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Estimation of specific yield in bedrock near-surface zone of hilly watersheds by examining the relationship between base runoff, storage and groundwater level

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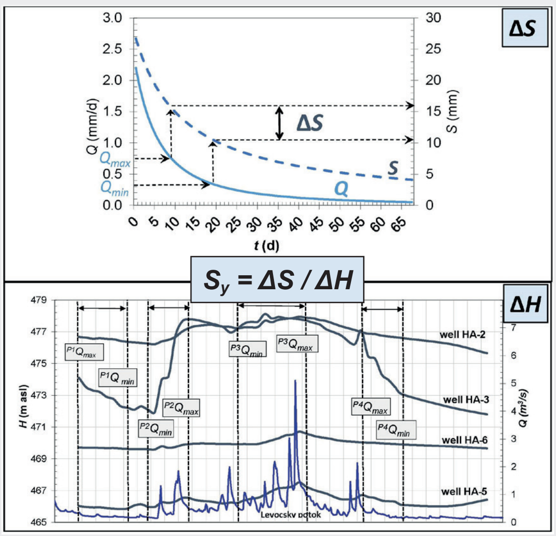
Abstract: A catchment-scale method for estimation of specific yield (S_y) in the zone of groundwater level fluctuation is proposed. It is applicable to hilly watersheds, where deep groundwater discharge – not drained by local streams as baseflow – is small and can be neglected. Therefore, it is mostly employable for bedrock flow systems, dominated by shallow unconfined fractured rock aquifers. Method provides an estimate of specific yield (S_y) by combined analysis of streamflow recession, storage/runoff relationship and groundwater level fluctuation (*Q-S-H*). For groundwater storage (*S*) values evaluation, river discharge (*Q*) master recession curves are constructed and interpreted. The method produces as more reliable results as number of groundwater level observation sites increases. As example, it is demonstrated at the Levočský potok Brook watershed (Western Carpathians, Slovakia), built by fracture porosity dominated Paleogene sediments. Estimated characteristic S_y value is from the interval 0.001–0.002 and 0.002–0.005 for low and medium storage/runoff conditions – or bottom and middle part of GWL fluctuation zone – respectively.

Key words: specific yield, groundwater storage, groundwater table fluctuation, base runoff, bedrock flow systems

Graphical abstract



Highlights

- Regional estimation of specific yield ($S_y = \Delta S\ /\ \Delta H$) based on a combination of stream flow recession – storage/runoff – groundwater level fluctuation (*Q-S-H*) analysis is suggested. It is applicable to aquifers in hilly watersheds, where deep groundwater discharge (not drained by local streams as baseflow) is small and can be neglected.
- For watershed groundwater storage (S) evaluation, river discharge (*Q*) master recession curves are constructed and storage-discharge relationship $S = aQ^b$ for nonlinear reservoirs is accepted. Groundwater level (*H*) measurements are required to determine a representative ΔH value.
- Demonstration of the method at the Levočský potok Brook watershed (Western Carpathians, Slovakia), which is built by fracture porosity dominated Paleogene sediments, suggest that the characteristic S_y value at studied watershed is from the interval 0.001–0.002 and 0.002–0.005 for low and medium storage/runoff conditions (or bottom and middle part of GWL fluctuation zone), respectively.

1. Introduction

The specific yield (S_y) of a rock or soil, with respect to water is defined as ratio of the volume of water which, after being saturated, it will yield by gravity, to its own volume (Meinzer, 1923).

$$S_y = V_W / V_T \tag{1}$$

where V_W is the volume of drainable water and V_T is the total volume of porous rock or sediment.

Specific yield is part of the total porosity of a porous rock or sediment. Total porosity includes the fraction of pore space that is interconnected (called "effective porosity") and porosity of isolated pores. The effective porosity consists of specific retention and specific yield. Specific retention is the ratio of the volume of water that a given body of rock or soil will hold against the pull of gravity to the volume of body itself. Specific yield is the amount of water that is actually available for groundwater pumping, when sediments or rocks are drained due to the lowering of the water table. The specific yield is used to determine how much water can be produced from an unconfined aquifer per a unit decline in the water table (Harter, 2019). Therefore, it is very important parameter, needed for evaluation and management of groundwater resources.

Methods commonly used to determine S_y are field methods (interpretation of aquifer hydrodynamic tests, water-budget methods, geophysical methods, methods based on recession curve analysis) and laboratory methods. A large variability exists in both laboratory and field- determined S_y values (Varni et al., 2013). Interpretation of aquifer tests in fractured-rock systems is often difficult because of ambiguity issues. Specific yield values determined from these tests are usually unreliable (Bardenhagen, 2000). Methods based on recession curve analysis derive hydraulic (including S_y) and geometric characteristic of aquifer from the recession curve shape. Their mathematical formulations represent various conceptual models developed from differential Boussinesq (1877) equation and they differ each other by ways of input data analysis (Brutsaert & Nieber, 1977; Brutsaert & Lopez, 1988; Parlange et al., 2001; Mendoza et al., 2003). Their applicability for S_y calculation in bedrock groundwater flow system is complicated by the need to know the aquifer thickness – however this aquifer datum is practically indeterminable in weathered fractured bedrock due to its vertical inhomogeneity of permeability. The water budget method is the most widely used technique for estimating specific yield in fractured-rock systems, probably because it does not require any assumptions of concerning flow processes (Healy & Cook, 2002). Several approaches, expressed by different forms of a simple water budget equation for basin

$$P + Q_{on} = ET + \Delta S + Q_{off} \tag{2}$$

and different ways of their members` determination (P – precipitation plus irrigation, Q_{on} and Q_{off} – surface and subsurface water flow into and out off the basin, ET – sum of bare soil and open water evaporation and plant transpiration, ΔS – change in water storage), were used (Walton, 1970; Gerhart, 1986; Hall & Risser, 1993; Rasmussen & Andreasen, 1959; Gburek & Folmar, 1999). However, given the current stage of the science, it is extremely difficult to assess the accuracy of any method. For this reason, it is highly beneficial to apply multiple methods of estimation and hope for some consistency in results.

This paper introduces a new method of regional estimation of specific yield based on a combination of stream flow recession – storage/runoff – groundwater level fluctuation (*Q-S-H*) analysis, which can be assigned to the group of field methods. It is applicable to aquifers in hilly watersheds, where deep groundwater discharge (not drained by local streams as baseflow) is small and can be neglected. Therefore, it is mostly employable for bedrock flow systems, dominated by shallow groundwater circulation. However, these aquifers often occupy the vast majority of mountain regions, which play a strategic role for water resources management at the regional and global scales (Aureli, 2002; Viviroli & Weingartner, 2004). Their study is difficult due to the complexities of the geology, the geomorphology and the climate patterns (Espinha Margues et al., 2013). Therefore, it entails challenges, concerning both input data collection and interpretation methods. The specific yield is a key parameter not only for groundwater resources evaluation, but also for estimation of recharge, using world-wide used water-table fluctuation (WTF) method (Schicht & Walton, 1961).

To demonstrate applicability of proposed method, here it is used to estimate average/representative value of S_y in the Levočský potok Brook watershed (the Hornádska kotlina basin / the Levočské vrchy Mts., Slovakia), which is built by fracture porosity dominated Paleogene sediments. Obtained results are compared to published values of S_y representing bedrock flow systems in hilly watersheds.

2. Method

Proposed method provides an estimate of specific yield (S_y) by combined analysis of stream flow recession – storage/runoff – groundwater level fluctuation (*Q-S-H*) at hilly watersheds. The method is based on the assumption that a rise in water-table elevation measured in shallow boreholes is caused by the addition of recharge across the water table at watershed and its following recession is caused by groundwater storage (S) loss due to baseflow (Q) generation. Supposing that storage loss (depletion) reflects baseflow recession, the recession curve extracted from the continuous multi-year hydrographs can be used to derive the storage loss value ΔS corresponding to average groundwater level decline ΔH in studied watershed. Thus, average Sy value in groundwater fluctuation zone is given as $S_y = \Delta S / \Delta H$.

Determination of the storage loss value ΔS is based on the construction of the master recession curve (MRC) for studied watershed, followed by specifying of recession coefficient a and recession exponent b from equation (1), by means of MRC interpretation.

MRC construction approach tries to find a solution to usual problem with application of individual recession curves, derived from selected time periods of recession – that in most cases they describe the process only partially, depending on the limiting water stages of these periods. To cover all possible solutions, different methods (Lamb & Beven, 1997; Rutledge, 1998; Posavec et al., 2006; Gregor & Malík, 2012b) of composing individual curves into a single master recession curve (with the longest course and covering all documented water stages) were created. In this study, the approach developed by Gregor and Malík (2012a) supplemented by computational tool (the RC 4.0 module in the freely accessible HydroOffice software: http://www.hydrooffice.org) is used. It is based on genetic algorithm (main principles applied within the genetic algorithm procedures is explained by Hynek, 2008), which allows creation of the most probable natural, unaffected recessional discharge sequences in time, from which the master recession curves can be constructed. Such assembling of recessional discharge time series can help to avoid obstacles such as limited time-series datasets, incomplete recessions, too many segments in many recessional successions, complicated hydrograph shape, different time intervals of observations, short time-series intervals, imprecise measurements, different types of datasets (averaged or instantaneous data) or even rough (inaccurate) measurements of discharges.

Construction of MRC is based on extraction of recession periods from hydrograph. As a way to focus on the true natural storage-discharge relationship, the influence of the unknown factors as evapotranspiration (ET), snow melt and low permeability due to frozen soil, can be minimized by extraction only recessions occurring during appropriate seasons (for example, autumn months in temperate climatic range of northern hemisphere). To avoid the influence of overland flow, the beginning of the baseflow recession must be assumed not earlier than certain time interval, depending on watershed size and morphology. For example, Wittenberg (1999) starts baseflow recession two days after the inflection point of the MRC and Ye et al. (2014) record only 70 % of a falling limb as a recession period.

Constructed MRC is believed to represent the true natural storage-discharge relationship, which is unique for each watershed. It can be simulated using various model equations (recession functions) derived by many authors, which are incorporated in the RC 4.0 tool. The storage loss between two time points on MRC can be computed as sum of amounts discharged in chosen time steps using any of them. But only two have advantage of mathematically defined *Q–S* relationships. The exponential function (Maillet, 1905)

$$Q_t = Q_0 e^{-t/k} \tag{3}$$

is used to describe the recession of baseflow, where Q is discharge at time t, Q_0 the initial discharge and k the retention constant that supposedly represents storage lag-time (Wittenberg, 1999). This concept of single linear reservoir uses constant reaction factor, so storage is proportional to baseflow: $S = kQ$. Nonlinear reservoirs have reaction factors that increase with increasing storage, thus the storage-discharge relationship was modified by adding an exponent b (Wittenberg, 1999)

$$S = aQ^b \tag{4}$$

to define recession curve equation

$$Q_t = Q_0 \left[1 + \frac{(1-b)\,Q_0^{1-b}}{ab}\,t\right]^{1/(b-1)} \tag{5}$$

In this study, concept of nonlinear reservoir and the notional value of $b = 0.5$ is accepted. The notional value of b = 0.5 is suggested by Wittenberg (1999), based on recession curves from more than 80 gauging stations in Germany. It is also confirmed by other authors who, by adopting more theoretical approaches, found storage-outflow relationship corresponding to $S = aQ^{0.5}$ or $Q = S^2$ for discharge from springs (Drogue, 1972) and unconfined aquifers (Werner & Sundquist, 1951; Schoeller, 1962; Roche, 1963; Fukushima, 1988). Corresponding value of a is determined by fitting procedure, using the RC 4.0 module in HydroOffice software, customized for this purpose. Within this procedure, the a value together with initial discharge Q_0 value has been manually alternating until model curve – generated automatically according to equation (4) on the graph window – visually fitted to the section of MRC no influenced by overland flow (Fig. 4). The a value determined in this way makes it possible to calculate the actual storage for arbitrarily chosen datum of studied period, using equation (1). However, baseflow must be used as Q value in this calculation. This requirement may be met by selection of days when only baseflow occurs in the stream, or by correction of correspondent recorded Q values, influenced by surface runoff, onto baseflow values. An appropriate hydrogram separation method may be used for this correction. Among these, the envelope line method (*ELM*) for groundwater table-discharge (*H-Q*) relationships, proposed by Kliner and Kněžek (1974) for runoff separation from hydrograph, seems to be the most suitable for this study. The method is based on assumption that close relationship between groundwater and stream water level should exist, considering hydraulic connections between rivers and aquifers. The upper limit of the points in the *H-Q* graph usually makes it possible to draw an envelope line representing the flux formed by groundwater runoff (Fig. 7). This line can be used to calculate groundwater runoff for any measured groundwater table, in different types of natural conditions (Holko et al., 2002).

Finally, calculation of *Sy* is possible after selecting two $S – H$ value pairs, using equation (6). One represents a high (S_{max}, H_{max}) and the other a low (S_{min}, H_{min}) stage of storage / baseflow. The H_{max} and H_{min} values are calculated as the average of the time relevant data from available observation boreholes.

$$S_y = (S_{max} - S_{min}) / (H_{max} - H_{min}) \tag{6}$$

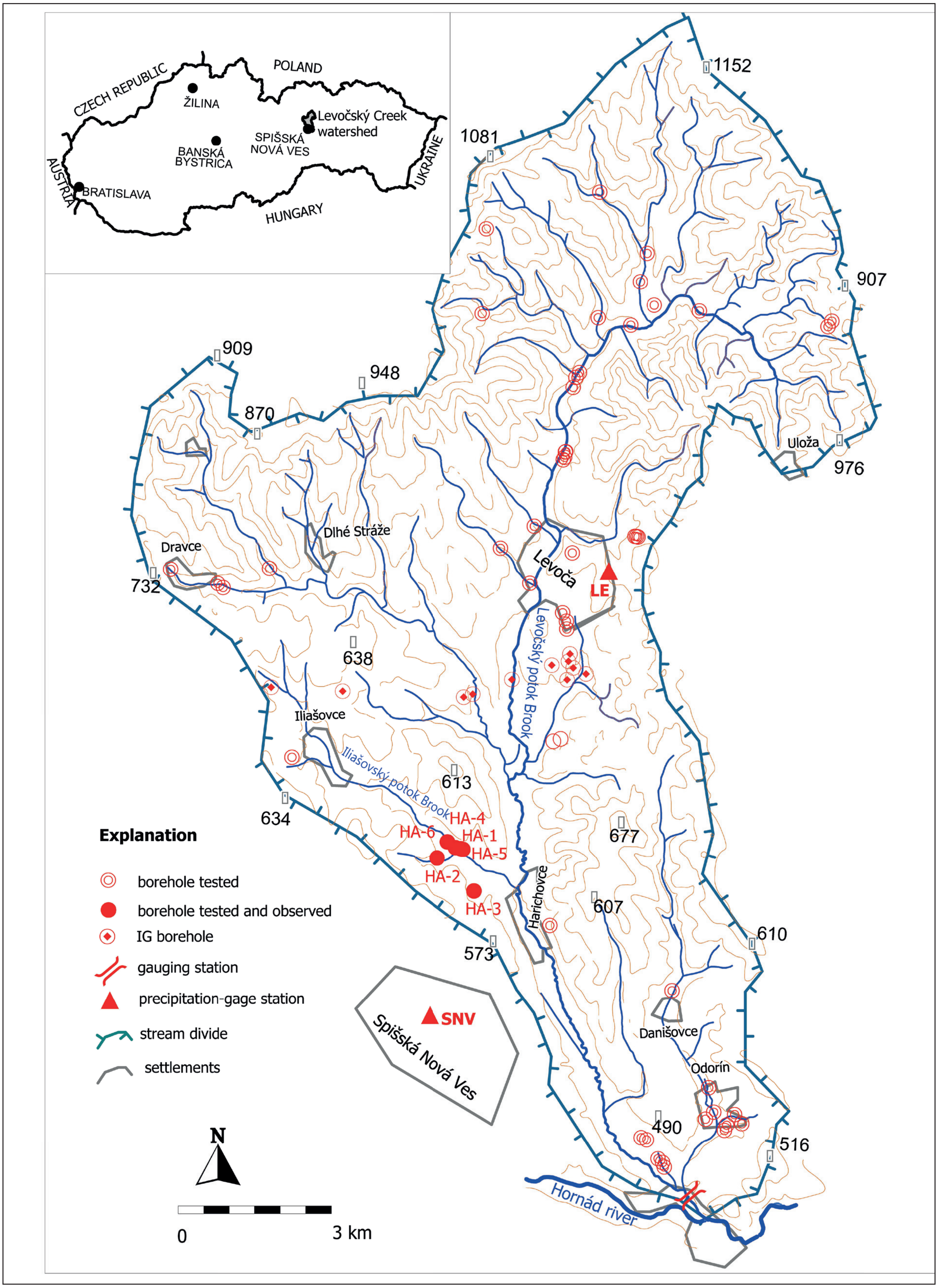


Fig. 1. Situation of the Levočský potok Brook watershed.

3. Example – S_Y estimation for the Levočský potok Brook watershed

3.1. Study area

The study site shown in Fig. 1 is a 154.81 km² watershed located in central part of Slovakia. Hilly landscape lies in altitude of 422–1 215 m (656 m in average). Slopes are of variable length (0.5–1 km) and generally moderate (10.1 % in average) with local relief of 100–250 m. Density of channel network in watershed reaches 1.38 km/km². It is dewatered by 25.9 km long Levočský potok Brook into the Hornád river.

Hillslope soils are characterized as cambisols – mostly saturated (eutric, stagnieutric and calcaric) cambisols prevails in southern part of studied watershed, whereas oligobasic (dystric cambisoils, cambic umbrisoils and stagni-dystric cambisoils) occur in its northern part (Šály & Šurina, 2002). Average hydraulic conductivity of soil at studied watershed is $5.44.10^{-6}$ m.s^{-1} (Malík et al., 2007). It is covered by coniferous forest (39.9 %), prevailingly in higher altitudes. Lover parts of land are cultivated (27.1 %) or covered by meadows.

Bedrock is represented by flysch rocks with a predominance of layers fractured sandstones over silts and claystones of Paleogene age (Mello et al., 2000; Map server of SGIDŠ, 2016), belonging to geological unit of the Central-Carpathian Paleogene: Biely Potok Formation (55 % of the catchment area) and Zuberec formation (45 %). The regional hydrogeological research (Jetel, 2000) revealed that the permeability of the Central-Carpathian Paleogene flysch rocks is distinctly controlled by actual depth position below ground surface. Regular decrease of mean permeability in particular formations with depth can be described by exponential functions of the depth. The mean permeability in depths of 0–100 m decreases on average to 26–59 % of the initial value per every 10 m of depth increase. Primary differences in permeability between sandstones and argillaceous rocks fade away as a result of diagenetic changes, reducing intergranular permeability. Fissure permeability is of decisive importance. The maximum permeabilities and transmissivities are found in tectonically predisposed joint zones without any unequivocal relation with lithology. Consequently, hydrogeological function of stratiform aquifers and intergranular permeability in the flysch complex is of rather little importance. The main aquifer here is represented by the near-surface zone of increased permeability in first tens of meters below ground surface. Deeper circulation of groundwater occurs predominantly in subvertical joint zones.

After data selected from database containing pumping test reinterpretation results (Malík et al., 2016) of 48 boreholes (Fig. 1), hydraulic conductivity of flysch rock ranges between $2.59.10^{-8}$ and $3.57.10^{-4}$ m/s (geometric mean $4.87.10^{-6}$ m.s^{-1}) at the Levočský potok Brook watershed. Borehole depth is 5.7–150.2 m (52.78 m in average). Groundwater level (GWL) was recorded in the depth of 25.4–0.1 m below ground surface (3.74 m in average) and its areal distribution does not significantly depend on geomorphology. This fact can be demonstrated on graphs when recorded GWL depth H is plotted against vertical elevation of borehole head above regional drainage base (RDB), (Fig. 2a), or against local drainage base (LDB) (Fig. 2b). The H values usually don't exceed 10 m not only in valleys bottom, but also on slopes, what supports above mentioned opinion that near-surface zone represents

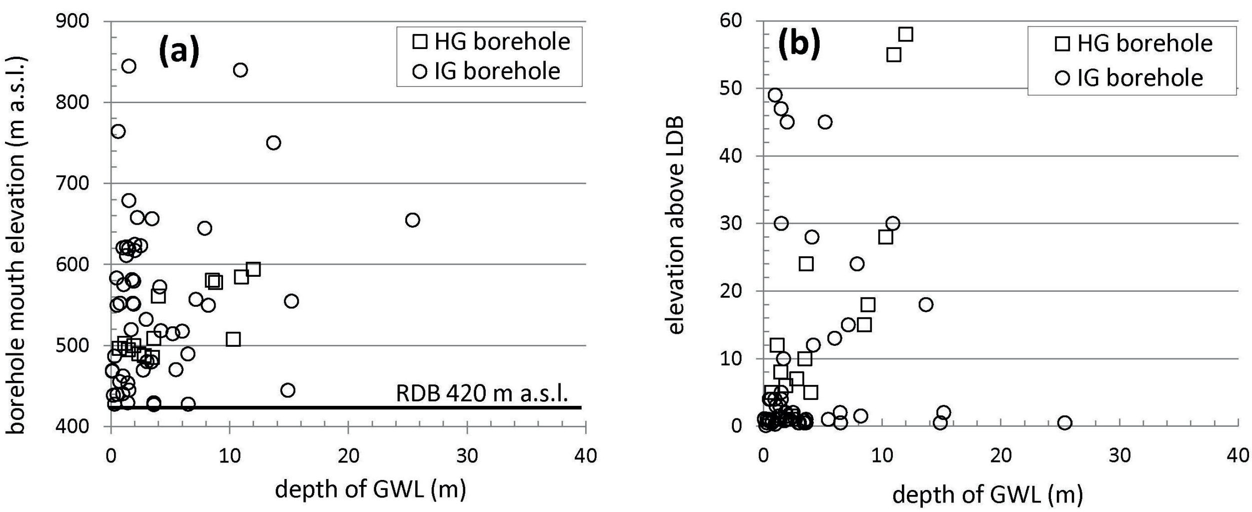


Fig. 2. Depth of groundwater level (GWL) in studied watershed recorded in boreholes versus: **a** – height of borehole mouths above sea level with the regional drainage level (RDB) marked; **b** – elevation of borehole heads above local drainage level (LDB); hydrogeological (HG) boreholes from database of Malík et al. (2016) and engineering geological (IG) boreholes from database of SGIDS are distinguished.

the main aquifer in flysch rocks of Central-Carpathian Paleogene. Therefore, this environment is dominated by bedrock groundwater flow system, in which baseflow forms substantial part of total groundwater discharge from watershed (Welch & Allen, 2014). Thus, cross-boundary groundwater flow is considered to be negligible.

Within the area of the Levočský potok Brook watershed, GWL fluctuation was observed on six boreholes, in frame of local hydrogeological research (Bajtoš & Michalko, 2003). HA-1, HA-4 and HA-6 boreholes are located 50–70 m from the Iliašovský potok Brook (tributary of the Levočský potok Brook). All three catch permeable fault zone, which is naturally dewatered by the Zimná Studňa Fissure Spring. Nearby situated HA-5 borehole is tectonically separated from this fault zone and it captures aquifer bound to near-surface zone. HA-2 and HA-3 boreholes are located in greater distance from the Iliašovský potok Brook (Fig. 1), their heads are 9.5 m and 16.7 m above local drainage base, respectively. Both boreholes are situated in a near-surface zone aquifer. GWL fluctuation on boreholes was measured during hydrological year 1995 (November 1994 – October 1995) on weekly frequency. Minimum GWL did not exceed depth of 7.1 m below ground surface in any borehole and range of GWL fluctuation ΔH (difference between minimal and maximal recorded GWL, $\Delta H = H_{min} - H_{max}$) reached values 0.491 – 6.170 m in individual boreholes (Tab. 1).

Rainfall precipitation events in studied area are distributed between March and October, whereas in period from November to February snow precipitations prevail. Highest monthly precipitation totals occur in June to August period, most dry conditions terms since January to March (Tab. 2). Annual precipitation total varies around 630 mm.

Discharge of the Levočský potok Brook is observed by SHMI on gauging station no. 8 424, situated close to

Tab. 1

Groundwater level fluctuation recorded at the Harichovce site during hydrological year 1995 (Bajtoš & Michalko, 2003)

Borehole	HA-1	HA-2	HA-3	HA-4	HA-5	HA-6
H_{avg}	–0.062	2.326	3.502	0.500	2.332	3.002
H_{max}	0.043	3.133	7.100	0.808	2.906	3.357
H_{min}	–0.448	1.408	0.930	-0.227	1.227	2.242
ΔH	0.491	1.725	6.170	1.035	1.679	1.115
Borehole depth	100	100	100	60	60	80
Borehole screen interval	7.5 – 59.0	6.6 – 54.3	9.4 – 45.0	10.0 – 55.2	10.0 – 55.6	10.0 – 54.0

Explanation: Havg, Hmin, Hmax – average, minimum and maximum GWL depth in meters below ground surface; ΔH = Hmax – Hmin

Tab. 2

Long term monthly and annual averages of precipitation total in mm recorded in period 1951–1980 by Slovak Hydrometeorological Institute on stations in Levoča (LE) and Spišská Nová Ves (SNV)

Station	I.	II.	III.	IV.	V.	VI.	VII.	VIII.	IX.	X.	XI.	XII.	Year
LE	26	25	26	43	70	97	90	82	50	42	41	32	624
SNV	23	25	27	46	73	99	90	79	48	46	46	31	633

Tab. 3

Long term monthly and annual averages (A), standard deviations (STD), minimum (MIN) and maximum values (MAX) of the Levočský potok Brook discharge in m^3/s recorded in period 1990–2012 by Slovak Hydrometeorological Institute on gauging station No. 8 424 in Markušovce

	I.	II.	III.	IV.	V.	VI.	VII.	VIII.	IX.	X.	XI.	XII.	Year
A	0.421	0.405	1.018	1.239	0.939	0.925	0.968	0.779	0.647	0.514	0.487	0.457	0.735
STD	0.424	0.313	1.259	1.098	0.738	1.229	1.328	0.753	0.840	0.360	0.436	0.458	0.893
MIN	0.131	0.101	0.072	0.152	0.150	0.168	0.098	0.105	0.119	0.135	0.144	0.075	0.075
MAX	5.723	2.383	13.313	8.212	9.272	19.488	15.823	7.011	11.900	3.486	4.448	4.501	19.488

its effluent into the Hornád river (Fig. 1). Highest average discharges connected with spring snow melting occur during March and April (Tab. 3). During winter season with little or no recharge, discharge is lowest. More than 80 % of observed time discharge not exceeded 1 m^3/s (Fig. 3), median value is 0.480 m^3/s.

3.2. Results

For S_y estimation at the Levočský potok Brook watershed, the record of stream discharge for period 1990–2012 and GWL data from 6 boreholes for hydrological year 1995 are disposable (Tab. 1). Based on recorded discharge Q, master recession curve (MRC) was constructed (Fig. 4). The use of MRC allows to simulate watershed groundwater storage (S) at different baseflow (Q) and corresponding GWL stages (H). Two different ways of S_y estimation are presented: 1) by comparing two hydrological stages selected on hydrograph (Fig. 6) and 2) by comparing two hydrological stages selected on Q-H graphs, using envelope line method (ELM; Kliner & Kněžek, 1974) (Fig. 7). Based on obtained S and H interdependent pairs of values, S_y is calculated in accordance with equation 6. Boreholes H-1 and H-4 was excluded from average H calculations because of their close distance to and their high hydraulic interconnection with H-6 borehole. Moreover, theirs GWL's are affected by the drainage effect of the Zimná studňa spring.

Master recession curve (MRC) for the Levočský potok Brook

MRC constructed for the Levočský potok Brook watershed through selection of 30 individual recessions from recharge record 1990–2012 and result of their automatic processing in genetic algorithm based procedure using RC 4.0 module in HydroOffice software is shown on Fig. 4. It is supposed that baseflow is exclusively present in discharge recession one day after inflection point or six days after its beginning (Fig. 4). This part of MRC was used for calibration of nonlinear model curve, in which the recession coefficient a = 15.5

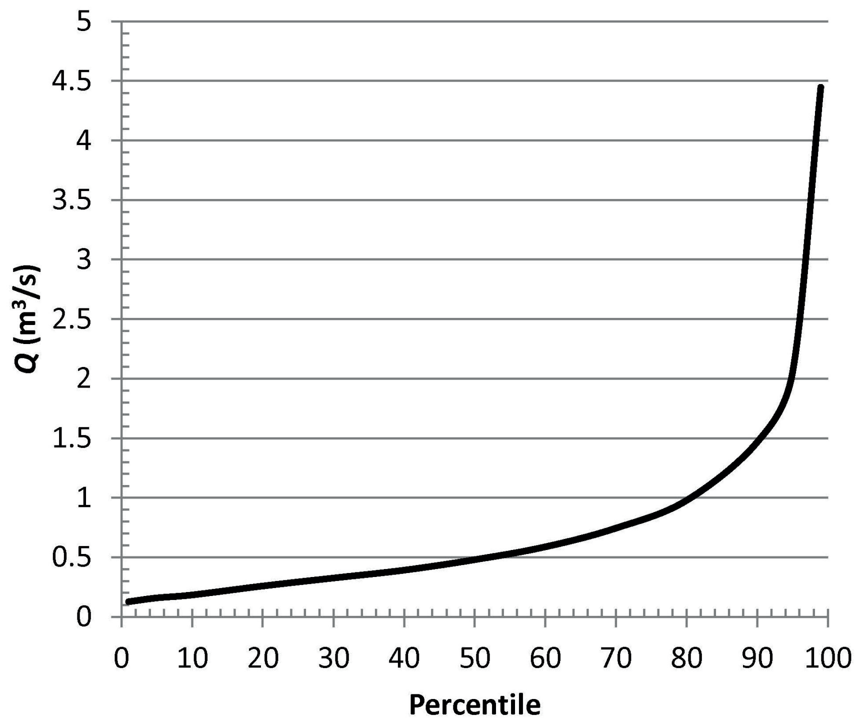


Fig. 3. Percentiles of the Levočský potok Brook discharge recorded in period 1990–2012 by Slovak Hydrometeorological Institute on gauging station No. 8 424 in Markušovce.

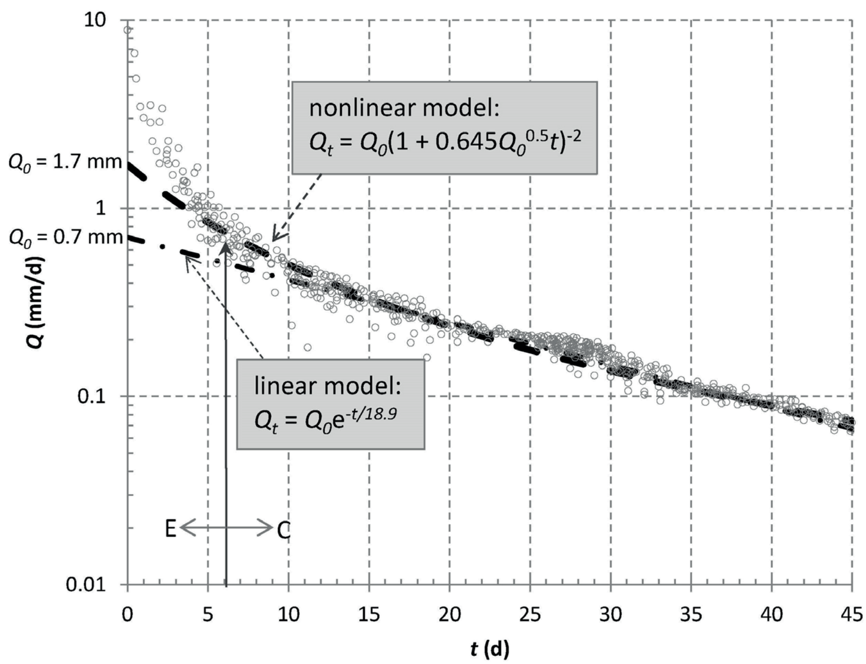


Fig. 4. Values of master recession curve (*circles*) generated using the RC 4.0 module in HydroOffice software by procedure based on genetic algorithm (Gregor & Malík, 2012a), with nonlinear model (dashed line) and linear model (dash-and-dot line) marked. Calibrated (C) and extrapolated (E) sections of nonlinear model line are distinguished.

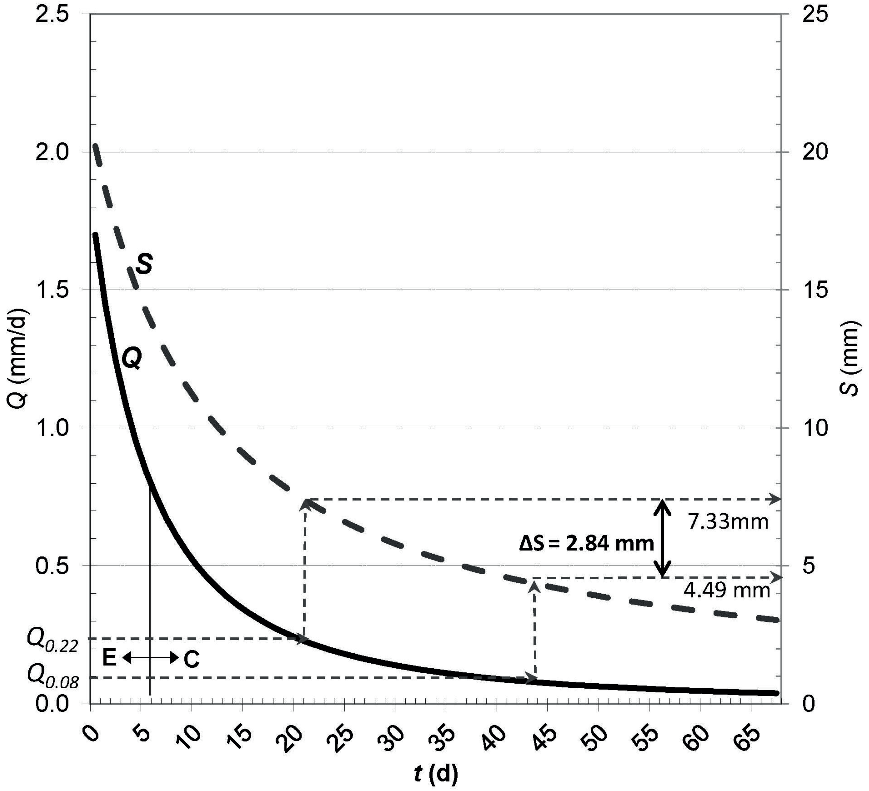


Fig. 5. Master recession curve (Q) and corresponding nonlinear changes of storage (S) for the Levočský potok Brook watershed. Calibrated (C) and extrapolated (E) sections of model MRC are distinguished. Further explanation in the text.

and initial discharge value Q_0 = 1.7 mm/d were found as fitted parameters. Therefore, storage-outflow relationship in studied watershed is described as $S = 15.5Q^{0,5}$. Both curves – model MRC with Q-values in mm/d and curve of corresponding S-values in mm – are depicted in Fig. 5, to illustrate their range and also to demonstrate the determination way of groundwater storage change for certain change of discharge. Expected maximum S value corresponding to Q_0 is 3.12 million m^3 (20.21 mm). Minimum S value corresponding to Q_{66} (Q in 66th day after Q_0 on MRC), and also equal to minimum recorded Q value, is 0.48 million m^3 (3.08 mm). Therefore, maximum amount of water being able to release from underground storage in given natural conditions is ΔS = 2.65 million m^3 (17.12 mm).

Estimation of S_y by comparing two hydrological stages selected on hydrograph

In the frame of this approach, the following procedure was applied: (a) One or more appropriate hydrologic periods are selected on the hydrograph. (b) For each period, two hydrological stages are selected, represented by river discharge values ${}^{Px}Q_{min}$ and ${}^{Px}Q_{max}$ together with their time relevant GWL values ${}^{Px}H_{min}$ and ${}^{Px}H_{max}$ for all disposable boreholes, whereby river discharge values are treated to represent baseflow (hydrological stages without surface flow were selected). (c) Using selected ${}^{Px}Q_{min}$ and ${}^{Px}Q_{max}$ values, ${}^{Px}S_{max}$ and ${}^{Px}S_{min}$ values are calculated (after storage-outflow equation $S = 15.5Q^{0,5}$ in this case, Fig. 5). (c) ${}^{Px}H_{max}$ and ${}^{Px}H_{min}$ are determined as average representatives of GWL in watershed for respective time. (d) S_y value is calculated following equation (5).

Five hydrological periods (P1–P5) are selected. Days bounding these periods are marked on hydrograph (Fig. 6). P1 period represents GWL decline with no or little recharge (winter time) and with low storage stage changed from 5.64 to 4.97 mm (Tab. 4). Period P2 is characterized by GWL rise and storage increase (from 5.01 to 7.96 mm) due to spring snow melting and soil thawing, combined with rain. During summer P3 period, GWL was rising due to repeated rains and storage increased from 7.07 to 8.71 mm. The decrease following this relatively high storage stage from 7.92 to 5.99 mm defines the P4 period. The autumn P5 period is characterized by low storage stage depletion from 1.37 to 0.69 m^3. Computed S_y values for selected hydrological periods vary from 0.0015 to 0.0045. Highest value belongs to high storage period P4, lowest one to winter dry period P1. Almost identical to the value for period P1 is the S_Y value of 0.0016 determined for spring period P2. For periods P3 and P5, mutually similar values of 0.0033 and 0.0027 was determined, respectively.

S_Y estimation based on comparison of two hydrological stages selected on Q-H graphs.

With this approach, Q-H graph is constructed for each disposable borehole (Fig. 7a–d) to obtain their characteristic H_{max} and H_{min} values corresponding to chosen low and high baseflow Q_{min} = 0.15 m^3/s and Q_{max} = 0.4 m^3/s (Tab. 5). In time of such hydrological stages, 0.69 or 1.13 million m^3 of groundwater is stored in watershed, respectively. Calculated value of S_Y = 0.0019 is very similar to those, obtained using previous approach (Tab. 4). Closest Q-H dependence was found for HA-3 borehole (Fig. 7b), suggesting that aquifer type observed by this borehole should represent the most important source of baseflow in studied watershed. Calculation of S_y using only GWT fluctuation in HA-3 borehole gives value of 0.0005.

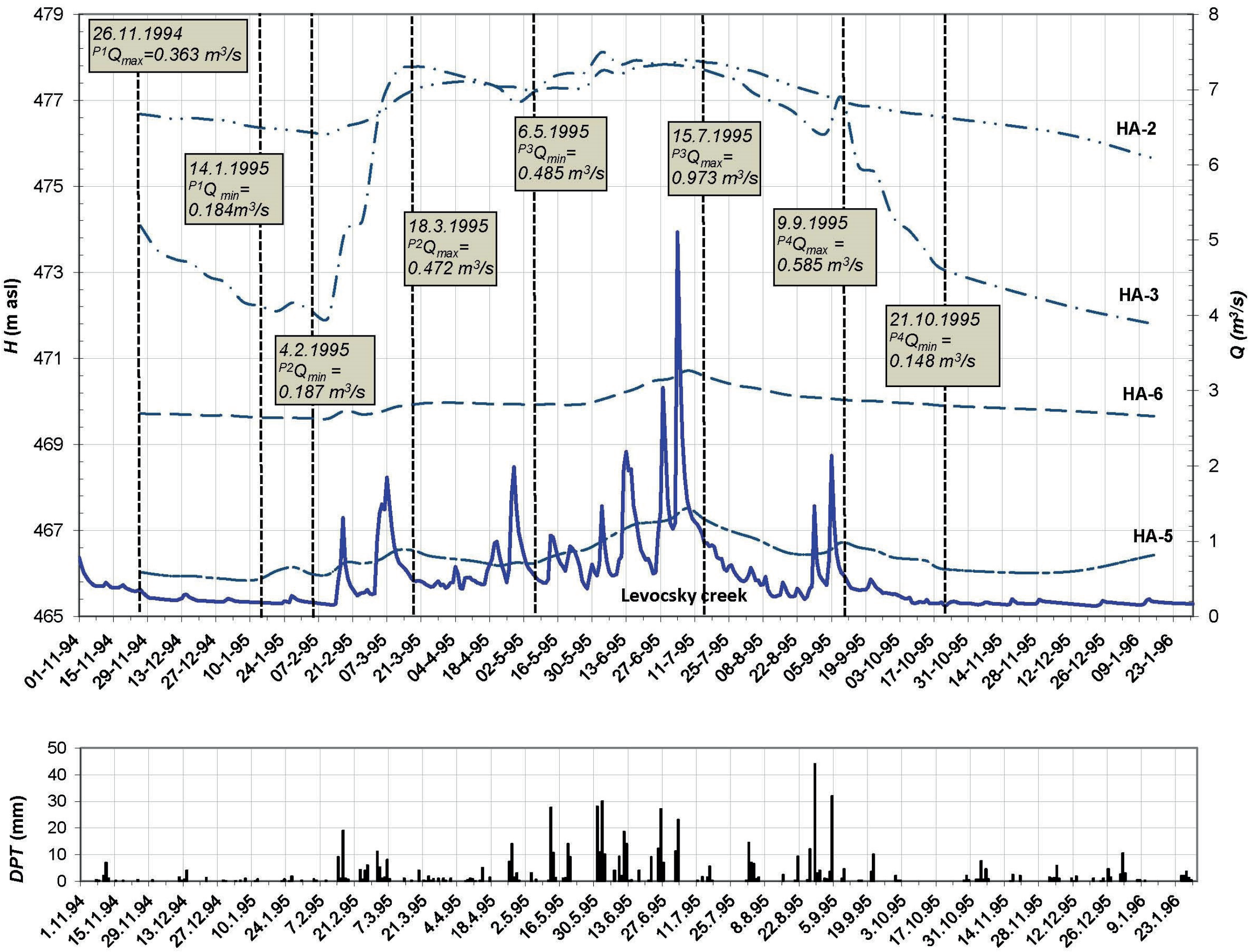


Fig. 6. Hydrograph of the Levočský potok Brook discharge, groundwater level in boreholes HA-2, HA-3, HA-5, HA-6 with selected hydrological stages chosen for specific yield calculation and daily precipitation totals measured at rainfall station Spišská Nová Ves.

Tab. 4

Altitude of groundwater level in boreholes and discharge of the Levočský potok Brook recorded with groundwater storage in watershed S_{LP} calculated and specific yield (Sy) values for selected time periods P1–P4

Period	GWL [m a. s. l.]					Discharge [m³/s]	Storage [m³]	Storage [mm]
		HA-2	HA-3	HA-5	HA-6	Q	S	S
P1	3. 12. 1994	476.630	473.520	465.972	469.706	0.237	872 384	5.64
	14. 1. 1995	476.365	472.220	465.872	469.626	0.184	768 675	4.97
	ΔH:	0.265	1.300	0.100	0.080	$\Delta^{P1}S =$	103 710	0.67
	ΔH_{avg} = 0.436 m					$^{P1}S_y$ =	**0.001 5**	
P2	4. 2. 1995	476.252	472.100	465.989	469.606	0.187	774 916	5.01
	18. 3. 1995	477.240	477.780	466.532	469.928	0.472	1 231 132	7.96
	ΔH:	0.988	5.680	0.543	0.322	$\Delta^{P2}S =$	456 216	2.95
	ΔH_{avg} 1.883 m					$^{P2}S_y$ =	**0.001 6**	

Tab. 4 – continuation

Period	GWL [m a. s. l.]					Discharge [m^3/s]	Storage [m^3]	Storage [mm]
P3	27. 5. 1995	477.608	477-643	466.248	469.942	0,372	1 092 962	7.07
	24. 6. 1995	477.855	477.705	467.188	470.474	0,565	1 346 970	8.71
	ΔH:	0,502	0,212	0,640	0,532	$\Delta^{P3}S$ =	254 008	1.64
		ΔH_{avg} = 0.472 m				$^{P3}S_y$ =	0.003 5	
P4	29. 7. 1995	477.737	477.448	466.926	470.362	0.267	925 954	5.99
	19. 8. 1964	477.360	476.787	466.49	470.141	0.467	1 224 594	7.92
	ΔH:	0.377	0.661	0.457	0.221	$\Delta^{P3}S$ =	298 640	1.93
		ΔH_{avg} = 0.429 m				$^{P3}S_y$ =	**0.004 5**	
P5	9. 9. 1995	476.988	477.077	466.718	470.041	0.585	1 370 603	8.86
	9. 12. 1995	476.222	472.245	466.030	469.783	0.176	689 389	4.46
	ΔH:	0.766	4.832	0.688	0.258	$\Delta^{P4}S$ =	681 214	4.40
		ΔH_{avg} = 1.636 m				$^{P4}S_y$ =	**0.002 7**	

Previous interpretation of *Q-H* graphs could be supported by documented relationship between local spring yield and GWL (Fig. 7e–h). The significant linear correlation is recorded for the Zimná studňa spring (ZSS) discharge and GWL in HA-6 borehole (R^2 = 0.955, R = 0.977, Fig. 7h), situated in the distance of 300 m from this spring. On the other hand, GWL in more closely located HA-5 borehole depend less significantly on spring discharge (R^2 = 0.863, R = 0.929, Fig. 7g), by reason that it do not intercepts the fault aquifer dewatered by ZSS. Dependency between ZSS discharge and GWL fluctuation in more remote boreholes HA-2 and HA-3 (Fig. 7e–f) is even more complex, comparing to HA-5 borehole. Slope of enveloping line changes from steep to sub-horizontal close to ground surface, suggesting the presence of upper vertical limit of GWL rise in given local conditions. This means that only steep section of enveloping line can be used for detection of baseflow in case of HA-2 and HA-3 boreholes (Fig. 7a–b). On the other hand, steeper sections of enveloping line in case of HA-5 and HA-6 boreholes can be regarded to shallow groundwater plus soil water (Kliner & Kněžek, 1974) or groundwater flow to shallow drains or ditches (Querner, 1997).

4. Discussion

Presented *Q-S-H* method of S_y estimation is based on comparison of groundwater storage change ΔS to respective difference of GWL ΔH (Eq. 5). Whereas values of ΔS are calculated by means of MRC constructed as unique one for studied watershed and therefore they characterize all watershed area, ΔH is averaged from as many site values as possible. In reality, the disposable number of observed

Tab. 5

Representative groundwater levels in boreholes determined for the Levočský potok Brook discharge Q_{LP} of 0.4 or 0.15 m^3/s with calculated groundwater storage in watershed *S* and expected specific yield (S_y) value

GWL [m a. s. l.]					Discharge [m^3/s]	Storage [m^3]	Storage [mm]
Borehole	**HA-2**	**HA-3**	**HA-5**	**HA-6**	***Q***	***S***	***S***
$H_{0.40}$	477.35	477.28	466.74	470.19	0.400	1 133 349	7.33
$H_{0.15}$	476.30	472.08	466.16	469.72	0.150	694 032	4.49
ΔH:	1.05	5.20	0.58	0.47		ΔS = 439 317 m3	ΔS = 2.84 mm
ΔH_{avg} = 1.46 m					S_y = **0.001 9**		

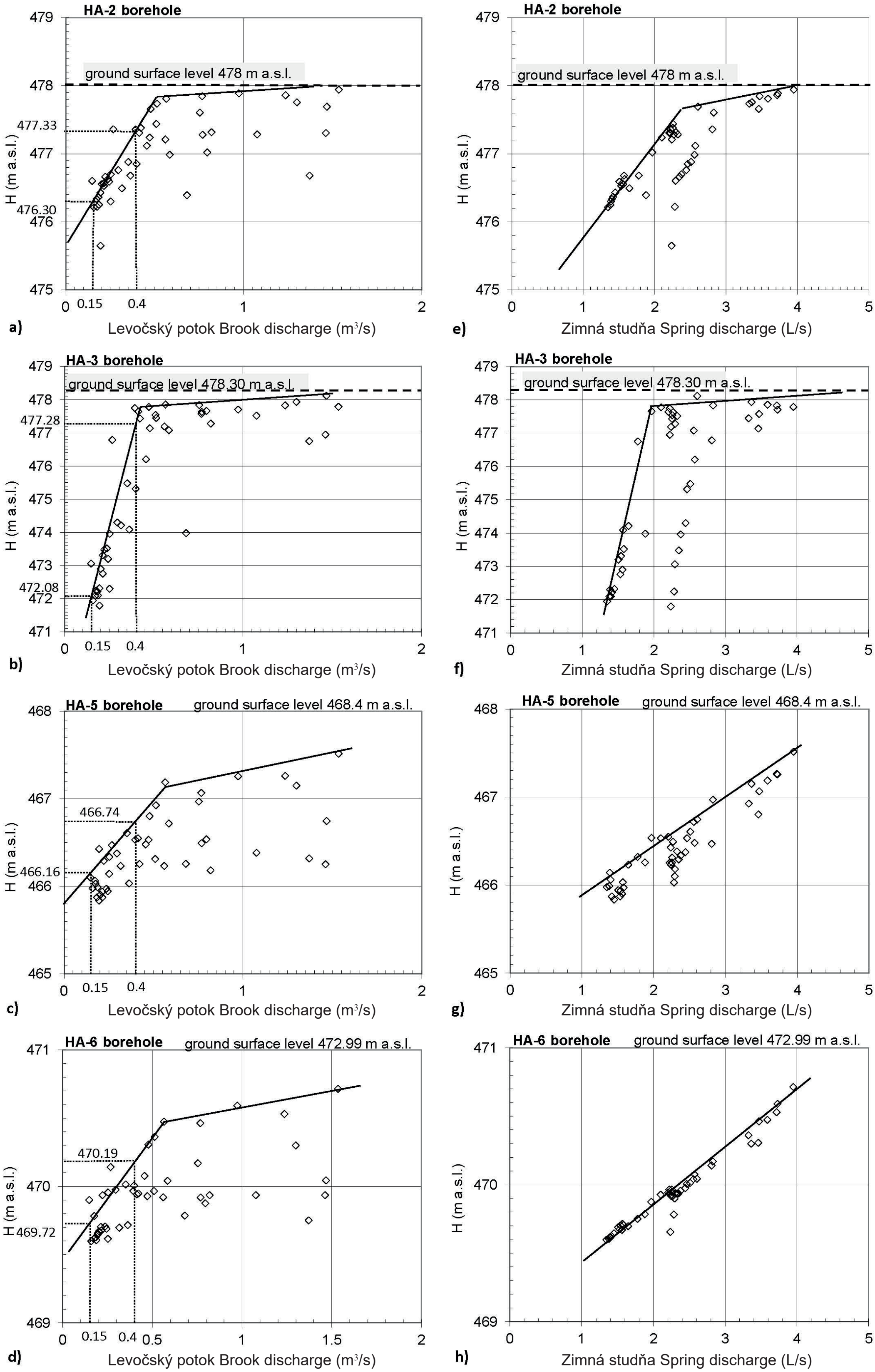


Fig. 7. Groundwater level H in observed boreholes versus the Levočský potok Brook discharge (a–d) and the Zimná studňa spring discharge (e–h). Explanation is given in the text.

boreholes (boreholes) usually is not very high therefore reliability of ΔH evaluation is crucial for the accuracy of S_y estimation.

In case of this study, four from six observed boreholes are usable for ΔH calculation – moreover they are situated within relatively small area comparing to overall watershed. The ΔH values recorded in them during hydrological year 1995 are from interval 1.04–6.17 m, giving average of 2.67 m. By comparing these values with ΔH values recorded on 11 boreholes situated in Paleogene sediments in the territory of Slovakia, which are observed by SHMI in frame of Slovak state monitoring program (0.72–4.28 m and 1.83 m in average, Tab. 6), it can be concluded that they are very alike in their range and also average values. The presented data suggest that difference of average ΔH value used for S_y estimation from real one

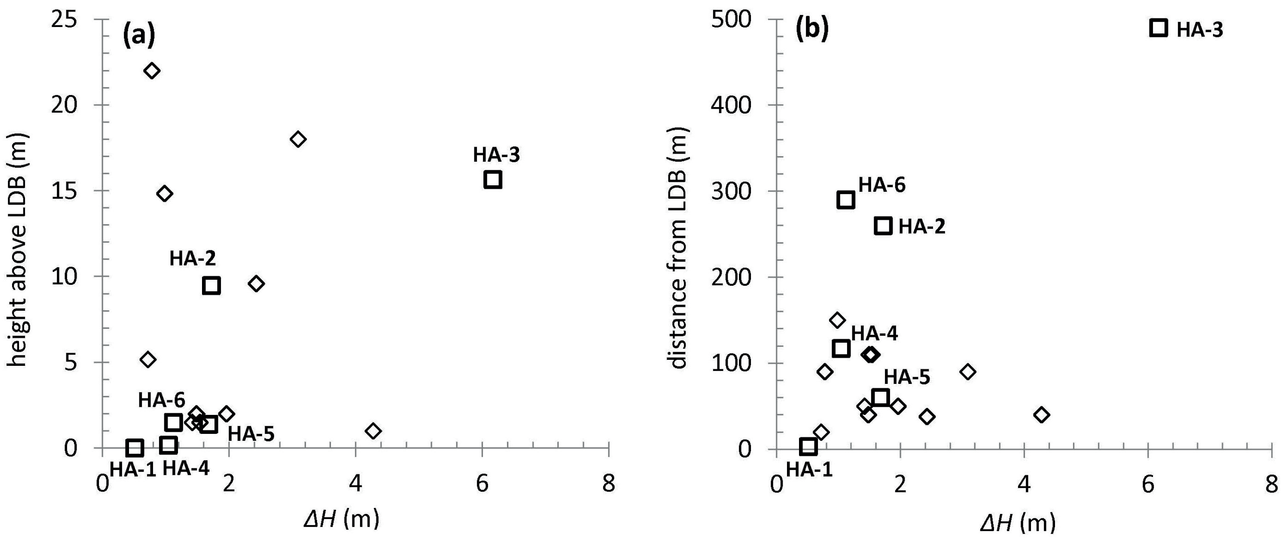


Fig. 8. Groundwater table fluctuation (ΔH) recorded in boreholes at the Harichovce site (squares) and in boreholes of state monitoring system (diamonds) versus: **a** – height of those borehole heads above local drainage base (LDB); **b** – distance of those boreholes from local drainage base (LDB).

Tab. 6

Groundwater level fluctuation on sites of state monitoring network situated in Paleogene sediments recorded by Slovak Hydrometeorological Institute (SHMI)

No.	Locality	Observed period	n	Elevation [m a. s. l.]	Hight above LDB [m]	H_{avg} [m]	H_{min} [m]	H_{max} [m]	ΔH [m]
5 211	Oravský Biely Potok	1991–2006	520	659.59	9.6	7.19	8.43	6	2.43
5 215	Jarabina	2004–2006	156	570.16	5.2	1.24	1.39	0.67	0.72
5 216	Ľubovnianske kúpele	2004–2006	156	566.18	22	5.32	5.58	4.8	0.78
5 219	Čirč	1982–2000	988	559.3	1.5	5.84	6.36	4.94	1.42
5 220	Livov	1982–2006	1 300	497.34	2	3.62	4.1	2.14	1.96
5 221	Olejníkov	1966–2006	2 132	514.21	1	5.23	6.5	2.22	4.28
5 222	Chminianske Jakubovany	1987–2005	988	401.21	1.5	1.53	2.22	0.68	1.54
5 223	Vyšné Raslavice	1983–2006	1 248	367.22	2	1.49	2.35	0.87	1.48
5 224	Dlhá Lúka	1989–2006	936	296.96	2	1.58	1.87	0.38	1.49
5 225	Hažlín	1989–2006	936	361.02	18	5.29	5.86	2.77	3.09
5 231	Zuberec	1990–2006	884	864.84	14.8	10.87	11.18	10.2	0.98

should be no higher than one meter in studied bedrock type. Underestimation of this calculated value is more likely than its overestimation.

Another question of *ΔH* values reliability is theirs possible affection by drainage effect and it relates to observation points located in zones of natural aquifer dewatering. Such kind of affection would cause underestimation of *ΔH*, compared to those observed in unaffected flow conditions on slope. In case of existence in regional scale, it could be revealed by positive correlation between *ΔH* recorded on individual boreholes and theirs height above local drainage base (LDB) – or theirs distance from LDB. However, existing data do not suggest existence of such correlation (Fig. 8a–b), so it can be supposed that *ΔH* estimation error due to drainage effect is not significant in this study.

The algorithm of the single nonlinear reservoir is used for the modeling of flow recession in this study, based on interpretation of MRC for the Levočský potok Brook. Although it does not fully correspond to the physical nature of ongoing processes, it describes the MRC more precisely as the algorithm for single linear reservoir (Fig. 4) and avoids the difficulty of dealing with multiple reservoirs. The problem of nonlinearity could also be solved using the assumption that baseflow is the outflow of two or more parallel linear reservoirs representing components of different response time (Moore, 1997; Schwarze et al., 1997). However, aquifers in the studied watershed are not linear reservoirs, as interpretation of the Zimná studňa spring recession curve reveals (Fig. 9). Depletion regime of local fault aquifer dewatered by this spring consists of two flow components. Simple groundwater flow component with laminar flow (described by exponential equation $Q_t = Q_0 e^{-\alpha(t-to)}$, Drogue 1967) is combined with turbulent flow component (linear equation $Q_t = Q_0[1 - \beta(t - t_0)]$, Mijatovič, 1972) occurring during highest recharge. Therefore, use of nonlinear model in this study is reasonable.

The two presented ways of using the *Q-S-H* method give comparable results. Using the first one gives the option to estimate *Sy* for hydrological seasons chosen. Latter one – making use of *Q-H* graphs – represents conservative estimate as envelope line

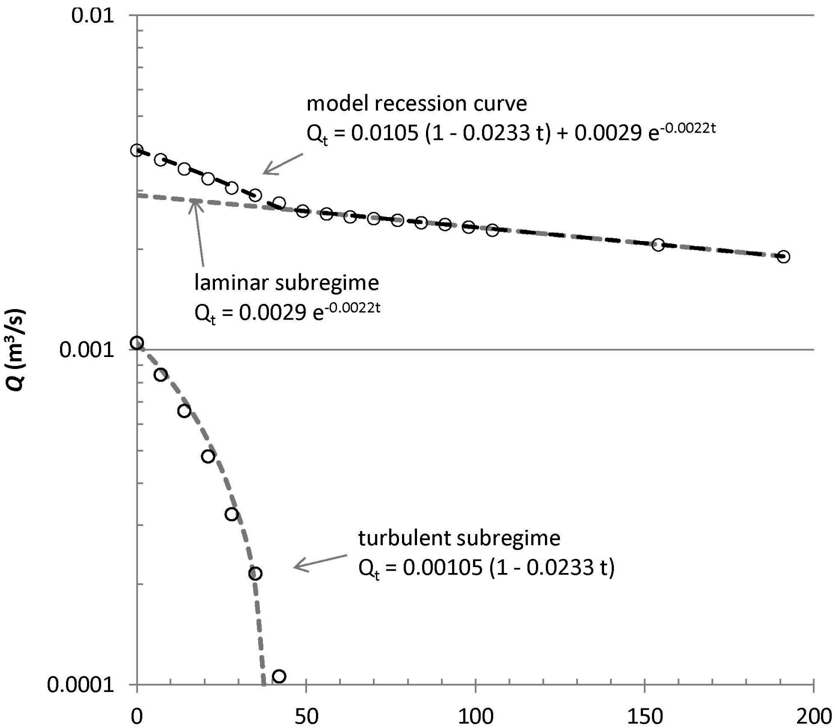


Fig. 9. Interpretation of the Zimná studňa spring recession curve.

Tab. 7

Results of S_y determination using five different methods based on recession curve analysis at the Odorica Brook watershed above gauging station no. 8 423 (Černák, 2016)

Considered aquifer thickness	LRM	KM	B [a1 – a3]	B [a2 – a3]	P
10	0.26	0.27	0.23	0.23	0.24
20	0.13	0.13	0.11	0.11	0.12
30	0.09	0.09	0.08	0.08	0.08
50	0.05	0.05	0.05	0.05	0.05
100	0.03	0.03	0.02	0.02	0.02

Explanation: Methods: LRM – linear reservoir model of Boussinesq (1877), KM – quadratic model of Boussinesq (1903), B(a1–a3) – method after Brutsaert, enveloping curves a1–a3, B(a2–a3) – method after Brutsaert – enveloping curves a2–a3, P – method after Parlange.

represents minimum baseflow that occurred for a particular measured groundwater table (Holko et al., 2002). Its advantage is given by ability to avoiding the error due to incorrect H data selection for calculation, which do not fully represent baseflow conditions. The S_y regional value obtained in this way for the Levočský potok Brook watershed of 0.0019 is very close to those, determined by pumping tests interpretation at the Harichovce site, where Bajtoš and Michalko (2003) states storativity coefficient of 0.0017 and 0.0014 for fissured fault zone captured in borehole HA-4 and HA-6, respectively (location of boreholes is shown on Fig. 1). As it is very probable that storativity in bedrock fissured/fault zones and in its weathered (near-surface) zone are very similar each other – like their permeability and transmissivity (Jetel, 2000) are – we can conclude here the consistency in results obtained by two independent methods.

The average computed from five S_Y values determined for different hydrological periods P1–P5 reaches 0.0028. More than this value, S_Y values computed for driest period P1 is similar to the value of 0.0019 determined using envelope line method. On the other hand, S_Y derived for wetter periods P3 and P4 are significantly higher (Fig. 10). These indications suggest that storativity of rocks forming upper parts of GWL fluctuation zone is significantly higher than those at its bottom. Based on data obtained the S_Y characteristic value at studied watershed is from the interval 0.001–0.002 and 0.002–0.005 for bottom and medium parts of GWL fluctuation zone, respectively. The existence of such vertical zonality of storativity is consistent with the previously described nature of studied bedrock aquifer.

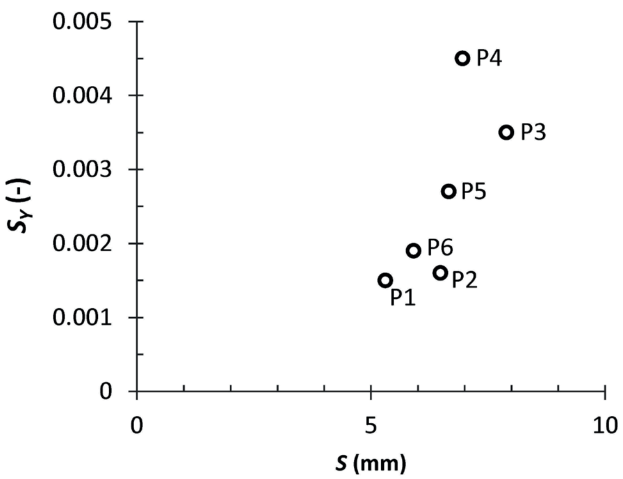


Fig. 10. Specific yield (S_Y) versus average storage (S) determined for different hydrological periods (P1–P5) and using envelope line method (P6) at the Levočský potok Brook watershed.

Černák (2016) used 5 different methods to determine S_y for the Odorica Brook watershed (Tab. 7), which creates the SE part of the Levočský potok Brook watershed (Fig. 1). Among an alternative S_y estimates for different considered aquifer thickness (real aquifer thickness cannot be exactly determined in this environment due to its vertical inhomogeneity), the best match with Sy = 0.002 value characterizing entire studied watershed is reached for considered thickness of 100 m, for all used methods.

Size of S_y values estimated by this study for the Levočský potok Brook watershed is also similar to those, evaluated by hydrographs analysis for fractured sedimentary rock aquifers at different sites. For highly fractured zones in shales and interbedded shales, siltstones and sandstones from Pennsylvania, USA, Gburek et al. (1999) compared the recession of borehole hydrographs with the base flow recession curve over a 40-day period for a stream draining the aquifer. Through calibration of a groundwater flow model, S_y was estimated to be 0.01 in the overburden, 0.005 in the highly fractured rocks at shallow depths, and 0.0001 in poorly fractured material below 22-m depth. Gburek and Folmar (1999) used a water-budget method and estimated S_y to range from 0.007 to 0.01 for the highly fractured zone at the same site. Moore (1992) compared stream-flow hydrographs with groundwater hydrographs from shale and limestone aquifers on the Oak Ridge Reservation, Tennessee, USA, and estimated S_y of approximately 0.001 from slopes of the recession curves. Using an approach analogous to hydrograph separation, Shevenell (1996) estimated S_y of 0.003, 0.001 and 0.0001 for conduits, fractures, and matrix elements, respectively, of the limestone and dolomite Knox Aquifer at Oak Ridge, by apportioning segments of borehole recession curves to these different flow regimes.

5. Conclusion

Proposed *Q-S-H* method for S_y estimation at fracture rock watersheds is based on the comparison of different groundwater storage (S) stages in watershed to corresponding groundwater levels. It is based on the premise that a rise in water-table elevation measured in shallow boreholes is caused by the addition of recharge across the water table at watershed and its following recession is caused by groundwater storage loss due to baseflow generation. Another assumption is that deep groundwater discharge – not drained by local streams as baseflow – is as small as can be neglected. Not but what this assumption restrict applicability of the method almost exclusively for bedrock flow systems dominated by shallow fractured rock aquifers, it can be broadly used as they are worldwide abundant. This approach is a gross simplification of many complex phenomena, however it makes the method simply and ease of use. Demonstration of the method at the Levočský potok Brook watershed (Western Carpathians, Slovakia), which is built by fracture porosity dominated Paleogene sediments, suggest that the characteristic S_y value at studied watershed is from the interval 0.001–0.002

and 0.002–0.005 for low and medium storage/runoff conditions (or bottom and middle part of GWL fluctuation zone), respectively. These findings showed consistency of achieved representative estimate with S_y values previously stated by local aquifer tests and also with the range of published S_y values, determined worldwide by other methods for shallow fractured rock aquifers.

The method can be applied for watershed where runoff and GWL fluctuation is observed within the same time period. Whereas runoff daily recorded data are needed for long enough period to construct MRC (usually 2 or more years in moderate climate), GWL fluctuation data can be observed on lower frequency. Even two GWL measurement campaigns could be sufficient, being performed in appropriate time regarding hydrological regime. More than time frequency, number of observed objects and theirs appropriate location are important in case of GWL data.

S_y belongs to very important parameters characterizing hydraulic properties of rocks. Since its knowledge is necessary for non-steady state groundwater flow modeling and groundwater storage balance, correct S_y values lead to better quality of practical hydrogeological issues concerning proper management and protection of valuable groundwater resources. Despite of multitude of known S_y determination methods, there is still lack of characteristic values describing specific rock types in the literature. Another question is quality and reliability of S_y values obtained by different methods and its correlative consistency. Proposed *Q-S-H* method can estimate S_y on local or regional scale (it is average or characteristic value for entire watershed), so it is very useful for studies at such scales. Additionally, it could be valuable to use it in combination in other (laboratory or aquifer pumping test) methods in site scale studies.

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Odhad vododajnosti pripovrchovej zóny skalných hornín v horských povodiach skúmaním vzťahu medzi základným odtokom, zásobou a úrovňou hladiny podzemnej vody

Vododajnosť (S_y) – schopnosť horniny nasýtenej vodou uvoľňovať ju voľným vytekaním pod vplyvom gravitácie – je dôležitý hydraulický parameter potrebný pri hodnotení a využívaní zdrojov podzemnej vody. Je aj kľúčovým parametrom pri odhade infiltrácie zo zrážok metódou WTF (Schicht a Walton, 1961). Na stanovenie vododajnosti sa používa viacero metód, či už terénnych alebo laboratórnych. Ich výsledkom sú však značne variabilné hodnoty S_y (Varni et al., 2013) a presnosť týchto metód sa hodnotí ťažko. Preto je veľmi prospešné aplikovať viacero metód stanovenia vododajnosti a sledovať konzistenciu dosiahnutých výsledkov.

Na regionálny odhad vododajnosti horninového prostredia v horských povodiach so zanedbateľným prestupom podzemnej vody do susedných povodí sa navrhuje aplikovať metódu založenú na kombinácii analýz výtokovej krivky povrchového odtoku, vzťahu zásoby vody v povodí k veľkosti podzemného odtoku a kolísania hladiny podzemnej vody (metóda *Q-S-H*). Za predpokladu, že pokles zásoby podzemnej vody odráža pokles základného odtoku v povodí podľa rovnice 4, reprezentatívna výtoková krivka (MRC; obr. 4) odtoku z povodia skonštruovaná z čiary prietoku z dostatočne dlhého obdobia môže byť využitá na určenie zmeny zásoby vody v povodí ΔS (obr. 5) zodpovedajúcej zmene úrovne hladiny podzemnej vody v povodí ΔH. Keďže hodnota ΔS zodpovedá množstvu odtečenej vody z povodia, môže sa určiť aj priamo z reálnej výtokovej čiary zaznamenanej vo zvolenom období neovplyvnenom rušivými vplyvmi ako súčet denných odtečených množstiev vo zvolenom časovom úseku. Priemerná vododajnosť v zóne kolísania hladiny podzemnej vody je potom daná vzťahom $S_y = \Delta S/\Delta H$.

Použitie metódy *Q-S-H* je demonštrované na povodí Levočského potoka, budovaného paleogénnymi sedimentmi s dominujúcou puklinovou priepustnosťou (obr. 1). Využili sa záznamy prietoku Levočského potoka zo stanice SHMÚ č. 8 424 v Markušovciach z rokov 1990 – 2012 a merania kolísania hladiny podzemnej vody v 6 vrtoch v hydrologickom roku 1995 (tab. 1). Na základe interpretácie zaznamenaných hodnôt prietoku bola skonštruovaná reprezentatívna výtoková krivka (MRC; obr. 4). Použila sa na kalibráciu nelineárnej modelovej krivky pri kalibračnom parametri získanom testovaním – koeficiente vyprázdňovania a = 15,5 (pri zvolenej hodnote parametra b = 0,5 reprezentujúcej nelineárny rezervoár). Jeho dosadením do rovnice 4 bol definovaný vzťah medzi zásobou podzemnej vody v študovanom povodí a jeho podzemným odtokom, vyjadrený rovnicou $S = 15{,}5Q^{0{,}5}$. To umožnilo simulovať zásobu podzemnej vody v povodí (S) pri rôznych úrovniach podzemného odtoku (Q) a korešpondujúcich úrovniach hladiny podzemnej vody (H). Prezentované sú dva odlišné spôsoby odhadu S_y: 1. porovnaním dvoch hydrologických stavov zvolených na hydrograme (obr. 6), 2. porovnaním dvoch hydrologických stavov zvolených na grafe *Q-H* s využitím metódy obalovej čiary (Kliner a Kněžek, 1974) (obr. 7). Na základe získaných časovo relevantných párov hodnôt S a H je hodnota S_y vypočítaná podľa rovnice 6. Prezentované dva prístupy dávajú porovnateľné výsledky. Použitie prvého z nich poskytuje možnosť výberu preferovaného hydrologického stavu v rámci sezónneho režimu, je však potrebné zvažovať rušivé vplyvy. Druhý z nich umožňuje konzervatívny odhad, keďže obalová čiara reprezentuje minimálny základný odtok pri určitej úrovni hladiny podzemnej vody.

Hodnoty S_y charakteristické pre pripovrchovú zónu hydrogeologického masívu v tomto povodí sú uvedeným postupom odhadnuté v intervale 0,001 – 0,002 pri nízkom základnom odtoku a v intervale 0,002 – 0,005 pri strednom základnom odtoku. Tieto hodnoty sú konzistentné s hodnotami koeficientu voľnej zásobnosti, zistenými v tomto povodí hydrodynamickými skúškami vrtov. Sú tiež v rozsahu hodnôt *Sy* určených inými terénnymi metódami v prostredí hydrogeologického masívu (Gburek, 1999; Gburek a Folmar, 1999; Moore, 1992; Shevenell, 1996). Pre potreby tejto štúdie bolo k dispozícii 6 vrtov s pozorovaním úrovne hladiny podzemnej vody (tab. 1). Len 4 z nich však bolo vhodné zaradiť do hodnotenia, a to kvôli vylúčeniu vplyvu drenážneho účinku miestnej eróznej bázy. Hodnoty *ΔH* za hydrologický rok 1995 v nich dosahovali 1,04 – 6,17 m s priemerom 2,67 m. Porovnanie týchto hodnôt s hodnotami *ΔH* zaznamenanými v 11 vrtoch situovaných v paleogénnych sedimentoch na území Slovenska, ktoré pozoruje SHMÚ v rámci štátneho monitoringu podzemnej vody (0,72 – 4,28 m a priemer 1,83 m; tab. 6), ukazuje ich značnú podobnosť. Toto porovnanie naznačuje, že rozdiel priemernej hodnoty *ΔH* použitej na odhad *Sy* oproti reálnej hodnote by nemal byť väčší ako 1 m. Podhodnotenie použitej vstupnej hodnoty *ΔH* je pritom pravdepodobnejšie ako jej nadhodnotenie.

Použitie tejto metódy je limitované na povodia, v ktorých hlbší podzemný odtok do susedných povodí je taký nízky, že ho možno zanedbať. Ide najmä o povodia budované hydrogeologickým masívom – komplexmi spevnených hornín s puklinovou priepustnosťou bez významnejších súvislých vrstvových kolektorov, s obehom podzemnej vody sústredeným do pripovrchovej zóny, prípadne uzavreté hydrogeologické štruktúry. Údajová báza na použitie tejto metódy pozostáva z čiary prietokov toku zo záverečného profilu hodnoteného povodia a časovo korešpondujúcich údajov o kolísaní hladiny podzemnej vody v pozorovacích objektoch situovaných v tomto povodí. Na konštrukciu reprezentatívnej výtokovej krivky (MRC) sú potrebné denné záznamy prietoku z dostatočne dlhého obdobia (zvyčajne 2 roky a viac). Frekvencia meraní úrovne hladiny môže byť nižšia – pri vhodnom načasovaní vzhľadom na hydrologický režim postačuje niekoľko opakovaní merania. Najdôležitejší je počet vhodne situovaných pozorovacích objektov. Hoci tento prístup predpokladá značné zjednodušenie zložitých prírodných procesov prebiehajúcich pri tvorbe podzemného odtoku, je výhodný z hľadiska možnosti získania potrebných podkladových údajov a nenáročnosti ich vyhodnotenia.

Vododajnosť patrí k dôležitým parametrom charakterizujúcim hydraulické vlastnosti hornín. Keďže predstavuje vstupný parameter pri modelovaní neustáleného prúdenia podzemnej vody, korektné hodnoty S_y sú potrebné na kvalitné riešenia praktických hydrogeologických úloh týkajúcich sa správneho manažovania a ochrany zdrojov podzemnej vody. Napriek početným známym metódam ich určovania v odbornej literatúre pretrváva nedostatok dostupných charakteristických hodnôt S_y reprezentujúcich špecifické horninové typy. Kvalita a spoľahlivosť dostupných hodnôt S_y a ich konzistentnosť pri ich získavaní rôznymi metódami sú predmetom diskusií. Preto je potrebné rozširovať existujúcu údajovú bázu a testovať spoľahlivosť použitých metód. Navrhovanou metódou *Q-S-H* možno odhadovať S_y v lokálnej alebo regionálnej mierke ako hodnotu priemernú, resp. charakteristickú pre študované povodie. Pri regionálnych štúdiách a lokálnych prieskumoch je vhodné kombinovať ju s inými dostupnými metódami, najmä hydrodynamickými a laboratórnymi skúškami.

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