Accepted Manuscript

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PII: S0022-1694(14)00126-7
DOI: http://dx.doi.org/10.1016/j.jhydrol.2014.02.019
Reference: HYDROL 19411

To appear in: Journal of Hydrology

Received Date: 9 August 2013
Revised Date: 15 January 2014
Accepted Date: 4 February 2014


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Characterisation and modelling of conduit restricted karst aquifers – example of the Auja spring, Jordan Valley

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Abstract

The conduit system of mature karstified carbonate aquifers is typically characterised by a high hydraulic conductivity and does not impose a major flow constriction on catchment discharge. As a result, discharge at karst springs is usually flashy and displays pronounced peaks following recharge events. In contrast, some karst springs reported in literature display a discharge maximum, attributed to reaching the finite discharge capacity of the conduit system (flow threshold). This phenomenon also often leads to a non-standard recession behaviour, a so-called “convex recession”, i.e. an increase in the recession coefficient during flow recession, which in turn might be used as an indicator for conduit restricted aquifers. The main objective of the study is the characterisation and modelling of those hydrogeologically challenging aquifers. The applied approach consists of a combination of hydrometric monitoring, a spring hydrograph recession and event analysis, as well as the setup and calibration of a non-linear reservoir model. It is demonstrated for the Auja spring, the largest freshwater spring in the Lower Jordan Valley. The semi-arid environment with its short but intensive precipitation events and an extended dry season leads to sharp input signals and undisturbed recession periods. The spring displays complex recession behaviour, exhibiting exponential (coefficient $\alpha$) and linear (coefficient $\beta$) recession periods. Numerous different recession coefficients $\alpha$ were observed: $\sim0.2–0.8 \text{ d}^{-1}$ (presumably main conduit system), $0.004 \text{ d}^{-1}$ (fractured matrix), $0.0009 \text{ d}^{-1}$ (plateau caused by flow threshold being exceeded), plus many intermediate values. The reasons for this observed behaviour are the outflow threshold at $0.47 \text{ m}^3 \text{ s}^{-1}$ and a variable conduit–matrix cross-flow in the aquifer. Despite system complexity, and hence the necessity of incorporating features such as a flow threshold, conduit–matrix cross-flow, and a spatially variable soil/epikarst field capacity, the developed reservoir model is regarded as relatively simplistic. As a number of required parameters were calculated from the hydrogeological analysis of the system, it requires only six
calibration parameters and performs well for the highly variable flow conditions observed. Calculated groundwater recharge in this semi-arid environment displays high interannual variability. For example, during the 45-year simulation period, only five wet winter seasons account for 33% of the total cumulative groundwater recharge.

Keywords

Spring hydrograph analysis; Flow threshold; Convex recession; Linear recession; Reservoir model; Jordan Valley

1 Introduction

At the catchment scale, the hydraulics of karstified carbonate aquifers are controlled by a network of highly permeable flow features (i.e. karst shafts, karst conduits) embedded in a less permeable fractured rock matrix (e.g. Atkinson, 1977; Király, 2002). The karst conduit system is generally believed to be the result of the coupled process of groundwater flow and carbonate dissolution (e.g. Klimchouk et al., 2000; Liedl et al., 2003). The spatial aquifer structure is furthermore influenced by landscape development and climate conditions during aquifer evolution (i.e. paleo-base level and paleo-climate). Due to the large contrast in the hydraulic parameter field in karst aquifers and the presence of turbulent flow conditions in the highly permeable structures, the characterisation and modelling of those systems is very complex (Király, 2002; Goldscheider and Drew, 2007).

For groundwater resources assessment and management, especially in the more arid environments, prerequisites are the quantification of groundwater recharge and the prediction of
groundwater resources, e.g. during periods of drought (Simmers, 1990; Tallaksen and van Lanen, 2004). Groundwater resources prediction usually requires the application of a mathematical model. Linear reservoir models do not imply detailed knowledge about catchment boundaries and internal aquifer structure and are widely applied for the simulation of karst water resources (e.g. Rimmer and Salingar, 2006; Fleury et al., 2007; Geyer et al., 2008; Hartmann et al., 2012a).

Since flow in karstified carbonate aquifers frequently converges to a distinct number of springs, the time series of spring discharge (spring hydrograph) provides integral information about the whole groundwater catchment of the spring. However, the hydrograph is the result of spatially and temporally diverse and superimposed processes (recharge, storage, and flow; Smart and Hobbs, 1986) within the system and therefore the relative contribution of the different processes is often difficult to deconvolute. According to the presence of integrated conduit networks and hence karst system maturity, karst aquifers have been classified into diffuse flow and conduit flow (free flow) end members (White, 1969; Atkinson, 1977). For mature karst aquifers, spring discharge displays a fast response to precipitation events in the recharge area, i.e. a sharply rising limb, an often sharp peak, followed by a strong flood recession and subsequently a more moderate baseflow recession (e.g. Atkinson, 1977; Geyer et al., 2008). Displayed as a semi-logarithmic graph, the recession part of the hydrograph can often be described by a sequence of three or more decreasing recession coefficients (Forkasiewicz and Paloc, 1967; Milanovic, 1981; Sauter, 1992). The above authors attributed each recession coefficient to the physical properties of a distinct aquifer section, i.e. conduit system, “intermediate system”, and matrix. However, e.g. Smart (1983) and Eisenlohr et al. (1997) showed that intermediate values can also be the result of transient flow conditions.

In contrast to the standard recession behaviour, a number of karst spring hydrographs reported in literature display a shoulder in the recession part of the hydrograph, a so called “convex recession” (Smart, 1983; for further examples see below). Different reasons for such discharge
behaviour have been discussed (Fig. 1). The limited water conveying capacity of the conduit system can lead to a plateau in discharge, regardless of excess groundwater recharge or high aquifer storage (Fig. 1a, Smart, 1983; Bonacci, 2001; Herman et al., 2008). The surplus in groundwater can either contribute to aquifer storage or be discharged by an overflow mechanism within the catchment, e.g. an intermittently active overflow spring (Smart, 1983; Bonacci, 2001; Barberá and Andreo, 2012). Alternatively, convex recession behaviour is attributed to an upgradient reservoir of water, such as infiltration from a flooded Polje (Bonacci, 1993, Fig. 1b), infiltration from an alluvial aquifer drained by the karst aquifer (Soulios, 1991) or from high porosity sections within the karst aquifer (“cave”, Bonacci, 1993; “constricted reservoirs”, Wong et al., 2012).

The interaction and water exchange between the conduits and the fractured matrix under variable hydraulic conditions is increasingly considered in karst hydrogeology (e.g. Bauer et al., 2003; De Rooij, 2008; Maréchal et al., 2008; Bailly-Comte et al., 2010; Reimann et al., 2013). Kovacs et al. (2005) distinguished a matrix restricted and a conduit influenced flow regime. Bailly-Comte et al. (2010) showed that the drainage of a conduit system without matrix contribution may produce a linear recession hydrograph (Fig. 1c), similar to the outflow of a reservoir of infinite hydraulic conductivity. At the point where the flow regime transition from conduit flow to matrix restricted flow is taking place, the recession becomes distinctively exponential (Fig. 1c). This way, the shape of the hydrograph displays a shoulder, too. In sum, a convex hydrograph recession (as displayed on linear axes) can indicate conduit restricted discharge behaviour. If spring discharge at the convex break point is close to the maximum observed discharge, conduit restricted behaviour can be assumed.

The main objective of the study is the characterisation and modelling of conduit restricted karst aquifers. Those aquifers are expected to be challenging regarding the conceptual and mathematical modelling approach. A combination of hydrometric monitoring, hydrogeological
data analysis, and minimally parameterised modelling was applied. To obtain unambiguous
aquifer parameters, a karst aquifer with low signal convolution (e.g. displaying a sharp recharge
input signal and a long recession period) was preferred, which could be expected for semi-arid
environments with short and intensive precipitation events and an extended dry season. The
approach was executed at the example of Auja spring, the largest fresh water spring in the West
Bank (Nuseibeh and Nasser Eddin, 1995). According to the shape of the hydrograph, a
pronounced outflow restriction could be assumed. Long-term hydrometric time series were
available. In addition, within the context of this study, several high-resolution monitoring stations
were installed.

2 Case study area

2.1 General hydrogeologic setting

The Jordan Valley is part of an endorheic basin with the Dead Sea as terminal lake that developed
along the Dead Sea transform (e.g. Garfunkel and Ben-Avraham, 1996). In the southern part of
the western margin of the Lower Jordan Valley, eight larger karst springs ($Q_{\text{mean}}$: 0.02–0.3 m$^3$ s$^{-1}$)
emerge from a ca. 800 m thick Cretaceous carbonate aquifer complex (Fig. 2). The springs
discharge from its upper sections (Rosenthal and Kronfeld, 1982; Guttman, 2007), constituted by
Cenomanian and Turonian dolomite and limestone. Towards the east, the aquifer complex is
covered by Coniacian to Maastrichtian sediments, predominantly chalk, which constitute an
aquiclude. The lateral boundaries of the spring catchments are unknown. According to stable
water isotope data (Kroitoru et al., 1985; isotope data e.g. Gat and Dansgaard, 1972; Ayalon et al.,
2004), the recharge area of the springs is located in the highland region along the surface water
divide (Fig. 2). The divide is located at a topographic elevation of ca. 800 m ASL, reaching up to
1000 m ASL. Recharge is predominantly autogenic, because the Coniacian–Maastrichtian cover is only present in the Southeast (Fig. 2). The long-term mean recharge fractions for the catchments of individual springs were estimated to be between 25% and 50% of the precipitation, based on a chloride mass balance (Schmidt et al., 2013).

2.2 Characteristics of Auja spring

Auja spring (also sometimes written Uja spring) is located at the banks of an ephemeral stream valley (Wadi Auja, Fig. 3). The outflow is located in alluvial sediments, which cover the wadi bottom. Spring water flows in the natural wadi course for about 750 m, after which it is captured and conveyed in a concrete channel to Auja village. Mean spring discharge was about 0.3 m$^3$ s$^{-1}$ during the hydrological years 1968–2000 (Palestinian Water Authority, 2000). The most prominent systematic feature of the Auja spring hydrograph is a plateau during discharge maximum. Usually, spring discharge does not exceed ca. 0.5 m$^3$ s$^{-1}$.

Auja spring is normally a perennial spring, however, it has been occasionally observed to dry up. For example, dry periods were reported for the consecutive autumn seasons of 1933–1936 (Department of Land Settlement and Water Commissioner, 1947), presumably because of a precedent low precipitation period (data in Rosenan, 1955). During the period of 1967–2000, spring discharge ceased in seven autumn/winter seasons: 1979, 1986, 1987, 1990, 1991, 1999, and 2000 (Palestinian Water Authority data). Also, recent years are characterised by a frequently observed dry-up in summer/autumn. Despite these documented observations, even some months after a considerable flow stopped, water is still observable in the spring tapping structure and abundant green vegetation can be found in the wadi for some hundred meters downstream of the spring. Those observations indicate continued low discharge within the thin wadi alluvial sediments. The response of the spring discharge following intensive precipitation events is rapid.
Lag time between the onset of precipitation in the highland area and an increase in spring discharge is less than one day (Schmidt et al., 2012).

2.3 Karst aquifer development and paleo-base levels in the Lower Jordan Valley region

Karst aquifer development started in the study area probably during the late Eocene (Frumkin and Fischhendler, 2005). With the subsidence of the Jordan Valley since the late Miocene, the steep incision of the valleys took place (Frumkin and Fischhendler, 2005). During the Pleistocene, the base level in the Dead Sea catchment area was very unstable due to the frequently changing level of the water bodies in the basin (e.g. Waldmann et al., 2009). Kafri and Yechieli (2010) determined paleo-base levels around −210 m, −90 m, and −30 m ASL by cave mapping in the western rift escarpment. Those levels match the elevation of the springs in the area: Sultan (−215 m), Duyuk and Fasayil (−110 m), Auja (+20 m), and Qilt (+10 m), which display either a relatively stable discharge or a discharge plateau (data courtesy of the Palestinian Water Authority). Lisker et al. (2009) provided evidence that the level of Lake Lisan ranged in a highstand elevation of −220 m to −190 m (corresponding to the elevation of the Sultan spring) between 40 ka to 17 ka before present. The two higher horizons are possibly related to paleo-base levels from the Pliocene or early Pleistocene (Kafri and Yechieli, 2010). Here it may be speculated that the Auja spring and other springs are reactivated remnants of those paleo-base levels, which were rapidly undercut by a deeper system. The flow system which adjusted to this base level could have been subsequently blocked by the fine grained sediments of the Lake Lisan (see Kafri and Yechieli, 2010).

It is assumed that the flow restriction is located in the terminal part of the flow system due to the frequent change of base level, resulting in a limited time for carbonate dissolution. Moreover, the
dolomite lithology of large parts of the aquifer and frequent thin intercalations of marl and chalk (Begin, 1975) might cause a relatively slow karstification.

3 Data collection and analysis

3.1 Spring discharge

The discharge of Auja spring was measured at monthly intervals during the hydrological years of 1968–2000 by the West Bank Water Department (Palestinian Water Authority, 2000). Measurements, e.g. by current meter, were conducted in the concrete channel conveying the spring water to Auja village (personal communication, M. Nuseibeh). These time series cover the broad range of hydrological conditions characteristic for the semi-arid environment. Additionally, a permanent V-notch weir was constructed in 2008 directly at the spring, to interpret discharge dynamics with a high accuracy and a high temporal resolution (Fig. 3; Schmidt et al., 2012). A rating curve was established by a number of discharge measurements (salt dilution method) covering the whole range of flow conditions. Stage was automatically recorded every 5 to 15 minutes (measurement devices: MPS-D and Dipper-TEC, SEBA Hydrometrie GmbH; Micro-Diver, Schlumberger Water Services). The measurements of stage are influenced by recreational activities at the spring (e.g. bathing in the approach channel) and need to be corrected manually before discharge can be computed. The processed discharge hydrograph is displayed in Fig. 4.

3.2 Recession and event analysis of the Auja spring hydrograph

Recently continuously recorded discharge and long-term point data of 1967–2000 provide the basis for a recession analysis using Maillet (1905): \( Q_t = Q_0 e^{-\alpha t} \), with \( Q_0 \) the discharge at the start...
of recession segment \((m^3 \text{ s}^{-1})\) and \(Q_t\) the discharge at time \(t\) (d). Recession coefficients of the long-term data were evaluated by analysing the parts of the hydrograph between the last major precipitation event of the preceding winter season and the first larger event of the succeeding season to avoid the influence of recharge events on the recession hydrograph. The hydrographs were checked for extended straight-line segments (length usually 2–6 months). The histogram of the data displayed distinct clusters and the data were grouped accordingly (Table 1). The high temporal resolution of the newly obtained hydrograph permitted a detailed determination of recession coefficients (Fig. 4 and Table 2).

Above a flow threshold at ca. 0.47 \(m^3 \text{ s}^{-1}\) (range: 0.46–0.48 \(m^3 \text{ s}^{-1}\), period 2009–2012), a slightly inclined discharge plateau was observed. For the plateau time period, recession displayed a median value of 0.0009 \(d^{-1}\) (Table 1). After the large recharge event of February/March 2012, an initial peak at 0.53 \(m^3 \text{ s}^{-1}\) superimposed on the plateau was observed, displaying a comparatively rapid recession coefficient of about 0.007 \(d^{-1}\) (Figs. 4 and 5). Discharge returned to the plateau value (at this time about 0.51 \(m^3 \text{ s}^{-1}\)) within seven days after the peak. Below the threshold, different recession coefficients were often derived for individual years (Table 1). The various recession coefficients are attributed to a variable hydraulic head and stored groundwater volume in different aquifer compartments. Conceptually, two reservoirs are postulated: a highly permeable reservoir, comprising the conduit system and other highly permeable features (fractures, karst voids, caves, etc.) and a low permeability reservoir, comprising a rock matrix with small-scale fractures. Figure 5 displays sketches for the different recession behaviour observed.

During years with a presumably high fill level of the low permeability reservoir (LPR), recession below the threshold was rather slow as well (scenario 1). A prominent baseflow recession coefficient derived is 0.004 \(d^{-1}\), estimated to be representative for the drainage of the LPR. In many years, a recession coefficient of about 0.008 \(d^{-1}\) was observed instead, at least for the early part of the post-plateau recession (scenario 2). During discharge conditions, when the LPR can be assumed as largely depleted, the recession coefficients are even higher during the early parts of
the recession, with 0.011 and 0.020 d$^{-1}$ being typical values. This period was again followed by periods of intermediate coefficients of ca. 0.008–0.011 d$^{-1}$ (scenario 3). During periods when the LPR is almost completely drained, the recession even retained the convex shape over the whole period, viz. recession coefficients increased with time. A typical sequence is 0.02—0.035—0.05...0.07 d$^{-1}$ until the spring “stops” flowing (scenario 4).

During the 2009–2012 discharge monitoring, a period of presumably very low matrix storage prevailed. Because no baseflow was measured at the onset of each individual winter flood, single recharge events often led to individual spring flow periods. Those events were analysed for discharge quantity and recession behaviour, in order to assess event groundwater recharge and system properties (Table 2). For small to medium events — the flow threshold at 0.47 m$^3$ s$^{-1}$ was not reached — event recession was usually very rapid (Fig. 4; Table 2). For certain events, the falling limb of the hydrograph even took a linear form (as displayed with linear axes; Figs. 1c and 4β), indicating that only the conduit reservoir was active. Therefore, it can be expected that conduit properties can be estimated from the data. For small to medium recharge events, a highly non-linear correlation of $\alpha$ with event discharge could be observed (Fig. 6). For most small events, the recession coefficient ranged at ca. 0.2 d$^{-1}$. For two low discharge events, coefficients of 0.7 and 0.9 d$^{-1}$ were evaluated. Since linear recession periods are not well described by semi-logarithmic coefficients, those recession periods were further analysed for coefficient $\beta$ (m$^3$ s$^{-1}$ d$^{-1}$) according to Bailly-Comte et al. (2010): $Q_t = Q_0 - \beta * t$ (Table 2, Fig. 4).

In sum, during the detailed monitoring period of 2009–2012, the observed range of recession coefficients $\alpha$ covered three orders of magnitude (0.0008–0.9 d$^{-1}$). The recession coefficients derived from the high-resolution data (especially those for the rainy winter season of 2011/2012) were in a good agreement with the coefficients derived for the long-term monitoring data, despite the different gauging methods and locations. As similar values were obtained for the flow threshold, the long-term data can be assumed to be largely correct.
3.3 Flow and recharge dynamics of the Auja spring system

A monitoring well in an unconfined aquifer (same hydrostratigraphic unit as Auja spring outlet) is located northwest of Auja spring (Fig. 2). A low local hydraulic conductivity of ca. $1 \times 10^{-6} \text{ m s}^{-1}$ was calculated from pumping test recovery data (Mekorot internal files). Therefore, the hydrograph is indicative of the hydraulic behaviour of the low permeability fractured rock matrix. However, due to unknown aquifer boundaries, it is not certain if the well is located in the catchment of Auja spring. Groundwater head records were available for the period of 1983–1991 (Fig. 7) displaying two long-term periods with extended increasing groundwater potentials and recession limbs. During the same period, the spring hydrograph displayed a recession behaviour covering several years, with superimposed peaks for individual winter seasons. The discharge amounts attributed to the individual winter seasons were graphically separated and are assumed to approximate the seasonal total recharge in the catchment area (Fig. 7). For the analysed nine hydrological years, more than half of the total discharge of about 70 million m$^3$ was estimated to be recharged during the two wet winter seasons of 1982/1983 and 1987/1988. These two winter seasons caused a substantial rise in the groundwater level observed in the monitoring well.

3.4 Meteorological data and calculation of potential evapotranspiration

Jerusalem central meteorological station (Israel Meteorological Service, 800 m ASL) is regarded as the reference station for the region (ANTEA, 1998), because of a continuous daily precipitation record starting 1950 (European Climate Assessment and Dataset, http://eca.knmi.nl; Klein Tank et al., 2002). The Aqraba gauging station is located at the surface water divide ca. 40 km north of Jerusalem (Fig. 2) and exhibited a relatively complete monthly precipitation time series for the period of 1963–1997 (SMART project database, http://www.ufz.de/daisy/). The monthly
precipitation record shows good correlation with the Jerusalem data (ANTEA, 1998 and own calculations), however some data gaps are observed. Within this study, several automatically logging rain gauges (RG3-M, Onset Computer Corporation) were installed around the surface water divide (Fig. 2). The stations represented ca. 80–110% of the relative precipitation depth observed at Jerusalem station during the measurement period.

For the calculation of potential evapotranspiration ETp, the Hargreaves-equation (Hargreaves and Samani, 1985) was selected. It was developed for the Davis lysimeter station in California (Hargreaves and Allen, 2003), a location with comparable Mediterranean semi-arid climatic characteristics (i.e. mean temperature, mean annual precipitation depth, precipitation distribution during the course of the year). Its suitability for the calculation of ETp, especially in semi-arid climates, was verified by Jensen et al. (1997), Droogers and Allen (2002), and Weiß and Menzel (2008). In the study of Weiß and Menzel (2008), the Hargreaves-equation calculations were in exact accordance with corrected pan evaporation data for the central part of the West Bank. ETp is calculated by the equation:

\[ ETp = 0.0023 \times Ra \times (T_{\text{mean}} + 17.8) \times (T_{\text{max}} - T_{\text{min}})^{0.5} \times \lambda^{-1} \]  

(1)

where Ra is extraterrestrial solar radiation (MJ d\(^{-1}\)), calculated on a daily time step by equations cited in Allen et al. (1998), \(T_{\text{mean}}\) is the mean air temperature during the respective time interval (average of daily minimum and maximum temperature, °C), \(T_{\text{max}} - T_{\text{min}}\) is the daily temperature range, and \(\lambda\) is the latent heat of vaporisation in order to obtain ETp in units of mm d\(^{-1}\).

Jerusalem central meteorological station was also the long-term reference station for the air temperature data for this study. The available time series covered the period 1964–2010 (European Climate Assessment and Dataset). Within this study, an automatic meteorological station (H21-001, Onset Computer Corporation) was installed in Kafr Malik in the highland region. Air temperature was measured during the period of 2008–2012 with a time resolution of 5 to 10 minutes. From this data, daily maximum and minimum air temperatures were derived. The
correlation with the Jerusalem data was very good (linear regression between parallel measurements; $R^2-T_{\text{min}} = 0.97$, $R^2-T_{\text{max}} = 0.99$). Because of its higher elevation, Kafr Malik station displayed on average a 0.7 °C lower air temperature.

4 Reservoir model

As a main tool for the quantitative analysis of the system, a conduit restricted reservoir model was developed. The model needed to simulate the long-term recession and simultaneously the observed fast discharge response, the plateau in discharge, and the water budget. The aim was to employ the least possible number of calibration parameters, in order to keep ambiguity low (e.g. Kirchner, 2006). A further aim was to develop a modelling concept that could be generalised and transferred to other karst groundwater systems. The model is based on the conservation of fluid mass and is split into two sub-modules, a soil/epikarst module SEM and an aquifer module AM. Both modules operate at a daily time step ($t_{1} - t_{0} = 1 \text{ d}$), assuming instant head equilibrium within the reservoirs. The SEM provides the input to the AM by deep percolation (Fig. 8). Deep percolation is regarded as groundwater recharge at the arrival of water at the water table. Because of the fast spring response to recharge events, a negligible time lag is assumed in this modelling study.

4.1 Soil/Epikarst Module (SEM)

The SEM consists of a soil water balance model (Sauter, 1992; Rushton et al., 2006; Geyer, 2008) with precipitation $P$ as the input parameter for the soil moisture storage SMS, which is varied between the permanent wilting point $PWP$ and field capacity $FC$ (all values in mm). To simplify
the approach, soil moisture storage at the PWP is set to zero. Output is by actual evapotranspiration $ET_a$ and by deep percolation $DP$, once the soil moisture storage exceeds field capacity:

$$SMS_{t_1} = SMS_{t_0} + P_{t_1} - ET_a_{t_1} - DP_{t_1} \quad \text{with}$$

$$DP_{t_1} = SMS_{t_0} + P_{t_1} - ET_a_{t_1} - FC \quad \text{if} \quad (SMS_{t_0} + P_{t_1} - ET_a_{t_1}) > FC \quad (2)$$

In hydrological models, $ET_a$ is often derived from $ET_p$ as a function of the relative water saturation in the soil store (e.g. Perrin et al., 2003; Hartmann et al., 2012a). In this study, a different approach is applied based on two considerations: (1) In the highland part of the study area, the predominant vegetation type is grass-/shrubland sustained by winter precipitation. Ryu et al. (2008) measured the $ET_a$ of a semi-arid grassland site with very comparable climatic characteristics (located close to the Davis lysimeter station; section 3.4). They showed that for most of the winter months with considerable precipitation, $ET_a$ nearly equaled $ET_p$ ("energy-limited period"). During the spring months, usually starting from April or May, $ET_a$ became considerably lower than $ET_p$, because of the low precipitation amounts but already high $ET_p$ (start of "water-limited period"). (2) Rushton et al. (2006) argue that especially fine grained soils retain a part of the infiltrated water temporally near the surface ("near surface soil storage") and hence provide local nearly saturated conditions, enabling ET at the potential rate. The soils in the study area are of silt loam texture and near surface soil storage was observed (Hanf, 2010). For those reasons, $ET_a$ is set equal to $ET_p$ until the wilting point is reached:

$$ET_a_{t_1} = \begin{cases} ET_p_{t_1} & \text{if} \quad ((SMS_{t_0} + P_{t_1}) \geq ET_p_{t_1}) \\ SMS_{t_0} + P_{t_1} & \text{if} \quad ((SMS_{t_0} + P_{t_1}) < ET_p_{t_1}) \end{cases} \quad (4)$$

It is apparent from the spring hydrograph that groundwater recharge occurs in all monitored years, even during winters with very low precipitation depth (e.g. 1998/1999). These observations cannot be reproduced by assuming a uniform field capacity across the entire area, as usually
applied in karst reservoir models (e.g. Rimmer and Salingar, 2006; Fleury et al., 2007; Hartmann et al., 2012a). A spatially variable soil thickness and epikarst storage capacity is frequently observed in karst systems (e.g. Williams, 1983; Arbel et al., 2010). Consequently, Hartmann et al. (2012b) used a soil/epikarst graduation consisting of 15 cells. In this study, the catchment area A was divided into two SEMs with different field capacities (SEM 1 with FC 1 and SEM 2 with FC 2) to account for soil/epikarst variability with a moderate number of parameters:

\[ A_{SEM_1} = x \times A \quad \text{with} \quad 0 < x < 1 \quad \text{and} \quad A_{SEM_2} = (1 - x) \times A \quad (5) \]

with x as the fraction of SEM 1 of the total catchment area A. This way, deep percolation and subsequent groundwater recharge is enabled for a part of the study area at lower precipitation totals.

### 4.2 Aquifer Module (AM)

The aquifer module consists of a highly permeable reservoir HPR and a low permeability reservoir LPR (Fig. 8). System outflow occurs exclusively from the HPR via spring discharge \( Q_{HPR} \). The aquifer module base areas \( A_n \) for reservoir n are intended to display the effective porosity \( n_{eff} \) of the respective aquifer compartment: \( A_n = A \times n_{eff} \) (Fiorillo, 2011). This way, as the base areas are calibration parameters, an estimate of \( n_{eff} \) for the different aquifer porosity compartments is derived from the calibrated model. For example, in case of a spring catchment area of 1 km\(^2\) and a calibrated matrix reservoir base area of 10 000 m\(^2\), an effective matrix porosity of 1% would be derived. The deep percolation calculated by the SEM is converted from mm d\(^{-1}\) to m\(^3\) d\(^{-1}\) according to the respective SEM size. A lateral flow concentration of infiltrated water towards preferential percolation pathways (shafts, main fractures) can be generally assumed for epikarst zones (Williams, 1983). Hence, in this model, all deep percolation is added to the HPR, which is assumed to be connected to the main points of recharge, such as the ephemeral
surface drainage network and preferential flow paths from the epikarst zone and soil pockets.

The water volume stored in the HPR is calculated by the following equation:

\[ V_{HPR, t1} = V_{HPR, t0} + DP_1 t_1 + DP_2 t_1 + Q_{EXC, t0} - Q_{HPR, t0} \] (6)

with \( Q_{EXC} \) being the water exchange between both reservoirs depending on the head difference multiplied by a lumped exchange parameter \( LEP \) (m² d⁻¹) (Bauer et al., 2003; De Rooij, 2008):

\[ Q_{EXC} = LEP \cdot \begin{cases} h_{LPR} - h_{HPR} & \text{if } (h_{HPR} > h_i) \\ h_{LPR} - h_i & \text{if } (h_{HPR} < h_i) \end{cases} \] (7)

The \( LEP \) is the product of the interfacial area between both reservoirs \( A_{EXC} \) (m²) and an exchange coefficient \( \alpha \) (d⁻¹):

\[ LEP = A_{EXC} \cdot \alpha \] (8)

Based on section 2.3, the flow restriction for the HPR is assumed to be located in the terminal part of the main conduit system, which connects the recharge area with the current discharge point. It is conceptualised as a conduit with low discharge capacity. Similar approaches were used e.g. by Smart (1983), Halihan et al. (1998), and Covington et al. (2009). The flow restricting pipe is assumed to be always fully saturated. Theoretical considerations regarding the expected pipe diameter and the flow velocity indicate that flow will be always turbulent. Turbulent flow in fully saturated pipes can be calculated by the Darcy–Weisbach and Colebrook–White laws (e.g. Liedl et al., 2003; Reimann et al., 2013):

\[ Q_{HPR} = -2Y \log \left( \frac{2.51 \Pi D}{4Y} + \frac{\varepsilon}{3.71D} \right) \] with \[ Y = \sqrt{\frac{h_{HPR} g D^3 \Pi^2}{8L}} \] (9)

Where \( \nu \) is the kinematic viscosity of water (m² s⁻¹), \( D \) the conduit diameter (m), \( \varepsilon \) the conduit surface roughness (m), \( g \) the gravitational acceleration (m s⁻²), and \( L \) the conduit length (m). Discharge is driven by the head difference between the conduit outlet and the head in the HPR:
with $V$ being a conceptual parameter (see Section 4.3.3 below). In order to simulate the non-linear discharge behaviour, an approach relating the reservoir base area to the hydraulic head within the reservoir is required. Therefore, the base area $A_{HPR}$ of the HPR is decreased below $h_1$ to $A'_{HPR}$ (Fig. 8). A similar approach was applied by Barrett and Charbeneau (1997) for modelling discharge at *Barton springs* (USA), a spring displaying a convex recession behaviour.

### 4.3 Parameter estimation

Three parameters of the model could be estimated by the analysis of the flow system (Sections 4.3.1–4.3.3). Additionally the constricting conduit parameters were partly estimated from the flow system and literature values in order to constitute a combined calibration parameter (Section 4.3.4).

#### 4.3.1 Spring catchment recharge area $A$ (by a combined water and chloride mass balance)

For the 33 hydrological years of 1968–2000, total spring discharge volume was evaluated at ca. 305 million m$^3$. This is equivalent to an average of 9.2 million m$^3$ a$^{-1}$. Precipitation measured at Jerusalem meteorological station was on average 556 mm a$^{-1}$ for the period. According to our own precipitation measurements (section 3.4) this value was assumed to be representative for the recharge area of the springs on a long-term basis. For the Auja spring catchment, a mean long-term average groundwater recharge fraction of precipitation of 34% was derived independently by a chloride mass balance (Schmidt et al., 2013). Accordingly, the average groundwater recharge

\[ h_{HPR} = \begin{cases} 
V_{HPR} / A'_{HPR} & \text{if } (V_{HPR} \leq V) \\
(V_{HPR} - V) / A'_{HPR} + h_1 & \text{if } (V_{HPR} > V) 
\end{cases} \]
for the period was 189 mm a\(^{-1}\). With these data, the catchment recharge area \(A\) was calculated at:

\[
A = \frac{9.2 \times 10^6 \text{ m}^3 \text{ a}^{-1}}{0.189 \text{ m a}^{-1}} = 49 \text{ km}^2.
\]

4.3.2 SEM partitioning coefficient \(x\) (by event analysis)

The surface area of module SEM 1 \(A_{\text{SEM 1}}\) could be estimated by the results of the event analysis (section 3.2), based on three assumptions: (1) after a preceding recharge event, there is no soil moisture deficit in SEM 1, therefore the following precipitation event \(P_{\text{event, corrected}}\) is exclusively converted to groundwater recharge, (2) for the subsequent small to medium recharge events, groundwater discharge \(Q_{\text{event}}\) measured at the spring equals groundwater recharge, and (3) for those small to medium events, only the module SEM 1 is contributing to recharge.

\[
A_{\text{SEM 1}} (\text{m}^2) = \frac{Q_{\text{event}} (\text{m}^3)}{P_{\text{event, corrected}} (\text{m})}
\] (11)

Nevertheless, between the events evapotranspiration occurs (\(ET_{\text{a pre-event}}\)), leading to a small depletion in SMS which has to be accounted for. This depletion might be partly compensated by small precipitation events \(P_{\text{pre-event}}\), which are likewise considered:

\[
P_{\text{event, corrected}} = P_{\text{event}} - ET_{\text{a pre-event}} + P_{\text{pre-event}}
\] (12)

The small to medium rainfall-discharge events of 2010 appeared to be particularly suitable for the calculation, because the time elapsed between events was small and the events “2010-2” and “2010-3” were the two largest medium recharge events of the high-resolution time series (Table 2). The calculated surface area for SEM 1 is ca. 17 km\(^2\) (event 2010-2) and 12 km\(^2\) (event 2010-3), respectively; the combined event reveals a value of nearly 14 km\(^2\). This value is equivalent to 28\% of the total catchment recharge area of 49 km\(^2\) (previous section), therefore \(x = 0.28\).
4.3.3 Conceptual volume $V$ of the highly permeable reservoir (by combined event and recession analysis)

The volume $V$ of the part of the HPR below $h_1$ is a conceptual parameter from which $A'_{\text{HPR}}$ can be calculated, once the parameter $h_1$ is calibrated ($A'_{\text{HPR}} = V / h_1$). It could be estimated from the discharge hydrographs of the years 2009 and 2010. Since both years displayed comparatively low recharge amounts, the discharge condition above the threshold was only maintained for about one month each year, therefore the LPR is assumed to receive only little recharge by cross-flow from the HPR. Accordingly, with gradients reversed, only a small discharge contribution from the LPR to total spring flow is assumed. Therefore, the discharged volume after the point where discharge is falling below the flow threshold (post-plateau recession) is indicative of $V$. The integrated discharge volume was ca. 1.1 million $m^3$ for 2009 and 1.3 million $m^3$ for 2010, respectively (Fig. 4, $V_{2009}$ and $V_{2010}$). The duration of post-plateau recession was very comparable for the two years with 82 and 81 days, respectively. In contrast, integrated post-plateau discharge for 2012 was nearly 3 million $m^3$ with a considerably longer post-plateau recession period, presumably due to high LPR–HPR cross-flow. The mean value of 1.2 million $m^3$ calculated for the years 2009 and 2010 was used.

4.3.4 Constricting conduit properties

The restricting conduit length $L$ was approximated from the maximum possible length of the catchment area of about 19 km (from Auja spring till the groundwater basin divide to the northwest) at 10 km. The conduit roughness was intended to be close to the value of $\varepsilon/D = 0.25$ as reported by Jeannin (2001). The conduit parameters can then be constrained by the observed outflow threshold at 0.47 $m^3$ s$^{-1}$ (Section 3.2).
4.4 Model calibration and input data provision

Model calibration was based on three objectives/constraints: (1) to retain the water budget during the calibration period of October 1967 to September 2000 (calculated total deep percolation = measured total discharge volume of 305 million m³), (2) to match the simulated spring hydrograph to the measured data, and (3) to optimise the Nash–Sutcliffe efficiency criterion (Nash and Sutcliffe, 1970). A criterion based on square root transformed values (NSE$_{sqrt}$) was applied, because it is sensitive for both, low and high flow periods (e.g. Perrin et al, 2003; Oudin et al., 2006; Pushpalatha et al., 2012). The period of January 1964 to September 1967 was considered as a “warm-up period” for the model. The simulation was extended until September 2012. For the period of 2001–2008, no reliable flow data were available. The high-resolution data of 2009–2012 provided the validation period.

Daily precipitation data of Jerusalem station served as the input to the SEM for the period of 1964–2007. The daily arithmetic average precipitation depth of three highland area stations (Kafr Malik, Taybeh, Mazra’a ash-Sharqiya) was used as the input for the period of 2008–2012. The average value represents ca. 99 % of the precipitation depth measured at Jerusalem. For the calculation of ETp during the period of 1964–2007, the air temperature data of Jerusalem station were used. For the period of 2008–2012, data from Kafr Malik station, adjusted to the Jerusalem temperature data, were selected. For a data gap during the hydrological year 2009, the Jerusalem data were used. For the 1st of January 1964, initial values for all four sub-modules were provided.

Six parameters of the reservoir model needed to be calibrated: two of the soil/epikarst module (FC 1 and FC 2) and four of the aquifer module ($A_{HPR}$, $A_{LPR}$, LEP, and the combined restricting conduit parameter represented by $h_1$). The first step of the calibration strategy comprised the selection of plausible field capacity values of the SEM in order to generate deep percolation also for years with low precipitation depth (e.g. winter of 1998/1999). The second step comprised the calibration of the parameters for the restricting conduit to an outflow rate of 0.47 m³ s$^{-1}$. 


Different combinations of $L$, $D$, $\varepsilon$, and $h_1$ were tested and model results appeared to be insensitive to combined changes in the parameters $L$, $D$, and $\varepsilon$ (Fig. 11). In contrast, $h_1$ appears to be a more sensitive parameter and is regarded as the main calibration parameter for the conduit restriction. The model parameters were optimised iteratively, applying the criteria stated above. The “optimum” parameter set was found at a $\text{NSE}_{\text{opt}}$ of 0.8. For the validation period, a $\text{NSE}_{\text{val}}$ of 0.77 was obtained.

4.5 Modelling results

The modelled and measured data generally match well (Fig. 9). In particular, the model is able to reproduce: (1) the plateau in discharge, (2) the initial peak superimposed on the plateau, and (3) the highly variable discharge conditions observed. From the values of $V$ (estimated, 1.2 million m$^3$) and $h_1$ (calibrated, 100 m), the base area of the lower part of the HPR ($A_{\text{LPR}}$) was calculated at 12 000 m$^2$. $A_{\text{HPR}}$ was calibrated to 0.2 km$^2$, and $A_{\text{LPR}}$ to 1 km$^2$, respectively. The LEP was calibrated to 3500 m$^2$ d$^{-1}$. Hence, the interface area $A_{\text{EXC}}$ between the two reservoirs was calculated at ca. 0.9 km$^2$ applying Eq. (8) with the recession coefficient $\alpha$ of the low permeability reservoir (0.004 d$^{-1}$).

The 45-year modelling results reveal that on average 96% of the groundwater recharge occurs during the four month period of December to March. The combined average recharge for April and November is 4%. Annual precipitation ranged from 210 to 1130 mm. Calculated annual ETa in general ranged between 300 and 420 mm (mean 360 mm). The hydrological year of 1999 displayed only 209 mm, due to limited precipitation. Two hydrological years with considerable precipitation amounts during the middle to late April, replenishing soil moisture storage, provided 450 and 460 mm of ETa. The calculated annual recharge range varied between 1 mm and 780 mm during the modelling period (Fig. 10). The five hydrologic years with the highest annual precipitation during the modelling period (1974, 1980, 1983, 1992, and 2003; all >800 mm
a−1) also displayed the highest calculated groundwater recharge. They provided 33% of the integrated modelled recharge. In contrast, the 14 hydrologic years with the lowest recharge amounts caused less than 10% of the integrated groundwater recharge (Fig. 10).

4.6 Sensitivity analysis

The sensitivity of single model parameters on model results was analysed to investigate the predictive power of the model and to identify those model parameters, the discharge variability is most sensitive to and that have to be determined independently (if possible). NSE\textsubscript{sqrt} was used as a quantitative criterion (Figure 11). FC 1 was varied with accompanying changes in FC 2 to fulfil the water budget constraint. Varying the conduit parameters L and h1 required adjustment of the other conduit parameters to match a flow threshold of 0.47 m\textsuperscript{3} s\textsuperscript{−1}. As an “optimum” parameter set, not the set with the highest NSE\textsubscript{sqrt} was selected. For example, a FC 1 value of 90 mm, despite overall good simulation results, would result in no recharge for 1999 and 2011 which does not correspond to field observations. Similarly, an A\textsubscript{LPR} of 1.3 km\textsuperscript{2} would provide a higher NSE\textsubscript{sqrt} but would result in a less optimal match of the recession behaviour.

5 Discussion

5.1 Hydrograph and recession analysis

The hydrograph of Auja spring shows a complex discharge and recession behaviour. A flow threshold at ca. 0.47 m\textsuperscript{3} s\textsuperscript{−1} was observed. Above the threshold, the flow regime is considered as conduit restricted, resulting in an inclined plateau in the observed spring hydrograph. Below the threshold, small to medium recharge events lead to marked spring flow events closely succeeding
precipitation input (Fig. 4). According to the considerations of Covington et al. (2009), this contrasting discharge behaviour might be classified into the flow regimes “(aquifer) geometry dominated” (flow threshold exceeded) and “recharge dominated” (below threshold).

The initial higher discharge peak above the plateau, recorded during March 2012 (Figs. 4 and 5), is also recognisable in some other years, e.g. during early 1974, 1983, and 1992 (Fig. 9). This phenomenon is recognised in other conduit restricted flow datasets as well (e.g. Herman et al., 2008, stage data). Here it is interpreted as an initial high water level within the highly permeable system (i.e. karst shafts) following extensive recharge (compare Fiorillo, 2011). Due to transfer of water from the highly permeable structures to other aquifer compartments, the initial local high hydraulic head is more evenly distributed in the system. The groundwater head in the highly permeable reservoir subsequently declines due to outflow of water by spring discharge and water transfer to the low permeability matrix, leading to a slightly inclined discharge plateau (\( \alpha \approx 0.0009 \) d\(^{-1}\)). In other conduit restricted aquifers, “excess” recharge is discharged by overflow springs, which is expected to produce a more level plateau (see e.g. Figure 4.11 in Smart, 1983). In contrast, the inclined plateau in the Auja flow system indicates that “excess” recharge is retained within the system and only gradually released.

Only a part of the large number of observed recession coefficients \( \alpha \) below the plateau is assumed to reflect physically based system values, e.g. the baseflow recession coefficient of the low permeable rock matrix of 0.004 d\(^{-1}\). This value is very similar to the baseflow recession values evaluated for other karst spring systems: 0.004 d\(^{-1}\) (median) for Foux de la Vis (Forkasiewicz and Paloc, 1967), 0.006 d\(^{-1}\) for Ombla spring (Milanovic, 1981), 0.002 d\(^{-1}\) for Gallusquelle (Sauter, 1992), and 0.006 d\(^{-1}\) for Fontaine de Vaucluse (Fleury et al., 2007). Most of those studies relate this lowest recession coefficient to the drainage of (small) fractures in the saturated zone. Fleury et al. (2007), state that this water may also originate in part from the epikarst or the vadose zone. The recession coefficient of the main conduit system was estimated
at about 0.2 d\(^{-1}\). The higher coefficients around 0.8 d\(^{-1}\) are attributed to the only partial filling of the main conduit system due to limited recharge. Comparable values were derived for other systems: 0.5 d\(^{-1}\) (median) for Foux de la Vis (Forkasiewicz and Paloc, 1967), 0.13 d\(^{-1}\) for Ombla spring (Milanovic, 1981), and 0.25 d\(^{-1}\) for Gallusquelle (Sauter, 1992).

Some small discharge events displayed a linear flow recession, assumed to reflect the drainage of a conduit system without matrix contribution (Bailly-Comte et al., 2010; Fiorillo, 2011). For those events, \(\beta\)-values around 0.02 m\(^3\) s\(^{-1}\) d\(^{-1}\) were derived (Table 2). Bailly-Comte et al. (2010) calculated a considerably higher value of about 1 m\(^3\) s\(^{-1}\) d\(^{-1}\) for the Vene spring in France. They relate high values of \(\beta\) to a large flow section of the conduit, hence to a high degree of karstification. The \(\alpha\) and \(\beta\) recession coefficients may be used in subsequent works as independent parameters for pipe flow. The intermediate recession coefficients (Table 2) are assumed to be the result of the variable hydraulic head and cross-flow between the conceptualised two aquifer compartments.

5.2 Reservoir model and water balance

A reservoir model was set up and calibrated in order to refine the conceptual model and used as the main quantitative evaluation tool. In contrast to modelling studies for which calibration is based on the shape of the hydrograph alone, in this study, the long-term water balance is also a primary calibration constraint. Since the water balance of a short period, especially in semi-arid environments, can be biased with respect to extreme hydrometeorological conditions (wet and dry), in this study an extended (33-year) calibration period was used. In this study, also the recession coefficients for the reservoirs are not fixed or calibrated (e.g. Fleury et al., 2007; Geyer et al., 2008) but are simply the result of model geometry and variable cross-flow. The outflow restriction was incorporated by using a physical pipe representation in combination with a groundwater head-dependent reservoir base area. The simulated drainage behaviour of the highly
permeable reservoir without cross-flow below $h_1$ is linear ($\beta \approx 0.008 \text{m}^3 \text{s}^{-1} \text{d}^{-1}$). This behaviour resembles the observed shape of some small discharge events, (Table 2, Fig. 4), and especially the straight line segments of the early post-plateau recessions of 2009 and 2010 with an observed $\beta$ of ca. 0.01 m$^3$ s$^{-1}$ d$^{-1}$ (Fig. 4).

Despite the simplified representation of nature, it is assumed that the calibrated parameters of the reservoir model reflect physical properties of the system. The two calibrated soil/epikarst “field capacities” of 70 and 190 mm are in the range of values observed for other sites in the Eastern Mediterranean karst region with very similar precipitation characteristics compared to the study site. Shaffer et al. (2011) assessed a combined soil/epikarst field capacity of about 100 mm from an irrigation experiment above a small vadose cave (drip onset = indicator of recharge). Arbel et al. (2010) assessed very variable seasonal drip onset threshold values at precipitation totals between 120 mm and 340 mm for different cave drip sites located in one cave within a lateral distance of only 20 m. Those values are the sum of the soil/epikarst field capacity and evapotranspiration from the soil store before field capacity is reached (assessed for the winter season 2005/2006). Hartmann et al. (2012a) calibrated lumped soil/epikarst field capacity values of 126 mm and 166 mm for five different lumped parameter models for Faria spring, located ca. 37 km north of Auja spring.

The base areas of the conceptual reservoirs in relation to the total recharge catchment area can be regarded as estimates of effective porosity (Section 4.2; Fiorillo, 2011). The calibrated values of 1 km$^2$ for $A_{LPR}$ and 0.2 km$^2$ for $A_{HPR}$ are roughly 2.0% and 0.4% of the total catchment area of 49 km$^2$, respectively. The value of 2% for the LPR is a realistic effective “matrix” porosity for carbonate aquifers (compare e.g. Atkinson, 1977; Sauter, 1992; Kiraly, 2002). The porosity value of the HPR of 0.4% is about one order of magnitude higher than the porosity generally attributed to karst conduit systems, which were however assessed by analysing the transport signals of the systems (e.g. Atkinson, 1977; Sauter, 1992). Nevertheless, this high HPR-value contradicts a strict
matrix-conduit differentiation, as widely applied in karst hydrogeology. In this modelling approach, it can be expected that the calibrated porosity of the HPR also incorporates higher permeable fractures. A considerable storage and hence porosity attributed to the highly permeable system in karst aquifers is also supported by recent hybrid continuum-discrete pipe flow modelling studies (e.g. De Rooij, 2008; Reimann et al., 2013) and termed “conduit-associated drainable storage” in Reimann et al. (2013).

Due to its presumed large spatial coverage, as well as the redistribution and focussing of infiltrated water within the soil/epikarst, the HPR receives the bulk of the recharge. This is supported by the delayed response of the groundwater head in the matrix (Fig. 7), attributed to transient conduit–matrix cross-flow. The effect of providing recharge to the LPR directly (“fraction to LPR” in Fig. 8) was assessed by a sensitivity analysis (Fig. 11). Adding up to 50% of recharge from DP 2 into the LPR had little effect on NES but decreased the observed discharge dynamics. Hence, DP 2 was added to the HPR exclusively. The calculated interface area $A_{\text{EXC}}$ between both reservoirs of ca. 0.9 km$^2$ is only an approximation because processes such as an unsteady-state head distribution inside the matrix or a possible head-dependent size of the interface area are not considered.

Despite the long-term equal precipitation amounts of Jerusalem gauging station and the recharge area of the springs, some larger discrepancies may be observed for individual storms, which are expected to influence model results. For example, during the large precipitation event of February 2010 (Table 2), cumulative precipitation was 147 mm for the highland area close to the spring (average of three rain gauges) and 196 mm for Jerusalem gauging station. Since the antecedent soil moisture storage was relatively high during February, a precipitation event of such magnitude leads to a massive area-wide groundwater recharge event. For a catchment area of 49 km$^2$, a difference of 50 mm is equivalent to 2.5 million m$^3$ of groundwater recharge. Those differences are expected to cause some deviations between measured and modelled spring
hydrographs, e.g. during the period of 1969–1973. For example, during the winter season of 1968/1969, annual precipitation depth was 100 mm larger for Aqraba than for Jerusalem, indicating a generally higher amount in the north.

Calculated annual actual evapotranspiration (ETa) displayed a comparatively low variability for the modeled 45 year period (range 300–420 mm a\(^{-1}\), excluding three outliers). Ryu et al. (2008) also observed, in a climatically similar region that the range of annual ETa was considerably lower than the range in annual precipitation. Flood runoff leaving the autogenic surface catchment of Wadi Auja was evaluated at 1.2% of cumulative precipitation during the monitoring period of 2010–2013 (Ries, 2013). Therefore, the amount of precipitation in excess of the evaporative demand will largely contribute to groundwater recharge. Because annual precipitation is highly variable in this semi-arid environment, annual groundwater recharge is even more variable. Ranges were from nearly zero to about 500 mm a\(^{-1}\) (excluding one unusually wet year). Seasons with a low to average precipitation depth will normally not provide a large recharge amount. However, during wet years the system is considerably recharged.

5.3 Implications for conduit restricted flow in the Jordan Valley region

In the case study area, also the springs Sultan and Duyuk (Fig. 2) are expected to be characterised by conduit restricted flow. Due to its very stable discharge, Rosenthal and Kronfeld (1982) regarded Sultan spring as an underflow spring. Both springs are likely fault-controlled springs. A limited hydraulic conductivity of the fault zones located in the Coniacian–Maastrichtian sediments (chalk and chert), which are overlying the carbonate aquifer, can be assumed. Also the Dan spring, the main spring of the Upper Jordan River (see inset map in Fig. 2 for location) may be fault-controlled (Figure 1 in Hartmann et al., 2013). Its hydrograph displayed a convex recession behaviour (Rimmer and Salingar, 2006), alternatively attributed to a large aquifer storage volume and continuing percolation from the vadose zone (Rimmer and Salingar, 2006).
In contrast to those springs, the Auja spring orifice is located in the carbonate aquifer unit itself, hence the conduit restricted behaviour is attributed to a limited developed terminal conduit segment (Section 2.3). An alternative line of argument is a change in seasonal recharge variability, e.g. a shift to a more extreme pattern following a change in climate from humid to semi-arid conditions. This might lead to a backing-up of groundwater flow within the conduit system. According to these examples and considerations, different mechanisms might be the reason for conduit restricted flow in the study area.

6 Conclusions

In this study, a characterisation and modelling approach for conduit restricted karst aquifers was proposed and executed on the example of the Auja spring (Lower Jordan Valley). A combination of hydrometric monitoring, hydrogeological data analysis, and non-linear lumped parameter modelling was applied.

Classical hydrogeological methods, especially spring hydrograph monitoring and analysis as well as water budget calculations, provide the conceptual framework for the characterisation of conduit restricted aquifers and enable the estimation of a number of system parameters. For those purposes, the acquisition of a precise discharge hydrograph is obligatory. High-resolution data are preferable. For example, in the presented case, the presence of the flashy discharge peaks below the flow threshold is not well resolved by monthly data. Especially in semi-arid environments, prolonged drought and wet periods are present and groundwater flow and storage can be very transient. Therefore, long-term time series need to be obtained. In this study, a 33-year period was used for the calibration of the model, and since the bulk discharge amount could be assessed by monthly data, high quality data of this kind are very valuable. Using only the
recently obtained high-resolution time series for analysis would easily lead to a misinterpretation of the system, i.e. the usually observed slow recession behaviour of the Auja spring during pluvial periods would have been fully missed. Here, the long-term low-resolution data and short-term high-resolution time series were successfully combined.

Semi-arid environments with a distinct seasonality in precipitation are particularly suited for a recession analysis, because prolonged hydrograph recessions, not interrupted by recharge, can be obtained. As precipitation is occurring as intensive and infrequent storms, sharp input signals are available. Furthermore, because precipitation depth displays a high interannual variability, input signals are also highly contrasting.

The reservoir model developed in this study can be regarded as relatively simplistic, despite the hydrogeological complexities encountered, which include: (1) the non-linear system behaviour at the flow threshold, (2) the considerable cross-flow between the highly permeable structures and the matrix of the aquifer (caused by the expected considerable hydraulic head differences in conduit restricted aquifers), and (3) spatially variable soil/epikarst properties. However, the last two features are also applicable for standard karst reservoir models. As a number of required parameters could be estimated from the analysis of the flow system the model only requires six calibration parameters, reducing the parameter ambiguity. The model can form example be used to predict spring discharge during the summer and autumn seasons for the planning of irrigated agriculture.

From a water management perspective, conduit restricted karst aquifers, if not drained by overflow springs, seem advantageous to the ones displaying free discharge. The conduit restricted flow mechanism prevents water from rapidly flowing out as soon as the threshold is reached. In this semi-arid region with prolonged drought periods, conduit restricted flow functions as a natural management instrument for aquifer discharge.
Acknowledgements

This study was conducted within the framework of the multi-lateral research project “SMART – Sustainable Management of Available Water Resources with Innovative Technologies” funded by BMBF (German Federal Ministry of Education and Research), references Nos. 02WM0802 and 02WM1081. Special thanks go to: Abu Ashraf, Abu Shadi, Ayman Shawana, Steffen Fischer, Mathias Toll, Andrea Hanf, Klaus Haaken, and many more people from Auja village for help in spring site constructions and fieldwork; teachers, school and water authority personnel for help in meteorological instrumentation; Mustafa Nuseibeh and coworkers for more than 30 years of superb spring gauging; the Palestinian Water Authority for permission for spring site constructions and data provision; the Israel Nature and Parks Authority, particularly Asaf Gottfeld for permission for spring site constructions; Akiva Flexer, Anat Yellin-Dror, and Nimrod Inbar for logistical support and discussion; Joe Nadolski for proofreading the manuscript; Jannes Kordilla for programming support. The detailed and helpful comments of the anonymous reviewers were highly appreciated.

References


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**Fig. 1:** Examples of conceptual models with convex discharge recession behaviour and representative spring hydrographs (displayed with linear axes). Conceptual models modified after Smart (1983), Bonacci (1993), and Bailly-Comte et al. (2010).

**Fig. 2:** Map of the study area displaying the outcrop areas of the geological strata (modified after Schmidt et al., 2013; outcrop areas modified after Sneh et al., 1998; groundwater divide after ANTEA, 1998; Auja monocline after Lenz, 1969). The numbers related to the rain gauges indicate the mean precipitation depth in percent relative to Jerusalem central meteorological station (=100 percent) for different periods of parallel measurements.

**Fig. 3:** Auja spring in March 2009 (discharge ca. 0.5 m$^3$ s$^{-1}$). The thalweg of the ephemeral Wadi Auja is indicated by black arrows. In this section, the wadi is incised more than 300 m into the rather flat paleorelief. The outcropping rock formation in the surroundings of the spring is composed of well bedded dolomitic of Cenomanian age (Begin, 1975). A buried large diameter tube connects spring and gauging station.

**Fig. 4:** High resolution hydrograph of Auja spring during 2009–2012, displaying the convex upward recession behaviour. Examples for different recession coefficients $\alpha$ (in units of d$^{-1}$) and $\beta$ (in units of m$^3$ s$^{-1}$ d$^{-1}$) are highlighted (see section 3.2). The last six weeks of the 2012 flow period, the hydrograph is linearly interpolated between two discrete discharge measurements. Precipitation data measured at meteorological station Kafr Malik.

**Fig. 5:** Schematics of the complex recession behaviour of the Auja spring flow system according to the water storage in the assumed low permeability reservoir.
**Fig. 6:** Relationship between recession coefficient $\alpha$ and the amount of water discharged for the small to medium recharge events (Table 2). For linear recession events, either the middle or mean $\alpha$ value was used. The number of event data with a very high $\alpha$ is still insufficient to derive a sound correlation, however, the intersection (the point of nonlinearity) is assumed to be located between 50 000–100 000 m³ of event discharge.

**Fig. 7:** Hydrographs of Auja spring discharge and water level in a monitoring well, together with the monthly and yearly precipitation sums for Jerusalem. Note the 1.5 year time lag between precipitation and maximum groundwater level (discharge data: Palestinian Water Authority, Ramallah; water level data: MEKOROT, Tel Aviv).

**Fig. 8:** Schematic representation of the reservoir model. The various parameters are specified in the text and in Table 3.

**Fig. 9:** Comparison of measured and modelled spring discharge (discharge data 1967–2000: Palestinian Water Authority, Ramallah).

**Fig. 10:** Annual groundwater recharge (in fact: deep percolation), calculated by the reservoir model for 45 individual hydrological years, sorted in ascending order. Groups of 29, 11, and 5 hydrological years each account for about one third of total modelled recharge.
Fig. 11: Sensitivity of selected model parameters. The arrows indicate the values of the “optimum” calibrated parameter set, based also on the subjective judgement of the simulation results.
Conceptual model  Spring hydrograph

a) Limited capacity and overflow spring

Smart, 1983

b) Upgradient reservoir of water

high porosity zone, "cave", Polje

Bonacci, 1993

c) Flow regime transition

CFR  MRFR

Conduit flow regime (linear)
Matrix restr. flow reg.

Bailly-Comte et al., 2010
Figure 3
Figure 6

Recession coefficient $\alpha$ (d$^{-1}$) vs. Spring discharge due to recharge event (m$^3$)

Mean (0.22; 0.64)
Figure 10

cumulative recharge (45 a)

mean recharge

Calculated recharge (mm a⁻¹)

Sequential year

33.5 %

67 %
Table 1: Characteristic recession coefficients derived from the analysis of the long-term monitoring data 1967–2000. A few higher α-values such as 0.05 or 0.07 were recognised as well.

<table>
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<th>Recession characteristics</th>
<th>α median (d⁻¹)</th>
<th>α mean (d⁻¹)</th>
<th>α range (d⁻¹)</th>
<th>sample n</th>
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<td>0.00092</td>
<td>0.00096</td>
<td>0.0005–0.0021</td>
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<tr>
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<td>0.0039</td>
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<td>13</td>
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<tr>
<td>medium</td>
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<td>0.0080</td>
<td>0.0068–0.0098</td>
<td>15</td>
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<td>medium–fast</td>
<td>0.0111</td>
<td>0.0118</td>
<td>0.0101–0.0153</td>
<td>6</td>
</tr>
<tr>
<td>fast</td>
<td>0.0199</td>
<td>0.0201</td>
<td>0.0183–0.0218</td>
<td>5</td>
</tr>
<tr>
<td>very fast</td>
<td>0.0353</td>
<td>0.0353</td>
<td>0.0349–0.0357</td>
<td>5</td>
</tr>
</tbody>
</table>
Table 2: Recharge event and recession analysis of the high resolution monitoring data 2009–2012 for individual storms.

Hydrograph recessions for small to medium events were occasionally linear. Therefore, more than one recession coefficient $\alpha$ is displayed with a logarithmic $y$-axis.

### Small to medium events

<table>
<thead>
<tr>
<th>No.</th>
<th>Date of precipitation</th>
<th>P event (mm)$^a$</th>
<th>Q peak ($l \cdot s^{-1}$)</th>
<th>Q event ($m^3$)$^b$</th>
<th>$\alpha$ (d$^{-1}$)</th>
<th>$\beta$ ($m^3 \cdot s^{-1} \cdot d^{-1}$)</th>
<th>Spring flowing before event?</th>
</tr>
</thead>
<tbody>
<tr>
<td>2009</td>
<td>20–22. Feb</td>
<td>48</td>
<td>320</td>
<td>180.000*</td>
<td>0.22</td>
<td>no</td>
<td>no</td>
</tr>
<tr>
<td>2010-1</td>
<td>17–18. Jan</td>
<td>53</td>
<td>170</td>
<td>110.000*</td>
<td>0.19</td>
<td>yes</td>
<td>yes</td>
</tr>
<tr>
<td>2010-2</td>
<td>24–25. Jan</td>
<td>27</td>
<td>405</td>
<td>380.000*</td>
<td>0.15</td>
<td>no</td>
<td>no</td>
</tr>
<tr>
<td>2010-3</td>
<td>03–05. Feb</td>
<td>42</td>
<td>370</td>
<td>290.000*</td>
<td>0.16</td>
<td>yes</td>
<td>yes</td>
</tr>
<tr>
<td>2011-1</td>
<td>30. Jan–02. Feb</td>
<td>76</td>
<td>190</td>
<td>40.000*</td>
<td>0.68</td>
<td>no</td>
<td>no</td>
</tr>
<tr>
<td>2011-2</td>
<td>04–05. Feb</td>
<td>25</td>
<td>260</td>
<td>120.000*</td>
<td>0.24</td>
<td>yes</td>
<td>yes</td>
</tr>
<tr>
<td>2011-3</td>
<td>20–21. Feb</td>
<td>22</td>
<td>160</td>
<td>70.000</td>
<td>0.22/0.64</td>
<td>0.027</td>
<td>no</td>
</tr>
<tr>
<td>2011-4</td>
<td>09–11. Mar</td>
<td>40</td>
<td>270</td>
<td>160.000</td>
<td>0.17/0.26/0.43</td>
<td>0.022</td>
<td>no</td>
</tr>
<tr>
<td>2011-5</td>
<td>24. Mar</td>
<td>29</td>
<td>190</td>
<td>140.000</td>
<td>0.075/0.19/0.26</td>
<td>0.014</td>
<td>yes</td>
</tr>
<tr>
<td>2012</td>
<td>13–14. Jan</td>
<td>29</td>
<td>175</td>
<td>50.000</td>
<td>0.89</td>
<td>no</td>
<td>no</td>
</tr>
</tbody>
</table>

### Large events

<table>
<thead>
<tr>
<th>No.</th>
<th>Date of precipitation</th>
<th>P event (mm)$^a$</th>
<th>Q peak ($l \cdot s^{-1}$)</th>
<th>Q event ($m^3$)$^b$</th>
<th>$\alpha$ plateau (d$^{-1}$)</th>
<th>$\alpha$ post-plateau section (d$^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>2009</td>
<td>27. Feb–03. Mar</td>
<td>129</td>
<td>500</td>
<td>not separable</td>
<td>0.0025</td>
<td>0.029 / 0.041 / 0.073</td>
</tr>
<tr>
<td>2010</td>
<td>25. Feb–01. Mar</td>
<td>147</td>
<td>480</td>
<td>2.600.000*</td>
<td>0.0014</td>
<td>0.029 / 0.043 / 0.120</td>
</tr>
<tr>
<td>2012</td>
<td>29. Feb–04. Mar</td>
<td>148</td>
<td>530</td>
<td>not separable</td>
<td>0.0008</td>
<td>0.0086 / 0.020 / 0.031</td>
</tr>
</tbody>
</table>

$^a$ Mean of the three precipitation gauging stations Kafr Malik, Taybeh, and Mazra‘a ash-Sharqiya, located in the highland area.

$^b$ Integrated discharge due to precipitation/recharge event. In case of composite events, at the hydrograph minimum between events, the recession limb of the first event was reconstructed. The calculated recession discharge value was added to the first event and subtracted from the subsequent event. Values are marked with an asterisk (*).
Table 3: Parameters used for the setup and calibration of the reservoir model. Calibration parameters marked in bold typeface.

<table>
<thead>
<tr>
<th>Soil-Epikarst Module (SEM)</th>
<th>Value</th>
<th>Unit</th>
<th>Source / calculation method</th>
</tr>
</thead>
<tbody>
<tr>
<td>R</td>
<td>Long-term mean recharge fraction of P</td>
<td>0.34</td>
<td>-</td>
</tr>
<tr>
<td>A</td>
<td>Catchment recharge area size (for mean R)</td>
<td>49</td>
<td>km²</td>
</tr>
<tr>
<td>P</td>
<td>Daily precipitation</td>
<td>mm</td>
<td>Measured</td>
</tr>
<tr>
<td>ETp</td>
<td>Daily potential evapotranspiration</td>
<td>mm</td>
<td>Calculated (Hargreaves equation)</td>
</tr>
<tr>
<td>ETa</td>
<td>Daily actual evapotranspiration</td>
<td>mm</td>
<td>Calculated (soil water balance)</td>
</tr>
<tr>
<td>PWP</td>
<td>Permanent wilting point</td>
<td>0</td>
<td>mm</td>
</tr>
<tr>
<td>FC 1</td>
<td>Field capacity of SEM 1</td>
<td>70</td>
<td>mm</td>
</tr>
<tr>
<td>FC 2</td>
<td>Field capacity of SEM 2</td>
<td>190</td>
<td>mm</td>
</tr>
<tr>
<td>x</td>
<td>SEM 1 area relative to total catchment area A</td>
<td>0.28</td>
<td>-</td>
</tr>
<tr>
<td>DP 1</td>
<td>Deep percolation from SEM 1</td>
<td>m³</td>
<td>Calculated</td>
</tr>
<tr>
<td>DP 2</td>
<td>Deep percolation from SEM 2</td>
<td>m³</td>
<td>Calculated</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Aquifer Module (AM)</th>
<th>Value</th>
<th>Unit</th>
<th>Source / calculation method</th>
</tr>
</thead>
<tbody>
<tr>
<td>h₀</td>
<td>Conduit outflow level</td>
<td>0</td>
<td>m</td>
</tr>
<tr>
<td>h₁HPR</td>
<td>Head in HPR</td>
<td>m</td>
<td>Calculated</td>
</tr>
<tr>
<td>h₁LPR</td>
<td>Head in LPR</td>
<td>m</td>
<td>Calculated</td>
</tr>
<tr>
<td>h₁</td>
<td>Head in HPR at flow threshold</td>
<td>100</td>
<td>m</td>
</tr>
<tr>
<td>L</td>
<td>Length of restricting conduit</td>
<td>10 000</td>
<td>m</td>
</tr>
<tr>
<td>D</td>
<td>Diameter of restricting conduit</td>
<td>0.8</td>
<td>m</td>
</tr>
<tr>
<td>ε</td>
<td>Surface roughness of restricting conduit</td>
<td>0.196</td>
<td>m</td>
</tr>
<tr>
<td>V</td>
<td>Volume of lower part of HPR</td>
<td>1 200 000</td>
<td>m³</td>
</tr>
<tr>
<td>A₁HPR</td>
<td>Base area of lower part of HPR</td>
<td>12 000</td>
<td>m²</td>
</tr>
<tr>
<td>A₁LPR</td>
<td>Base area of HPR</td>
<td>200 000</td>
<td>m²</td>
</tr>
<tr>
<td>A₂LPR</td>
<td>Base area of LPR</td>
<td>1 000 000</td>
<td>m²</td>
</tr>
<tr>
<td>LEP</td>
<td>Lumped exchange parameter LPR↔HPR</td>
<td>3 500</td>
<td>m² d⁻¹</td>
</tr>
<tr>
<td>z</td>
<td>Exchange coefficient</td>
<td>0.004</td>
<td>d⁻¹</td>
</tr>
<tr>
<td>AEXC</td>
<td>Interface area between LPR and HPR</td>
<td>~900 000</td>
<td>m²</td>
</tr>
<tr>
<td>Q₁HPR</td>
<td>Spring discharge</td>
<td>m³ s⁻¹</td>
<td>Measured / calculated</td>
</tr>
</tbody>
</table>
Highlights

• Complex recession behaviour caused by flow threshold and conduit–matrix cross-flow

• Spring hydrograph and recession analysis of long-term and high-resolution time series

• Simulation of spring hydrograph with a non-linear reservoir model

• Water balance modelling to assess groundwater recharge in a semi-arid karst area