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Instability driven flow and runoff formation in a small catchment

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ABSTRACT

Two anomalous phenomena were observed in a small catchment: 1) In some situations, the water supplied by rain caused a pronounced decrease in the soil water content. 2) In these periods, the soil water movement could be explained only by assuming an irregularly oscillating outflow of soil water into lower horizons. In these situations a large volume of water flows through the soil; therefore, on the hydrological scale, this phenomenon forms a great part of the outflow from a watershed. These phenomena are described in the frame of the instability driven flow theory and explained as consequences of the porous soil body's capacity to become conductive as a result of a very little change of its moisture content. Therefore the soil profile can attenuate or amplify the rainfall pulses during their transformation to the outflow below the soil profile. If the soil water content is lower than the threshold value, the rainfall pulses can be suppressed down to zero. If the soil profile contains more water, the soil does not attenuate the rainfall pulses and it can even amplify them by adding the released soil water. This is the mechanism of rapid growth of rising hydrograph limb during a storm event. The rapid transport of the soil water can occur in any part of the porous soil body regardless of the pore size and can be caused by any rainfall event with any intensity, duration or total volume.

KEYWORDS | Hydrology. Rainfall. Runoff. Soil. Catchment. Preferential flow.

INTRODUCTION

In the past, the so-called rainfall – runoff relationship was amply studied. Its formulation was based on the finding that in large catchments (the area exceeding 100 km^2), a well describable relationship between the actual discharge in the closure profile and the precipitation total for a given antecedent period can be found. Gradually it became evident that models conceived in this way are unable to describe the reality of runoff formation from a small catchment with area up to 10 km^2. Hydrodynamic aspects of water movement are fully omitted in these models (Burnash, 1995).

Another way how to study runoff formation is the hydrodynamic approach. Its ambition is to explain the rainfall – runoff relationship in terms of hydrodynamics (Robinson, 1993). Three distinct types of transport processes are used in order to describe water transport in a catchment: 1) the channel flow – in brooks and rivers, 2) the surface flow – on the soil surface covered by vegetation, and 3) the flow in a porous medium – in the soil and subsoil. Every one of these processes is quite well understood. However, the key problem is to separate an infiltrating precipitation into the part directly flowing in the soil and the part that is stagnant on the soil surface. This stagnant water may be a source of the surface runoff if the soil surface is sloped. A second separation of flowing water in the soil may occur – into a part flowing vertically and a part flowing in the direction of the sloping soil horizons or the subsoil layer. The sloping flow in a horizon near the soil surface is called a subsurface flow.

The storm runoff generation in a small catchment is characterized by three effects: 1) The rising hydrograph limb grows very quickly and its duration is short – a few minutes or hours. 2) The falling hydrograph limb lasts for many days or weeks. 3) The greatest value of the soil water content is reached as a rule before the rain ends. The way how to determine the proportion of transport processes in runoff generation is the runoff separation method. In this article the channel flow and the surface flow are not studied. Therefore, two components are separated from the runoff: the vertical soil-water flow and the water flow in the sloped drainage layer. The attenuation of precipitation in the soil, drainage layer and whole catchment is also studied. The theory of saturated source areas (Hewlett and Nutter, 1970) is adapted for the conditions characterizing the Liz experimental catchment.

HYDRODYNAMIC APPROACH TO RUNOFF FORMATION

The principal problem of the hydrodynamic approach to runoff formation resides in the assessment of the rules governing the separation of flowing water; the separation of precipitation into infiltrating water and surface runoff, and the separation of infiltrating water into a vertical and subsurface flow. The solving of these questions is the main goal of soil hydrology (Kutílek and Nielsen, 1994). Theories concerning soil water flow, created in the frame of soil hydrology, strongly inspire the research on runoff formation on the catchment scale. On the other hand, the key problem in runoff hydrology –rapid runoff generation during a storm event– is always a challenge for soil hydrologists.

It is possible to distinguish four processes concerning the hydrodynamic mechanisms attributed to rapid runoff formation:

- surface runoff in the whole catchment area,
- subsurface runoff in the variable contributing areas,
- macropore flow in parts of soil pores, and
- instability driven flow in the prevailing part of soil pores.

The changes of view on runoff formation are inspired by the developments in soil physics. During the period 1930–1990, the theory of soil water transport was dramatically changed. Corresponding changes in runoff hydrology are linked to the role of soil water movement in runoff formation. It is visible that the soil cover plays an ever greater role in the hydrodynamic theory of the rainfall – runoff transformation.

In the past, soil water movement was studied under the strong influence of the theory presented by Richards (1931). This theory, in its original form, explains only the slow water movement in the soil. Therefore, the conclusion was that the rapid delivery of water into the stream during the stormflow could not be attributed to the soil water flow. This is why the rapid growth of rising hydrograph limb was commonly attributed to the surface or overland flow (Horton, 1940). But in natural conditions, when surface runoff was not observed, the mechanisms of rapid flow were not known.

A “Copernican revolution” in hydrology (Bonnel, 1993) is the variable contributing area hypothesis (Hewlett and Nutter, 1970; Beven and Kirkby, 1979). “Rapid delivery of water into the stream during the stormflow is attributed to a shrinking and expanding of the saturated area that can occur anywhere in the catchment where the infiltrated water cannot be transferred through the soil. Nevertheless, the majority of saturated areas occur near streams” (cited by Kostka and Holko, 1997). This hypothesis explains why no surface runoff is observed and just the rapid growth of the rising hydrograph limb is recorded. The source of water delivered into the stream is the subsurface flow in the saturated soil layer near the catchment surface. If the flow paths are short (some tens of meters) and the saturated soil layer is sloped, the velocity of saturated flow is enough to generate a rapid outflow wave in the stream.

In the field of contaminant hydrology, new ideas about the soil water movement have appeared. It was experimentally demonstrated that a part of water flows through the soil more quickly than can be explained by Richards’ theory (Lichner, 1986). On the basis of this fact, the macropore flow hypothesis was formulated: The rapid flow of water in the soil occurs in the greater non-capillary pores. One of several possible hydrodynamic mechanisms of rapid flow in macropores is a dissipation of momentum – e.g. kinematic wave based on the boundary-layer flow theory (Germann, 1985). On the other hand, the slow flow is attributed to the diffusion of the potential energy in smaller pores in the soil matrix

(Richards' flow). In this approach, the macropores make pathways linking the soil surface and the underground water table. If, as a consequence of rapid rainwater delivery by macropores, the level of the underground water table rises, the discharge into the stream increases and the rapid rising hydrograph limb is generated. This rapid transport in larger pores is possible, from the hydrodynamic point of view, only if the rainfall is heavy. Therefore, this theory explains the rapid growth of hydrograph rising limb only in larger rain situations. It means that the causal quantitative relation between the volume of rain water and the volume of runoff is preserved in the scope of the macropore flow hypothesis.

INSTABILITY DRIVEN FLOW

Efforts aimed at elucidating runoff formation in the Liz catchment revealed the existence of transport phenomena which, on principle, cannot be described by any composition of macropore and Richards' flow (Pražák et al., 1992). Two coupled anomalous phenomena were observed: 1) In some situations, the water supplied by rain caused a pronounced decrease in the soil water content. 2) In these periods, the soil water movement could be explained only by assuming an irregularly oscillating outflow of soil water into lower horizons. In these situations a large volume of water flows through the soil; therefore, on the hydrological scale, this phenomenon is responsible for great part of the outflow from a watershed (Šír et al., 2000). In contrast to the macropore flow hypothesis, as mentioned above, the rapid transport of soil water can occur in any part of the porous soil body independent on the pore size and can be caused by any rain regardless of its intensity, duration and total volume. Therefore, it is probable that the rise of the outflow wave is a manifestation of a qualitative break in transport processes called forth by a small input of water into a soil. This then implies that, from a quantitative point of view, water transport in a soil depends on its moisture in a jump-like way. It means that the causal quantitative relation between the volume of rain water and the volume of runoff is not preserved during the soil water transport generating the rainfall–runoff transformation.

In order to describe the anomalous transport phenomena in porous media the instability driven flow hypothesis was formulated and experimentally verified in laboratory and field conditions (Pražák et al., 1992; Šír et al., 1996, 2000). In the terms of the capillary displacement theory, "The liquid fed into a porous body fills up its pores. At the moment when the water-filled pores form a water body of a certain critical height in the gravitational field, this body flows through the pore labyrinth and leaves the original position. Owing to an influx of liquid, new water bodies are incessantly formed and the process is reproduced. In dependence on the volume of the water body retained in the pores, a perceptible outflow wave arises or does not arise. When dropping down through the pore labyrinth, the water body can also be broken up into bodies smaller than critical: these can be retained in the pores (i. e. the water is stabilizing) and become a rudiment of a new outflow. Then the inflow of only a small volume of liquid may prove sufficient for these bodies, which are closely below the critical height, to unite and form a powerful outflow wave (i. e. the flow instability) conducing to the drainage of a substantial part of pores originally filled with liquid." In the scope of this approach, the above mentioned anomalous transport phenomena can be explained as consequence of the porous body's capacity to become conductive as a result of a very little change in its moisture. The oscillating outflow is a sequence of two alternate processes: the water stabilization and the water flow driven by instability. During the instability driven flow, the volume of water contained in the porous body can decrease, even if the inflow continues.

In soil hydrology terms, the network of pores is variously filled and emptied in the course of water movement. Depending on the stage of filling of the network, two different water transport regimes alternate. The water transport in drier soil is approximately describable by the diffusion analogy (diffusion type flow DTF). In the DTF water flows mainly through smaller pores while the larger pores are filled with air. The water movement in more moist soil is describable by the instability driven flow IDF, as the so-called capillary displacement. In this regime larger pores play an important role; it is where the volumetrically greater part of the flow takes place. Water in the smaller pores is either almost motionless or it is set into motion due to a suddenly established hydraulic interconnection of large and small pores. The beginning or the end of the hydraulic interconnection of large and small pores is very sensitive to small changes in soil moisture content. If a hydraulic interconnection of small and large pores takes place in a sufficiently high layer, it results in an outflow of a considerable amount of water to the substratum. In consequence, larger pores (and sometimes also a great part of smaller ones) are emptied and the DTF flow with an insignificant outflow sets in again. In this way, the outflow oscillations can arise causing a hydrologically significant amount of water flow out from the soil, which surpasses causal rainfall by the volume of water stabilized in the soil for a long time and suddenly released from smaller pores.

In runoff hydrology terms, the soil profile can attenuate or amplify the rainfall pulses during their transformation to the outflow below the soil profile. If the soil water content is lower than the threshold value, the rainfall pulses can be suppressed down to zero. If the soil profile contains more water, the soil does not attenuate the rainfall pulses, it can even amplify them by adding the released soil water. This is the mechanism of rapid growth of rising hydrograph limb during a storm rain event. The soil water flows through the whole soil profile, so that no macropore flow has to be a priori postulated.

PREFERENTIAL FLOW

The essential non-homogeneity of water transport in the pore's microscale (capillary displacement) is the reason of the non uniform (i.e.preferential) flow in the macroscale of porous medium (Pražák et al., 1988). The word "preferential" means that the flowing water does not flow through all the pores but it prefers certain pathways in the porous body instead. Preferential flow can be heterogeneity or instability driven (Steenhuis et al., 1996). Heterogeneity driven preferential flow is the result of the existence macro- and mesopores (Beven and Germann, 1982) - the preferential pathways are large pores. Instability driven flow (either gravity or viscous) occurs mainly in a sandy or coarse soil with or without layers, and in water repellent soils (Dekker and Ritsema, 1996). In this case, the preferential pathways have no cause in the pore structure but in the physics of capillary displacement (Pražák et al., 1988; Šír et al., 1996).

Based on experimental experience it is apparent that the preferential flow is an even more universal event than supposed. The existence of the preferential flow in the structurally homogeneous porous media was experimentally verified in the space scale from tenths of cm^3 to m^3 up to a small watershed (about 1 km^2). The theory of preferential flow in structurally homogeneous porous media was developed both on the microscopic and macroscopic level: (1) at the microscopic level as an analogy to Brown's movement (Pražák et al., 1988), (2) at the macroscopic level, making use of percolation theory of infiltration, redistribution, free drainage, air pressure displacement of water and isothermal evaporation (Šír et al., 1996).

TABLE 1 **Charecteristics of the Liz catchment.**

Drainage area (km^2) DA	0.99
Mean discharge (m^3/s)	0.01 (1976–1997)
Runoff coefficient (-)	0.38 (1976–1997)
Mean annual air temperature (ºC)	6.30 (1976–1997)
Average slope (%)	17
Basin length (km)	1.45
Channels length (km) CL	1.43
Elevation (m a.s.l.)	828–1074
Precipitation sum (mm/year)	851 (1976–1997)
Runoff depth (mm/year)	324 (1976–1997)

At the macroscopic level, the differences between heterogeneity driven preferential flow and instability driven flow are seemingly negligible. In both cases the flowing water forms a typical mosaic pattern of wet pathways and dry islands. These patterns are visualized with the help of various techniques (computer controlled angiography or tomography, neutron radiography, etc.). In the case of the real soil, it is no simple matter to distinguish if the wet pathways are really formed by interconnected larger pores. Therefore, it is very easy to mistake the heterogeneity and instability driven flow. Perhaps this is why that the macropore flow is very often reported in the literature now. The proportion of the preferential and the matrix flow can be estimated with the help of the so called bypassing ratio (Lichner, 1997). A symptom of the instability driven flow is the oscillating value of the bypassing ratio during the water flow, whereas the non-oscillating one indicates the heterogeneity driven flow.

EXPERIMENTAL CATCHMENT

The experimental catchment Liz is located in the Šumava Mts. in the Czech Republic (Fig. 1). The fully forested watershed is covered by mature spruce forest. The soil cover (acid brown soil) is composed of several horizons with different hydraulic properties, but the infiltrated water largely flows downwards through the soil, so that surface and subsurface runoff are rare phenomena. Highly permeable subsoil forms a shallow drainage layer transporting water from the soil to a small brook. This layer is not fully filled with water, so that no significant areas with ground water table are in the catchment. Geological bedrock (paragneiss) forms an impermeable layer.

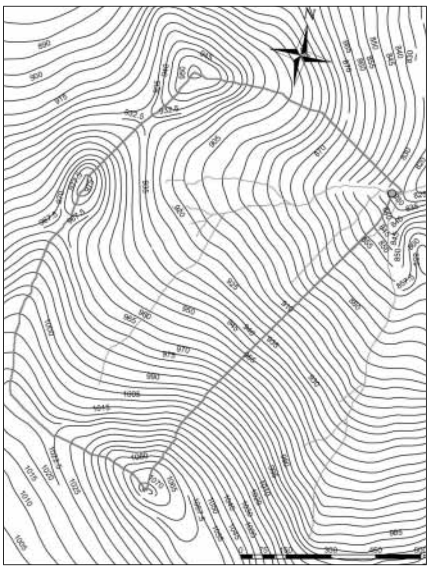

FIGURE 1 **Experimental catchment Liz.**

The experimental area is described elsewhere (Pražák et al., 1994; Tesař et al., 2000). Some catchment characteristics are given in Table 1. Air temperature, precipitation, global radiation, tensiometric pressure at the depths of 15, 30, 45, 60 and 90 cm, and discharge in the closing profile are measured in the catchment. A scheme of the soil profile is shown in Fig. 2.

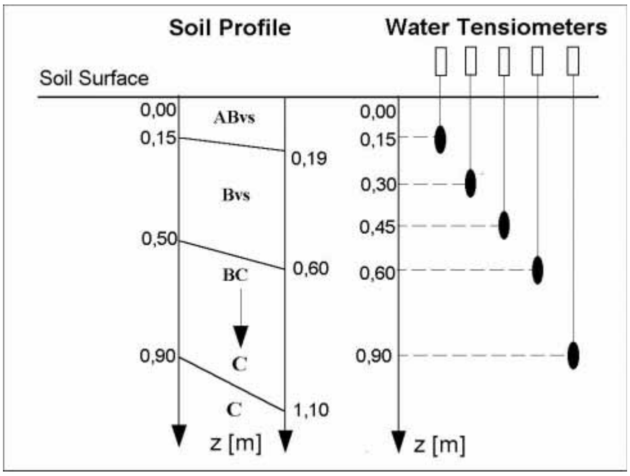

FIGURE 2 **Scheme of the soil profile.**

OUTFLOW FROM THE SOIL INTO THE DRAINAGE LAYER

Soil water balance method

Outflow from the soil cover into the drainage layer can be evaluated using a soil water balance equation in a one-day step.

$$O = P - ET - \Delta Z \qquad (1)$$

O: outflow from the soil into the drainage layer (mm/day), P: daily precipitation sum (mm/day), ET: daily evapotranspiration sum (mm/day), ΔZ: daily change of the soil water content (mm/day). Time series of tensiometric pressures, daily precipitation and evapotranspiration totals are measured. Actual evapotranspiration is evaluated as the water requirement for plant cooling (Pražák et al., 1994; Tesař et al., 2000, 2001). The outflow from the soil into the drainage layer is the only unknown value in the balance equation. It is a computed value. That is why the outflow is a fuzzy value loaded with the inaccuracies of all measured values in Eq. 1.

The soil water content is the sum of water contained in the particular soil layers. Retention curves are used in the recalculation of tensiometric pressures on the soil water content in the particular soil layers (Pražák et al., 1994). During the instability driven flow (Pražák et al., 1992; Tesař et al., 2001), typical when heavy rain infiltrates, this recalculation is inaccurate so that the soil water content is uncertain. The daily change of the soil water content is determined as the difference between the soil water content at the end and at the beginning of the current day. If the instability driven flow lasts more than one day, both values of the soil water content are uncertain, so that the outflow value estimated by Eq. 1 is unreliable.

The mass curve yields a better possibility to estimate the outflow from the soil. The mass curve of the outflow ΣO is calculated as a difference of the mass curve of precipitation ΣP and mass curve of evapotranspiration ΣET reduced every day by the actual deficit ΔZ_i of soil water content with respect to the content in the first day of the balanced period.

$$\Sigma O = \Sigma P - \Sigma ET - \Delta Z_i \qquad (2)$$

The mass curve of precipitation is a cumulative addition of the daily precipitation totals since the season's beginning. Similarly the mass curve of evapotranspiration is constructed. If the first day is not rainy the starting value of the soil water content is estimated with sufficient accuracy.

Simulation of the soil water movement

Simulation modelling is another tool to estimate the outflow from the soil cover into the drainage layer. RETU (Retention – Evapotranspiration Unit) is a simulation model of water transport in the soil based on the solution of the Richards' equation (Tesař et al., 2001). In principle, this model is based on the hydraulic characteristics of the soil (retention curve and hydraulic capacity function, saturated and unsaturated hydraulic conductivity for each genetic soil horizon) and with the help of the limiting (threshold) value of tensiometric pressure hindering the withdrawal of water for transpiration. Parameters characterizing plants are given by the value of optimum temperature (Pražák et al., 1994). The identification of the RETU model parameters and calibration consists in obtaining parameter values leading to a sufficient agreement of the measured data with the model outputs, i.e. the time series of soil water content and the mass curve of the outflow from the soil to the drainage layer. The calibration of soil hydraulic characteristics is based on small changes of measured retention curves and saturated hydraulic conductivities (Bayer at al., 2000).

The calibration of the RETU model is illustrated in the Liz catchment in the vegetation season 1999. The limiting (threshold) value of tensiometric pressure inhibiting water withdrawal is about –60 kPa. The optimum temper-

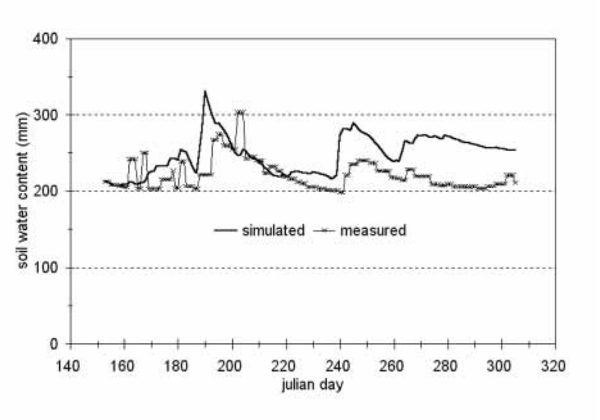

FIGURE 3 | **Measured and simulated soil water content.**

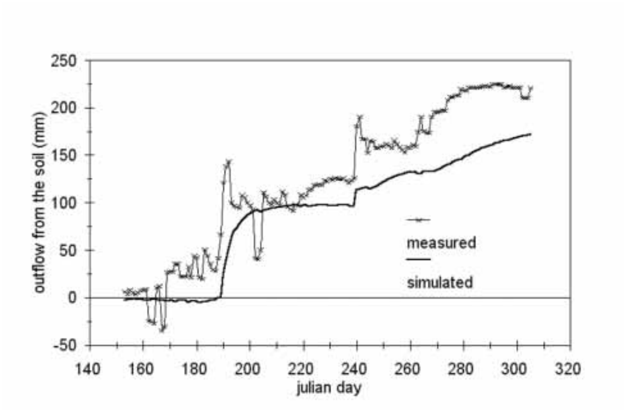

FIGURE 4 | **Mass curves of measured and simulated outflow from the soil into the drainage layer.**

ature of plant is 25 ºC. Figure 3 shows the attained agreement between the measured and simulated soil water content. Fig. 4 presents the agreement of the mass curves of the outflow derived from measured tensiometric pressures (Eq. 2) and simulated by RETU.

VEGETATION SEASON OF 1999

Vegetation season 1999 was in the long-term scale typical regarding the rainfall – runoff relationship and was also rather warm (Table 2). During the season two significant precipitation events were recorded (Table 3). Daily precipitation sums in both situations were lower than the threshold value of ca. 60 mm/day. It means that surface runoff does not appear. A warm and dry period separated these two precipitation events. Runoff characteristics of this period are given in Table 4.

RUNOFF FORMATION

Water transport in the Liz catchment can be divided into two parts: 1) the water movement in the soil cover, i.e. precipitation – outflow transformation, and 2) the flow of water through the subsoil (drainage layer) into the stream, i.e. outflow – runoff transformation. Following

TABLE 2 | **Climatic and runoff characteristics of the vegetation season 1999.**

Vegetation season (May 1 – Sept. 30)	1999	avg. of 1983–2000
Precipitation sum (mm)	406	416
Potential evapotranspiration sum (mm)	214	258
Runoff sum (mm)	116	117
Average air temperature from 5 to 20 hours (ºC)	13.6	12.6
Global radiation sum (kWh/m^2)	490	503

TABLE 3 | **Precipitation events in the vegetation season 1999.**[1]

Precipitation event	S1: July 7–12	S2: Aug. 26–29
Julian days	188–193	238–241
Duration (days)	6	4
Initial tensiometric pressure (kPa) in the depth of 100 cm	–55	–73
Precipitation sum (mm) *PS*	137	70
Outflow sum (mm) *OS*	65	19
Runoff sum (mm) *RS*	17	3
Precipitation peak (mm/day) in the day 190 (240) *PP*	54	52
Outflow peak (mm/day) in the day 190 (240) *OP*	24	18
Runoff peak (mm/day) in the day 191 (241) *RP*	6	2

(1) Precipitation sum is the total precipitation within the precipitation duration. Runoff sum is the volume of the discharge wave in the stream decreased by the base flow. Outflow sum is the volume of the outflow from the soil. Peak values are the maximum daily sums.

TABLE 4 **Runoff characteristics of the accumulative phase.**

Duration of the accumulative phase July 12 – Aug. 28	193 – 240, i.e. 48 days
Precipitation sum (mm)	91
Evapotranspiration sum (mm)	80
Decrease of the soil water content (mm)	27
Outflow sum (mm) *O*	38
Runoff sum (mm) *R*	33

the proposed scheme, water transport in the soil can be excluded from the hydrologic cycle in the catchment and investigated separately. The outflow from the soil is the only inflow into the drainage layer. Water contained in the layer does not flow back into the soil cover. In principle, the soil water movement is one-dimensional and vertical. Water in the drainage layer moves anglewise downhill to the watercourse. The scheme of water transport in the soil and the drainage layer is given in Fig. 5.

The precipitation – outflow transformation is shown in Fig. 6. The second step, the outflow – runoff transformation, is shown in Fig. 7. The outflow from the soil in Figs. 6 and 7 was computed using the RETU simulation. Both precipitation peaks S1 and S2 caused corresponding outflow and runoff peaks, as is demonstrated in Figs. 6, 7 and 8. Precipitation peaks are consequently attenuated in the soil and in the drainage layer.

The peak attenuation in the soil *PAS*, in the drainage layer *PAD*, and in the whole catchment *PAC* can be estimated with the help of Eq. 3.

$$PAS = PP/OP \quad PAD = OP/RP \quad PAC = PP/RP \quad (3)$$

The total attenuation *TAS*, *TAD* and *TAC* are similarly defined.

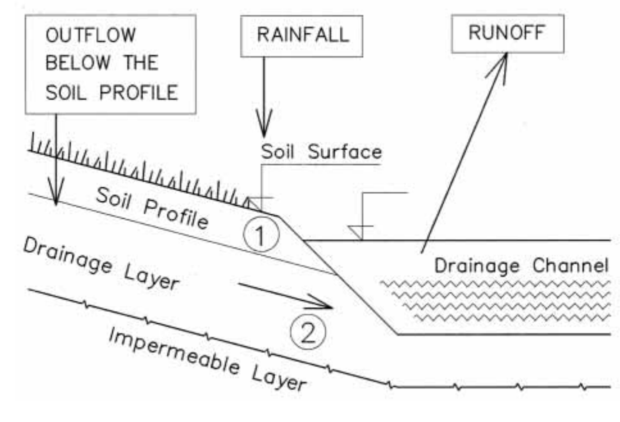

FIGURE 5 **Scheme of the water transport in the soil profile and drainage layer. 1: rainfall- outflow transformation; 2: outflow - runoff transformation.**

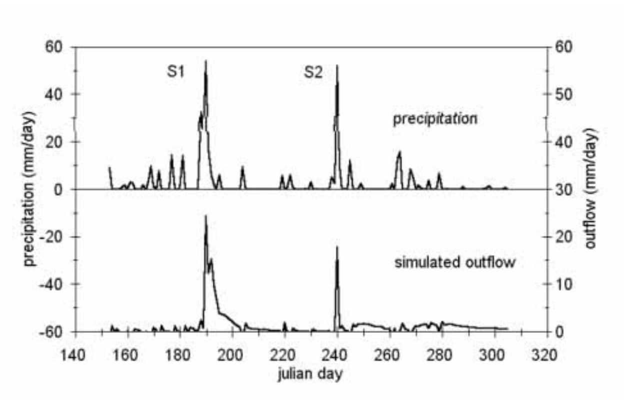

FIGURE 6 **Precipitation – outflow transformation.**

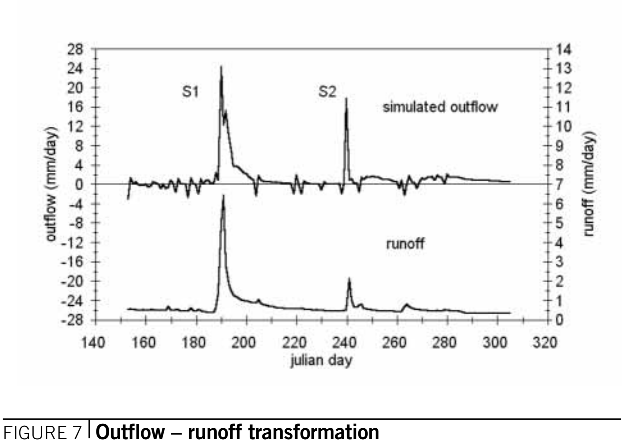

FIGURE 7 **Outflow – runoff transformation**

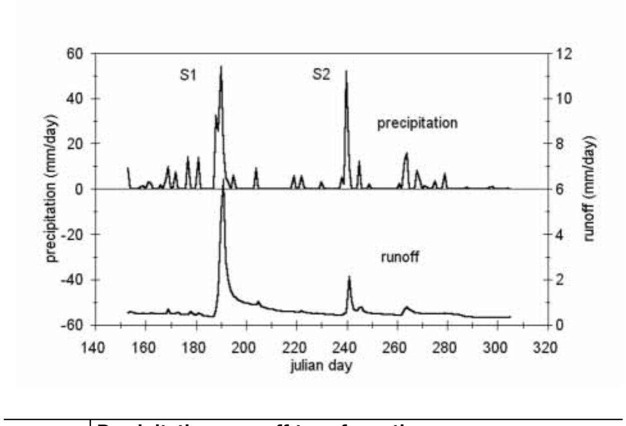

FIGURE 8 **Precipitation – runoff transformation.**

TABLE 5 **Attenuation of precipitation in the soil, drainage layer and catchment.**

	peak attenuation		total average attenuation
	S1	S2	in both situations
soil cover	*PAS* = 2	*PAS* = 3	*TAS* = 3
drainage layer	*PAD* = 4	*PAD* = 9	*TAD* = 5
catchment	*PAC* = 8	*PAC* = 27	*TAC* = 15

$$TAS = PS/OS \quad TAD = OS/RS \quad TAC = PS/RS \qquad (4)$$

The values of *PP*, *OP*, *RP*, *PS*, *OS* and *RS* are given in Table 3. The attenuation of precipitation in the soil, drainage layer and catchment are summarized in Table 5.

Another view on the role of the soil cover in the runoff generating process gives the concept of variable source (or contributing) areas. Eq. 5 is a modification of the concept published by Hewlett and Nutter (1970). The estimation of the source area (Eq. 5) is based on the finding that the outflow from the soil cover is synchronous and equal on the whole catchment area (Tesař et al., 2002). The source area *SA* shows what part of the catchment contributes in the runoff wave in the stream within the precipitation duration.

$$SA = RS/OS \qquad (5)$$

The values of *RS*, *OS*, *RP*, *OP* are given in Table 3. The source areas in both precipitation events are presented in Table 6.

Supposing that the source area has the form of a source belt along the channels in a catchment, the breadth of it can be estimated with the help of Eq. 6.

$$L = 0.5 * DA * RA/CL \qquad (6)$$

The values of the drainage area and the channel lengths are given in Table 1, *SA* in Table 6. The breadth of the source belt in both precipitation events is shown in Table 6.

TABLE 6 **Source area and source belt**

Precipitation event	S1	S2
Source area (-) *SA*	0.26	0.16
Breadth of the source belt (m) *L*	90	55

DISCUSSION

The precipitation – runoff transformation is realised in two steps in the Liz catchment. In the first step, the soil behaves as a reservoir, filled with rainwater, and emptied by the water uptake for plant transpiration. In the course of the vegetation season, the soil water content oscillates between two typical values – maximum and minimum (approx. 200 and 300 mm in Fig. 3). The maximum value corresponds to such soil water content where the infiltration of further rain causes percolation of water to the drainage layer. The minimum value corresponds to such soil water content where insufficient soil moisture renders a further withdrawal of water for plant transpiration impossible. This feedback is a source of complicated problems concerning water and energy transport in the soil – plant – atmosphere system. Thus the soil water regime is a very complex matter.

A hypothesis concerning the nature of the soil water regime on the hydrological scale was published in Tesař et al. (2001). In the article, two types of soil water movement are discussed: the diffusion type flow (DTF) in drier soils, and the instability driven flow (IDF) in soils with a higher soil water content. This corresponds to two phases of soil water flow: the percolation phase (when IDF is taking place), and the accumulation phase (when DTF is taking place). In the percolation phase, the rainwater percolates through the soil into the drainage layer (situations S1 and S2 in Figs. 6 and 7). In the accumulation phase (between S1 and S2), the rainwater accumulates in the soil and does not outflow into the drainage layer.

The second step of the precipitation – runoff transformation, water movement in the drainage layer, is the purely hydraulic matter, because the link to the plant transpiration is missing. The drainage layer forms a reservoir in which water is stored for a long time. In the percolation phase, this reservoir is filled by water percolating from the soil and simultaneously emptied to the stream (rising hydrograph limb). In the accumulative phase, when the inflow is negligible, the water content in the drainage layer decreases during time and the inflow into the stream decreases as well (recession or falling hydrograph limb).

Following the running phase, different mechanisms act in runoff formation. In the percolation phase, water storage recharges in the drainage layer and outflows into the stream in discharge waves immediately reacting to precipitation (situations S1 and S2 in Fig. 8). In the accumulation phase, water slowly outflows from the drainage layer and forms the base flow (between S1 and S2). The altering of the accumulation and the percolation phases can be described with the help of these rules: (1) The percolation of soil water to the drainage layer sets in if the threshold value of the soil water content is overstepped (approx. 270 mm in Fig. 3). In this situation, the water supplied by rain causes a pronounced water outflow and consequently decrease the soil water content. (2) In a situation where the soil water content is less than the threshold value, the percolation is negligible.

The first and second step of the precipitation – runoff transformation are concurrent processes. This means that the water in the soil, drainage layer, and in the stream is a mixture of water originating from series of antecedent rainfalls. In the case of the vegetation season 1999, the second precipitation situation S2 follows the previous one after a 48-day dry period. Runoff characteristics of this accumulative phase certify that the runoff in this period is formed mainly by water originating from the precipitation situation S1 ($R = O - 5$ mm, see Table 4). This fact confirms that the second step, water movement in the drainage layer, is independent of the plant transpiration. This means that 1) the plant roots do not grow into the drainage layer, and 2) the water contained in the drainage layer does not rise into the soil.

The values of breadth of the source belt in both precipitation events support the hypothesis that the source belt is formed in the drainage layer rather than in the soil cover (Table 6). It means that in the percolation phase the source area is the narrow belt only, but in the accumulation phase it is the whole area of the catchment. Similar results were presented in the corresponding natural conditions (Kostka and Holko,1997; Stehlík, 2000; Šanda and Císlerová, 2000).

CONCLUSIONS AND PROSPECTS

Analysis of the relation between the soil water regime and runoff from the Liz catchment shows that in a small catchment the runoff can be investigated as two transformations: 1) rainfall to outflow from the soil into the drainage layer – rainfall water movement through the soil in a vertical direction, and 2) outflow from the soil to runoff – water transport in the drainage layer on the sloping impermeable horizon to the stream. The proportion of both transformations during runoff formation changes following the running phase of the soil water regime: 1) in the percolation phase, water flows through the soil into the drainage layer and outflows into the stream in discharge waves immediately reacting to precipitation, and 2) in the accumulation phase, the rainwater accumulates in the soil and does not outflow into the drainage layer.

In the percolation phase, the rising hydrograph limb grows very quickly and its duration is short – a few minutes or hours. It is due to the rapid vertical transport of water through the soil into a very permeable drainage layer caused by overfilling the soil with rainwater. This so called instability driven flow can occur in any part of the porous soil body regardless of the pore size, and can be caused by any rain regardless of its intensity, duration and total volume. The greatest value of the soil water content is reached during the rain and simultaneous instability driven flow. After it, the soil water content and consequently the outflow and runoff decreases, and therefore, the accumulation phase begins and the falling hydrograph limb is generated. Its source is the water stored mainly in the porous drainage layer. The runoff decreases according to the decline of the water content in the drainage layer.

The precipitation peaks are attenuated in the soil and in the drainage layer. The biggest attenuation takes place in the drainage layer. Roughly speaking, the runoff peak reaches to 10% of the precipitation ones. The source area, in which the runoff peak is generated, covers about 25% of the whole catchment area. The breadth of the source belt bordering the flow channels is no more than about 90 m. During the whole vegetation season, 50% of seasonal precipitation sum is used for plant transpiration, 25% makes the runoff, and 25% is stored in the drainage layer.

In this article it is demonstrated that the soil cover plays an ever greater role in the hydrodynamic theory of the rainfall – runoff transformation. Therefore, the detailed description of soil water transport offers a promising way for further research activities (Kostka and Holko, 2001; Tesař et al., 2001). Namely, the variable contributing area hypothesis may be modified in order to describe the growth of contributing areas in the drainage layer (Beven and Kirkby, 1979). The kinematic wave theory (Germann, 1985) seems to be a source of ideas for further development of the instability driven flow theory.

The creation of early warning systems is the urgent task in mountainous countries – in the Czech Republic and Slovakia - in order to reduce losses caused by floods. Results presented in this article show that the rise of rapid runoff is catalysed by the combination of the rain and the soil water content. It means that the early warning system has to be based on the measurement of both values.

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