

IMPLEMENTATION OF CONCEPTUAL LINEAR STORAGE MODEL OF RUNOFF WITH DIURNAL FLUCTUATION IN RAINLESS PERIODS

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The aim of this paper is to set up the model describing the depletion runoff process when a catchment is supplied only from subsurface water sources and also to describe a transpiration effect when depleted mostly by riparian vegetation considering it as a diurnal process in hot and rainless period. This model is conceived as a simple linear description of all forms of subsurface runoff supplying the stream flow in fully forested the Teply Brook small catchment (district Zdar nad Sazavou, the Czech Republic) for the dry seasons of 2003 to 2007 when the discharges were continuously recorded at the outlet. There are three interconnected soil water zones, represented by the corresponding subsurface storages in the model. Actual evapotranspiration from the riparian part of the catchment is approximated by a special form of the Fourier series. The model simulates a runoff process using seven independent parameters, evapotranspiration is simulated through another four parameters. The paper points out how the model can be successfully implemented in the simulation of diurnal discharge fluctuation on the small forested experimental catchment in dry periods.

KEY WORDS: Linear Runoff Model, Most Upper Soil Layer, Unsaturated and Saturated Zone, Hot and Dry Rainless Period, Transpiration of Riparian Vegetation.

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Na základě kontinuálně naměřených průtokových dat z období let 2003–2007 byl pro Teplý potok (okres Žďár nad Sázavou) sestaven lineární odtokový model. V článku je podrobně rozebírána hydrologická charakteristika povodí a na příkladech diskutována vhodnost nasazení tohoto třinádržového systému pro popis chování průtoku korytem v bezesrážkových obdobích. Navrhovaný model charakterizuje povodí s pomocí sedmi hydrologických nezávislých parametrů a dalších čtyř parametrů charakterizujících evapotranspiraci. Článek ukazuje, že řešení navrhovaného modelu může s dobrou přesností aproximovat chování odtoku tohoto malého, převážně lesnatého povodí.

KLÍČOVÁ SLOVA: lineární odtokový model, vrchní hypodermická půdní vrstva, nenasyčená a nasycená zóna, denní variabilita odtoku, horké a suché bezdešťové období, transpirace břehových porostů.

Introduction

The basic aim of the paper is to verify possibilities to describe a linear storage model (LSM) of a catchment that respects a particular hydrologic phenomenon – diurnal discharges fluctuation which can reduce during hot summer rainless periods almost one third of daily runoff on small catchments. We started our research comparing some earlier

results from international studies and from simulation approaches using various mathematical models.

The interaction between vegetation and hypodermic zone or shallow groundwater is an important part of ecosystem dynamics in the riparian zone. Where the catchment area supplies the alluvial groundwater, there can be a significant diminution of water volume actually reaching the stream,

though the flow direction is in the direction of the stream (Troxell, 1936).

Troxells' purpose was to better understand the transfers between streamflow, hypodermic zone or groundwater and transpiration and thereby demonstrate that the assumptions adopted by most hydrologists of the time were wrong, leading to an underestimation of the actual transpiration loss (Burt, Pinay, Sabater, 2010).

Burt (1979) was obviously one of the first who described the hydrological response of Bicknoller Combe, small brook in Somerset, England, during the summer drought of 1976. He noted that the observations of a short time step discharges showed a clear diurnal fluctuation and characterised this phenomenon as response to intense evapotranspiration during the day and a resumption of downslope water movement at night. Other author (Bren, 1977) published the effect of slope vegetation removal on the diurnal variations of a small mountain stream Cropper Creek draining a 46 ha catchment in Australia. He concluded that the amplitude of the variation is diminishing in time and it is insensitive to changes in slope hydrology. The group of authors (Bond et al., 2002) described the time lag between base flow defined by the daily maxima, and actual flow, and use it to estimate the zone, or area, of riparian vegetation that influences daily streamflow patterns. Other authors (Loheide, Butler and Gorelick, 2005) explored evapotranspirative consumption of groundwater using the concept of readily available specific yield based on diurnal discharge fluctuations. They described the method of specific yield determination based on soil texture and the "master recession curve". A big progress in unsaturated and saturated zones dynamics and also in evapotranspirative consumption by phreatophytes respecting diurnal water table fluctuations was made one year later when other authors (Fenicia, Savenije, Mathgen and Pfister, 2006) published their research. They implemented a simple reservoir model determining the storage-discharge relation that simulated the slow hydrograph component. This component was computed using a linear reservoir, which describes best the observed groundwater behaviour. The results were used to trigger a discussion of the general use a linear groundwater reservoir in hydrological modelling. This has been the starting point of our investigations.

The subject of our interest is the conceptual linear storage model (LSM) of a small natural catchment, where a continuing measurement of water discharges has been recorded within the years 2003

to 2007. The model aim is a hydrologic simulation of the daily dynamics of a runoff process. Our intention was not to solve direct rainfall-runoff process, and due to a small geometric and hydraulic parameter values (small cross-section profiles and a short channel length) no channel transformation solution process was planned. However, the aim of this paper is to develop and calibrate the model describing the recession runoff process when supplied just from subsurface water resources, and also to describe the effect of actual evapotranspiration when depleted by riparian canopy, considering it as a diurnal process.

This model is conceived as a simple linear description of all substantial forms of subsurface runoff supplying the stream flow in largely forested the Teply Brook small catchment (district Zdar nad Sazavou, Czech Republic) during the dry periods. The major idea behind is that the model presents three mutually interconnected water storages (most upper hypodermic soil zone, unsaturated, and saturated zones). Each of them represents a certain subsurface soil zone when its storage is time dependant on actual water available content participating in the runoff process. The actual evapotranspiration from the riparian part of the catchment is approximated by a special, simplified Fourier series formula.

Materials and methods

We have recently studied similar problems in the Teply Brook catchment southern Moravia, Czech Republic, where we utilised the rainfall-runoff data and data for potential evapotranspiration computation (Dvořáková, Zeman, 2010a). The streamflow high resolution data series showed also the diurnal fluctuation during hot and dry summer rainless period on the Teply Brook.

Catchment description and measurement technology

The Teply Brook catchment is located about 20 km South West from the city of Nove Mesto na Morave, the Czech Republic (Fig. 1, catchment outlet ordinates: x: -619069.767, y: -1130440.339). The catchment characteristics have been derived from a detailed GIS analysis and they are summarised in Tab. 1.

The altitude ranges from 417 m to 602 m a.s.l. with a mean elevation slightly over 500 m a.s.l. (Fig. 1a, Tab. 1). The mean catchment slope is about 22% with maximum slopes more than 55%

occurring in the middle part of the catchment (Fig. 1b). The geology of the catchment is mainly metamorphic rocks (gneiss, amphibolite, migmatite) and in a lower altitude of the catchment area they are peridotite and serpentinite (Fig. 1c). Alluvial deposits of loamy sands and gravels occupy the valley bottom.

As far as the soils are concerned, typical for the catchment area are forms of Cambisols and Leptosols (Fig. 1d), as there are presented hapliccambisols, alcalic or cambicleptosols. In the streams valley are fluviccambisols and gleysols (FAO-WRB, 2006).

The prevailing land use is a mixed semi-natural forest of spruce (*Piceaabies*), pine (*Pinussylvestris*), hornbeam (*Carpinusbetulus*) and alder (*Alnusincana*). Only in the southern part of the catchment area is arable land. The climate is moderately warm and humid, with an estimated mean annual precipitation of 650–750 mm and mean annual temperature 7 °C.

T a b l e 1. Characteristics of the Teply Brook catchment.

A	Catchment area	1.56	[km ²]
L _{th}	Length of thalweg	1.41	[km]
L _{MF}	Length of main river	1.10	[km]
P	Length of waterdivide	5.13	[km]
B	Average width of catchment	1.10	[km]
I _R	Average slope of river	13.0	[%]
I _S	Average catchment slope	15.1	[%]
H _{max}	Maximum catchment elevation	595	[m a.s.l.]
H _{min}	Minimum catchment elevation (outlet)	417	[m.a.s.l.]

Measured data

By the submersible level gauge Vega Vegawell 71 with a range of application from 0.0 m to 1.0 m water column with a sapphire membrane was measured at 1.0 m upstream the V-notched weir (Fig. 2) water level height. For the digitalization AD converter Drak3 was used. Data were stored in the classic PC installed in the Technical Maintenance building of the alternative storage for fuel of the Nuclear Power Plant Skalka, Czech Republic. Measurements were made in the period of 1.56 s (Zeman, 2007).

The theoretical distinction of water level was 0.1 mm, which corresponds to the discharge of 6 cm³ s⁻¹, (i.e. 0.4 %). Discharges were determined from water level measured 1.0 m upstream from the weir and calculated according to Eq. (1):

$$Q = 1.415 \cdot h^2 \cdot \sqrt{h}, \quad (1)$$

where Q – discharge [m³ s⁻¹] and h – water level [m]. Eq. (1) is known as the Thomson weir equation which is broadly used for discharge measurement.

It can be reasonably assumed that the evaporation, i.e. transpiration of the forest, is both a function of light intensity, and also a function of temperature and relative air humidity. Time of sunrise and sunset are shown in Fig. 3. Air humidity during the night is close to saturation (as shown on the continuous measurement in 2004–2005).

Model simulation

We tried to write out the simplest linear model (LSM) possible, which would describe, at least approximately, all the stream flow characteristics during rainless periods (Dvořáková, Zeman, 2010b). We created the matrix coefficients $B_{(3,3)}$, which characterize the catchment as a whole. Matrix elements include the characteristic constant parameters of the catchment, i.e. most upper soil layer, unsaturated storage and groundwater storage. Furthermore, permeability coefficients between each layer and permeability between layers and the brooks were used, as shown in the explanation to the Eq. (2).

Multiplying this matrix by the zonal water saturation vector h , we get a time change in vector h and the right side of the equation, formed by evaporation, approximated by the simplified Fourier series, which is a sine function of evapotranspiration time series. Such approximation conforms to the diurnal variation on leaf water potential in a forest canopy (Jarvis, 1976). The time series of this function are close to the time series of sun radiation, and albedo values of the stomatal vegetation resistance (Dickson et al., 1986). In such a model we search for matrix B elements which correspond to the measured runoff data.

The data were selected from those measured on the Thomson weir next to the estuary of the Teply Brook in the period from spring 2003 to autumn 2005. We tried to find linear coefficients, which would properly characterise the runoff process if the Teply Brook during rainless period, when the stream flow is determined by the progressive water depletion from all three soil zones. It has been shown, that in order to describe this process it would be necessary to construct at least three water reservoirs (Fig. 4), in which the available water always moves toward lower saturation levels, proportionally to the volume saturation variations.

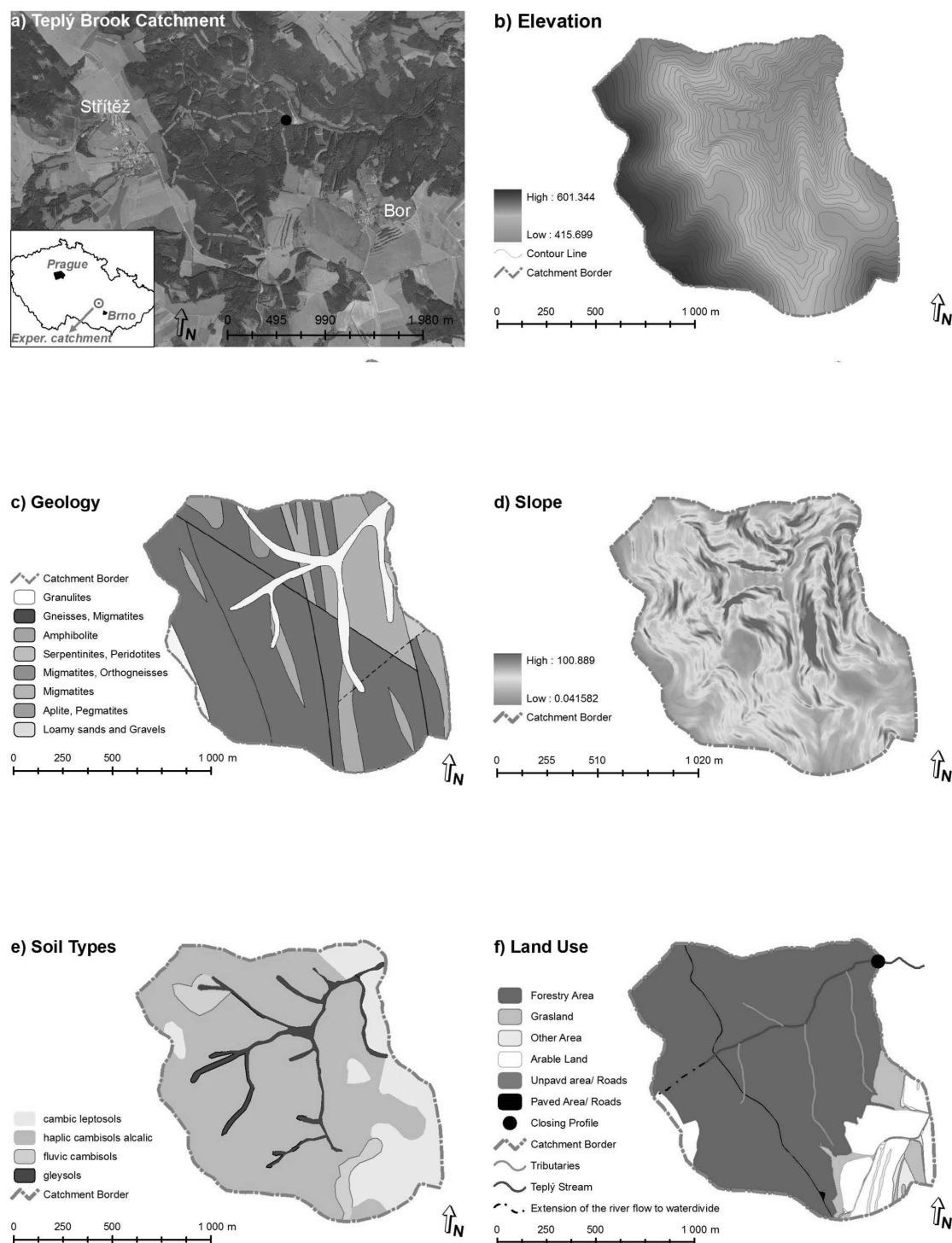


Fig. 1. GIS Maps of the Teplý Brook catchment characteristics.



Fig. 2. Discharge gauging profile on the Teply Brook catchment.

Also, the depth of runoff from each zone to the stream flow is directly proportional to the available water in the zone. Simpler models could not be used for describing all types of influences on the flow by the catchment, which appear during rain-

less periods. In Fig. 5 we were mainly concerned with the description of the most common flow processes.

Results

The analysed version of the simulated process, described above, assumes evaporation characterised by the sine harmonic function, and is independent from days, cloudy formations, air humidity and above all the volume of available water in the most upper hypodermic layer. Despite these drastic limitations, which have been adopted for better analytical solutions of the linear model, the consistency between measured discharges and computed ones is quite strong. Their consistency is expressed by statistical criteria, most commonly by determination and variation coefficients (e.g. *Beven*, 2006).

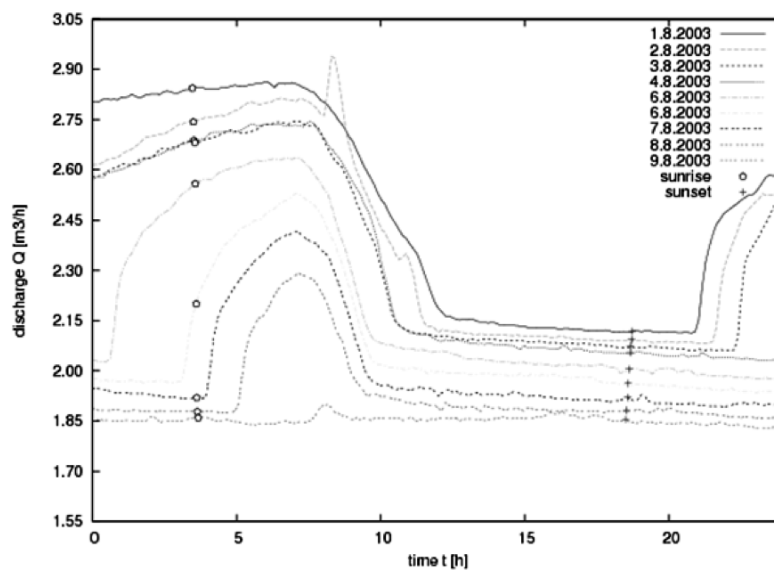


Fig. 3. Hydrograph of the Teply Brook catchment in period 1.–9. 8. 2003 with the marked time of sunrise and sunset.

As stated in the introduction, we are trying to find a description of the catchment runoff by means of differential equations, with a non-nil right side, in order to describe, at least in principle, the runoff process, as seen in Fig. 4 and 5. We described the hydrological system as a system of three interconnected reservoirs, each having the capacity to transfer water to the other reservoir and at the same time in to the brook basin. The flow between reservoirs or between reservoir and catchment is directly pro-

portional to the difference in water storage in the zones. The drained water storage from one zone to other places than to the stream flow, or to a neighbouring zone, is defined by the harmonic function of time. These premises can be worked out as follows:

$$\frac{d\vec{h}}{dt} = B\vec{h} + \vec{v}, \quad (2)$$

$$\text{where } \vec{v} = \begin{pmatrix} 0 \\ 0 \\ \beta(\sin(\omega t + \varphi) + 1) \end{pmatrix} \quad \text{and}$$

$$B = \begin{pmatrix} \frac{-a_{12}}{V_1} & \frac{a_{12}}{V_1} & \frac{a_{13}}{V_1} \\ \frac{a_{12}}{V_2} & \frac{-(a_{12} + a_{22} + a_{23})}{V_2} & \frac{a_{23}}{V_2} \\ \frac{a_{31}}{V_3} & \frac{a_{23}}{V_3} & \frac{-(a_{23} + a_{33})}{V_3} \end{pmatrix}$$

a $\vec{h}(t)$ is storage vector of available water in zones.

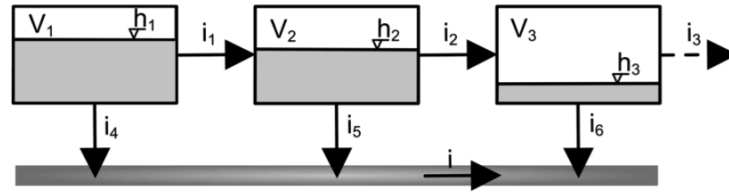


Fig. 4. Model scheme of streamflow dynamics in rainless periods.

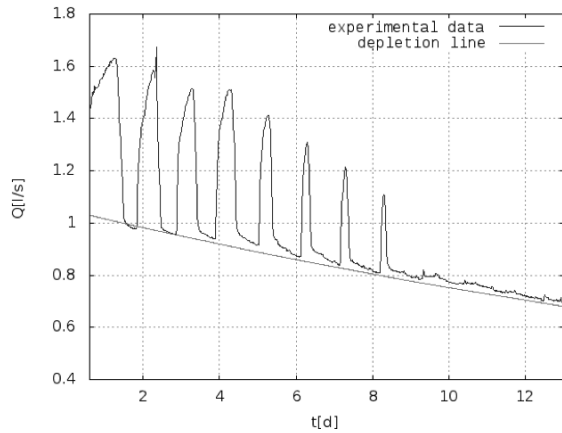


Fig. 5. Fluctuation of discharges in the Teply Brook catchment in rainless period 1.-13. 8. 2003, indicating the depletion line.

water zone, h_2 – water saturation of the unsaturated zone, h_3 – water saturation of the most upper soil layer, β – evaporating coefficient, ω – periodicity of the main evaporation element, which we assume as equal to 24 h, φ , ψ – time shift between maximum evapotranspiration and midnight UTC [day].

This system with right side nil value has the solution, $\vec{h}_0(t) = e^{\lambda_1 t} \vec{k} + e^{\lambda_2 t} \vec{l} + e^{\lambda_3 t} \vec{m}$, where the eigenvalues λ_1 to λ_3 correspond to negatively con-

The meaning of coefficients used in matrix B are as follows: a_{11} – groundwater runoff into the streamflow, a_{22} – runoff from the unsaturated zone to the streamflow, a_{33} – runoff from the most upper soil (upper hypodermic) layer to the streamflow. In the matrix we consider (except of diagonal) $a_{ij} = a_{ji}$, a_{12} – runoff from the groundwater zone to the unsaturated zone, a_{13} – equals 0, because the groundwater does not flow directly to the most upper soil layer, a_{23} – runoff from the unsaturated zone to the most upper soil layer, V_1 – storage of the groundwater reservoir [m^3], V_2 – storage of the unsaturated reservoir [m^3], V_3 – storage of the most upper soil reservoir [m^3], h_1 – water saturation of the ground-

sidered reciprocal relaxation time of emptying zones. The solution with non-nil right side has the form

$$\vec{h}(t) = e^{\lambda_1 t} \vec{k} + e^{\lambda_2 t} \vec{l} + e^{\lambda_3 t} \vec{m} + \vec{n} \quad (3)$$

$$\text{where } \vec{n} = \begin{pmatrix} \beta_1 \sin(\omega t + \gamma_1) \\ \beta_2 \sin(\omega t + \gamma_2) \\ \beta_3 \sin(\omega t + \gamma_3) \end{pmatrix} \quad \text{and}$$

$\vec{k}$, $\vec{l}$, $\vec{m}$ – eigenvectors of hydrological system where the individual components have a meaning of the corresponding storage contributions to discharges in given delayed time,

β_1 , β_2 , β_3 – contribution amplitudes of individual storages to periodical components to stream flow discharges,

γ_1 , γ_2 , γ_3 – phase delays of maximal contributions from individual to total streamflow discharges [day].

The coefficients of this solution are found through the method of smallest squares divergences measured and calculated flow in selected flow sections.

For coefficients defined by the relation (2) supply of each reservoir in the stream flow is taken as

$$i_4 = h_1 \cdot a_{11} = 0, \quad i_5 = h_2 \cdot a_{22}, \quad i_6 = h_3 \cdot a_{33}.$$

The total discharge is therefore their sum:

$$i = i_4 + i_5 + i_6.$$

Flow between each reservoir is considered equal to:

$$i_1 = (h_1 - h_2) a_{12}$$

$$i_2 = (h_2 - h_3) \cdot a_{23}$$

From these relations and relation (3), our approximation indicates the change of form in stream flow Q is as follows:

$$Q(t) = p \cdot e^{-\frac{t}{\tau_1}} + q \cdot e^{-\frac{t}{\tau_2}} + r \cdot e^{-\frac{t}{\tau_3}} + \delta \sin(\omega t + \psi). \quad (4)$$

The coefficients in this equation were found by interpolation of measured data *ab initio*. Resultant values are indicated in Tab. 2. The coefficient ω equals 2π . The results of these coefficients, compared with the measured stream flow *ab initio*, are indicated in Figs. 6 – 8.

Table 2. Approximate function coefficients during the investigated period.

Event date	$P [m^3 \text{ day}^{-1}]$	$\tau_1 [\text{day}]$	$Q [m^3 \text{ day}^{-1}]$	$\tau_2 [\text{day}]$	$R [m^3 \text{ day}^{-1}]$	$\tau_3 [\text{day}]$	$\delta [m^3 \text{ day}^{-1}]$	$\Psi [\text{day}]$
9.–17. 10. 2003	145.811	1.983	213.592	0.179	133.341	155.969	3.454	–0.196
10.–11. 9. 2003	63.489	3.495	26.649	0.182	14.612	235.181	4.291	0.429
18.–22. 5. 2005	81.305	0.140	119.701	0.869	332.05	29.746	17.789	–0.662

Table 3. The values of statistical indicators for the assessed period.

Event date	Coefficient of determination	Coefficient of variation
9.–17. 10. 2003	0.9943	0.0175
10.–11. 9. 2003	0.9983	0.0086
18.–22. 5. 2005	0.9903	0.0091

The accuracy of the implemented model calibration, i.e. consistency between measured flow values with calculated flow values, is expressed by two statistical criteria, i.e. determination coefficient, called the Nash-Sutcliffe efficiency coefficient (Nash, Sutcliffe, 1970), and a variation coefficient (Tab. 3).

$$CD = 1 - \frac{\sum (Q_m - Q_v)^2}{\sum (Q_m - \bar{Q}_m)^2} \text{ and } CV = \frac{\sqrt{\frac{\sum (Q_m - Q_v)^2}{n}}}{\bar{Q}_m}$$

In Tab. 2 herein, the coefficients have the following meaning: P , Q , R – contribution of each soil layer per day: P (V_1), Q (V_2), and R (V_3) (see Fig. 4), τ_i – halve-time to reach 1/e of the equivalent layer (V_i – compartment), e – the Euler constant, δ – estimate of evapotranspiration, Ψ – time lag between transpiration peak time and midnight time.

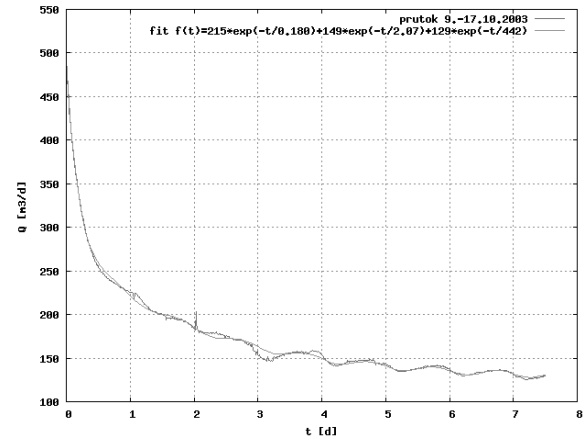


Fig. 6. Measured and computed discharges comparison on the Teply Brook catchment in time period 9.–17. 10. 2003, with approximate functions (3) and coefficients indicated in Tab. 2.

where Q_m – measured discharges [l s^{-1}], Q_v – calculated discharges [l s^{-1}], $\bar{Q}_m$ – mean value of measured discharges [l s^{-1}] n – number of measured values.

The coefficient of determination CD in best fit conditions equals 1.0; on the other hand, the coefficient of variation CV in ideal conditions equals 0.0 (Chow et al., 1988).

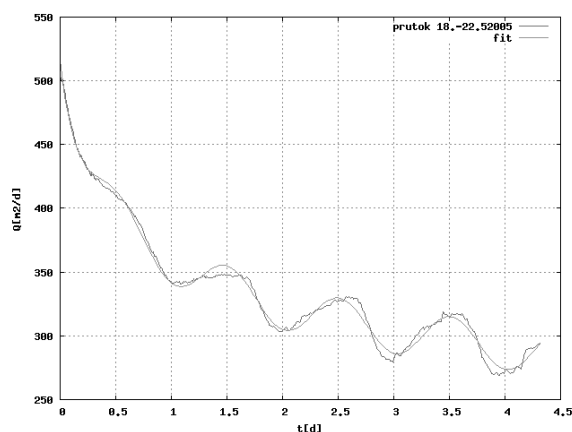


Fig. 7. Measured and computed discharges comparison on the Teply Brook catchment in time period 18.–22. 5. 2003, with approximate functions (3) and coefficients indicated in Tab. 2.

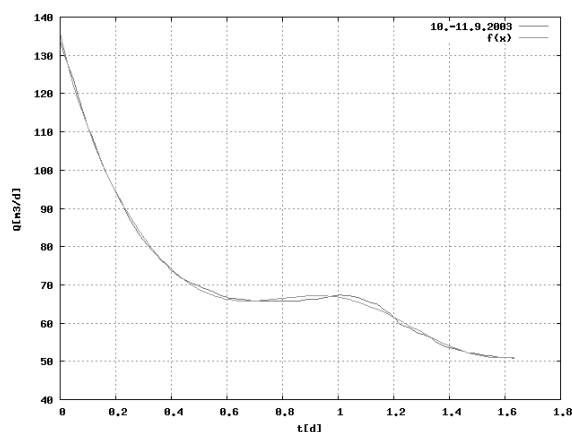


Fig. 8. Measured and computed discharges comparison on the Teply Brook catchment in time period 10.–11. 9. 2003, with approximate function (3) and coefficients indicated in Tab. 2.

Discussion

The first author who described a diurnal fluctuation of flow was American hydrologist Troxell who stated three assumptions of this phenomenon in his excellent paper (Troxell, 1936):

- Discharge cycle immediately identifies the time of minimum and maximum transpiration loss.
- Curve connecting the points of maximum discharge represents the stream discharge if there is no transpiration.
- Difference between this quantity (i.e. maximum stream discharge) and the actual discharge equals the transpiration loss).

Troxell's approach combines records of stream discharge and indirectly actual transpiration in rainless periods.

The model under discussion is coherent with Troxell's assumptions. From collected data and

observations it has been assessed, that in the hot and dry periods the mentioned linear model was a good tool for runoff simulation. In the simulated periods, the chance that the use of the model will be impossible, does not occur, because the situation, where a linear model cannot be organically integrated and one of the sub surface zone dries up completely, also does not occur. The linear model in this situation continues by making the vector $\vec{h}$ components negative leading to the situation, that even from emptied (dried up) zones there is water runoff. This problem can be dealt with only partially, using the so called threshold, which means during complete drought the function of the upper most layer replacing matrix $B_{(3 \times 3)}$ by matrix $B'_{(2 \times 2)}$, thus solving the problem of only two interconnected zones. This is a general problem with so called threshold models. This threshold represents a non-linear element in a linear model and in fact is combination of threshold and linear reservoir. Winsemius et al., (2006) describe the implementation of two threshold models (LEW and STREAM), which describe non-linear behaviour of a part of the river Zambezi, based on a combination of a threshold and linear model. Mul et al. (2007) simulated the rainfall-runoff process in the South Pare Mountains, Northern Tanzania, using a conceptual model STREAM which generates the runoff from the catchment. This model has been extended with an extra storage sub-model LEW which simulates a drainage of riparian zone. Such a storage unit as the LEW sub-model, is likely simpler than our linear model LSM with three compartments, however it might be possible, like in the case of the LEW sub model, to associate it with a conceptual model, e.g. KINFIL (Vrana et al., 2012), HEC-HMS (US ACE, 2000), etc. Connecting LSM linear model with the KINFIL conceptual model enables detailed assessment of the extent of a runoff process and simulation of diurnal transpiration process in a riparian zone. There are a number of similar examples how to use our model LSM as either the isolated tool for a simulation of diurnal base flow fluctuation or as an added tool joint to a generic water balance model running usually in time step one day or longer to describe an extra transpiration phenomenon of riparian zone in hours.

Concerning the analysis of base flow fluctuation due to diurnal transpiration it may be mentioned that logically in hot and dry days this phenomenon manifests more expressively with a higher amplitude and a shorter delay (Bond et al., 2002). This

observation is coherent with our measured data. The same authors and also *Mul et al.* (2007) and *Winsemius et al.* (2006) found that the time lag between the potential evaporation and reduction of discharges is larger during the cooler period than during the hot period. Our conclusion is the same.

In general, it can be said that our LSM model can be applied regardless of geographical position and soil hydrology conditions. However, a significant evaporative conditions and dense riparian vegetation along the stream provide usually more results. A threshold model is used for non-linear models, specifically in case of extremely non homogenous catchment structures (e.g. *Kovář, Dvořáková, Kubátová*, 2006; *Budagovskiy and Novak*, 2011).

The LSM model has eight major parameters, which are interrelated. Their physical meanings are clear from the Tab. 2 legend. We assume that actual evapotranspiration is proportional to the most upper soil layer and it has a periodical character. It is also related to the diurnal conditions during the period. From this periodicity we consider its major harmonic component which corresponds to evapotranspiration consumed from the loss of discharge (see the Troxell's assumptions).

When the storage of each three layers (i.e. reservoirs V_1 to V_3) has different degrees of saturation, their coefficients differ too and the model ceases to be linear (quasi-linearity phenomenon). The LSM model can simulate a runoff event when a saturation process changes slowly and continuously. However, when the individual events are compared, the parameter value can differ more significantly. Usually, in one isolated event the parameter error might represent less than 5%, but in case several events occur, the error might be as much as 20%.

It is interesting to note, that the time delay between midnight and maximum transpiration impact on the stream flow, varies according to the season (see last column Tab. 2). This is caused by the occurrence of high values of potential evapotranspiration. From the point of view of intense transpiration of phreatophytic trees (willow, poplar, alder etc.) during the summer months, these trees create in favourable conditions unsaturated flow of groundwater into the root system. In case of an isolated tree it is radial, in serial plantation it is lineal (*Meyboom*, 1966). This physiological effect can increase even more the difference between minima and maxima daily fluctuation (*Anderson*, 2001).

Conclusions

Diurnal base flow fluctuation has been observed during hot and dry periods with high potential evapotranspiration rates. Similarly, as other authors (e.g. *Mul, Savenije, and Uhlenbrook*, 2007), we did not find on the Teply Brook such diurnal discharges fluctuation when water level is not extremely low and, at the same time, the evaporation conditions are highly exhaustive. Actual transpiration of riparian vegetation along the stream plays a significant role under these conditions. *Bond et al.* (2002) indicates that the groundwater flow is disconnected from the riparian zone and therefore „that trees do not tap out of this reservoir“. We also agree with this assumption arguing the area of riparian vegetation is smaller than 5% of the catchment area. Nevertheless, about 0.3 to 0.6 l s⁻¹ is the actual loss of discharges during a series of days.

It may be said, that the approximate functions, derived as a consequence from the linear triple reservoir, when total drying up of one of the soil layer does not occur, illustrate the measured data quite accurately, despite the fact that for the amount of transpiration we have used only the first dominant member of Fourier series. Such coefficients are relevant input values for determining model parameters of the Teply Brook. The results of comparing measured and calculated stream flow in three investigated time periods corroborate this fact. It will be useful to use the harmonic fluctuation of change (diurnal/nocturnal) not only for measuring or modelling of stream flow or current evapotranspiration of a coastal zone, but also for increasing the alkalinity in water and concomitant processes during diurnal stream flow fluctuation (*Tetzlaff et al.* 2007). These phenomena open further possibilities for detailed research. Therefore, we are confident that the proposed model, with its system of parameters, is applicable for similar catchments. This, however, will be investigated in the next papers.

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