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Impact of Evapotranspiration on Diurnal Discharge Fluctuation Determined by the Fourier Series Model in Dry Periods

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Abstract

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Precise measurements of discharges at the outlet of a small catchment, using high resolution sensing equipment, can currently be done without difficulty. In particular, measurements can take place even during dry periods, when high temperatures increase actual evapotranspiration on the catchment and diurnal streamflow fluctuation changes occur in a harmonic wave at any time of the day. Some 10–15 years ago, a current runoff measurement record based on a high resolution equipment clearly recognizing a diurnal wave-shape fluctuation could hardly be available. The measurement of discharge ordinates from the catchment, and from free water pan evaporation, showed an undulating fluctuation tendency. However, the discharge minima appeared at day time and their maxima at night. The measured discharge data are represented not only by a fluctuating form, but also by a mild form, an even straight line, or by a flat depletion curve. For the purpose of analyzing the wave shape of discharge we implemented the Fourier series model, simulating the measured data through the Fourier input, output, and transformation coefficients. The purpose of this analysis was to use the Fourier equations in order to substitute the missing data (when the discharge or evaporation measurements collapsed). Due to very sensitive data, when the measured discharge series are jagged, the equation can be smoothed by the harmonic approximation or by the polynomial approximation. Our study was carried out on the small experimental catchment of the Starosuchdolsky Brook, in the vicinity of the campus of the Czech University of Life Sciences Prague. The harmonic analysis provided an interesting outcome, as well as innovative methodology.

Keywords: catchment depletion curve; Fourier series; harmonic coefficients; high resolution sensing; rainless periods

Long and dry depletion events on small catchment provide valuable data for the assessment of the impact of actual evapotranspiration on runoff reduction at various scales. Solar radiation and temperature variations cause streamflow diurnal cycles, which also can be used to assess the impact of climate change on catchment behaviour (Mutzner *et al.* 2015). The first indication of the streamflow fluctuation caused by evapotranspiration was based on observations of a small catchment in the dry year 1976 (Burt 1979). Diurnal streamflow cycles were characterized through their amplitudes and timing of the minimum and maximum streamflow. The harmonic process of baseflow delay was described, tracing lower discharge values in the daytime hours and higher values in the night hours, due to the same process of evapotranspiration. The delay of the wave-shaped depletion curve was caused by the evaporation conditions and partly also by hydraulic roughness (Dvořáková *et al.* 2014).

In the last two decades, thanks to extremely accurate measurement of water discharges, many relevant books have been published describing the impact of evapotranspiration on catchment runoff (Zhang *et al.* 2001; Brown *et al.* 2004; Loheide *et al.* 2005; Fenicia *et al.* 2006; Winsemius *et al.* 2006). In his paper “Catchment as simple dynamic systems”, Kirchner (2006, 2009) formulated the mass-conservation equation:

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$$\frac{dS}{dt} = P - E - Q \tag{1}$$

where:
S – water storage
P – precipitation
E – evapotranspiration
Q – discharge

In this equation, only the discharge is an aggregated measurement for the entire catchment. In the case of a dry period we could neglect precipitation to get the sum (E–Q) on the right side of this equation, which would then express the storage time series showing a drying process on the left side of Eq. (1).

Hydrologic processes in small catchments started to be analyzed and described using a modern systems approach in the late 1960s, soon after the systems engineering linkages and their feedbacks were explained and published (Kraijenhoff & O'Donnell 1966). Systems hydrology nowadays takes into account not only the rainfall-runoff correlation, but also the correlation between runoff and evaporation (Kirchner 2009). All this became possible thanks to high-resolution measuring equipment. Both links of measurement, rainfall–runoff or evapotranspiration–runoff represent important hydrological processes that can be described by Fourier series (Hardy & Rogosinski 1971; Kovář *et al.* 2014a).

Extreme droughts are often estimated through streamflow discharge measurements. They show harmonic evapotranspiration rates that have orders of magnitude, which are smaller than the levels assessed on typical catchments (Kirchner 2006, 2009; Langhammer & Vilimek 2008; Deutscher & Kupec 2014). The problem is thus clearly delimited. It can be resolved through the Fourier series, based on the systems theory.

MATERIAL AND METHODS

Fourier series model. The starting point of the Fourier model introduction can be Eq. (1) and its transfer into Eq. (2) to express the input–output analysis:

$$\frac{dS}{dt} + y(t) = x(t) \tag{2}$$

where:
$x(t)$ – straight line/curve input of the depletion curve
$y(t)$ – output in the form of undulated streamflow discharges

For the computed discharges: the $yc(t)$ in Eq. (3) from the measured variables $x(t)$ and $y(t)$ during a rainless period we need a transformation function $u(t - \mathrm{t})$ in the convolution integral to be the theoretical alternative of the discharges computation. The rational computation of the dry events requires to substitute the integral by the summation of $x(t)$ and $u(t - \mathrm{t})$ multiples within the certain limits corresponding to the duration of the event in Eqs (3) and (4):

$$yc(t) = \int_0^t x(\tau) \times u(t - \tau)d\tau \tag{3}$$

$$yc(t) = \Delta t \sum_{i=1}^{n} (x(i) \times u(n - i)) \tag{4}$$

The term of addition in Eq. (4): $x(i) \times u(n - i)$ for the finite limits expresses the convolution procedure when $x(t)$ is not zero, then the computed runoff $yc(t)$ can be expressed by Eq. (5) which is the Fourier expansion:

$$g(t) = yc(t) = A_0 + \sum_{r=1}^{n-1} (A_r \times \cos r\frac{2\pi t}{n} + B_r \times \sin r\frac{2\pi t}{n}) \tag{5}$$

The Fourier series describe the harmonic periodic process as an orthogonal function (Hardy & Rogosinski 1971). The function $g(t)$ in the interval $1 < t < n$ can be represented in any time t of the interval. The cosine and sine functions are orthogonal to one another yielding a K value that equals to $n/2$.

The harmonic coefficients A_0, A_r and B_r are the output coefficients for the $yc(t)$ runoff computation, where r is the index for harmonic coefficients, n is the length (i. e. the number of discharge ordinates) of the time series. The output function $y(t)$ transformed by the evapotranspiration process has the following coefficients:

$$A_r = \frac{n}{2}(a_r \times \alpha_r - b_r \times \beta_r), \text{ but } A_0 = n \times a_0 \times \alpha_0$$
$$B_r = \frac{n}{2}(a_r \times \beta_r - b_r \times \alpha_r) \tag{6}$$

The other coefficients a_0, a_r, b_r are the input coefficients $x(t)$, and α_0, α_r, and β_r are the transformation coefficients $u(n - t)$ – see Eq. (4) (O'Donnell 1960; Kraijenhoff & O'Donnell 1966).

The Fourier Series Model (FSM) has been adapted from the classic Fourier series expansion, which was developed earlier for simulation of rainfall-runoff events. However, instead of rainfall hyetograph as an input function, the depletion curve function $x(t)$ is used either in the form of straight line or in the form of an exponential curve (Boussinesq) approximating

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a depletion process. Thus for the input function $x(t)$, the Fourier series can be written as follows:

$$x(t) = a_0 + \sum_{r=1}^{n-1}\left(a_r \times \cos r\frac{2\pi t}{n} + b_r \times \sin r\frac{2\pi t}{n}\right) \tag{7}$$

where the input coefficients a_0, a_r, and b_r:

$$a_r = \frac{2}{n}\sum_{r=1}^{n-1}\left(x(t) \times \cos r\frac{2\pi t}{n}\right),\ \text{but: } a_0 = \frac{1}{n}\sum_{r=1}^{n-1} x(t)$$
$$b_r = \frac{2}{n}\sum_{r=1}^{n-1}\left(x(t) \times \sin r\frac{2\pi t}{n}\right) \tag{8}$$

The transformation process $x(t)$ to $y(t)$ is based again on the Fourier series expansion for the transformation function with the coefficients α_0, α_r, and β_r. The basic equation for this function $u(t)$ is:

$$u(t) = \alpha_0 + \sum_{r=1}^{n-1}\left(\alpha_r \times \cos r\frac{2\pi t}{n} + \beta_r \times \sin r\frac{2\pi t}{n}\right) \tag{9}$$

Then we solved the coefficients as follows:

$$\alpha_r = \frac{2}{n} \times \frac{\alpha_r \times A_r + b_r \times \beta_r}{a_r^2 + b_r^2},\ \text{but } \alpha_0 = \frac{1}{n} \times \frac{A_0}{a_0}$$
$$\beta_r = \frac{2}{n} \times \frac{a_r \times \beta_r + b_r \times A_r}{a_r^2 + b_r^2} \tag{10}$$

Now, after all coefficients are complete, we can finish this procedure and go back to Eq. (5) and substitute all coefficients to it.

The prevailing physiographic characteristics on the Starosuchdolsky Brook catchment are given in Figure 1 and Table 1. The meteorological station is

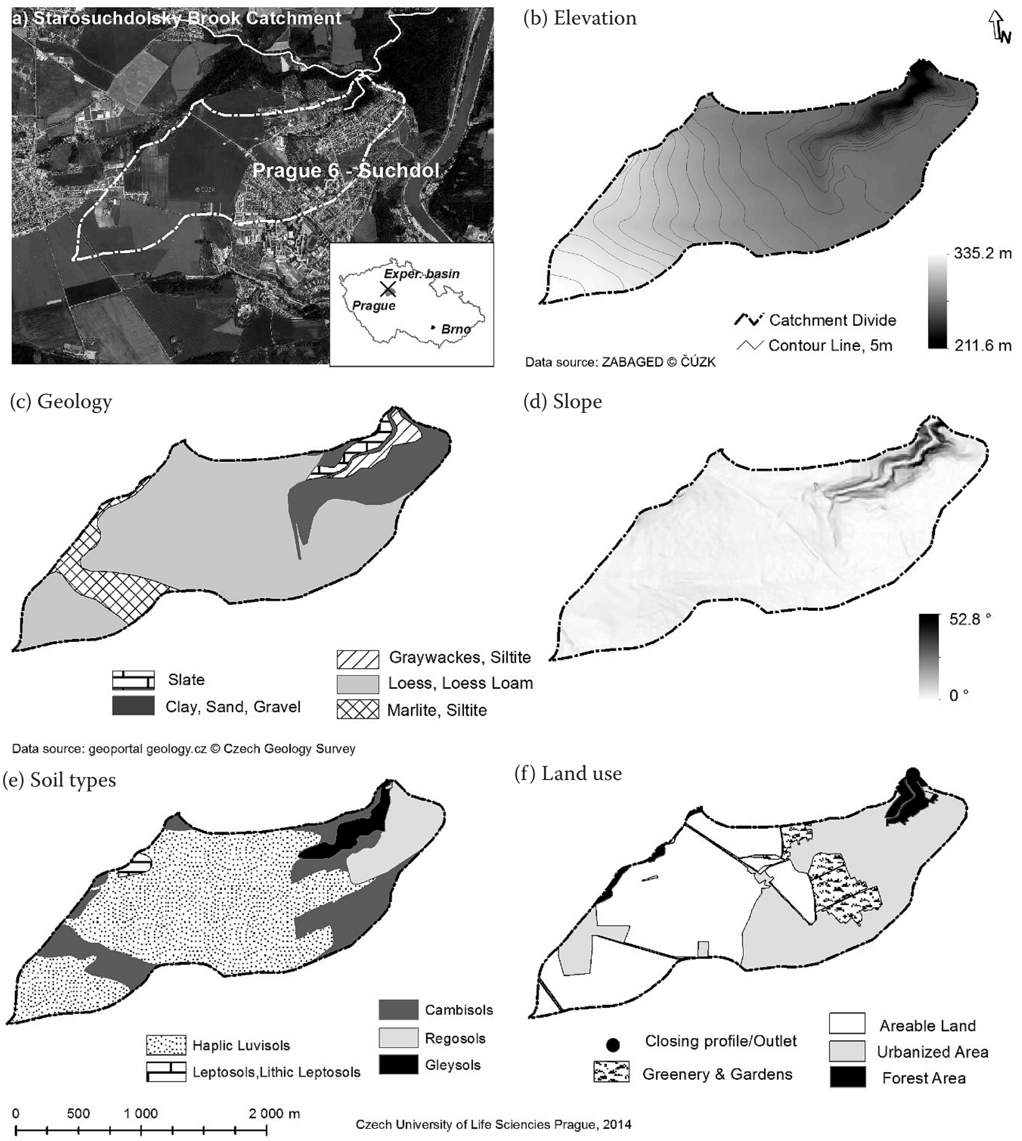

Figure 1. Selected characteristics of the Starosuchdolsky Brook catchment

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Table 1. Characteristics of the Starosuchdolsky Brook catchment

Physiographic factors of the Starosuchdolsky Brook			
Catchment area (A, km^2)	2.95	maximum catchment elevation (H_{max}, m a.s.l.)	335
Length of thalweg (L_{th}, km)	3.7	minimum catchment elevation (outlet) (H_{min}, m a.s.l.)	211
Length of brook (L_b, km)	0.58	river network density (R_d, –)	0.33
Length of water divide (P, km)	9.1	annual average precipitation (mm)	496
Average slope of brook (J_b, %)	5.4	annual average runoff (mm)	158
Average catchment slope (J_s, %)	20	annual average temperature (°C)	8.8
Land use categories of the Starosuchdolsky Brook			
Arable land (%)	50.2	urbanized area (%)	37.9
Forest (%)	3.5	permanent grassland/greenery (%)	8.4

The daily automatic measurement of precipitation and temperature: 2004–2015, our own measurement of runoff in the outlet profile "Spaleny Mlyn": 2011–2015

located close to the Czech University of Life Sciences Prague campus, longitude 14°22', latitude 50°08', time zone CET (GMT + 1 hour). The precipitation and temperature have been automatically measured daily since 2004, our own monitoring of runoff in the outlet profile "Spaleny Mlyn" takes place also on daily basis, since 2011.

The land use is represented by arable land (50% of the catchment area) and urbanized areas (including gardens 38%). The middle part of the Brook which flows through the urbanized area within a close circular profile is only interrupted by inspection shafts and a few junction wells. These possibilities can hardly allow to pump water for irrigation. Forested area is a mixture of semi-naturals. The downstream part of the catchment is environmentally protected in its riparian belts by a valuable canopy. These river belts, situated on both sides of the Starosuchdolsky Brook, contain typical local forest species represented by *Alnus glutinosa*, *Fraxinus excelsior*, *Quercus robur*, and rarely *Carpinus betulus*. During a hot and dry summer period, diurnal discharge fluctuation occurs in the studied catchment. Soil moisture and groundwater measurements in the catchment indicate the presence of a soil layer stretching at a depth over 1.0 m, which is always partially water-saturated due to deep valley morphology along the downstream part of the catchment . Therefore, a little groundwater contributes to the streamflow even in dry seasons.

RESULTS

The Starosuchdolsky Brook catchment has been monitored since 2011. The discharge data is obtained from the water level data, taken from measurements that are carried out every ten seconds at the outlet of the catchments, using a V-notched Thomson weir equipped with a cable sensor type Vegawell 71 (Vega, Grieshaber KB, Germany) submersible water level gauge. The gauge measures the water level using a pressure transducer with a high resolution sensitivity.

Discharge Q represents an aggregated measurement for the entire catchment (Kirchner 2006, 2009). Runoff processes are evidently the most important components of the hydrological cycle that can be conveniently measured. That is why many papers cite this experience (Tallaksen 1995; Beven 2006) etc. Our team also shared this experience.

Among many dry episodes measured on the catchment of the Starosuchdolsky Brook, we have selected the Event 1: since 08/08 (2:00 h) to 16/08 (20:00 h), 2012 (n = 211 h), where n is the number of time steps Δt, which was set to one hour.

The Event 1 discussed above is presented in this paper. First, the linear regressions and also the exponential correlation are presented in Table 2 to

Table 2. Linear and exponential regressions of the depletion curves of the Starosuchdolsky Brook catchment in the rainless Event 1

Linear regressions	
Approximated equations: $y = a \times x + b$	
$a = -0.001822$	$b = 0.959497$
$R^2 = 0.022853$	
Exponential regression	
Approximated equations (Boussinesq):	$y = y_0 \times e^{-\alpha x}$
$y_0 = 0.900967$	$\alpha = -0.001064$
$y = 0.900967 \times e^{-0.001064x}$	
$R^2 = 0.044559$	

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Table 3. Transformation coefficients α_r and β_r of Fourier Series Event 1 (n = 211, rr = 15)

	0	1	2	3	4	5	6	7	8	9	10	11	12	13	14
α	0.005	0.004	0.004	0.005	0.003	0.004	0.002	0.001	–0.003	0.039	0.009	0.011	0.008	0.010	0.010
β		0.001	–0.001	0.001	0.002	0.001	0.003	0.004	0.011	–0.026	–0.008	–0.003	0.000	–0.002	0.002

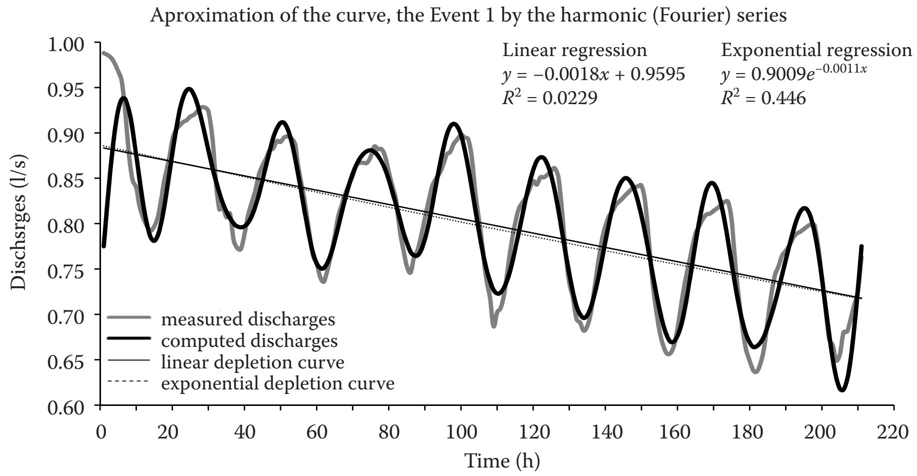

Figure 2. Discharges in the dry period from 8/8 /2012 to 16/8/2012 measured on the Starosuchdolsky Brook catchment

illustrate the Fourier input series, and then their coefficients a_r, b_r are computed (see Eq. (8)). These input series simulate a smooth depletion process in the form of either a line (a linear function) or an exponential curve by the Boussinesq equation. The difference between these functions is very small.

The measured discharges in their wavy curve were used to compute the output Fourier coefficients A_r, B_r, (Eq. (6)). Then it was easy to solve the transformation harmonic coefficients α_r, β_r (Eq. (10), Table 3), to compute the transformation function (Eq. (9)), and to solve the simulated Fourier series as the computed model response of the Event 1 (Eq (5)). Figure 2 shows the approximation of Event 1. A comparison of the measured discharges and their computed pairs is presented there.

The efficiency coefficient (Nash & Sutcliffe 1970) computed for their goodness of fit is derived as:

$$EC = 1 - (\Sigma_{i=1}^{N} (Q_i - QC_i)^2)/(\Sigma_{i=1}^{N} (Q_i - \bar{Q})^2) \qquad (11)$$

where:

Q_i – measured discharge ordinates (l/s)

QC_i – computed discharge ordinates (l/s)

$\bar{Q}$ – mean value of the measured discharges (l/s)

The Nash-Sutcliffe coefficient EC and its value for good acceptance should be EC > 0.75 (WMO 1992).

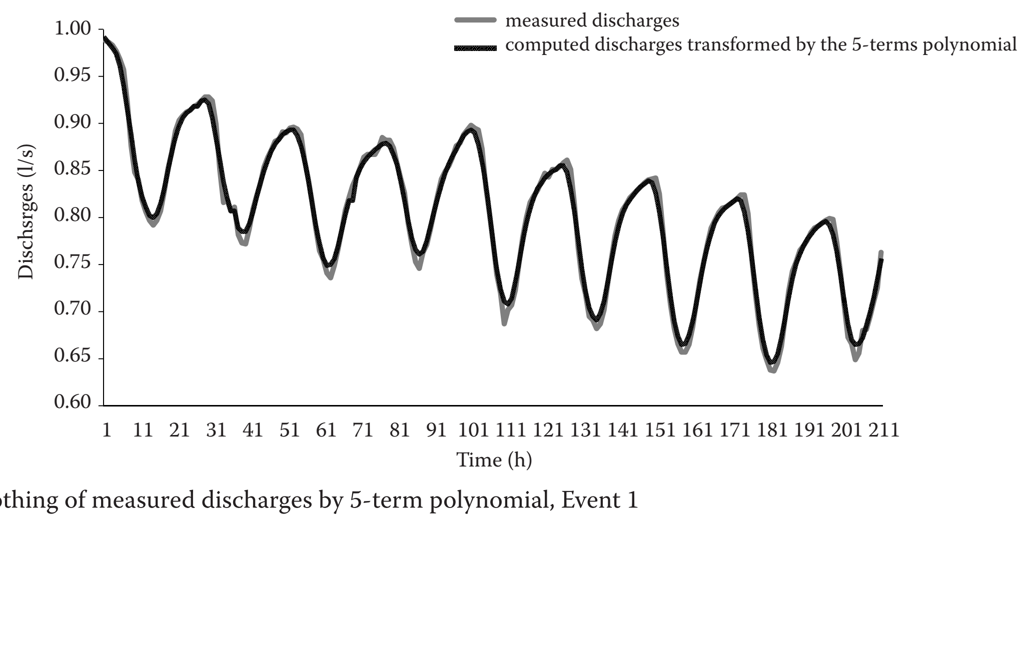

Figure 3. Smoothing of measured discharges by 5-term polynomial, Event 1

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The Event 1 has achieved EC = 0.860 when the number of harmonics RR equals to 15. If the measured discharges were smoothed by 5-term polynomial, the goodness of fit EC approaches the value 0.93. It is presented in Figure 3.

DISCUSSION

Diurnal discharge fluctuation during a hot and dry summer period is often observed in the measured discharge records. The discharges show a declining trend in their runoff depletion curve when the catchment becomes dry. The soil zone of the catchment is always partially saturated, due to the deep valley morphology of the Starosuchdolsky Brook plain.

A dynamic catchment system offers useful information on how streamflow hydrographs may be applied for reconstructing evapotranspiration records. In such a dynamic system, precipitation and evapotranspiration have comparable but opposite effects on the catchment storage, and therefore on the streamflow. As the fluctuations in the streamflow reflect the precipitation to the catchment, it is natural to conclude that these fluctuations also reflect the evapotranspiration losses. In the past decades, hydrologists studied the ways of using discharge measurements during streamflow recession to show harmonic evapotranspiration rates or a hypothetic approach (Beven 2010) and/or use of kinetic equation coefficients (Banasik *et al.* 2014; Krajevski *et al.* 2014). The smaller the catchment, the more significant the fluctuations are (Brutsaert & Nieber 1982; Boronina *et al.* 2005; Szilagyi *et al.* 2007). Similar results were found in the delay time and water-use related fluctuations by evapotranspiration for 1–3% in mid-Wales, in the Severn and the Wye (Kirchner 2009). The differences were evidently caused by different climatic and geographical conditions.

Figure 2 presents interesting data illustrating what happens when evapotranspiration influences the flow. It is practical to implement FSM based on the Fourier series where the input coefficients (a_r, b_r) and output coefficients (A_r, B_r) are used for computing the transformation function coefficients (α_r, β_r). However, the results might be loaded with some noise from the subsurface processes, which delay the surface discharge mainly due to hydraulic roughness (Dvořáková & Zeman 2010; Dvořáková *et al.* 2012, 2014; Kovar *et al.* 2014b). There is some uneven spatial and temporal distribution of hydraulic and hydrologic factors and situation influencing the impact of evapotranspiration on extra-irregularities of discharges. First of all, there is a variability of soil moisture content, groundwater storage, daily weather data, etc. Still, these changing factors distributions do not affect the physical principles of the coherence between the convolution and orthogonal (Fourier) principles on the evapotranspiration–runoff processes. One more remark for discussion: the approximation of the transformation function $u(n–t)$ (see Eqs (4) and (5)) for the computed discharges by the Fourier series model (FSM) offers higher goodness of fit than other similar mathematical models, i.e. Laguerre functions or matrix inversion model. The FSM can be improved through the choice of the period length $n = t$ (Kraijenhoff and O'Donnell 1966). Herein, the number of the Fourier's harmonics rr can be increased up to the number of the discharge

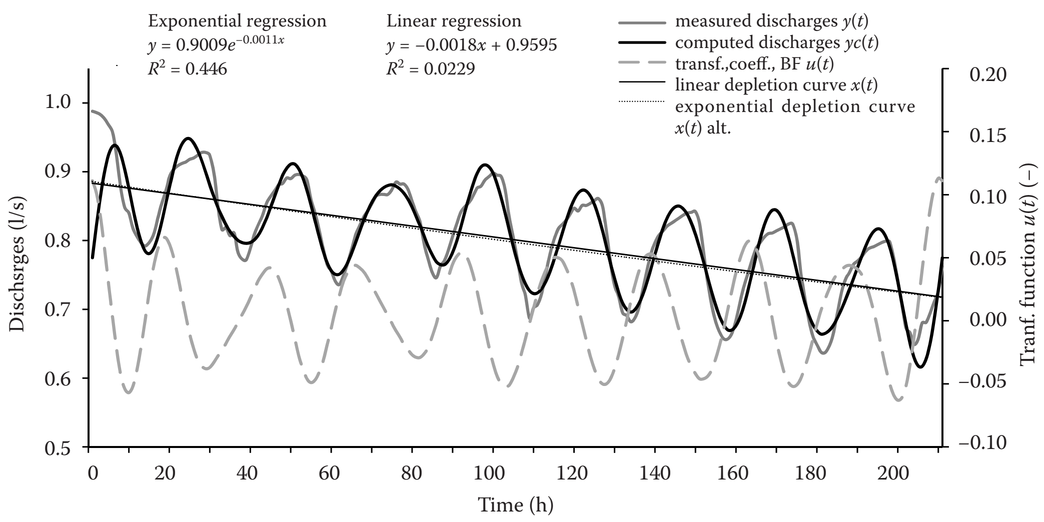

Figure 4. Approximation of the transformation function $u(t)$, Event 1 by the Fourier Series Model

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ordinates n. Figure 4 shows an optimal choice of the parameters rr and n. The method for computing discharge ordinates undoubtedly has an excellent mathematical background (O'Donnell 1960; Hardy & Rogosinski 1971). Beven (2010) inferred the stages in testing small basins hydrological models as hypotheses with limits of acceptability approach within the GLUE methodology.

Interpretation of the results should acknowledge that catchment runoff formation is influenced also by hydraulic resistance which slows down water percolation. Compared to evapotranspiration, the influence of hydraulic resistance is smaller. Our computation indicates that contribution of hydraulic resistance to reduction of catchment runoff is less than 10%. The major effect of hydraulic resistance is the delay evapotranspiration influence on catchment discharge. The Fourier model coefficients can also be used for computing missing discharge data due to a measurement failure. In this case, we can use the input coefficients and the transformation coefficients from the time series just before the discharge measurements collapse.

CONCLUSIONS

The impact of evapotranspiration on catchment runoff is an interesting but little studied hydrological phenomenon. Water use by riparian vegetation is closely linked to diurnal streamflow variability. The FSM model used in this study is based on the Fourier series, and it takes full advantage of its mathematical properties, such as harmonic functions, convolution principles, and strong convergence. All of these can be used in hydrology, not only for rainfall-runoff relations but also for evapotranspiration-runoff relations. Water exhaustion by actual evapotranspiration in dry periods is a long-term continuous process that lasts for many weeks, depending of course on the initial water supply.

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