Characterization of subsurface fluxes at the plot scale during flash floods in the Valescure catchment, France

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Abstract

This study focuses on a 10-m^2 plot within a granitic hillslope in Cevennes mountainous area in France, in order to study infiltration and subsurface hydrological processes during heavy rainfalls and flash floods. The monitoring device included water content at several depths (0-70cm for the shallow soil water; 0-10m for the deep water) during both intense artificial and natural rainfall events, chemical and physical tracers, time-lapse electrical resistivity tomography. During the most intense events, the infiltrated water was estimated to be some hundreds of millimeters, which largely exceeds the topsoil capacity (≤40 cm deep in most of the cases). The weathered/fractured rock area below the soil clearly has an active role in the water storage and sub-surface flow dynamics. Vertical flow was dominant in the first 0-10m, and lateral flow was effective at 8-10 m depth, at the top of the saturated area. The speed of the vertical flow was estimated between 1 and 10 m/h, whereas it was estimated between 0.1 and 1 m/h for the lateral flow. The interpretation of the experiments led to a local pattern of the 2D-hydrological processes and profile properties. It suggests that fast triggering of floods

at the catchment scale cannot be explained by a mass transfer within the hillslope, but should be due to a pressure wave propagation through the bedrock fractures, which allows exfiltration of the water downstream the hillslope.

Keywords: subsurface fluxes, soil hydraulic properties; plot experiment; flash floods, Mediterranean climate

1. INTRODUCTION

Understanding hillslope runoff response in extreme rainfall conditions has an essential place in hydrology for flash floods predictions. Among the main hydrological processes, the subsurface flow was recognized as a first-order control of the runoff generation on most of the hillslopes in the world (Uchida et al., 2004), whether it contributes directly to the flooding (Beven and Germann 2013, Gotkowitz et al., 2014, Gonzalez-Sosa et al., 2010, Pinault et al., 2005) or it governs the water content of the soil, which is determinant in the response of the catchment (Borga et al., 2007, Le Lay and Saulnier, 2007, Tramblay et al., 2010). An increasing number of authors argue that subsurface water could contribute very fast to the streamwater flow, faster than the hydraulic conductivity of the material and a classical Darcy model would allow. This is due to the propagation of a pressure wave instead of a mass transfer, which needs to consider the celerity of the pressure wave instead of the velocity of the water flow (Beven and Gemann, 2013; McDonnell and Beven, 2014).

Sub-surface flow is controlled by the hydrodynamic properties of soils/bedrock, i.e. hydraulic conductivity and water retention; soil geometry (slope, layer thickness); bedrock permeability and topography (Tromp-Van Meerveld et al., 2006b; Gabrielli et al., 2012), but these properties are often poorly known, namely in mountainous areas, where the lack of agricultural activity did not make so urgent the study of soils.

Several studies of the sub-surface processes have been conducted at small scale under natural rainfalls (Tromp-van Meerveld and McDonnell, 2006a, Graham et al., 2010, Wenninger et al., 2008) or artificial rainfalls (Kienzler and Naef, 2008, Fu et al., 2013, Fu et al., 2015, Gevaert et al., 2014, Schneider et al., 2014; Scaini et al., 2018). For example, Kientzler and Naef (2008) conducted controlled sprinkling experiments over plots of 13m x 8m with high precipitation intensities to observe the subsurface stormflow formation. Graham et al. (2010) combined irrigation and excavation measurements in the Maimai Experimental Watershed in New Zealand. Fu et al. (2013) showed that in a granitic catchment in southern China (plot experiment of 5m x 10 m), subsurface flow occurs at the soil-bedrock interface when certain thresholds of precipitation and initial soil moisture conditions were exceeded. All these works proved that the role of the sub-surface flow is determinant when exploring runoff processes. However, most of them are site-specific and climate-specific and there is a need to expand the results to different climatic, pedological and geological conditions.

Our region of interest is the mountainous Cevennes-Vivarais region in south-eastern France, which is known to be prone to flash floods. The study site is the Valescure granitic catchment, one of the densely monitored catchments of the FloodScale project (Braud et al., 2014), which is part of the HyMeX (Hydrological Cycle in the Mediterranean Experiment, Drobinski et al., 2013 and Ducrocq et al., 2014).

In this region, the subsurface flow was found to be relevant especially in forested and granite catchments (Cosandey and Didon-Lescot, 1990, Tramblay et al., 2010, Adamovic et al., 2015). The infiltration tests (Ayral et al., 2005; Le Bourgeois et al., 2015) of a topsoil show high infiltration capacity (some hundreds of mm per hour) and scarcity of surface runoff. However, no data were yet available up to now at small scale, in this area, and our experiment is the first one to give real physical insight into hydrological processes.

In this study, the sub-surface flows were monitored under artificial and natural rainfalls over a 10-m^2 plot using water content and pressure head probes, time-lapse electrical resistivity tomography (ERT), chemical tracers and piezometers (one of them located downstream of the plot). This kind of device can be compared to others like Kientzler et Naef, (2008) for example. The artificial rainfalls are useful because they can be considered as extreme events, which cannot be observed frequently by definition. It is also straightforward to check the estimated soil properties with natural rainfalls, to control that the estimated parameters are robust from the normal to the extreme, and independent of specific boundary conditions (e.g. no upstream flux in artificial conditions).

The objective of this paper is thus to study both vertical and lateral fluxes by means of experimental work. We focus on the following questions: (1) what is the depth of the active area, where water flows through topsoil and weathered/fractured bedrock, (2) which are the respective vertical or lateral fluxes through the different layers of the active area, (3) what are the main characteristics of these fluxes (e.g. velocity), (4) what kind of hydrological process does it suggest for the generation of the flash floods at the catchment scale? To answer those questions, we first introduce the experimental device (section 2), then the results leading to the characterization of both vertical and lateral fluxes which occurred along the hillslope during the rainfalls (section 3). Finally, a local pattern of the 2D-hydrological processes is proposed, which suggests that pressure wave along the hillslope could be the main process generating flash floods on this kind of catchment (section 4).

2. DATA AND METHODS

2.1 The Valescure catchment

The study area is the Valescure catchment, in South-eastern France (Figure 1). This small catchment (3.9 km^2) is typical of the mountainous area in the south of the Cevennes region. The catchment ranges from 200 to 800 m in elevation and is drained by one tributary. The mean slope of the catchment is 0.36 m.m^{-1}. The catchment is characterized by granitic geology made of chaotic and heterogeneous near-surface structures (mixture of consolidated rocks, weathered areas and disaggregated sandy arenas). The topsoil is detailed in section 2.2. The land cover is mainly green oak, chestnut and pine forests.

The mean annual precipitation in this area is 1580 mm/year (over 2003-2014 period at *Château* rain gauge, see Figure 1), and daily rainfalls exceeding 200 mm are frequent (~2 years return period), occurring mainly in fall season. During the period 2003-2014, the maximum peak flow measured at the outlet of the catchment was 16 m^3/s, and was supposed to be more than 40 m^3/s in October 2006, after a 400 mm rainfall in 36h (the most intense part occurring at the end of the event) which damaged the recorder. Extreme specific peak flow discharges are usually estimated to reach 20 m^3/s/km^2 in such catchments (Delrieu et al., 2005; Brunet et al., 2012).

The response times of the floods (i.e. the differences in time between the maximal intensity of the rainfall and the peak flow) range between 1 and 3 hours at the outlet of the Valescure catchment. The lag-times of the floods (i.e. the difference in time between the centres of inertia of both rainfalls and discharges) at the Valescure outlet range between 4 and more than 70h (25h in average), and are strongly correlated to the antecedent moisture conditions (Tramblay et al., 2010). These high values (for this small area with steep slopes) suggest that subsurface flow is an important component of the flood.

[Insert Figure 1]).

2.2 Soil and sub-soil properties

The soil is classified as a brunisol with a rich Ah humic horizon in the first 10 centimetres overlaying first a 0-1m thick root layer with loamy texture, small rocky fragments, and second the weathered or consolidated granite bedrock. Granulometry analysis was derived from the Hymex database (http://mistrals.sedoo.fr/HyMeX/) containing 40 samples at depth 30 cm in average, in the Valescure catchment : it resulted in 54% coarse fragment > 2 mm (s.d. 10%), 9% coarse sand between 0.2 and 2 mm (s.d. 3%), 17% sand and silt (s.d. 10%), 5% clay < 2 µm (s.d. 2%) and 15% organic matter (s.d. 8%). Using 100 cm^3 steel cylinders, 18 soil samples have been taken from surface to 40 cm every 10 cm deep. The measured density was 1.1 $g.cm^{-3}$ at the surface, 1.35 at 40 cm depth; porosity was 0.60 at the surface, and 0.50 at the 40 cm depth.

In order to determine the saturated hydraulic conductivity (*Ks*) of the soil, 10 measurements were conducted in Valescure catchment with disk infiltrometer at the soil surface or near-surface (depth of 0 m, 0.10 m and 0.15 m, Le Bourgeois, 2015). These measurements resulted in *Ks* exceeding 500 mm/h near saturation, which reflects the high hydraulic conductivity in these soils. Water retention data were obtained from 25 measurements using tensiometers installed across the Valescure catchment (including near the experimental plot) at the depth of 30 cm. The Mualem-Van Genuchten equation was fitted to the observed data, by using nonlinear least squares analysis (Figure 2). The following parameters were obtained: $\theta_r = 0.05\, cm^3/cm^3$, $\theta_s = 0.50\, cm^3/cm^3$, $\alpha = 0.04\ mm^{-1}$ and $n = 1.3$. It resulted in low RMSE value of 0.03 and coefficient of determination (R^2) of 0.97.

[Insert Figure 2]

2.3 Experimental set-up

Experimental plot

The experimental plot (lat = 44°05’39”N, long = 3°49’29”E) was set up from the 18/04/2014 to the 12/09/2016. It was located in a 30° hillslope, over a 10m^2 rectangular area (2.50 m x 4m, the longer along the slope), see the location in Figure 1. The area of the hillslope upstream the plot is unknown, it could be some hundreds, even thousands of square meters.

The sprinkling device was made of four 2.50 x 1 m^2 panels through which the rainfall was delivered by a drop by drop device (photo 1). The water was injected by 2 pipes for each panel, and the rainfall intensity delivered in each pipe was measured by a gauge. At the total, 8 gauges worked independently, controlling the rainfall intensity of both the right and left part of each panel.

Piezometers

Piezometric levels were investigated using 5 piezometers. Four of them were placed along the central longitudinal axis of the plot as shown in Figure 3 with respective depths: P1 (90cm), P2 (70cm), P3 (80cm) and P4 (70cm), diameter 40mm, draining the water along the whole depth of the piezometer. An additional 12m-deep P5 piezometer was installed in November 2015, 4 m downstream the lower edge of the plot, in a 150 mm diameter borehole. The piezometer was made of a 52 (inner)-60 mm (outer) diameter PVC tube, and the borehole was filled around the tube with concrete (0-2.5m) and gravels (2.5-12m). The 2.5-12m depth was equipped with a trainer associated to a geotextile membrane, able to drain water all along this depth. The borehole log was made of a first 0-2m layer of silts and granite fragments, then a 2-10m layer of grey weathered granite without water at the moment of the borehole, then a 10-12m layer of brown weathered granite with small entry of water. The bottom of the P5 piezometer was nearly at the same elevation than the Cartaou stream, some tenth meters downstream the plot. This stream generally stops flowing at the beginning of summer.

The piezometric levels in P1 to P4 were monitored between May 2014 and September 2016, at a 1-mn time step during the artificial experiments and at a 15-min time step during other periods, with Paratronic CNR probes. The P5piezometer was monitored from 27/11/2015 up to now, at a 15-mn time step, with a OTT mini Orpheus probe. An additional CTD Solinst probe was settled in P5 before the beginning of the artificial rainfall, 14/04/2016 until 27/05/2016.

<u>Soil water probes</u>

Soil water content was measured by 10 Thetaprobe ML2x (Delta devices Ltd) humidity probes. Probes were located along 4 lines across the slope with the following depths, (Figure 3):

- **Line 1** (near P1) : 30, 30, 35 cm
- **Line 2** (near P2) : 30, 50 cm
- **Line 3** (near P3) : 45, 30, 70 cm
- **Line 4** (near P4) : 50, 30 cm

[Insert Figure 3]

The humidity probes were settled at the base of vertical cylindrical boreholes, of diameter 40 mm. The boreholes were perforated by using an electric-powered drill with threaded stem at the end, which allowed to dig not only through the soil, but also through the weathered area and sometimes (although slowly, ~5mm/min) through the rock. In most of the cases, the probes have been located in the topsoil layer. In the other cases, a small volume of soil has been added at the base of the borehole, in order to settle the probes in correct conditions for the contact of the electrodes with soil.

[Insert Photo 1]

Since May 2014, soil water content was recorded at a 1-mn time step during the artificial experiments and at a 15 or 30-min time step during other periods (15 mn after the 15/04/2016). The calibration given by the constructor for mineral soils was directly used here. This calibration proved to be satisfactory for Valescure soil: one sample of undisturbed soils was collected at the field, within a 1-liter cylinder; the sample was then saturated and weighted during its desaturation, during a 30-days period. The humidity volume difference determined by weighting and by the ML2 probe remained less than 0.03 $cm^3.cm^{-3}$, for a range of water content extending from 0.05 up to 0.40 $cm^3.cm^{-3}$ (Le Bourgeois et al., 2015).

During the installation of the piezometers and the soil water probes on the experimental plot, a specific attention has been paid to maintain a natural surface appearance in the neighborhood of the probes to avoid the formation of preferential flows (deep or surface) during artificial or natural rainfalls.

Electrical Resistivity Tomography (ERT)

ERT was also performed in order to estimate the water content and depth of the different layers in the vertical profile. Three electrodes lines have been put along the plot, two 24-electrodes lines across the slope direction, one 48 electrodes line along the slope direction (Figure 3). Thin (Ø 5mm) electrodes were 20 cm spaced, and a Wenner-Schlumberger device was used for apparent resistivity computation. Inverse modelling of the ERT was performed by using RES2DInv software.

Chemical and physical tracers

To assess the water flow directions in the sub-surface, salt was injected in the upstream part of the plot (see Tracer injection on Figure 3) during the artificial rainfalls. The flow direction was then tracked via time-lapse ERT measurement after the tracer injection. Twenty (20) liters of salt solution were injected by filling a PVC cylinder (diameter 20 cm) located near the electrode line and near the piezometer P1. The PVC cylinder was sunk 5-10 cm into the soil.

Fluorescein (150g, diluted in 5 liters of water) was also injected during the 15/04/2016 artificial rainfall, directly in the piezometer P4. Fluorescein was monitored in the deep piezometer P5, by using a flow-through field fluorimeter Albillia FL24, with a 4mn time step resolution during the artificial rainfall.

Electrical conductivity (EC) and temperature were monitored in the surface piezometers P1-P4 (Campbell CS 547 probes) and in the deep piezometer P5 (CTD Solinst probe) at a 15mn time step. The dates of the monitoring in P1-P4 were nearly the same than those of the water levels monitoring. EC and temperature were monitored in P5 from 14/04/2016 until 27/05/2016 (time step 1mn during artificial rainfalls, 15 mn else, temperature resolution 0.1 °C, EC resolution 0.001 $mS.cm^{-1}$), then from 06/07/2016 up to now (time step 15 mn, temperature resolution 0.1 °C, EC resolution 0.01 $mS.cm^{-1}$).

Rainfalls

Three artificial rainfalls were performed: on 26/03/2015, 27/03/2015 and 15/04/2016, the latter after that P5 was installed. During the 26/03/15, a total amount of 186 mm was delivered between 13:40 and 15:40 TU; during the 27/03/15, a total amount of 477 mm was delivered between 09:20 and 14:20 TU. During the 15/04/16, a total amount of 702 mm of rain was delivered between 8:10 and 14:10 TU. The different experiments gave similar results, which will be presented below only for the 15/04/16, the most complete experiment including deep piezometric levels in P5. The rainfall intensity was around 100 mm/h from 8.10 TU to 13.10 TU, and 200 mm/h from 13.10 TU to 14.10 TU. The soil was nearly semi-saturated at the beginning of each artificial rainfall. The fluorescein solution was injected into P4 at 9:21 TU, and the salt solution near P1 at 10:47 TU.

The natural rainfalls were recorded in the Chateau rain gauge (lat = 44°05’43”N, long = 3°49’46”E) with a tipping-bucket pluviometer (Précis Mécanique, 1000cm^2) located 400m in the east of the plot (see Figure 1). Rainfall data were collected during the whole period of the experience, and processed at a 15 minutes time step.

The natural intense rainfall recorded on May 2016, 8-11th (70 mm on the 8th, 44 mm on the 9th, 41 mm on the 10th, 25 mm on the 11th) was selected to study the relationship between rainfall and water content under natural conditions. This event was indeed one of the most intense after that the deep piezometer P5 was installed on the site.

3. RESULTS

During all the artificial rainfalls, there was no superficial runoff over the plot, which means that all the rainfalls were completely infiltrated in the near surface or deep layers of the hillslope. It is supposed that the natural rainfalls, less intense, were also infiltrated, unless the whole subsurface vertical profile would be saturated, that was not the case. At the plot scale, all the fluxes were thus subsurface fluxes, which we have to qualify and quantify from the experimental device.

3.1 ERT

Electrical resistivity was typically (typically means here that the same patterns were found from more than 50 sites in the same area) shown to increase first at some tenth cm deep, from some hundreds of Ω.m up to more than 3000 Ω.m (Figure 4, above), that shows a thin topsoil layer (0-40 cm) above weathered or consolidated bedrock. The time-lapsed ERT during artificial rainfall showed that the water fluxes remained essentially vertical through the high resistive deep layer. Such result was highlighted during the 27/03/2015 artificial rainfall, by using salt tracer to increase the electrical resistivity of the water fluxes. The vertical water fluxes under the electrodes line along the slope were effective down to more than 1 meter depth, after less than 1 hour (Figure 4, bottom). A decrease of the electrical resistivities was also observed in the topsoil layer near the surface, downstream the point of injection, but should be interpreted as a leak from the tank, which generated some surface runoff. Thus, in spite of thin topsoil layers and apparently consolidated bedrock at small depths, the infiltration processes typically affect more than 1 or 2 meters depth within the hillslope.

[Insert Figure 4]

3.2 Water contents in the 0-1m deep area

During the 15/04/2016 artificial rainfall, almost all the probes showed quick reactions (Figure 5) to the rain drop-by-drop experiment between 4 and 19 minutes after the beginning of the rainfall, either they were placed near the surface or deeper. The corresponding vertical velocities ranged from 1 to 7.5 $m.h^{-1}$ (3.2 $m.h^{-1}$ in average), if considering here the velocity as the ratio of the response time of the probe and its depth. It suggests a high permeability of the shallow organic topsoil but also of the other subsoil horizons. The probe P1-30 did not react to the artificial rainfall, probably because it was saturated the day before, due to an unwilling sprinkling. The near-saturated water content has been reached in all the cases, ranging from 0.18 to 0.60 $cm^3.cm^{-3}$ whereas the initial soil water content ranged between 0.08 to 0.37 $cm^3.cm^{-3}$. The high variability of those water contents (initial and near saturation) could be interpreted as the heterogeneity of the material, combining topsoil, weathered area and rock mass. Thus, P1-30, P1-35, P2-30 and P4-50 should be considered as more or less embedded within a coarse/rocky soil, whereas the other probes were settled within in the topsoil. The recession of the water contents almost started at the moment when the rain stopped. Note that most of the probes exhibited a decrease of the water content a few minutes after the salt injection.

[Insert Figure 5]

[Insert Figure 6]

During the 8-10/05/2016 natural rainfall, the maximal water content ranged from 0.13 $cm^3.cm^{-3}$ to 0.55 $cm^3.cm^{-3}$, and was smaller than in artificial conditions, which shows that the saturation could not be reached (Figure 6). The initial water contents ranged from 0.09 $cm^3.cm^{-3}$ to 0.28 $cm^3.cm^{-3}$, which were similar to those at the beginning of the artificial rainfall event. For the first rainfall (08/05), the response time of the probes ranged between less than hour to 2-3 hours when considering the difference in time between the maximal intensity of the rainfall and the maximal water content of the probe.

The comparison of the water contents under artificial or natural rainfalls showed that the higher the amounts and the intensities of rainfalls were, the higher the soil water contents were, for all the probes, suggesting mainly vertical fluxes in the 0-1 m layer.

3.3 Piezometric levels and tracers in the 0-1m deep area

During the 15/04/2016 artificial rainfall, the response time ranged from 5 to 10 mn for P2, P3, P4 and 24 mn for P1 (Figure 7). The maximal rises ranged from 48 cm (P1) to 78.0 cm (P4). The recessions were faster for P2 and P4 than for P1 and P3. These different behaviours highlight the heterogeneity of the layer 0-1 m, combining soil, weathered area and massive rock. The bottom parts of P1 and P3 were clearly embedded in rock or weathered area, which explained the slow recession rates under a given depth (at least -40 cm for P1, nearly -45 cm for P3); P4 seemed to be essentially surrounded by soil, which allowed fast recessions and quick recharges; P2 was probably partly embedded in a weathered area, limiting the alimentation of the piezometer, but the bottom seemed to be free drained into a more permeable material.

[Insert Figure 7]

Tracing with salt during the artificial rainfall showed a very quick increase of EC in the piezometers, between 1 and 3 minutes after the salt injection. However, each piezometer exhibited different EC dynamics and values. EC reached 3.9 $mS.cm^{-1}$ in P1, and started to decrease after 12h, a much longer time than in the other piezometers. EC was higher in P2 and P3, 45 $mS.cm^{-1}$, and started to decrease between 13-32 mn, whereas EC reached 3.8 $mS.cm^{-1}$ in P4, and started to decrease only 8 mn after the salt injection. EC shows that the flume was mainly concentrated along P2 and P3, but P1 and P4 also reacted because of the leak of the salt solution during the injection, as said above (the first EC peak in P4 was due to the injection of fluorescein). If considering the time between the beginning of the rain and EC max for the metric of the velocity, the water flew between 2 and 8 $m.h^{-1}$ in P2, P3, P4 (P1 was located upstream of the injection area, that could explain its slow reaction time). Note that before the salt injection, EC ranged between 0.06 and 0.09 $mS.cm^{-1}$ in the piezometers. This is coherent with the fact that the water used for the artificial rainfall came from the streamwater, of which sampling during the Floodscale project led to similar values (Bouvier et al., 2018).

The temperatures started to increase more than 1 hour after the beginning of the rainfall, except for P4. The maximal temperatures were reached between 5 and 7 hours after the beginning of the rain. Thus, in spite of the water levels increased after less than 10 mn in P1-P2-P3, the longer times of the changes of temperature show that the quick level rise in the piezometers was first due to the soil water, and the to a mix of soil/rain water.

During the 8-10/05/2016 natural rainfall, the piezometers reacted less (Figure 8), but some saturated areas seemed to have been effective in P2, P3, P4 (~ 10 cm from the bottom of the piezometers). EC and temperatures could not be efficiently measured, because of the low levels in the piezometers. The behaviour of the piezometers under artificial and natural rainfalls was same as in the water content probes : the more rainfall, the higher piezometric level, suggesting mainly vertical fluxes in the 0-1 m layer.

[Insert Figure 8]

3.4 Piezometric levels and tracers in the 10m deep area

Figure 9 shows the piezometric level at the deeper piezometer P5 during the year 2016. The piezometric signal was highly correlated to rainfall (period February-June), and increased up to 200 cm in April 2016 (131 mm between 04/04 15:00 and 05/04 14:15), up to 300 cm in November 2016 (270 mm between 20/11 4:00 and 22/11 15:00). In April 2016, it needed almost 1 month to come back to its initial level, before the next rise. If considering that the hillslope length upstream the plot would be 100-300 m, a mean velocity of the lateral fluxes along the slope should be nearly 0.15-0.5 $m.h^{-1}$. The piezometric level was rather stable between the rainfalls, close to -10m.

[Insert Figure 9]

During natural rainfalls, the piezometer fast reacted to the highest intensities, as shown for the 08-10/05/2016 event (Fig.10). The rainfall generated rise of the piezometer level from -9.6 m to -8.3 m, and different piezometer levels peaks were seen after the most intense rainfall in 15 mn. Time from the rainfall peak to the piezometer level peak was nearly 4 hours.

[Insert Figure 10]

For the natural rainfalls, the temperature exhibited a typical pattern, depending of the season (Figures 10 and 11). The temperature increased (resp. decreased) when the water level increased during the spring season (resp. the fall season).

For most of the natural rainfalls, EC decreased or remained unchanged after a rise of the piezometric level, but this was sometimes erratic. Note however that for the 22/11/2016 flood, the EC pattern was inverted and that EC increased with the water level (Figure 11). It suggests that in some cases, more mineralized water (i.e. deeper water) could fill the piezometer. As a matter of fact, the rise of the water table on 22/11/ 2016 was one of the most important of the period.

[Insert Figure 11]

During the artificial rainfall, the increase of the piezometric level was only 30 cm, despite of a much higher cumulated rainfall, 702 mm (Fig. 12). The piezometer started to react 3:30 hours after the beginning of the rain (i.e. a 2.5 $m.h^{-1}$ velocity), and the maximum level was reached 10 hours after the beginning of the rainfall. The recession time was about 2 days.

EC and piezometric level were positively correlated. This was due to the deep percolation and then the dilution of the salted water, after it was injected from the surface. The response time to the salted water was 3:35 hours, similar to the response time between the beginning of the rainfall and the beginning of the rise of the piezometric level. The maximal EC was 0.094 $mS.cm^{-1}$, which shows the dilution of the salt between the 0-1m layer and the deep layer.

There was no change of temperature in the P5 piezometer during the artificial rainfall 15/04/2016, probably because the volume of “new” water was not enough to impact the temperature of the pre-existing water (from -9.2 m down to 12.0), the temperature probe being located near the 12m-depth. Under natural rainfalls, when the rise of the water level was much more in the piezometer, the change of temperatures did not exceed some 1/10è of °C.

A first rise of fluorescein concentration (up to 300 ppm) was detected 5:30 h after its injection in piezometer P4, and a second major rise (up to 1500 ppm) 8:30 h after its injection. Such occurrence of several rises could be interpreted as the contribution of several deep fractures to the piezometer dynamics. The maximal concentration was reached 12:00 h after the injection. The response times associated to fluorescein were a little longer than those derived from water level and salt solution, which can be explained by the fact that the fluorescein injection was located in P4, and probably activated specific deep fractures, different than those associated to the salted solution (injection near P1) or partly included in those associated to the water level (at the whole plot scale).

[Insert Figure 12]

Comparing the deep piezometer levels occurring during both artificial and natural rainfalls showed that the rising time and time to peak were in the same order of magnitude, whereas the peak flows and the recession times were one order of magnitude larger during the natural rainfalls. The higher piezometric levels observed during the natural rainfalls, in spite of lower rainfall than during the artificial rainfall, can only be due to the fact that a lateral subsurface flow occurred all along the hillslope upstream the plot, and supplied an additional inflow during the natural rainfall. At the opposite, less subsurface flow occurred during artificial flows because the rain only fell over the 10-m^2 plot, and there was no contribution of the upstream hillslope. The higher recession time during the natural rainfall corroborates the hypothesis of a lateral flow along the hillslope.

4. DISCUSSION

At this step, main conclusions are that i) vertical fluxes occur not only through the topsoil, but also through deep layers down to 10 m depth, ii) lateral fluxes occur along the hillslope at the top of the saturated area, at 10m depth (at this plot site), iii) short times of reaction either in the shallow 0-1 m layer or in the deep 0-10 m layer suggest that the velocities of the vertical fluxes exceed several thousand mm/h, let say between 1 and 10 $m.h^{-1}$, iv) long travel times of the lateral fluxes under natural rainfalls suggest that the mean velocities could be estimated between 0.1 and 1$m.h^{-1}$.

This local pattern of vertical/lateral fluxes suggests indeed that the 0-10 m layer is highly weathered/fractured, which allows that water is able to flow quickly to the bottom of the layer, and then downstream to the main streams. Below this depth, fracturation should be effective, as shown by the decrease of the deep piezometric levels during the longest dry periods, but less dense than in the 0-10 m layer, for allowing saturated area. Anyway, the active area where water fluxes are important cannot thus be reduced to the topsoil layer. In addition, if considering that the lateral flow would have 0.1-1 $m.h^{-1}$ velocity and that the mean hillslope length would be at least some tenth of meters, it should take some days that the infiltrated water would reach the stream.

The hydrological processes which generate floods should be seen through the large gap between the travel times of the lateral deep subsurface flow, and the small response time of the floods at the Valescure catchment scale. The former makes indeed that several days, even weeks, are required to reach the bottom of the hillslopes, whereas the latter shows that the floods trigger some few hours after the rainfall. This paradox was already pointed out by Cosandey (1988) for another granite catchment under a different climate (Oceanic climate, Brittany, West France). The author claimed that the piston flow was responsible of the fast rise of the water table in the downstream part of the hillslope, which generated streamflow either directly from rain falling over saturated areas or from direct exfiltration due to piston flow itself. However, there was no explanation of such processes.

We propose an explanation of this pressure wave propagation. First, we consider that the 1-10m deep layer must be seen as the fractured bedrock, below the top soil. The main fractures can be wide and conductive enough to allow a water flux depending of their shape and their hydraulic conductivity. It is also supposed that the hydraulic transmissivity is not sufficient in the downstream part of the hillslope to allow a free drainage of the water flow in the fractures. This could be due to a least density or a smallest size of the fractures in the downslope, because the consolidated rock is less deep from the surface; another explanation could be that the water flux increases at the bottom of the hillslope, as the area of the hillslope increases, and exceeds the transmissivity capacity. When long periods with no rain occur, the water table is in equilibrium, and could have a slope similar to the slope of the hillslope, this is the case n°1 in Figure 13. When raining, the water levels rise in the main fractures, but they rise more in the downstream fractures than in the upstream fractures because the hydraulic transmissivity is overpassed downstream. This is the case n°2 in Figure 13. The more rainfall, the more the fractures fill, and some of the downstream fractures start to flow at the top of the bedrock, within the topsoil. It generates both runoff in the gullies and the streams of the hillslope, and also possibly extended saturated areas in the top soil. The rain falling on these saturated areas will also generate surface runoff. If more rainfall again, more upstream fractures will indeed flow and increase the surface water. This is the case n°3 in Figure 13.

In this interpretation, the surface runoff occurs each time that the water level in a main fracture reaches the surface. It means that the gullies network over the hillslope should be mainly generated by the geometry of the bedrock fractures. It also means that surface flow starts downstream, and progresses then more upstream. This is similar to the variable source area due to the saturation of the compete profile of the soil, but it really describes another kind of process. Here, only the gullies activate runoff because of the exfiltration of the subsurface fluxes, and the more effective rainfall, the more upstream length of the gullies is activated.

This is coherent with the fact that surface runoff was not observed over the hillslope, during the FloodScale program (Braud et al., 2014).

[Insert Figure 13]

5. CONCLUSIONS

In this work, time-lapse ERT, physical /chemical tracing and soil water content measurements were performed to characterize the vertical and lateral fluxes at the plot scale over a 30^0 steep forested hillslope, under artificial and natural rainfalls.

The ERT measurements have shown that the first 40 cm can be considered as topsoil, which overlays heterogeneous weathered and consolidated rocky materials. Time-lapse area during tracer experiment have however shown that at least the 2 first meters had high permeability, and that fluxes were mostly vertical in this area. All the water content measurements performed in the shallow 0-1 m layer also show that vertical fluxes are dominant in the first meters of the profile. The more rainfall, the higher the water contents. The vertical hydraulic conductivity at saturation could be estimated between 1-10 m/h in the 0-1m deep layer, either under natural or artificial rainfalls.

A 10m-deep piezometer located downstream the plot gave valuable information concerning the deep boundary condition and the dynamics of the 0-10m area. Contrasted dynamics of the piezometric level under natural and artificial rainfalls suggested that lateral fluxes are effective at the bottom of the permeable layers. At the contrary of the 0-1m deep layer, the water contents at 8-10m are higher for the natural rainfalls than for the more intense artificial rainfalls. This brings the evidence of a lateral flux along the slope, at the bottom of this deep layer. The mean velocity of this flux could be between 0.1-1 m/h. This is a rough estimation, due to imperfect knowledge of the hydrodynamic properties of the 0-10m layer, to the bedrock topography and to hillslope area drained by the plot. However, this device proved to be able to give valuable information concerning sub-surface fluxes within a steep Mediterranean hillslope, as well as patterns of hillslope soil/bedrock properties. Such “light” experiment could be developed in other points of the granitic area, as well as in other geological units of the Cevennes mountainous area or mountainous area in general.

The characteristics of those subsurface fluxes bring an interesting scope of the behavior of such Mediterranean catchments, of which the response time is no more than few hours in general. Due to the massive vertical infiltration through the 0-10m layer and the slow velocity of the lateral flow at the bottom of the deep layer, the fast response of the catchments during the heavy Mediterranean rainfalls cannot be due to water transport along the hillslope, but brings the evidence that water accumulates or exfiltrates downstream by pressure transport, e.g. piston flow. More efforts should be deployed to better understand this apparent paradox in the response time of the floods. Some more experiments should be carried out in order to estimate the representativeness of the plot, and modelling strategy should be performed in order to confirm the estimated values of the 0-10m layer characteristics and dynamics.

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DATA AVAILABILITY

All the data concerning rainfalls, water contents, piezometric levels, temperature and EC are available in the Hymex database https://mistrals.sedoo.fr/?editDatsId=1597&datsId=1597&project_name=HyMeX&q=valescure

REFERENCES

Adamovic M, Braud I, Branger F, Kirchner JW. 2015. Assessing the simple dynamical systems approach in a Mediterranean context: application to the Ardèche catchment (France). Hydrol. Earth Syst. Sci., 19: 2427-2449. DOI: 10.5194/hess-19-2427-2015.

Ayral P.A., Sauvagnargues-Lesage S., Bressand F. 2005. Contribution à la spatialisation du modèle opérationnel de prévision des crues éclair ALHTAÏR, *Etudes de Géographie Physique*, n°XXXII, pp. 75-97.

Beven K, Germann P. 2013. Macropores and water flow in soils revisited. Water Resources Research, 49: 3071-3092. DOI: 10.1002/wrcr.20156.

Borga M, Boscolo P, Zanon F, Sangati M. 2007. Hydrometeorological Analysis of the 29 August 2003 Flash Flood in the Eastern Italian Alps. Journal of Hydrometeorology, 8: 1049-1067. DOI: 10.1175/JHM593.1.

Bouvier C., Patris N., Freydier R., Guilhe-Batiot C., Seidel J.-L., Brunet P., Remes-Busiau A., 2018. The Floodscale experiment in the small catchment of Valescure, France: An overview of the isotopic and geochemical data base. Geosciences Data Journal, Volume5, Issue1, 14-27. https://doi.org/10.1002/gdj3.57

I. Braud, P.-A. Ayral, C. Bouvier, F. Branger, G. Delrieu, J. Le Coz, G. Nord, J.-P. Vandervaere, S. Anquetin, M. Adamovic, J. Andrieu, C. Batiot, B. Boudevillain, P. Brunet, J.

Carreau, A. Confoland, J.-F. Didon-Lescot, J.-M. Domergue, J. Douvinet, G. Dramais, R. Freydier, S. Gérard, J. Huza, E. Leblois, O. Le Bourgeois, R. Le Boursicaud, P. Marchand, P. Martin, L. Nottale, N. Patris, B. Renard, J.-L. Seidel, J.-D. Taupin, O. Vannier, B. Vincendon, and A. Wijbrans. Multi-scale hydrometeorological observation and modelling for flash flood understanding, 2014. Hydrol. Earth Syst. Sci., 18, 3733–3761

Cosandey C, 1988 (in French). Etude de la formation de l'écoulement rapide de crue dans un petit bassin versant forestier breton. La *Houille blanche*, 5-6, 381-383

Cosandey C, Didon-Lescot J. 1990 (in French). Etude des crues cévenoles: conditions d'apparition dans un petit bassin forestier sur le versant sud du Mont Lozère, France. Regionalisation in Hydrology, IAHS Publication, Ljubljana, Slovenia, 191: 103-115.

Delrieu, G., V. Ducrocq, E. Gaume, J. Nicol, O. Payrastre, E. Yates, P.-E. Kirstetter, H. Andrieu, P. A. Ayral, C. Bouvier, J. D. Creutin, M. Livet, A. Anquetin, M. Lang, L. Neppel, C. Obled, J. Parent-du-Chatelet, G. M. Saulnier, A. Walpersdorf, and W. Wobrock, 2005: The catastrophic flash-flood event of 8-9 September 2002 in the Gard region, France: a first case study for the Cévennes-Vivarais Mediterranean Hydro-meteorological Observatory. Journal of Hydrometeorology, 6, 34-52

Drobinski P, Ducrocq V, Alpert P, Anagnostou E, Béranger K, Borga M, Braud I, Chanzy A, Davolio S, Delrieu G, Estournel C, Boubrahmi NF, Font J, Grubišić V, Gualdi S, Homar V, Ivančan-Picek B, Kottmeier C, Kotroni V, Lagouvardos K, Lionello P, Llasat MC, Ludwig W, Lutoff C, Mariotti A, Richard E, Romero R, Rotunno R, Roussot O, Ruin I, Somot S, Taupier-Letage I, Tintore J, Uijlenhoet R, Wernli H. 2013. HyMeX: A 10-Year Multidisciplinary Program on the Mediterranean Water Cycle. Bulletin of the American Meteorological Society, 95: 1063-1082. DOI: 10.1175/BAMS-D-12-00242.1.

Ducrocq V., Braud I., Davolio S., Ferretti R., Flamant C., Jansa A., Kalthoff N., Richard E., Taupier-Letage I., Ayral P-A., Belamari S., Berne A., Borga M., Boudevillain B., Bock O., Boichard J-L., Bouin M-N., Bousquet O., Bouvier C., Chiggiato J., Cimini D., Corsmeier U., Coppola L., Cocquerez P., Defer E., Delanoë J., Delrieu G., Di Girolamo P.D., Doerenbecher A., Drobinski P., Dufournet Y., Fourrié N., Gourley J.J., Labatut L., Lambert D., Le Coz J., Marzano F., Montani A., Nuret M., Ramage K., Rison B., Roussot O., Said F., Schwarzenboeck A., Testor P., Van Baelen J., Vincendon B., Aran M., Tamayo J., 2014. HyMeX-SOP1, the field campaign dedicated to heavy precipitation and flash-flooding in Northwestern Mediterranean, Bulletin of the American Meteorological Society, Volume 95, Issue 7, 1083-1100.

Fu C, Chen J, Jiang H, Dong L. 2013. Threshold behavior in a fissured granitic catchment in southern China: 1. Analysis of field monitoring results. Water Resources Research, 49: 2519-2535. DOI: 10.1002/wrcr.20191.

Fu ZY, Chen HS, Zhang W, Xu QX, Wang S, Wang KL. 2015. Subsurface flow in a soil-mantled subtropical dolomite karst slope: A field rainfall simulation study. Geomorphology, 250: 1-14. DOI: http://dx.doi.org/10.1016/j.geomorph.2015.08.012.

Gevaert AI, Teuling AJ, Uijlenhoet R, DeLong SB, Huxman TE, Pangle LA, Breshears DD, Chorover J, Pelletier JD, Saleska SR, Zeng X, Troch PA. 2014. Hillslope-scale experiment demonstrates the role of convergence during two-step saturation. Hydrol. Earth Syst. Sci., 18: 3681-3692. DOI: 10.5194/hess-18-3681-2014.

Gonzalez-Sosa E, Braud I, Dehotin J, Lassabatère L, Angulo-Jaramillo R, Lagouy M, Branger F, Jacqueminet C, Kermadi S, Michel K. 2010. Impact of land use on the hydraulic properties of the topsoil in a small French catchment. Hydrological Processes, 24: 2382-2399. DOI: 10.1002/hyp.7640.

Gotkowitz M, Attig J, McDermott T. 2014. Groundwater flood of a river terrace in southwest Wisconsin, USA. Hydrogeol J, 22: 1421-1432. DOI: 10.1007/s10040-014-1129-x.

Graham C.B., Woods R.A, McDonnell.J.J., Hillslope threshold response to rainfall:(1) A field based forensic approach. Journal of Hydrology 393 (2010) 65-76

Kienzler PM, Naef F. 2008. Subsurface storm flow formation at different hillslopes and implications for the 'old water paradox'. Hydrological Processes, 22: 104-116. DOI: 10.1002/ hyp.6687.

Le Bourgeois O, Bouvier C, Brunet P, Ayral PA. 2015. Inverse modelling of soil water content to estimate hydraulic properties of shallow soil and the associated weathered bedrock. Journal of Hydrology, 541, 116-126.

McDonnell, J. J., and K. Beven (2014), Debates—The future of hydrological sciences: A (common) path forward? A call to action aimed at understanding velocities, celerities, and residence time distributions of the headwater hydrograph, Water Resour. Res., 50, 5342– 350, doi:10.1002/2013WR015141.

Pinault JL, Amraoui N, Golaz C. 2005. Groundwater-induced flooding in macropore-dominated hydrological system in the context of climate changes. Water Resources Research, 41: n/a-n/a. DOI: 10.1029/2004WR003169.

Scaini, A., Hissler, C., Fenicia, F., Juilleret, J., Iffly, J. F., Pfister, L. & Beven, K, 2018. Hillslope response to sprinkling and natural rainfall using velocity and celerity estimates in a slate-bedrock catchment, Journal of Hydrology, Volume 558, 366-379

Schneider P, Pool S, Strouhal L, Seibert J. 2014. True colors- experimental identification of hydrological processes at a hillslope prone to slide. Hydrol. Earth Syst. Sci., 18: 875-892. DOI: 10.5194/hess-18-875-2014.

Tramblay Y, Bouvier C, Martin C, Didon-Lescot J-F, Todorovik D, Domergue J-M. 2010. Assessment of initial soil moisture conditions for event-based rainfall–runoff modelling. Journal of Hydrology, 387: 176-187. DOI: http://dx.doi.org/10.1016/j.jhydrol.2010.04.006.

Tromp-van Meerveld HJ, McDonnell JJ. 2006a. On the interrelations between topography, soil depth, soil moisture, transpiration rates and species distribution at the hillslope scale. Advances in Water Resources, 29: 293-310. DOI: http://dx.doi.org/10.1016/j.advwatres.2005.02.016.

Tromp-van Meerveld HJ, McDonnell JJ. 2006b. Threshold relations in subsurface stormflow: 1. A 147-storm analysis of the Panola hillslope. Water Resources Research, 42: n/a-n/a. DOI: 10.1029/2004WR003778.

Uchida T, Asano Y, Mizuyama T, McDonnell JJ. 2004. Role of upslope soil pore pressure on lateral subsurface storm flow dynamics. Water Resources Research, 40: n/a-n/a. DOI: 10.1029/2003WR002139.

Wenninger J, Uhlenbrook S, Lorentz S, Leibundgut C. 2008. Identification of runoff generation processes using combined hydrometric, tracer and geophysical methods in a headwater catchment in South Africa / Identification des processus de génération de l'écoulement par combinaison de méthodes hydrométriques, de traçage et géophysiques dans un bassin versant sud-africain. Hydrological Sciences Journal, 53: 500-500. DOI: 10.1623/hysj.53.2.500.

FIGURE LEGENDS

Figure 1 : Catchment location and instrumentation

Figure 2 : Water retention data and fitted van Genuchten equation for depth of 30 cm

Figure 3 : Scheme of the experimental sprinkling set-up

Figure 4 : Time-lapse ERT along the vertical 48-electrodes line L1, using salt tracer injection, 27/03/2015 (slopes are not displayed here). The upper electrode lies on the left side of the section, at length 0.4 m. Salt was injected at length 3.80 m.

Figure 5 : Observed soil water content during the 15/04/16 artificial rainfall event

Figure 6 : Observed soil water content during the 8-10/05/16 natural rainfall event

Figure 7 : Piezometric levels and physical-chemical tracers in the 0-1m layer during the 15/04/16 artificial rainfall event

Figure 8 : Observed piezometric levels in the 0-1m layer during the 8-10/05/16 natural rainfall event

Figure 9 : Piezometric variability within the borehole P5 during the year 2016. The black dotted circle indicates the rise of the piezometric level during the 15/04/2016 artificial rainfall (700 mm in 6 hours). The artificial rainfall is not displayed on the figure.

Figure 10 : Piezometric level, temperature and EC in P5 during the 08-10/05/2016 rainfall event.

Figure 11 : Piezometric level, temperature and EC in fall season

Figure 12 : Piezometric levels and physic-chemical tracers in the P5 piezometer, during the 15/04/2016 artificial rainfall.

Figure 13 : Interpretation of the hydrological processes during rainfall within the hillslope. The water level in the downstream fractures is pushed up because of a piston flow from the upstream fractures, due to steep slopes and overpassed hydraulic transmissivity downstream. The extension of the saturated fractures depends on either the rainfall or the initial water content. Fractures flow can either exfiltrate in the gullies or main streams, or locally saturate the topsoil area. Both cases will generate surface runoff, either directly if exfiltration, or when rain occurs over saturated topsoil