.6	42.3	35.1	Clay loam	8.51	5.5

† Textural classification was according to the USDA soil taxonomy; n.a., not available.

shallow and the weathered bedrock starts at depths below about 0.30 m. The first few centimeters of grassland have higher clay contents (27–34%) than the forest soils (15–17%). The accumulation of the fine material on the soil surface is expected to be caused by the sediment transport and erosion of clay-rich material from uphill. In contrast, in forest soil, there is no or negligible erosion and sediment transport effects of the permanent forest vegetation cover. In addition, in the forest, local litter transformation, as observed in situ during the profiles excavation, promotes lower bulk density in the few centimeters topsoil layer in comparison to grassland topsoil layer.

Laboratory Analysis

The texture of samples (one sample per depth) was determined after H_2O_2 treatment to remove organic material (Konen et al., 2002). The sand fraction (2000–63 μ m) was obtained by wet sieving. The amounts of silt and clay fractions, 63 to 2 μ m and <2 μ m,

respectively, were measured on pretreated samples by sedimentation with a SediGraph 5100 (Micromeritics, Norcross, GA) using particle sedimentation rates in combination with X-ray absorption. Soil pH was measured 1:2 (soil/0.01 M $CaCl_2$) on a mass basis (Soil Survey Staff, 2004). Organic carbon content was determined by the mass loss on ignition. For all sites, only one sample per depth was used for the analysis of the pH and the organic carbon content.

For the analysis of total porosity and PSD, five undisturbed soil samples of 100-cm³ volume were taken from each soil horizon, and three horizons were considered (0.05-0.15, 0.35-0.45, and 0.55–0.65 m soil depths). Saturated hydraulic conductivity K_{sat} was determined on five undisturbed samples with a diameter of 55 mm and length of 42 mm taken between 0.10 and 0.20 m soil depth. K_{sat} was determined with a constant head permeameter (Klute and Dirksen, 1986). Total porosity was then determined directly for each undisturbed sample after drying at 105°C for 24 h assuming a particle density of 2.65 g cm⁻³. An aggregate of soil was then carefully taken from the dried, undisturbed sample that was selected to avoid desiccation cracks, root pores, or other discontinuities. The PSD of this aggregate was analyzed by mercury porosimetry (Fiès, 1992). Considered a nonwetting liquid, the mercury was forced into the dry aggregate by air pressure. The relationship between the equivalent pore diameter (D, μ m) and the applied pressure (*P*, kPa) was obtained according to the Jurin–Laplace equation (Fiès, 1992). The pressures P used in this operation varied between 4 and 2000 kPa, corresponding to a PSD varying from 360 to 0.006 μ m. Three clods of about 2 cm^3 in volume from each soil layer (0.15-and 0.25-m depths) were sampled and oven-dried at 105°C for 24 h before measurements. During the analysis, no shrinkage was observed. We defined the micropores as the pores smaller than 0.2- μ m equivalent diameter, mesopores between 0.2 and 50 μ m in diameter, and macropores larger than 50 µm (Sekera, 1951; Luxmoore, 1981). At the microaggregate scale, the upper limit of macropore diameters was found to range at about 120 μ m for both grassland and forest soil samples.

Total porosity $(n_{\rm T})$ was determined on 100-cm³ samples. Total volume of micropores, mesopores, and macropores was obtained from the microaggregate analyses with mercury porosimeter (c.f., Dexter et al., 2008; A. Alaoui and J. Lipiec, unpublished data, 2010). The structural porosity $(n_{\rm SP})$ is defined here as the difference between total porosity of the 100-cm³ sample and the total volume of the sum of the three pore classes as:

$$n_{\rm SP} = n_{\rm T} - (n_{\rm ma} + n_{\rm me} + n_{\rm mi})$$
 [1]

with $n_{\rm ma}$ being the volume of the macropores, $n_{\rm me}$ the volume of mesopores, and $n_{\rm mi}$ the volume of micropores.

In situ Field Irrigation Experiments

Six soil plots of 1-m² area (3 in grassland and 3 in forest hillslopes) were selected to carry out the sprinkling experiments. To

investigate the impact of the rainfall intensity on runoff generation, three different intensities were chosen: 24, 36, and 48 mm h^{-1} . Soil moisture was measured using time domain reflectrometry (TDR) (CR10X & TDR100, Campbell Scientific, Logan, UT), with 0.20-m-long wave guides (two parallel rods 6 mm in diameter). The calibration was performed according to Roth et al. (1990). This calibration consists of separating the impact of the wave-guide geometry from the soil properties, such as bulk density and the contents of clay and organic matter, on the dielectric constant. The TDR probes were inserted horizontally at four soil depths (0.12, 0.27, 0.52, and 0.72 m) in the three forest hillslopes (F1, F2, and F3), and at three soil depths in the three grassland hillslopes (0.07,0.17, and 0.40 m for G1 and G2 and 0.15, 0.25, and 0.35 m for G3). The instrument noise was set at $d\theta = 0.018 \text{ m}^3 \text{ m}^{-3}$, and any variation in θ measured using wave guides that exceeded the value of \pm d θ , was considered significant (Alaoui and Goetz, 2008). The TDR data were recorded every 60 s. On the 1-m² hillslopes, irrigation was supplied by a rainfall simulator consisting of a metallic disc with a surface of 1 m² perforated with 100 holes attached to small tubes that lead into a reservoir. Irrigation was applied from a height of 0.50 m from the ground. An electric motor moved the metallic disk, and a flow meter controlled the irrigation intensity (Alaoui and Helbling, 2006). Each irrigation event lasted 1 h.

Surface runoff out of each of 1-m^2 hillslope was measured during the irrigation, using a metallic sheet (1×0.50 m) inserted in the soil profile at 0.05 to 0.10 m depth to collect surface runoff along a width of 1 m. The volume of collected water was measured with a flowmeter and stored automatically in a datalogger (CR10X, Campbell Scientific).

In addition to G1 through G3 and F1 through F3, some hillslopes with various slopes were selected (within the circles, Fig. 1) to carry out additional irrigations to complete the paired data: surface runoff versus slope.

Dye Tracer Experiments

A dye infiltration experiment was performed at all forest (F1, F2, and F3) and grassland (G1, G2, and G3) hillslopes to visualize the flow pathways. A dye tracer solution of 110 L was prepared by diluting 440 g of Brilliant Blue (BB) FCF powder, known as food-dye E133 (Flury and Flühler, 1995), in ordinary tapwater to obtain the tracer solution with a concentration of 4 g L⁻¹. Being neutral or anionic, BB is not strongly adsorbed by negatively charged soil constituents. However, adsorption of BB may occur in clayed soils, which diminishes its suitability for tracing the travel time of water itself (Flury and Flühler, 1995; Ketelsen and Meyer-Windel, 1999). In our case, the adsorption of the BB is not relevant since we used it for tracing the flow paths and not for travel time or water infiltration depth.

At each site, 100 L of the BB solution was applied to the soil surface according to Alaoui and Goetz (2008) for 2 h and 47 min at a constant intensity of 36 mm h⁻¹. Twenty hours after irrigation, a soil pit was excavated, and six vertical profiles (0, 0.2, 0.4, 0.6, 0.8, and 1 m within a surface square of 1 m²) were prepared. Rubber string grids of 1 by 0.7 m and 1 by 1 m were attached in front of each profile in grassland and forest hillslopes, respectively. The photos of soil profiles were taken with a digital camera (Hp Photosmart 945; resolution: 5 megapixels). Resulting digital images had a resolution of approximately 2200 by 2200 pixels.

Image Analysis and Optical Calibration

The percentage of the coverage of the stained areas was determined with Photoshop CS2 image editing program using the procedure described by Alaoui and Goetz (2008). The maximization of the saturation of the blue stains resulted in three colors (yellow, green, and blue) and two tinges (light and dark blue). The brown color corresponding to the unstained soil areas was then removed, resulting in four BB related tints in the image. The colors were then separated by cropping them successively and pasting them into individual JPG files. The dyed areas in the pictures were then turned black by tone value corrections, while the rest stayed white. The resulting patterns were then cut into 30 to 35 horizontal strips, and the distribution of the dye coverage was calculated for each depth and each image by horizontally counting the pixels (e.g., Forrer et al., 2000).

A calibration was performed to link specific colors with corresponding ranges of dye concentrations to allow a more quantitative perception of the flow patterns. For this purpose, 10 standard solutions with different BB concentrations (0.1, 0.5, 1, 2, and 4 g L^{-1}) were prepared. The soil samples were taken in situ from top- and subsoil and saturated therein for 1 d. After letting them dry for one night, each sample was photographed with the same camera and under the same light conditions as in situ (soil samples illuminated with diffusive daylight using the same white parasol). The mean concentration of all six vertical profiles of each hillslope was calculated for each concentration class and each corresponding depth layer.

By using the scheme proposed by Weiler and Flühler (2004), five different flow types were categorized in respect with SPW profiles. Supposing an isotropic distribution of the stained objects at a given depth in a three-dimensional axis, the SPW can be deduced in a one-dimensional extension of the dye patterns (Weibel, 1979). Therefore, at a given depth in a vertical dye pattern, the width of the stained object may be used as a proxy for its size at this depth. The analysis of the flow patterns according to this method results in a frequency distribution of SPWs for each soil layer (Weiler and Flühler, 2004).

Results and Discussion

Surface runoff was observed at the soil surfaces of all three grassland hillslopes and of only a single forest hillslope, F3. This hillslope (F3) had runoff coefficients (RC) of 0.18, 0.13, and 0.19 for irrigation

intensities of 24, 36, and 48 mm h⁻¹, respectively (Fig. 2). In contrast, no runoff was collected on the other two forest hillslopes (F1 and F2). The values of RC for grassland were 0.38, 0.66, and 0.57 for site G1; 0.19, 0.25, and 0.44 for G2 (with irrigation intensities of 24, 36, and 48 mm h⁻¹, respectively); and 0.38 for G3 (with the intermediate irrigation intensity of 36 mm h⁻¹). While the forest hillslope with the steepest slope (Fig. 2) produced some runoff, the grassland hillslope with the steepest slope (G2, 14.8°) had a lower runoff than G1 (10.6°), which generated higher RC.

As confirmed in several studies, field slope is a key parameter influencing runoff generation (e.g., Markart et al., 1997; Scherrer and Naef, 2003). Accordingly, we analyzed the correlation between these parameters measured in the areas under consideration. Figure 3 (illustrating RC versus slope) shows that field slope alone does not control runoff generation especially for slopes with intermediate steepness of between 10 and 18°.



Fig. 2. Runoff coefficient (RC) of the plots in forest and grassland soils for different slopes and intensities (24, 36, and 48 mm h^{-1}).



Fig. 3. Runoff coefficient (RC) vs. field slope in grassland soils. Additional sites were considered in this correlation (see Fig. 1); the values of RC were measured for at least three different intensities (24, 36, and 48 mm h^{-1}).

Below, we aim to explain (i) why only forest hillslope F3 with the steepest slope generated surface runoff and (ii) what caused the high RC values in the grassland plot with the gentlest slopes, by analyzing the soil pore system of grassland and forest hillslopes.

The pore volume distribution obtained from the analysis of mercury porosimetry shows a notable difference between forest and grassland soils (Fig. 4). Compared to the forest hillslopes, the micropore volume of the grassland hillslopes is higher in the topsoil down to 0.35 m (Fig. 4A). With regard to mesopores and macropores, the differences between forest and grassland soils are minor (Fig. 4B and 4C). Examination of the soil textural analysis of all hillslopes (Table 1) reveals that clay content is higher in the topsoil layers of the hillslopes that generate surface runoff (F3, G1, G2, and G3) than in the topsoil layers of the hillslopes without surface runoff (F1 and F2). Clay content is linearly correlated with micropore volume for all 14 hillslopes (Fig. 5A). This relationship is in agreement with results obtained from examining PSD in the topsoil layer in the grassland hillslopes (Fig. 4A). One possible explanation is that the high clay content in the topmost 0.10 to 0.15 m promotes matrix flow and consequently delays water routing into macropores by sorptivity promoting such surface runoff (Fig. 5B).

Examination of $K_{\rm sat}$ measured at the depth of 0.10 m highlights a negative correlation: RC decreases with higher values of $K_{\rm sat}$. A threshold exists between 1200 and 1700 mm h⁻¹ beyond which no runoff occurs at these precipitation rates. This is the case with F1 and F2 (Fig. 6). Examination of $K_{\rm sat}$ yields another observation related to the fact that the hillslope with the gentlest slope (G1, 10.6°) generated larger runoff than G2 (14.8°) (Fig. 2): G1 has



Fig. 4. Pore volume distribution vs. depth in forest and grassland hillslopes for (A) micropores, (B) mesopores, and (C) macropores.

a lower K_{sat} value (180 mm h⁻¹) than G2 (1200 mm h⁻¹). This order-of-magnitude difference may explain the fact that G1 has a higher RC than G2 (Fig. 6).

In this study, structural porosity, n_{SP} was defined according to Eq. [1] for a topsoil layer of between 0.05 and 0.15 m for all hillslopes



Fig. 5. Relationship between (A) micropore content related to the soil volume (%) and clay content (%) and (B) runoff coefficient and micropores (mm³ g^{-1}) for topsoil (0.10–0.15 m depth).



Fig. 6. Runoff coefficient (RC) vs. saturated hydraulic conductivity (K_{sat}) measured between 0.1 and 0.2 m for forest and grassland soils and three irrigation intensities.

under consideration (Fig. 7). The values obtained for forest soil are higher than those for grassland soil, especially at F1 and F2. The $n_{\rm SP}$ values for the deeper soil layers did not differ to such an extent between grassland and forest or between the individual hillslopes. The results for the topsoil layer are plausible inasmuch as the root system in the forest soil is characterized by a dense and spatially variable structure, causing a well-developed pore system that has considerable effects on water movement (Jarvis, 2007; Legout et al., 2009). Whereas, in general, the upper decimeters of most agricultural soils are redistributed annually due to seasonal tilling operations, the structure of soil forest, in particular the macropores and biopores, may remain unchanged for decades, and may be continuous throughout entire soil profile.

For the following analysis, five different flow types we categorized using SPW profiles according to their dye coverage (DC) percentage (Table 2). Three of them are related to macropore flow, and two of them involve matrix flow.

The classification of flow types according to Weiler and Flühler (2004) suggests that homogeneous matrix flow exists at least down to a depth of 0.05 m in all hillslopes (Fig. 8). For the near-surface soil between 0 and 0.10 m depth, our analysis is unable to differentiate between different soils that generate surface runoff and those that do not. Below 0.10 m depth, there appears to be macropore flow with high interaction with the surrounding soil matrix in the forest hillslopes and with low to mixed interaction in the grassland hillslopes. This classification of flow types based on interpretation of the infiltration and runoff processes corresponds with qualitative observations from photo images of the cross-sections (Fig. 9). This indicates that macropore flow, which was more evident in the grassland than in forest soils, may be reduced by the combined effects of both slope and the finer and denser structure in the topsoil, which limits water flux into the underlying macropores. This hypothesis is derived from soil texture investigations and requires more detailed study, including confirmation of K_{sat} and $n_{\rm SP}$ measurements.

Using a more quantitative approach, we compared $t_{\rm SSF}$, the start of surface runoff (time of the first drop of water that arrives in the collector), with $t_{\rm w}$, the time of reaction of the first TDR probe, located at any depth of grassland hillslope that react to the irrigation (Table 3). In general, it can be stated that water arrives slightly earlier in the collector than at a certain soil depth on all plots of grassland except G2 during the first irrigation (intensity of 24 mm h⁻¹). In this case, only the uppermost layer down to a depth of 0.07 m becomes water saturated before the initiation of surface runoff, suggesting shallow saturation excess overland flow (0–0.07 m). These observations confirm that in grassland, water is in general routed laterally on the soil surface during the first 10 min of irrigation rather than flowing vertically downward. In forest hillslope F3, inverse phenomena occurred especially for the irrigations of the



Fig. 7. Structural porosity (n_{SP}) vs. slope angle for three depths (0.05–0.15, 0.35–0.45, and 0.55–0.65 m) in forest (F1–F3) and grassland (G1–G3) soils. TP is total porosity; values above the bars represent structural porosity.

Table 2. Classification of flow types in soil using dye pattern according to Weiler and Flühler (2004) based on the stained pathway width (SPW) and their dye coverage (DC).

Flow type	SPW 1 <20 mm	SPW 2 >200 mm
	%	ó ———
Macropore flow with low interaction with surrounding matrix	>50	<20
Macropore flow with mixed interaction (high and low)	20-50	<20
Macropore flow with high interaction	<20	<30
Heterogeneous matrix flow and fingering	<20	30-60
Homogeneous matrix flow	<20	>60



Fig. 8. Flow types in the forest (F1-F3) and grassland (G1-G3) soil profiles derived from the dye tracer analysis; classification according to Weiler and Flühler (2004).

Table 3. Differences between arrival times of the wetting front (t_w) corresponding to the reaction time of the first TDR probe that react to the irrigation and arrival times of the surface runoff at the collector (t_{SSF}) for soils on grassland (G1–G3) and on forest F3 hillslopes and three irrigation intensities.

	G1			G2			G3	F3		
Intensity (mm h ⁻¹)	24	36	48	24	36	48	36	24	36	48
$t_{\rm w}(s)^{\dagger}$	1020	1020	1140	900	600	660	2100	1140	600	480
$t_{\rm SSF}\left(s\right)$ ‡	900	840	510	1080	540	570	1200	600	880	1330
Difference (s)	120	180	630	-180	60	90	900	540	-280	-850

 $\dagger t_{w}$, the time of the first TDR that react in response to the irrigation (see Fig. 10). $\dagger t_{CSE}$, the arrival time of the first drop of shallow water to the collector.

intensities of 36 and 48 mm h^{-1} showing such shallow saturation excess overland flow.

In the forest hillslopes, below the layer of homogeneous matrix flow, it appears that mixed flow types (macropore flow with high interaction and heterogeneous matrix flow and fingering) dominate (Fig. 8). Vertically downward percolation prevents the initiation of runoff. This does not apply to plot F3 though, probably due to its steep slope (31.3°) in comparison to F1 and F2 and its relatively low structural porosity. Although macropore flow with low interaction occurs in grassland soils, there is no infiltration dominating flow processes (Fig. 8). The low efficiency of macropores can be explained by several facts: (i) the combined effects of the slope and the finer and denser structure of the top few centimeters of the soil layer caused by the land use, which may limit water flux into the underlying macropores, thus, giving more time to surface runoff to occur; and (ii) their restricted number and their tortuosity, preventing efficient vertical infiltration. Although the first point is supported by TDR measurements ($t_{SSF} < t_w$; Table 3) and soil texture (Table 1), there is a need for additional investigations (i.e., unsaturated hydraulic properties) of the soil, especially of the soil layer between 0 and 0.10 m, to confirm this hypothesis.

Analysis of the progression of the wetting front on the basis of TDR measurements indicates that the front propagates homogeneously from the topsoil to the deeper soil layers in the forest, as it is characteristic for matrix flow (Fig. 10). In contrast, an inverse phenomenon occurs in grassland soils, marked by a heterogeneous distribution of the wetting front as is characteristic for macropore flow. In fact, the wetting front arrived slightly earlier in deeper soil layers than in the top-most 10 cm. Analysis of t_w in different depths of between 0 and 0.40 m reveals that t_w is higher in F2 than in F1 and F3. This is in agreement with the type of flow indicated by the classification of Weiler and Flühler (2004). In F2, homogeneous matrix flow dominates down to a depth of 0.18 m, followed by heterogeneous matrix flow and fingering down to a depth of 0.27 m, thus delaying the progression of the wetting front. In F1 and F3, these types of flow are found only in the top-most 10 cm.



Fig. 9. Photo images of the horizontal cross-sections in (A) forest soil F2 at the 0.20-m depth and (B) in grassland soil G1 at the 0.20-m depth.

We analyzed the change in volumetric soil moisture content, $\Delta \theta$, in response to irrigation and relative to initial soil moisture, θ_{init} , measured before irrigation. The value of $\Delta \theta (= \theta_{max} - \theta_{init})$ is equivalent to the difference between maximum soil moisture, θ_{max} , during infiltration and θ_{init} measured at the same location and shows the magnitude of the increase of soil moisture in the soil layer under consideration.

The effect of θ_{init} on $\Delta \theta$ is studied in the top- and subsoil (Fig. 11). In reference to the high values of total porosity of forest soil (Fig. 7), results clearly show higher storage capacity of forest hillslope (Fig. 11A–D). The value of $\Delta \theta$ is negatively correlated with RC. This correlation may show—as expected for uniform flow—that the drier the soil, the greater its sorptivity, consequently generating less



Fig. 10. Vertical progression of the wetting front in the soils of the forest (F1-F3) and grassland (G1-G3) hillsopes. Dark zones indicate the first TDR probes that react to the irrigation, the white zones indicate the last TDR probes that react to the irrigation, and the numbers indicate the time (in seconds) of the TDR probes reaction.

surface runoff (Fig. 11E). Any influence of external weather factors can be ruled out because the infiltration experiments were performed during the same periods in the summer season, excluding any precipitation events during at least 1 wk. Although these observations are not directly linked with flow process, they allow the larger root uptake in forest soil to be considered, in addition to the structural porosity, as an additional factor enhancing storage capacity. For structural soil with preferential flow in macropores, however, the effect of a soil water saturation deficit on infiltrability and surface runoff generation can be limited by reduced interaction between macropoes and soil matrix as in grassland. In forest soil, the high interaction between macropoes and soil matrix creates homogeneous distribution of flow patterns (Fig. 8 and 9). These conditions may be favorable for enhancing the effect of soil infiltrability. Similarly, Badoux et al. (2006) reported that irrigation experiments performed on dry to humid Cambisols result in no or only low RC values varying between 0.01 and 0.16. They reported higher RC values after dry antecedent conditions than after wet antecedent conditions, probably due to water repellence. In contrast, high precipitation intensity on 1-m² hillslopes leads to high RC values (from 0.39 to 0.94) on humid to wet Gleysols (Badoux et al., 2006). The draining properties may also be due to the root water uptake in forest soils. These results (high values of RC)





Fig. 11. (A–D) Magnitude of the increase of soil moisture ($\Delta\theta$, m³ m⁻³) vs. θ_{init} (m³ m⁻³) for forest and grassland hillslopes at top- and subsoil and (E) runoff coefficient vs. $\Delta\theta$ (m³ m⁻³) in the topsoil; *R: significant at the 95% level for data obtained from all irrigation intensities.

are rather unexpected for forest soils, which have generally high infiltrability, preventing surface runoff (e.g., Schwarz, 1986; Kohl et al., 1997; Markart et al., 1997). The differences in RC values especially for forest hillslopes found in the literature indicate great variability in runoff generation that depends mainly on soil conditions and vegetation types. Taking into account these limitations, it would be necessary to repeat similar sprinkling experiments for different soil types and/or soil vegetations.

Conclusions

The objectives of the experimental investigation presented in this article were to qualitatively characterize the flow pathways on grassland and forest hillslopes and to quantitatively define the relevant parameters controlling surface runoff generation using both infiltration and dye tracer experiments supplemented with soil textural and structural data (i.e., saturated hydraulic conductivity K_{sat} , structural porosity n_{SP} , and pore size distribution PSD).

The marked differences in the textural and structural porosities especially of the topsoil between forest and grassland hillslopes appear to control runoff processes. Forest soil structure, characterized by relatively high values of K_{sat} and n_{SP} (F1 and F2), enhances the infiltrability of the soil and consequently prevents or at least reduces surface runoff. In this case, it is supposed that the top soil layer (between 0 cm and about 0.05 to 0.10 m) that has a lower density promotes vertical percolation to the underlying horizons. The dye patterns suggest that macropore flow with high interaction with the surrounding soil matrix occurs in forest soil, while macropore flow with low to mixed interaction with the surrounding soil matrix dominates in grassland soil (in the upper part of the subsoil). The low efficiency of grassland soil macropores in transporting all water vertically downward can be explained by (i) the combined effect of the slope and the finer and denser structure of the top few centimeters soil layer caused by land use that limits water flux into the underlying macropores, (ii) their restricted number, (iii) their tortuosity, and/or (iv) the restricted interaction between macropores and the matrix below the topsoil layer. Therefore, it appears that combined effects of both structural and topsoil textural porosities reduce the influence of field slope on runoff generation.

The results suggest an increased storage capacity of forest soil as compared to grassland soil, possibly caused by the more intense root water uptake by the trees. The enhancing effect of soil water storage capacity on infiltrability (e.g., of the drier forest soil) is only valid for homogeneous soil with uniform flow patterns. For structured soil with preferential flow in macropores, however, the effect of a soil water saturation deficit on infiltrability and surface runoff generation can be limited by reduced interaction between macropores and the soil matrix. Thus, for the grassland hillslopes, the quantification of the effects of low macropore–matrix interactions on surface runoff remains unclear and poses a challenging task for future studies.

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