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Estimation of baseflow residence times in watersheds from the runoff hydrograph recession: method and application in the Neversink watershed, Catskill Mountains, New York

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Abstract

A method for estimation of mean baseflow residence time in watersheds from hydrograph runoff recession characteristics was developed. Runoff recession characteristics were computed for the period 1993–96 in the 2 km^2 Winnisook watershed, Catskill Mountains, southeastern New York, and were used to derive mean values of subsurface hydraulic conductivity and the storage coefficient. These values were then used to estimate the mean baseflow residence time from an expression of the soil contact time, based on watershed soil and topographic characteristics. For comparison, mean baseflow residence times were calculated for the same period of time through the traditional convolution integral approach, which relates rainfall $\delta^{18}O$ to $\delta^{18}O$ values in streamflow. Our computed mean baseflow residence time was 9 months by both methods. These results indicate that baseflow residence time can be calculated accurately using recession analysis, and the method is less expensive than using environmental and/or artificial tracers. Published in 2002 by John Wiley & Sons, Ltd.

Introduction

Mean residence time of stream baseflow is a watershed variable that has proven useful to describe the mixing of waters and the contribution of groundwater discharge to streamflow in headwater catchments (Burns *et al*., 1998; McDonnell *et al*., 2000). Most estimates of baseflow residence time have been made by mathematical models using a convolution integral approaches that relates baseflow to rainfall isotopic or chemical data (Maloszewski *et al*., 1992; Vitvar *et al*., 1999, McGuire *et al*., 2002). These methods are cumbersome and expensive and rely on data that are rarely available routinely. Techniques in which baseflow residence time can be calculated in watersheds through inexpensive, readily available data are much needed. To this end, Wolock *et al*. (1997) used readily available topographic and soils data to estimate the 'contact time' of baseflow within soils of the Neversink River watershed in southeastern New York. They defined contact time as the residence time of water in the soil. The contact time estimate was based on the groundwater runoff generation routine of Beven and Kirkby (1979), and used mean values of soil characteristics (hydraulic conductivity and storage coefficient and the mean watershed topographic wetness index value of $\ln(a/\tan\beta)$, where a is the specific contributing area, and β is

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the slope angle). This paper extends the Wolock *et al.* (1997) approach by proposing a new method to estimate the residence time of baseflow within watersheds through hydrograph recession analysis. This new approach is based on the subdivision of the recession limb of a hydrograph into a set of linear segments (Tallaksen, 1995) from which aquifer parameters are calculated. The recession analysis technique is derived from methods commonly used to study groundwater dynamics and the exchange between groundwater and stream water in karst and fractured aquifers (Mijatovic, 1974, Bonacci, 1993, Powers and Shevenell, 2000). Previously, this approach has not been used to estimate hydraulic characteristics on a watershed scale.

We demonstrate the application of the estimated hydraulic characteristics as alternative values to the soil-derived parameters used in the approach of Wolock *et al.* (1997). This new approach is applied to a small watershed in the Catskill Mountains of New York and compared with results of mean baseflow residence time based on isotopic data using the convolution integral approach. We also show how mean residence time differs from soil water contact time computed from soil properties and topographic characteristics.

Study Site

Winnisook is a 2 km^2 watershed that forms the headwaters of the West Branch Neversink River in the Catskill Mountains. This watershed has been studied in numerous hydrologic and biogeochemical investigations over the past decade (Wolock *et al.*, 1997; Burns *et al.*, 1998; Lawrence *et al.*, 2000). Elevation in the watershed ranges from 817 to 1274 m. The watershed is underlain by Devonian sandstone and conglomerate and overlain by Pleistocene till (Way, 1972). Soils are Inceptisols with a mean depth of about 75 cm. The catchment is completely forested with northern hardwoods. Mean annual temperature is 4·3 °C, and mean precipitation as measured at the Slide Mountain station (<1 km from base of catchment) is 1570 mm, of which 23% falls as snow. The US Geological Survey (USGS) established a streamflow gauge for continuous measurement of discharge in 1991. Mean annual watershed evapotranspiration is 430 mm, and mean annual runoff is 1140 mm. Stream water samples for δ^{18}O analysis were collected during 1993–96 (as part of the study by Burns *et al.* (1998)) and were analysed by mass spectrometry at a USGS laboratory in Menlo Park, CA.

Methods: Derivation of Aquifer Parameters from Baseflow Recession

Our new approach is a modification, of sorts, to the Wolock *et al.* (1997) method of contact time calculation. Their approach was to compute T_c(s) as:

$$T_{\mathrm{c}} = \frac{S}{k}\,\mathrm{e}^{\lambda} \tag{1}$$

where λ is the mean watershed topographic wetness index ($\ln(a/\tan\beta)$) value, S is the storage coefficient and k is the soil hydraulic conductivity. A detailed description of this approach is given in Wolock *et al.* (1997) and we do not repeat those details here. Our new method essentially eliminates the need for a k value—something highly variable in space and difficult to quantify (even within a few orders of magnitude on a catchment basis).

Records of daily discharge in the Winnisook watershed from the period 1993–98 were analysed. The baseflow hydrograph was modelled by an exponential relation of the form (e.g. Fetter, 1988)

$$Q_t = Q_0\,\mathrm{e}^{-\alpha t} \tag{2}$$

where Q_0 is the baseflow at time $t = 0$, Q_t (m^3 s^{-1}) is baseflow at a later time t, and α is the recession coefficient expressed in inverse time (86 400 s in a day). The maximum storage volume was then described as:

$$V_{\mathrm{m}} = \frac{Q_0 t}{\alpha} \tag{3}$$

where V_m (m^3) is the transient storage of water that would be discharged during a recession from peak flow to zero if no additional recharge entered the stream. Q_0 represents the point on the recession line after the runoff peak at the beginning of the runoff recession. The recession line from Q_0 to the point at which the groundwater storage is empty represents the total possible recession time t_r. Since this time in Equation (2) is theoretically infinite, we arbitrarily defined the point at which $Q = 0{\cdot}001$ m^3 s^{-1} as zero. Thus, the quantity V_m/t_r gives the mean baseflow during the recession Q_m. As an analogy, runoff recession can be equated to a pumping test in an unconfined aquifer (Cooper and Jacob 1946), wherein the drawdown dH (m) is analogous to the depletion of the mobile water reservoir, and Q is analogous to stream outflow. The depletion can be quantified as:

$$\mathrm{d}H = \frac{2{\cdot}30 Q_\mathrm{m}}{4\pi T} \tag{4}$$

where T is transmissivity and Q_m is the mean baseflow during the recession. Solving this equation for T gives the mean transmissivity. Because the 'true' aquifer depth is generally unknown, the hydraulic gradient I of groundwater can be calculated from Darcy's Law as (Moore, 1992):

$$I = \frac{LQ_0}{AT} \tag{5}$$

where A is the watershed area, and L is the maximum flowpath length (Figure 1). If the Winnisook watershed is idealized as an area measuring 2×1 km^2 (based on geographic data in Lawrence *et al.* (2000)), the flowpath length a is equal to half of the watershed width, i.e. 500 m. The mean groundwater saturated flow Q_m (m^3 s^{-1}) from both hillslope sides towards the stream can then be written as

$$Q_\mathrm{m} = 2kIA \tag{6}$$

where k is the mean saturated hydraulic conductivity. If I is known, then the storage coefficient S can also be expressed in terms of volume (Moore, 1992):

$$S = \frac{V_\mathrm{m}}{LIA} \tag{7}$$

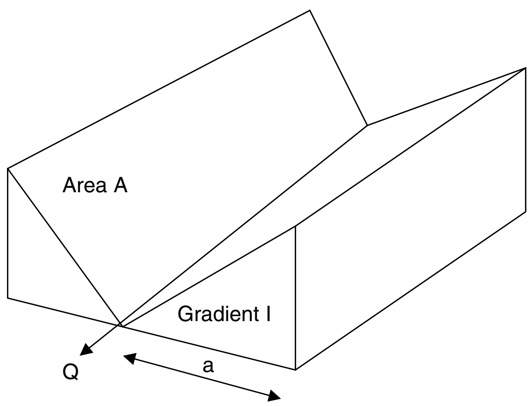

Figure 1. Idealized watershed geometry for the Winnisook watershed, where Q is discharge at the watershed outlet, A is watershed area, and I is mean gradient along the flowpath length (half of the watershed width)

Equations (6) and (7) yield mean values for hydraulic characteristics k and S of the dynamic groundwater reservoir. This then replaces the k term approach from Equation (1) in favour of a more spatially integrated measure of catchment hydrological control based on baseflow residence time.

The ^{18}O convolution integral approach provided an independent estimate of the mean residence time of baseflow. Available ^{18}O data in baseflow for 1993–96 at the Winnisook gauge was analysed, and input–output flow models were applied at a monthly time step using methods described in Maloszewski *et al.* (1992). The ^{18}O input record was derived from weekly samples of precipitation collected at Biscuit Brook, a US National Atmospheric Deposition Program site about 8 km from Winnisook watershed, during 1993–96. These values were projected back to 1990–93 from data recorded at the IAEA/WMO station in Ottawa to extend the data series further in time. The projection was based on an estimated difference of $\delta^{18}\mathrm{O} = +2‰$ SMOW from Ottawa to Biscuit Brook (derived from a linear regression relation based on latitude and elevation). We have made similar calculations in our previously published work from catchments in Switzerland (Vitvar *et al.*, 1999), New Zealand (Stewart and McDonnell, 1991) and New York State (Burns and McDonnell, 1998). Details of the approach are not presented in this Briefing, but the reader is referred to our earlier work for a more complete description of the procedure.

Finally, the total storage volume V_t (m^3) of the watershed was calculated as

$$V_\mathrm{t} = Q_\mathrm{b} A T_\mathrm{m} \tag{8}$$

where Q_b (mm) is the separated mean baseflow at the Winnisook gauge, T_m is the mean residence time, and A is the area. This volume is greater than the dynamic transient volume V_m because it also consists of water that does not participate in the recession.

Results

Our new method

Baseflow recessions for a 'typical' water year (we chose 1996) were selected from daily discharge data in the Winnisook watershed for the period 1993–98 (Figure 2). The shape of the hydrograph recessions

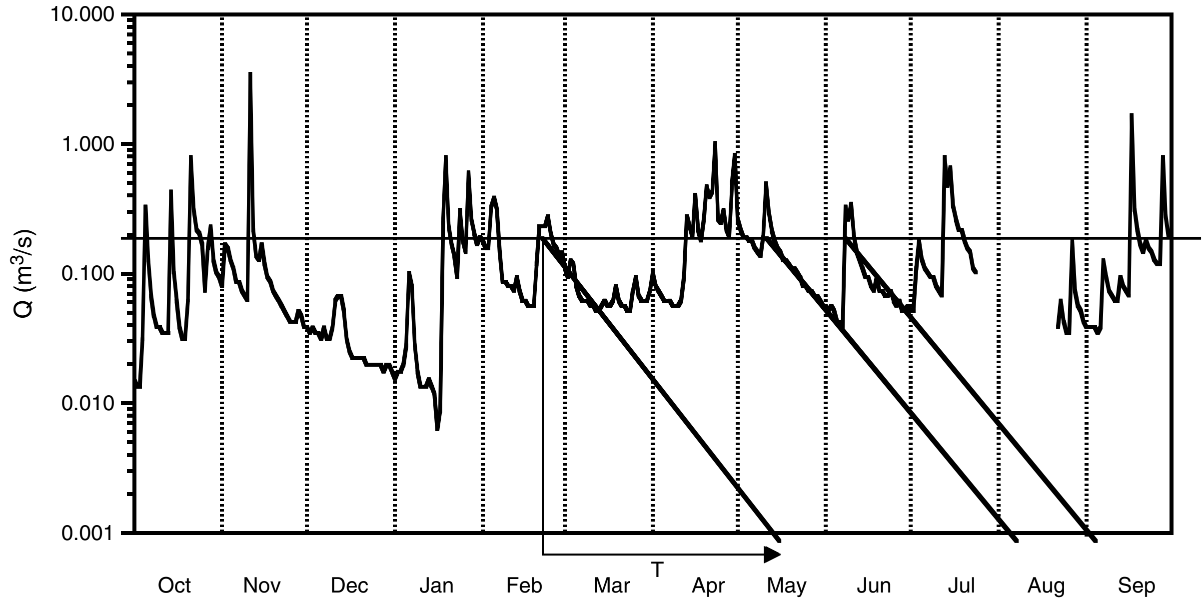

Figure 2. Baseflow recession segments and daily runoff record at the Winnisook gauge, water year 1996

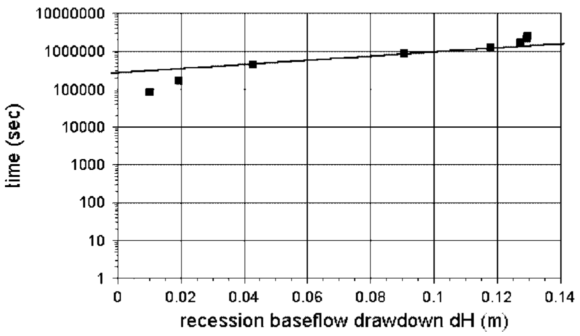

Figure 3. Semi-logarithmic drawdown–time relation for the defined baseflow recession segments. One logarithmic time cycle corresponds to a drawdown of about 0·18 m. The relation was fitted using the Cooper–Jacob solution defined in the text

was consistent for the whole period, except during snowmelt. Baseflow segments were determined to begin at a baseflow peak $Q_0 = 0{\cdot}20\ \text{m}^3\ \text{s}^{-1}$ and decline exponentially toward zero with a shape that is described by the recession coefficient 0·08 days^{-1}. A full recession–depletion to 0·001 $\text{m}^3\ \text{s}^{-1}$ would occur in 80 days (Figure 2). Thus, applying Equation (3) gives a transient volume of 216 000 m^3. The mean baseflow for the entire recession period t_r is $V_m/t_r = 0{\cdot}03\ \text{m}^3\ \text{s}^{-1}$. The drawdown–time relation for this recession segment is plotted in Figure 3. The drawdown is calculated as a reservoir change within a time interval $dH = dV/A\,dt$. Solving Equation (4) for the mean transmissivity T gives a value of 2 600 $\text{m}^2\ \text{day}^{-1}$. Then applying to Equation (5), peak baseflow $Q_0 = 0{\cdot}20\ \text{m}^3\ \text{s}^{-1}$, watershed area A of 200 ha, transmissivity T of 2 600 $\text{m}^2\ \text{day}^{-1}$ and the maximum flowpath length $L = 500$ m, yields a hydraulic gradient $I = 1{\cdot}7 \times 10^{-3}$. Applying this gradient to the flowpath length of 500 m and inserting the storage-capacity expression (Equation (7)), together with the transient volume $V_m = 216\,000\ \text{m}^3$ and watershed area $A = 200$ ha, gives a mean storage capacity $S = 0{\cdot}13$. Finally, applying Equation (6) to calculate the mean storage hydraulic conductivity k, from the mean recession baseflow $Q = 0{\cdot}03\ \text{m}^3\ \text{s}^{-1}$, gradient $I = 1{\cdot}7 \times 10^{-3}$, and area $A = 200$ ha, yields a value of $k = 0{\cdot}38\ \text{m day}^{-1}$. The mean baseflow residence time as obtained by Equation (1) is then 250 days.

The ^{18}O-derived mean residence time

The simulated and measured ^{18}O record in baseflow at the Winnisook gauge during 1993–96 is shown in Figure 4. The best fit was obtained through a dispersion model (parameter $D/vx = 0{\cdot}4$) as well as an exponential piston flow model (parameter $\eta = 1{\cdot}2$).

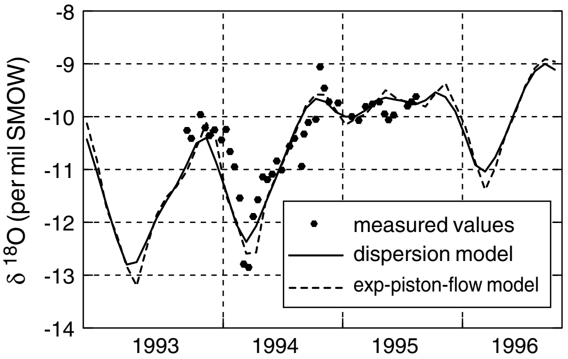

Figure 4. Simulated and measured ^{18}O output data for stream baseflow at the Winnisook gauge, New York, on a monthly step, 1993–96

The physical meaning of the parameters D/vx and η are described, for example, in Vitvar *et al.* (1999) and Stewart and McDonnell (1991). The computed mean residence time was 270 days. The parameters were calibrated with respect to the best fit (Nash–Sutcliffe efficiency criterion) with an uncertainty of 0·5 months. While we acknowledge that other "fits" could be obtained using different system response functions (other than the exponential piston flow model), field observations from Winnisook show the presence of both coarse alluvial material and bedrock rubble on hillslopes immediately adjacent to the stream, which are highly permeable and drainable on an event time scale. We conceptualize that these areas of alluvial material and bedrock rubble are the first to contribute to the hydrograph recession, followed by slower matrix flow through the soils, till, and bedrock fractures. This fast response is, therefore, a critical part of the shallow flow system, especially in the most upstream locations, where soils are thin and bedrock is near the ground surface. We hypothesize that rapid flow through this coarse material contributes to the 'piston flow' part of the groundwater flow system (water flowing with minimal mixing through channels and openings) based upon our visual observations, evidence from previous studies (e.g. Brown *et al.*, 1999; Welsch *et al.*, 2001), and the fact that the exponential piston flow model best approximates our isotopic system response function. Therefore, much of the fast near-surface contributions to streamflow may have their source in this bedrock rubble and not solely in the soils.

The original subsurface contact time approach

The subsurface contact time value obtained from soil and topographic parameters at Winnisook watershed was 15·3 months, or 460 days (Wolock *et al.*, 1997). This value is considerably higher than the values obtained from the other two methods. Mathematically, this results from the high values of the soil storage coefficient ($S = 0{\cdot}5$) that were used to solve Equation (1) in Fan (1995) and published by Wolock *et al.* (1997). The behaviour of the rubble, rock, shallow groundwater and till storages (discussed above) would suggest large heterogeneities in the hydraulic conductivity of soil, till, and bedrock rubble. Thus, application of Equation (1) with consideration of only 'soil' properties may not capture the first-order controls on flow.

Discussion and Conclusions

The new method and the standard isotope-based mean residence time calculation agree well. Both showed mean baseflow residence time at the Winnisook gauge to be on the order of 8–9 months. Additionally, a similar mean residence time of 11 months was obtained by Burns *et al.* (1998), who used a simplified sinusoidal ^{18}O transfer function approach based on a smaller data set from the same site. This provides some evidence that using hydraulic parameters obtained by our new method is reasonable. Applying Equation (8) for $T_m = 9$ months, $A = 200$ ha, and $Q_b = 0{\cdot}7$ m (as obtained by the program HYSEP; Sloto and Crouse, 1996) gives a total reservoir volume V_t of about 1×10^6 m^3. This volume is much greater than the transient volume V_m because it includes all of the water that appears in stream baseflow, including the relatively stable flow during dry periods. The volume V_m, on the other hand, includes only the depletion during the recession. This may be one reason for the discrepancy between our method and the original subsurface contact time approach, where the latter approach yielded mean baseflow residence times almost double that of the new method and ^{18}O-derived calculations.

This paper shows that using hydrograph recession analysis to calculate baseflow mean residence time can yield reliable values compared with the traditional (and more expensive and difficult) isotope-based residence time estimation technique. The new method is based on the assumption that mean aquifer parameters

k and S obtained by recession analysis are 'hydrologically applicable' in a manner similar to that used for mean soil parameters in the contact time approach (Equation (1); Wolock *et al*., 1997). Application of this method requires two assumptions: (1) that idealized watershed topography accurately represents the baseflow generation process; and (2) baseflow recession is log–linear and repeatable over a large number of recession periods.

The assumption that watershed topography and storage volumes equate to the pumping test approach to hydrograph recession analysis requires further testing before application to other more geomorphologically complex watersheds. Winnisook is a symmetrical watershed that can reasonably be idealized as a 1×2 km^2 area where the transient storage V_{m} is evenly distributed across both sides of the catchment. Additionally, the form of the recession hydrograph for Winnisook (Figure 1) indicates no extreme difference in hydraulic characteristics between transient storage and subsurface storage for the entire reservoir; thus, the mean residence time of transient storage can be considered similar to the entire storage V_{t}. If there were an extreme difference, the runoff peak from the transient storage would be extremely sharp, followed by a flat non-receding baseflow from deeper storage (such as that reported by Harr (1977)). In such a hypothetical situation, Equation (2) would not show linear baseflow recession segments and the hydraulic parameters of subsurface storage would require additional data sets for parameterization, such as a well hydrograph close to the stream (Moore, 1992; Powers and Shevenell, 2000). The similar shape of the recession segments at Winnisook through the entire period of analysis (except snowmelt) indicates stable depletion of the aquifer under stable hydraulic conditions. There are certainly seasonal effects on the form of the recession hydrograph, such as possible evapotranspiration from subsurface storage in the summer. Nevertheless, these effects are neglected when mean annual recession is analysed.

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