ELSEVIER

Journal of Hydrology 219 (1999) 20–33

Journal of Hydrology

Watershed groundwater balance estimation using streamflow recession analysis and baseflow separation

H. Wittenberg[a,*], M. Sivapalan[b]

[a]Department of Civil Engineering, Fachhochschule Nordostniedersachsen, University of Applied Sciences, 29556 Suderburg, Germany
[b]Centre for Water Research, Department of Environmental Engineering, University of Western Australia, Nedlands W.A. 6907, Australia

Received 27 April 1998; accepted 10 March 1999

Abstract

By the analysis of the observed time series of streamflow from catchments, the main components of the underlying groundwater balance, namely, discharge, evapotranspiration loss, storage and recharge, can be identified and quantified. This holistic (as opposed to reductionist) estimation method is demonstrated for the Harris River catchment in southwest Western Australia. The relationship between the groundwater discharge and the reservoir storage of shallow unconfined aquifers was found to be nonlinear based on the analysis of numerous streamflow recession curves. However, depletion of groundwater by evapotranspiration losses, through the water uptake of tree roots, was found to bias the recession curves and the estimated reservoir parameters. As a result of the seasonality of both rainfall and potential evaporation, analysis of the recession curves, stratified according to time of the year, allowed the quantification of evapotranspiration loss as a function of calendar month and stored groundwater storage. Time series of recharge to the groundwater aquifer were computed from the observed total streamflows, and the estimated discharge and evapotranspiration losses, by *inverse nonlinear reservoir routing*. Using traditional unit hydrograph methods unit recharge responses to rainfall were computed by least squares fitting. The shapes of the estimated unit response functions showed no significant seasonal variation.

Keywords: Groundwater balance; Baseflow separation; Nonlinear reservoir; Baseflow recession; Recharge response

1. Introduction to groundwater balance

It is well known that much of the observed streamflow of many rivers, in many different hydrological and climatic settings, is the outflow emanating from the shallow groundwater reservoirs of the associated catchments. Such groundwater reservoirs are also an important water resource both for the maintenance of the natural environment as well as for human needs. An understanding and quantification of the water balance of these shallow groundwaters, which take part in the seasonal water cycle, expressed in the form of the time series of storage, discharge and evapotranspiration (outflows), recharge (inflow), and the relationship of the latter to rainfall inputs, are important for their monitoring and management.

Fig. 1 provides a quantitative illustration of the main fluxes to and from a shallow groundwater aquifer. These were estimated for the Harris River catchment, a tributary of the Collie River in southwest Western Australia with an average annual

* Corresponding author.
E-mail address: hartmut-wittenberg@fbbwu.fh-lueneburg.de (H. Wittenberg)

PII: S0022-1694(99)00040-2

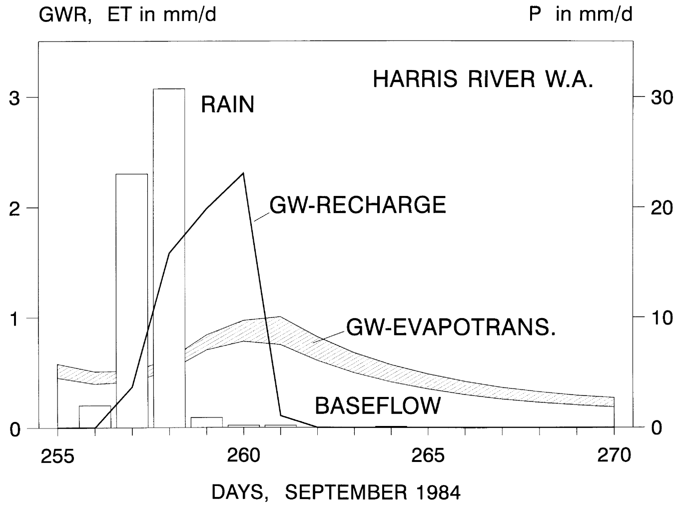

Fig. 1. Main fluxes to and from the groundwater aquifer of a catchment.

precipitation depth of about 950 mm. In this example a September (later winter) storm event, with an observed rainfall depth of 57 mm, yields a recharge to the saturated groundwater of less than 6 mm. Owing to the high transmissivity of the sandy and lateritic soils in the area, and especially due to infiltration through preferred pathways and macropores (Johnston, 1987; Bronstert and Plate, 1997), the groundwater recharge response is almost spontaneous, starting within hours and lasting a few days. The groundwater reservoir also reacts immediately to this recharge causing an immediate increase of the groundwater discharge, reaching a maximum at about the time of cessation of recharge; this is quite similar to what is seen in flow routing through surface water reservoirs, where inflows cause the increase of outflows due to associated rise of hydraulic head. The last remaining component of the groundwater balance is depletion due to evapotranspiration, which may be important depending on climate, soil properties, and especially vegetation (Nichols, 1994). In studies and practical work mostly considered negligible in zones of moderate climate, these losses may actually surpass baseflow under arid and semi-arid conditions, such as those which prevail in southwest Western Australia. Over long time periods, baseflow discharge added to rates of evapotranspiration should, and do, balance rates of recharge to groundwater.

1.1. Approaches to estimating groundwater balance

The more common way to estimate the fluxes to and from groundwater aquifers are the reductionist or ‘bottom-up’ approaches (Caro and Eagleson, 1981; Besbes and de Marsily, 1984; Renger et al., 1986; Wu et al., 1997) which start with measured rainfall at the surface and estimate, in order, infiltration, redistribution, evapotranspiration, percolation of the residual water through the unsaturated zone to the groundwater table, and discharge of the groundwater aquifer to streams as baseflow. However, only a small fraction of precipitation may actually recharge groundwater—while this may be up to 30% in the regions of moderate climate of Central Europe, it is less than 5% under the semi-arid conditions of Western Australia. In these circumstances, the errors in the measurement of precipitation and in the derivation of catchment-average rainfall, combined with the errors caused by the uncertainty of soil parameters used in the individual process models, can easily exceed both groundwater recharge and discharge.

It therefore appears sensible to consider the use of an alternative, holistic or ‘top-down’ approach to assess groundwater balance which relies entirely on the analysis of measured streamflow. The top-down approach is based on the separation of observed total (integrated) streamflow into quickflow and baseflow, and the identification of the baseflow as the outflow from a groundwater reservoir. This is the thrust of this article, which relies heavily on the integrative nature of natural catchments, avoiding the difficulties in the use of small-scale phenomenological equations. This approach has been previously utilised to compute groundwater recharge by baseflow separation (using linear reservoirs) by Chapman (1997); Chapman and Maxwell (1996); Fröhlich et al., 1994; Nathan and McMahon (1990) and others. This article follows a similar approach (but uses a nonlinear reservoir algorithm) to compute recharge time series, but goes further, and relates the estimated recharge time series to the rainfall inputs.

1.2. Outline of the article

The principal aim of the article is to estimate the shallow groundwater balance in a semi-arid catchment in Western Australia by streamflow recession analysis and baseflow separation. The methodology consists of four steps: (i) baseflow recession analysis to estimate a storage–discharge relationship for the groundwater aquifer; (ii) estimation of the seasonal variation of evapotranspiration losses based on the estimated baseflow recession curves; (iii) estimation of the recharge to groundwater based on the inverse baseflow separation; and (iv) estimation of a unit response function of the vadose zone to relate precipitation to groundwater recharge. The methodology for water balance proposed here is entirely empirical, and relies on the decoding of observed stream flow data, especially flow recessions, using a mixture of physical intuition and established groundwater flow theory.

Baseflow of rivers originates predominantly from the saturated zone, the shallow groundwater reservoir, which in most cases is unconfined. Discharge from this groundwater reservoir exfiltrates through river banks and the bottom of river beds. This discharge has a relatively fast response to rainfall due to the mobilization of ‘‘old’’, pre-event groundwater (Chapman and Maxwell, 1996; Herrmann, 1997) stored within a short distance to the river, by percolating rainwater increasing the level and hydraulic head of the groundwater reservoir. When recharge ceases and superficial influences such as surface depression storage, losses or abstraction are negligible, the resulting streamflow recession reflects the storage–discharge relationship of the groundwater aquifer alone, which can then be decoded by appropriate numerical analyses.

Theoretical (i.e. application of Darcy's law to unconfined aquifers) and numerical analyses of streamflow recession (Coutagne, 1948; Kubota and Sivapalan, 1995; Chapman, 1997; Wittenberg, 1994; Wittenberg, 1999 have shown that the storage–discharge relationship is nonlinear with, in the ideal case, discharge being proportional to the square of groundwater storage. This is in contrast to the traditional concept of the linear reservoir, first introduced by Maillet (1905), and still widely used following tradition and due to the ease of mathematical manipulation. One important application of the information gained from the recession curve analysis is the separation of the baseflow component from the time series of total streamflow (Nathan and McMahon, 1990; Fröhlich et al., 1994; Chapman and Maxwell, 1996), and the estimation of the corresponding time series of groundwater storage and recharge. Wittenberg (1999) proposed and implemented a nonlinear reservoir flow routing algorithm for this purpose (i.e. an inverse baseflow separation procedure), which is based on the previously estimated nonlinear storage–discharge relationship.

However, baseflow recession studies in the semi-arid catchments in the Collie River basin, in southwest Western Australia, suggest a strong seasonal variation of the storage–discharge relationship of the shallow aquifers, which can be attributed to biasing by seasonally varying evapotranspiration losses. One aim of this article is to quantify these losses through streamflow recession analyses, and to incorporate these in the nonlinear baseflow separation algorithm of Wittenberg (1999). Another aim of the article is to make inferences about the relationship between the estimated groundwater recharge rates and measured rainfall intensities, and to investigate the feasibility of characterising it in terms of simple time-invariant unit response functions (i.e. unit hydrographs) based on travel time distributions. These results will be compared against those found by other authors (e.g. Besbes and de Marsily, 1984; Wu et al., 1997) using theoretical (bottom-up) approaches, and by making inferences based on lysimeter measurements.

2. Baseflow recession analysis and estimation of storage–discharge relationship

Ever since Maillet (1905), the exponential function $Q_t = Q_0 \cdot \exp(-t/k)$ has been widely used to describe the baseflow recession, where Q_t is the discharge at time t, Q_0 the initial discharge, and k the recession constant which can be considered to represent average response time in storage. The exponential function implies that the groundwater aquifer behaves like a single linear reservoir with storage linearly proportional to outflow, namely $S = k \cdot Q$.

It is, however, evident that the parameter k fitted to different discharge ranges of the recession curves in actual rivers does not remain a constant but increases systematically with the decrease of streamflow (Wittenberg, 1994; Moore, 1997), which is a strong indication of nonlinearity. The convenient assumption that the baseflow may be the outflow from two or more, parallel (i.e. independent) linear reservoirs, representing components of different response times is often made (Moore, 1997), and does result in better fits to the observed recession curves. However, this is perhaps only because there are more parameters to be calibrated, giving more degrees of freedom for curve fitting. In most catchments it is unlikely that the dynamic groundwater aquifer can be divided so neatly into such independent storage zones; more likely it consists of a spatially variable (including layered) system of hydraulically communicating pore or fissure systems.

Thus, the use of a single but nonlinear reservoir is considered to be more physically realistic. Nonlinear reservoir algorithms have been proposed and implemented in a large number of catchments around the world (Wittenberg, 1994; Wittenberg, 1999; Chapman, 1997; Brutsaert and Lopez, 1998), and are used in the present study. To allow for nonlinearity the linear storage–discharge relationship is generalised by adding an exponent b as follows:

$$S = a \cdot Q^b \tag{1}$$

For S in m^3 and Q in m^3/s the factor a has the dimension $m^{3-3b}\,s^b$. If the volumes are expressed in depth units (i.e. volume per unit area) and the time step is a day (d), then S is in mm, Q in mm/d and a will be in $mm^{1-b}\,d^b$. The exponent b is dimensionless. The linear reservoir is a special case of Eq. (1), i.e. when $b = 1$. Combining Eq. (1) with the continuity equation for a nonlinear groundwater reservoir without inflow, i.e. $dS/dt = -Q$, yields Eq. (2) for the recession curve starting at an initial discharge of Q_0, namely:

$$Q_t = Q_0\left(1 + \frac{(1-b)Q_0^{1-b}}{ab}t\right)^{1/(b-1)} \tag{2}$$

This corresponds to the expression found by Coutagne (1948). Its derivation is given in detail in Wittenberg (1999).

Given the streamflow recession data the parameter values a and b can be determined by an iterative least squares fitting method (Wittenberg, 1994). By systematically varying the parameter b, the value of parameter a is solved at each iteration step, with the condition that the computed outflow volume during the considered time period is equal to that of the observed recession curve. The set of a and b values providing the best fit to the observed curve is considered as representing the properties of the aquifer. Eq. (2) has only one optimal combination of a and b and no restriction was imposed on the value of b. When fitting Eq. (2) to recession data, in almost all cases no significant variation of the parameters a and b was found over different parts of the recession curve, unlike$k$ in the linear case, which has been shown to exhibit strong systematic variation (Wittenberg, 1994).

Analysis of recession curves obtained from time series by simply taking sequences of daily discharges at more than 80 gauging stations in Germany (Wittenberg, 1999) achieved close fits to Eq. (2). The average deviation from the curves, expressed in terms of the coefficient of variation (standard deviation divided by the mean, corresponding to the least squares criterion) was $\overline{CV} = 4.4\%$. The start of the baseflow recessions had been assumed not earlier than two time intervals (days) after the inflection point of the total hydrograph recession. The skewed distribution of the exponent b is peaking between 0.3 and 0.4 with a mean value of $b = 0.49 \approx 0.5$ and a standard deviation of 0.25 This empirically estimated mean value of 0.5 (i.e. discharge proportional to the square of storage) has also been obtained theoretically for the unconfined aquifers by other authors (Werner and Sundquist, 1951; Schoeller, 1962; Fukushima, 1988). Chapman (1997) found the values between 0.3 and 0.4 for 10 out of 11 rivers in Eastern Australia, and explained this in

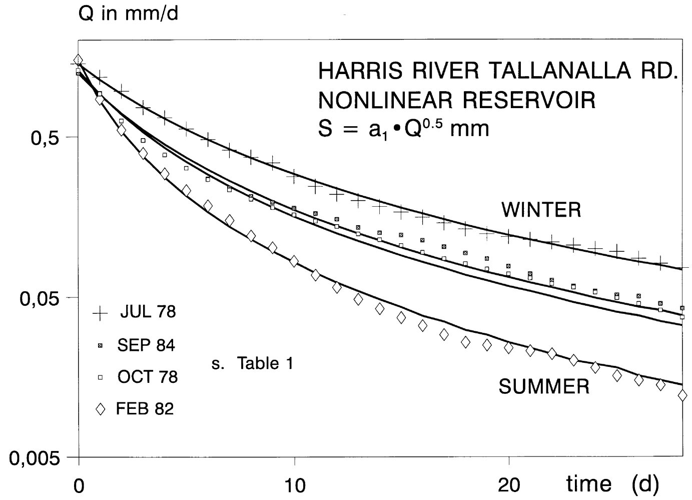

Fig. 2. Recession curve analysis and modelling by the nonlinear reservoir algorithm.

terms of the flow convergence in the groundwater system. Wittenberg (1994) showed for flow recessions of rivers in Germany and China that, on average, a value of 0.4 provided the best fit.

For most practical purposes, such as the regionalisation of the relationship given by Eq. (1), it seems reasonable to fix the exponent b at a mean or dominant value, and to allow the coefficient a to vary between catchments. A value of $b = 0.5$ is suggested here for this purpose. This is especially applicable to hillslope flow-strips (Kubota and Sivapalan, 1995) or to partial catchment areas as these are less subject to spatial heterogeneities. It is believed that even if the ''true'' value of b is not exactly reproduced, the assumption of $b = 0.5$ would be more physically realistic and would provide a better match to observed streamflows in a majority of river basins, than the linear reservoir. When fitting the model function with a fixed value of $b = 0.5$ to the German flow recessions an average variation coefficient of $\overline{CV} = 7.2\%$ was obtained instead of 4.4% for individual values of b (see above).

Table 1
Coefficients of the nonlinear reservoir, determined for flow recessions shown in Fig. 2

Month	$a(\text{mm d})^{1-b}$	b	CV (%)	a_1[a]	CV_1[a]
July 1978	9.5	0.53	3.8	9.9	4.1
September 1984	8.8	0.32	8.5	6.7	13
October 1978	7.1	0.38	5.2	6.2	9.6
February 1982	3.7	0.54	3.7	3.8	4.9

[a] For $b = 0.5$.

Wittenberg (1999) shows that the notional value of $b = 0.5$ for the unconfined aquifers is independent of the number of flow strips which make up the catchment. He also provides a discussion of the likely causes for the deviation of the field estimates of the exponent b (seen in many rivers) from the theoretical value of $b = 0.5$.

Normally a_1 (values of a for $b = 0.5$) will be determined from the observed flow recessions. Groundwater losses, e.g. by evapotranspiration, however, can have a biasing effect, lowering the computed a_1 so that this factor will no longer represent the true storage–discharge relationship of the aquifer.

2.1. Application to semi-arid catchment and seasonal variation of storage–discharge relationship

Fig. 2 shows four recession curves extracted for the different seasons from the observed daily flows (1976–1992) of Harris River (gauge at Tallanalla Road) located in the Collie basin in the southwest of Western Australia. The contributing catchment area at this location is 382 km^2, and is covered nearly entirely by native jarrah (*Eucalyptus marginata*) forest and shrubs. Data from the Collie basin have been

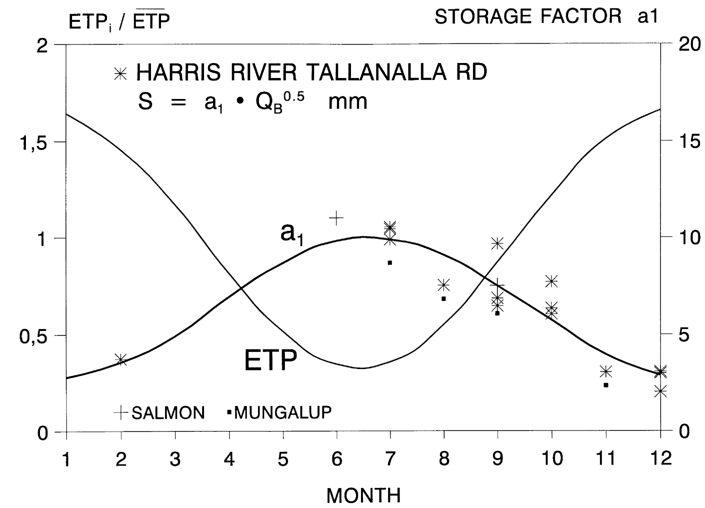

Fig. 3. Seasonal variation of parameter a_1 (dots represent values obtained for individual recessions), compared to variation of potential (pan) evaporation.

frequently used for research work and are considered of high quality. The gauging stations are equipped with V-notch measuring weirs. The coefficients a and b for the nonlinear reservoir equation (Eq. (2)) were calibrated as given in Table 1. The deviation between data and model function expressed by the coefficient of variation is in the range of a few percent. Estimates of the exponent b are reasonably close to 0.5. Assuming that their deviation from the theoretical value of 0.5 is due to random, unexplained spatial variability, the value of the exponent was fixed at $b = 0.5$ and the factor a_1 and the respective coefficient of variation $CV1$ determined again (Table 1). As shown in Fig. 2 the modelled recession curves based on the nonlinear reservoir with $b = 0.5$ (Eq. (2), lines) correspond well to the observed data (dots).

It is evident from Fig. 2 that, while the exponent $b = 0.5$ may be considered constant for all seasons, the

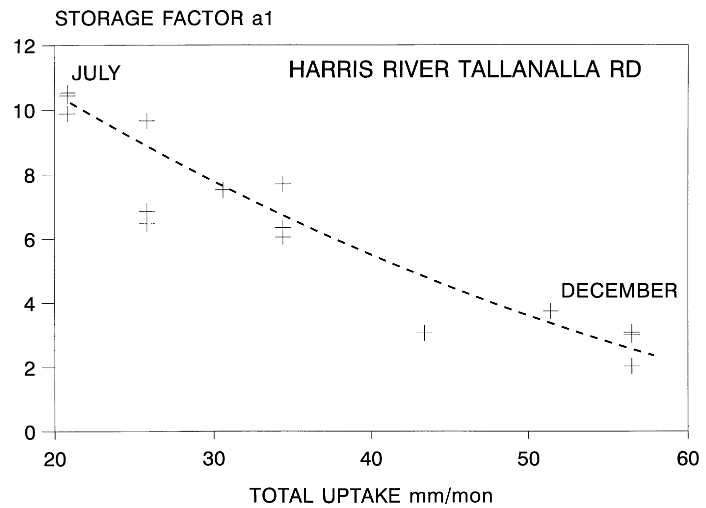

Fig. 4. Correlation of parameter a_1 to water uptake by trees (Marshall and Chester, 1992).

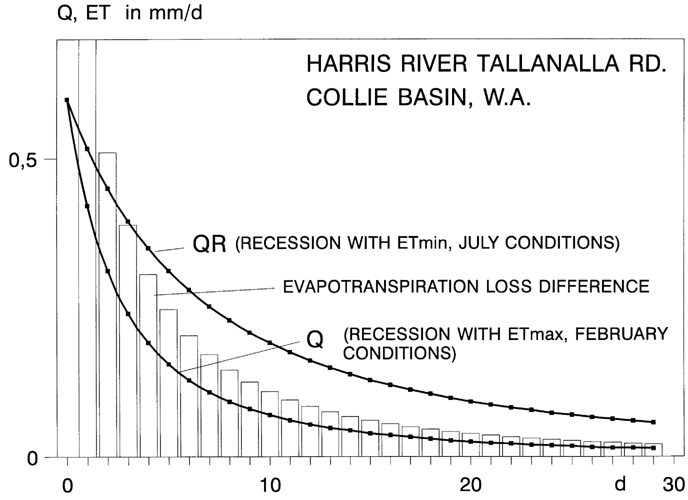

Fig. 5. Impact of groundwater evapotranspiration on flow recession.

actually determined coefficient a_1 has a significant seasonal variation, yielding a faster recession and a smaller value a_1 during summer, and a slower recession and a higher a_1 during winter. In Fig. 3 values of a_1, each of them empirically determined from a recession of 20–30 d length from different years, are plotted against the calendar month (dots) of their occurrence. There are 15 recessions analyzed for Harris River and additionally six for the small tributary Salmon River (0.82 km^2) and the Collie River at Mungalup Tower (2550 km^2), respectively. A tentative sinusoidal curve is fitted through these dots, showing the average seasonal variation of a_1.

As the storage–discharge relationship of the nonlinear reservoir (Eq. (1)) reflects the volumetric and hydraulic conditions of the groundwater aquifer, the true values $a = a_R$ and $b = 0.5$ are considered to be constants over time (as it is likewise assumed with the recession constant in linear reservoir models). Hence, the observed seasonal variation of the coefficient a_1 suggests that the baseflow is not the only outgoing water flux from the groundwater reservoir. A seasonally varying rate of evapotranspiration loss from the groundwater aquifer appears as the most probable and plausible cause for the changing steepness of the streamflow recession.

The mean seasonal variation of potential (pan) evapo-transpiration in the Collie region of southwest Western Australia is presented in Fig. 3 for comparison. The strong negative correlation of pan evaporation to the estimated variation of a_1 is clearly evident. A similar correlation is found for the values of a_1 with the average water consumption of trees in the respective calendar months which was estimated in the region by Marshall and Chester (1992) from measurements of water uptake in tree stems by independent heat pulse techniques, as shown in Fig. 4. Apart from some direct evaporation from areas with a saturated seepage face or having a shallow water table in the order of 10 m below the surface, much of the evapotranspiration losses in these parts can be attributed to the transpiration by deep-rooted (down to 35 m) eucalyptus trees. Indeed, the major overstorey species, *Eucalyptus marginata*, is able to maintain a substantial rate of transpiration throughout the arid summer despite considerable moisture deficit in the upper unsaturated soil profile (Carbon et al., 1980; Sivapalan et al., 1996).

3. Estimation of evapotranspiration losses from groundwater storage

The depletion of groundwater storage by evapotranspiration, or through fluxes other than baseflow, results in a biased streamflow recession curve which decreases at a faster rate than it would be expected with the ''true'' reservoir coefficient a_R. This is

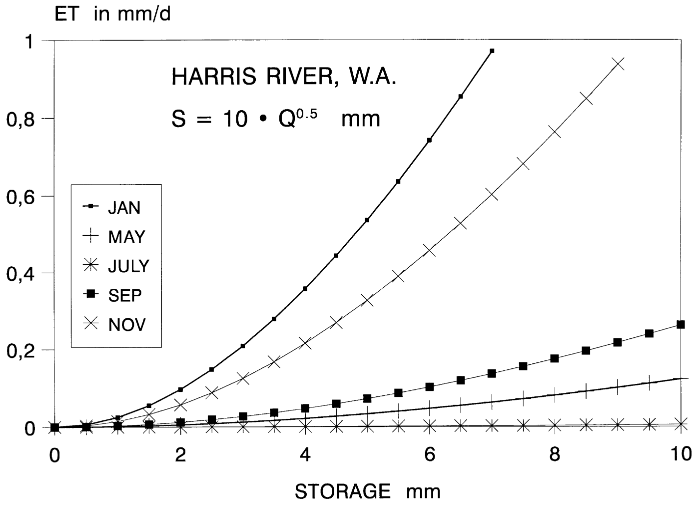

Fig. 6. Evapotranspiration loss from the groundwater as a function of season and storage, computed by Eq. (6).

demonstrated in Fig. 5 by two hypothetical recession curves for Harris River, starting from an arbitrary value Q_0. The upper recession QR would occur under winter conditions (July) which is subject to minimum losses, and here assumed to be zero, and Q under summer conditions (January) with maximum losses. For every time interval Δt evapotranspiration loss can be determined as the difference between the theoretical (i.e. potential) storage $SR = a_R \cdot QR^b$ which would have occurred at the end of the time interval with minimum evapotranspiration loss, and corresponding to theoretical baseflow discharge QR, and the actual storage $S = a_R \cdot Q^b$ (subject to increased losses).

Note that a_R, being the ''true'' unbiased reservoir coefficient, determines the true storage corresponding to outflow in any season. That is, $S = a_R Q^b$ or $SR = a_R QR^b$ are hydraulic-volumetric hence physical relationships for the reservoir. However, parameter a_1 is biased, and is smaller only because the additional evapotranspiration loss makes the baseflow recession curve steeper. Hence, if one wants to estimate this steeper recession curve, then a_1 must be used, as in Eq. (4).

In terms of a groundwater balance equation a preceding storage value S_{i-1} would become, after a time interval $\Delta t = 1$ d, at time i (i.e. on the ith day):

$$SR_i = S_{i-1} - \int_{i-1}^{i} Q\,dt \qquad \text{with only baseflow } Q$$

and

$$S_i = S_{i-1} - \int_{i-1}^{i} Q\,dt - \int_{i-1}^{i} ET\,dt$$

with base flow Q and evapotranspiration ET

For simplicity we define ET in terms of daily depth. Then the last equation becomes:

$$S_i = S_{i-1} - \int_{i-1}^{i} Q\,dt - ET_i$$

Combination of the aforementioned equations yields:

$$ET_i = SR_i - S_i = a_R QR_i^b - a_R Q_i^b \tag{3}$$

Note that both the terms on the right side relate to real (physical) storages, not biased ones. QR is discharge (during a recession) when ET is minimum, and Q is the discharge during recession which is influenced by evapotranspiration.

Starting from the preceding baseflow Q_{i-1}, the value QR_i is obtained according to Eq. (2) using the constant a_R (minimum ''no'' losses, June–July), while Q_i is computed with a_1 (increased evaporation

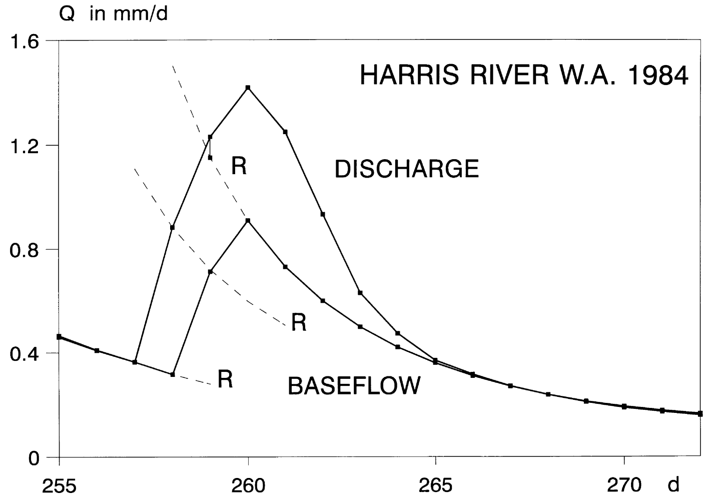

Fig. 7. Construction of transition curve between two recession curves, R: recession curves computed by Eq. 2.

losses). Eq. (3) thus becomes:

$$ET_i = a_R Q_{i-1}^b \left(1 + \frac{(1-b)Q_{i-1}^{1-b}}{a_R b}\right)^{b/(b-1)} - a_R Q_{i-1}^b \left(1 + \frac{(1-b)Q_{i-1}^{1-b}}{a_1 b}\right)^{b/(b-1)} \quad (4)$$

or, for $b = 0.5$,

$$ET_i = a_R Q_{i-1}^{0.5}\left[\left(1 + \frac{Q_{i-1}^{0.5}}{a_R}\right)^{-1} - \left(1 + \frac{Q_{i-1}^{0.5}}{a_1}\right)^{-1}\right] \quad (5)$$

Replacing $Q_{i-1}^{0.5} = S_{i-1}/a_R$ yields Eq. (6) which shows clearly that evapotranspiration losses from the groundwater depend on season via the factor a_1 and groundwater volume S which is related to groundwater depth:

$$ET_i = S_{i-1}\left(\frac{1}{1 + S_{i-1}/a_R^2} - \frac{1}{1 + S_{i-1}/a_R a_1}\right) \quad (6)$$

Fig. 6 shows the relationships between evapotranspiration loss and storage depth of the groundwater computed by Eq. (6) using the average values of a_1 for different months of the year as given by the sinusoidal curve in Fig. 3. In their general shape and order of magnitude these curves are comparable to the rates of total groundwater transpiration by phreatophyte shrubs in Nevada (24–50 mm/y), estimated by Nichols (1994) using a energy budget approach. The curves shown in Fig. 6 are extrapolated. Most computed daily depths of evapotranspiration loss from the groundwater for Harris River (1976–92) are in the lower part of the diagram, in the range of 0–0.25 mm/d (compare Fig. 8). It should be noted that the rates of evapotranspiration shown here represent the losses from the groundwater reservoir over and above the unknown minimum loss implied for winter conditions (for which we have assumed $a_1 = a_R$ during the computation of recession).

4. Inverse modelling through baseflow separation

The nonlinear reservoir algorithm was also applied for the separation of baseflow from time series of total daily streamflow. The procedure and application has been amply described by Wittenberg (1999), and will not be repeated here. The computation starts at the last value of the time series and proceeds backwards along the time axis. A flow recession at the time $t - \Delta t$ is determined from the flow at the time t using Eq. (7), which has been derived by inverting Eq. (2). The time step Δt is normally one day.

$$Q_{t-\Delta t} = \left(Q_t^{b-1} + \frac{\Delta t(b-1)}{ab}\right)^{1/(b-1)} \quad (7)$$

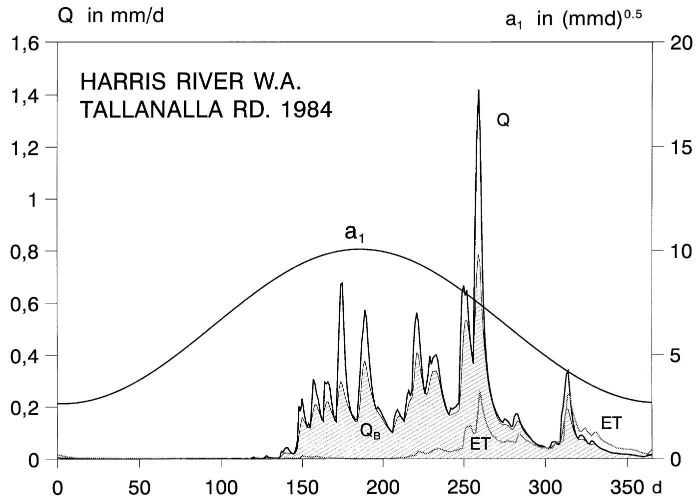

Fig. 8. Baseflow separation and evapotranspiration loss.

Recession periods of the flow hydrographs are disrupted by the recharge periods. Baseflow will then rise and we need to develop a scheme to connect the preceding lower baseflow values to the higher baseflow values which follow after the new storm event recharges the groundwater reservoir. Recharge volume and the duration of recharge can be determined from the difference between the aforementioned recessions.

As recharge is usually coincident with the rising and peaking of total flow, the following approach was adopted (Wittenberg, 1999). When the reverse computed baseflow recession curve intersects the rising limb of the total hydrograph (Fig. 7), a transition point which is at the next time step forward from the total flow is adopted as the peak of baseflow. Values of the rising limb of the baseflow hydrograph are then found as the computed recession curve for one time step forward for each given total flow value. This procedure is similar to the digital filter described by Chapman (1997) for baseflow separation for the linear reservoir.

From the thus obtained continuous time series of daily baseflow Q, effective groundwater recharge is computed for every time step as follows:

$$GWR_i = S_i - S_{i-1} + \int_{ti-1}^{ti} Q\,\mathrm{d}t + ET_i \qquad (8)$$

where S is the actual storage computed by Eq. (1) using the unbiased storage factor a_R. For practical computation the baseflow volume during this time interval is determined by the trapezoidal formula, thus $\int Q\,\mathrm{d}t \approx \Delta t(Q_{i-1} + Q_i)/2$. Evapotranspiration losses (ET) from the groundwater are computed using Eq. (6) with daily values of a_1. It is evident that during flow recession groundwater recharge (GWR) is zero.

A FORTRAN program, BNLP, was written for the automatic separation of baseflow from long time series of daily discharge using the algorithm described earlier. Fig. 8 shows the results for the year 1984 for the Harris River in the Collie River Basin in southwest Western Australia. Evapotranspiration loss from groundwater is low in the first part of the year as groundwater storage in the dynamic saturated zone is low. With higher storage volume and higher potential evapotranspiration and water uptake by trees, as reflected in decreasing values of a_1, this loss becomes significant.

In the time period 1976–1992, the average annual runoff depth for Harris River was 54 mm, varying between 12 and 141 mm. Average annual baseflow was estimated to be 40 mm (8–106 mm), and groundwater recharge as 50 mm (8–105 mm). The average evapotranspiration loss (above minimum winter loss) from the groundwater was computed to be 10 mm (2–22 mm). Fig. 9 shows estimates of the monthly water balance components for the years 1984–86.

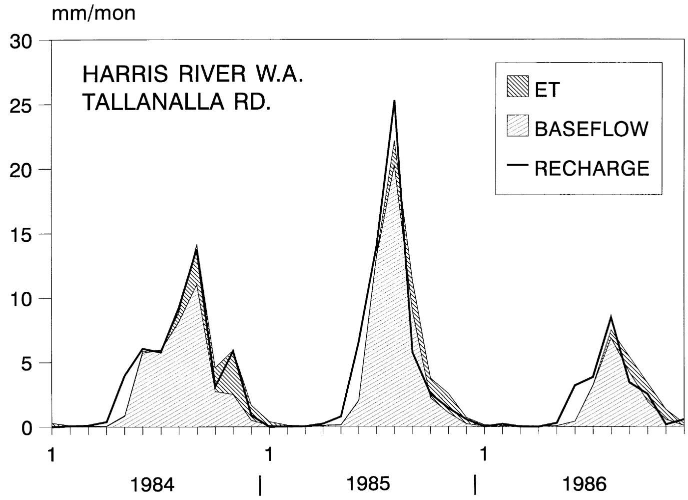

Fig. 9. Monthly groundwater fluxes.

5. Recharge response to rainfall

The increased flexibility of the nonlinear reservoir produces estimates of baseflow closer to the observed total discharge hydrograph that can be achieved using the one-parameter linear reservoir algorithm—this also leads to higher values of storage and recharge being obtained. In fact, it has been shown by measurement of concentration of tracers (Herrmann, 1997) such as ^{18}O (Chapman and Maxwell, 1996) and salt (Stokes, 1985), that even in flood periods discharge from shallow groundwaters is the major contributor to streamflow. The relatively fast response of the shallow groundwaters is due to an increase of hydraulic head caused by quick recharge. Macropore infiltration and preferential flows have been suggested to play an important role in this response (Johnston, 1987; Bronstert and Plate, 1997).

Groundwater recharge hydrographs computed from separated baseflow for the Harris River are shown in

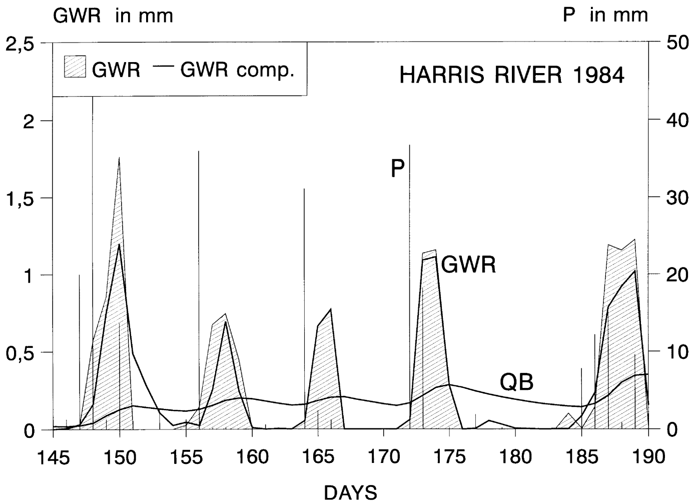

Fig. 10. Groundwater recharge (*GWR*) determined from baseflow *QB*, and computed from rainfall *P*.

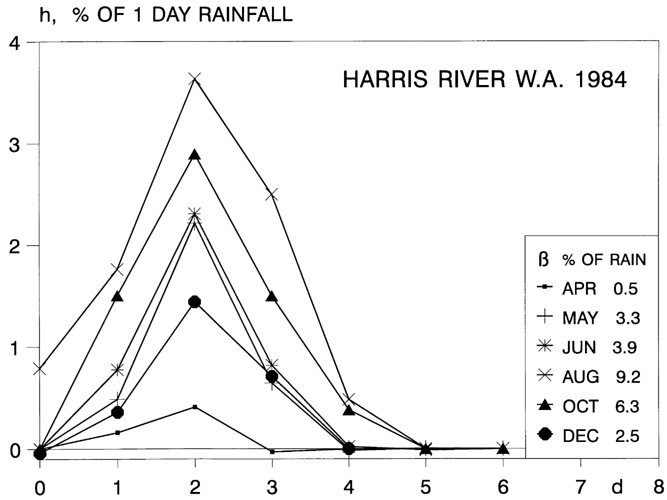

Fig. 11. Response functions of groundwater recharge to one day's rainfall.

Figs. 1 and 10. The comparison with daily rainfall depths at station M509082 in that catchment shows the evident relationship to rainfall, and suggests the feasibility of the application of a simple transfer function to transform effective infiltrating rainfall into recharge.

As every rainfall impulse appears to produce a similar response, differing of course in magnitude, it appears reasonable to apply a linear unit response function of the unit hydrograph type. Linear response functions to estimate recharge have been derived and applied by Besbes and de Marsily (1984) and Wu et al. (1997) using theoretical, bottom-up, approaches based on soil properties, and verified by well and lysimeter observations.

The assumption of linearity for the transfer of infiltrated rainwater, through the unsaturated vadose zone to the groundwater table, appears reasonable because the conductivity of that zone can be assumed not to vary significantly with time when water percolates through it (Besbes and de Marsily, 1984). For the same reason, time invariance can also be expected. Under these two conditions, recharge *GWR* from infiltrating rainfall *I(t)* can be estimated by the application of the convolution integral:

$$GWR_t = \int_0^t (I_{t-\tau} \cdot h_\tau)\, \mathrm{d}\tau \tag{9}$$

where h is the unit response function, which is defined as the theoretical recharge hydrograph which would occur for 1 mm of (effective) rainfall percolating through the groundwater surface. For practical computations, with digital data of a time interval Δt, the convolution integral becomes:

$$GWR_i = \sum_{k=1}^{i} (I_{i-k+1} \cdot h_k)\Delta t \tag{10}$$

where *GWR* and *I* are in mm. In this study effective rainfall *I* has been assumed proportional to measured rainfall throughout each recharge event. As the time interval for computation is $\Delta t = 1$ d, the response function h in Eq. (10) represents a travel time distribution in d^{-1}. For every sequence of n_i values of effective rainfall I there is a corresponding sequence of n values of recharge *GWR*, which could be computed by convolution, i.e. multiplication of the response function h with every value I and time-shifted superposition of the estimated recharge hydrographs. The length or number n_h of values of the response function h is thus $n_h = n - n_i + 1$. Eq. (10) thus represents a system of n linear equations with $n_h (< n)$ unknowns h, which can be resolved by the least squares method (Snyder, 1955). A least squares computer programme designed for traditional unit hydrograph computation has been applied in this study.

Fig. 11 shows response functions computed for rainfall-recharge events in different months of the year 1984 for Harris River. The functions are given in % of a day's rainfall or in 1/100 of mm for a unit event of 1 mm depth. The strong seasonal influence on the response functions, corresponding to the recharge-rainfall ratio β of each event, is evident. Hydrographs of groundwater recharge recomputed by convolution of the response function with measured rainfall, are given in Fig. 10 (lines) for comparison with recharges obtained from baseflow separation (areas).

The shape of the determined functions however, is very similar. The travel time distribution through the vadose zone thus appears rather time invariant not only within the events but also over all seasons. Peak recharge is reached at the second day after the rainfall event and recharge ends after 5–6 d. The similarity of the response functions allows the derivation of a typical mean unit response function. After dividing each response by their volume the mean unit response is obtained by a simple averaging of these normalized curves.

These results are confirmed by Johnston (1987) based on piezometric observations in the experimental Salmon catchment located near Harris catchment which is under study here. He found ''recharge reaching the water table within 12–14 h of the start of the rainfall'' of one day and a dissipation of the groundwater mound, i.e. the cessation of recharge, ''within a further four days''.

Though principally of similar appearance, the recharge response functions derived by Besbes and de Marsily (1984) for the upper Lys River in Northern France and by Wu et al. (1997) from lysimeters, are considerably slower and longer than those obtained here by the baseflow analysis. This can be partially explained by the very high transmissivity of the lateritic soils, and especially the effectiveness of preferred path ways and macropores in the Harris River catchment causing much shorter travel times than would be obtained for similar apparent transmissivities assuming Darcian behaviour.

Concerning the recharge functions obtained in this study, the shape will be influenced by baseflow modelling during the recharge phase. The length is restricted by the basic assumption of baseflow separation that there is no further recharge when the typical recession starts. The comparatively long recession of the recharge functions derived in other studies, however, may be partially due to the adoption of the cascade of linear reservoirs (three-parameter γ function) as a model function for which the long tail is to be expected.

6. Conclusions

The groundwater balance of a catchment and the processes of recharge, storage, evapotranspiration loss and discharge can be described by simple but physically based conceptual model components. The properties of these components can be identified and obtained from streamflow data. Observed streamflow data and especially flow recessions are considered as a very authentic database for a catchment, carrying a wealth of information about the foregoing hydrological processes. The decoding of some of this information was the main aim of this work. The results are backed by the field observations (Johnston, 1987).

The nonlinearity of the storage–discharge relationships found in previous studies to dominate groundwater outflow in rivers was also identified in the Collie River Basin in Western Australia. Depletion of the groundwater aquifer by evapotranspiration losses, however, biases the observed flow recession curves, depending on the storage, vegetation and potential evapotranspiration. Though these losses are known, and acknowledged in the literature (Tallaksen, 1995), in the past it was rarely considered in the recession analysis; as shown in this article, baseflow recession analyses also permit their quantification. Further studies comparing the estimated fluxes with those found by the evaluation of water uptake of deep-rooted trees should be carried out to confirm the accuracy of these estimates.

By including evapotranspiration flux in baseflow separation techniques, hydrographs of recharge to the aquifer were computed by inverse nonlinear flow routing. Linear time-invariant unit response functions were identified between the measured rainfall and the recharge hydrographs estimated by baseflow separation. These compare reasonably well to response functions estimated by other authors based on theoretical, bottom-up, approaches, and by making inferences based on lysimeter measurements.

Acknowledgements

This work was performed during the sabbatical visit of the first author at the Centre for Water Research, University of Western Australia. The stay was funded by the Deutsche Forschungsgemeinschaft, Germany. The first author is pleased to acknowledge this financial support, as well as the assistance and support provided by the Centre for Water Research. CWR Reference ED 1381 MS.

References

Besbes, M., de Marsily, G., 1984. From infiltration to recharge: use of a parametric transfer function. J. Hydrol. 74, 271–293.

Brutsaert, W.H., Lopez, J.P., 1998. Basin-scale geohydrologic drought flow features of riparian aquifers in the southern Great Plains. Water Resour. Res. 34 (2), 233–240.

Bronstert, A., Plate, E., 1997. Modelling of runoff generation and soil moisture dynamics for hillslopes and micro-catchments. J. Hydrol. 198, 177–195.

Carbon, B.A., Bartle, G.A., Murray, A.M., MacPherson, D.K., 1980. The distribution of root length, and the limits to flow of soil water to roots in a dry sclerophyll forest. Forest Sci. 4, 656–664.

Caro, R., Eagleson, P.S., 1981. Estimating aquifer recharge due to rainfall. J. Hydrol. 53, 185–211.

Chapman, T., 1997. A comparison of algorithms for streamflow recession and baseflow separation. In: McDonald, A.D., McAleer, M. (Eds.), Proceedings, International Congress on Modelling and Simulation MODSIM 97. Hobart, Tasmania, Australia, pp. 294–299.

Chapman, T., Maxwell, A., 1996. Baseflow separation—comparison of numerical methods with tracer experiments. Proceedings of the 23rd Hydrology and Water Resources Symposium. Hobart, Australia, pp. 539–545.

Coutagne, A., 1948. Etude générale des variations de débits en fonction des facteurs qui les conditionnent, 2ème partie: Les variations de débit en période non influencée par les précipitations, La Houille Blanche, Septembre–Octobre, pp. 416–436.

Fröhlich, K., Fröhlich, W., Wittenberg, H., 1994. Determination of groundwater recharge by baseflow separation—regional analysis in North–East China. IAHS Publ. No. 221, pp. 69–75.

Fukushima, Y., 1988. A model of river flow forecasting for a small forested mountain catchment. Hydrol. Process. 2, 167–185.

Herrmann, A., 1997. Global review of isotope hydrological investigations. In: Oberlin, G., Desbos, E. (Eds.), FRIEND third Report, UNESCO, Paris.

Johnston, C.D., 1987. Preferred water flow and localised recharge in a variable regolith. J. Hydrol. 94, 129–142.

Kubota, J., Sivapalan, M., 1995. Towards a catchment-scale model of subsurface runoff generation based on synthesis of small-scale process-based modelling and field studies. Hydrol. Process. 9, 541–554.

Maillet, E., 1905. Essai d'hydraulique souterraine et fluviale: Librairie scientifique. Hermann, Paris.

Marshall, J.K., Chester, G.W., 1992. Effect of forest thinning on jarrah (*Eucalyptus marginata*) water uptake. CSIRO, Division of Water Resource Report No. 92/24.

Moore, R.D., 1997. Storage-outflow modelling of streamflow recessions, with application to a shallow-soil forested catchment. J. Hydrol. 198, 260–270.

Nathan, R.J., McMahon, T.A., 1990. Evaluation of automated techniques for baseflow and recession analysis. Water Resour. Res. 26, 1465–1473.

Nichols, W., 1994. Groundwater discharge by phreatophyte shrubs in the Great Basin as related to depth to groundwater. Water Resour. Res. 30, 3265–3274.

Renger, M., Strebel, O., Wessolek, G., Duynisveld, W., 1986. Evapotranspiration and groundwater recharge—a case study for different climate, crop patterns, soil properties and groundwater depth conditions. Z. Pflanzenernaehr. Bodenk 149, 371–381.

Schoeller, H., 1962. Les eaux souterraines. Masson, Paris.

Sivapalan, M., Ruprecht, J., Viney, N., 1996. Water and salt balance modelling to predict the effects of land-use changes in forested catchments: 1. small catchment water balance model. Hydrol. Process. 10, 391–411.

Snyder, W.M., 1955. Hydrograph analysis by the method of least squares. J. Hydraul. Div., Amer. Soc. Civ. Engrs. 81, 793.

Stokes, R.A., 1985. Stream water and chloride generation in a small forested catchment in South Western Australia. Water Authority of Western Australia, Perth, Report No. WH7, p. 176.

Tallaksen, L.M., 1995. A review of baseflow recession analysis. J. Hydrol., 349–370.

Werner, P.W., Sundquist, K.J., 1951. On the groundwater recession curve for large watersheds. IAHS General Assembly, Brussels, IAHS Publ. 33, pp. 202–212.

Wittenberg, H., 1994. Nonlinear analysis of flow recession curves. IAHS Publ. 221, pp. 61–67.

Wittenberg, H., 1999. Baseflow recession and recharge as nonlinear storage processes. Hydrol. Process., Special Issue Process Interactions in the Environment.

Wu, J., Zhang, R., Yang, J., 1997. Estimating infiltration recharge using a response function model. J. Hydrol. 198, 124–139.