Hydrogeology Journal
DOI 10.1007/s10040-015-1233-6

Impact of managed aquifer recharge on the chemical and isotopic composition of a karst aquifer, Wala reservoir, Jordan

Julian Xanke · Nadine Goeppert · Ali Sawarieh · Tanja Liesch · Jochen Kinger · Wasim Ali · Heinz Hötzl · Khair Hadidi · Nico Goldscheider

Abstract Storm-water harvesting and storage via managed aquifer recharge (MAR) is a promising approach to combat water scarcity in semi-arid regions, but poses a challenge for karst aquifers and regions with highly variable water availability. The infiltration of low-mineralized surface water and its impact on highly mineralized groundwater of a karst aquifer was investigated at Wala reservoir in Jordan over a period of approximately 10 years. The results show significant groundwater-level rise in a wellfield, in response to the yearly average infiltration of about 6.7 million m^3. This corresponds to about 60 % of the yearly average abstraction of about 11.7 million m^3, confirmed by mixing calculations with tritium. A decreasing trend in infiltration due to sedimentation is observed. Mean groundwater residence times of several thousand years, derived from carbon-14 dating, indicate a large storage capacity of the aquifer. The heterogeneous distribution of the residence times is caused by strong groundwater withdrawals and artificial recharge along with karst-specific aquifer characteristics. Temporal groundwater salinity fluctuations in the wellfield are observed after the first MAR infiltration. Enhanced groundwater flow along the wadi course was demonstrated, which is an important aspect with regards to future MAR projects in similar wadis of the region.

Keywords Managed aquifer recharge · Hydrochemistry · Semi-arid climate · Karst · Jordan

Received: 29 May 2014 / Accepted: 29 January 2015

Electronic supplementary material The online version of this article (doi:10.1007/s10040-015-1233-6) contains supplementary material, which is available to authorized users.

J. Xanke (✉) · N. Goeppert · T. Liesch · J. Kinger · W. Ali · H. Hötzl · N. Goldscheider
Karlsruhe Institute of Technology (KIT),
Division of Hydrogeology, Adenauerring 20b, 76131, Karlsruhe, Germany
e-mail: julian.xanke@kit.edu
Tel.: +49-721-608-41968

A. Sawarieh
Natural Resource Authority, P.O.Box 711118, Amman, Jordan

K. Hadidi
Water Authority Jordan, P.O. Box 241211183, Amman, Jordan

Introduction

Access to safe drinking water is a major challenge in many regions of the world (WHO 2004), especially in countries with a fast-growing population and a natural deficit of water that is prone to high seasonal variability. In Jordan, these factors come together and limit the availability of freshwater resources and their management, hence, limiting the supply of domestic water. Furthermore, overexploitation of Jordan's groundwater resources has led to a lowering of the water tables in the last few decades (El-Naqa and Al-Shayeb 2009). Besides pumping from alluvial aquifers and deep sandstone aquifers, the major water sources are the prevailing carbonate formations found in large parts of the country (MWI 2004).

Managed aquifer recharge (MAR) is an important approach to provide people with water, especially in semi-arid and arid regions (Dillon 2005) and to counteract falling water tables (Bouwer 2000; Mays 2009). Furthermore, Wolf et al. (2007) and Daher et al. (2011) refer to its important role in integrated water resources management (IWRM). Common MAR techniques for underground storage of water are described by Dillon (2005), Dillon et al. (2009) and Bouwer (2002). However, the global number of MAR examples in karst aquifers is small compared to those in porous media (Daher et al. 2011) because MAR implementation in karst poses a particular challenge due to strong hydraulic heterogeneity of the underground (Bakalowicz 2005; Einsiedl 2005). This is characterized by the occurrence of variable storage and flow conditions, which are attributed to the varied porosities of the rock matrix, fractures and conduits (Goldscheider and Drew 2007). An intensified karstification along riverbeds is proposed by Green et al. (2014) for carbonate aquifers in semi-arid regions.

Published online: 17 March 2015

For the purpose of MAR in karst terrains, especially along wadis, high recharge infiltration rates, fast hydraulic response of the aquifers and carbonate dissolution are expected and often come with challenges. Thus, the reduction of reservoir infiltration due to sediment accumulation often causes significant problems (Bouwer 2002). Furthermore, Vanderzalm et al. (2010) describe the dominant influence of calcite dissolution on the change of the water chemistry, which can be intensified below dam sites and reservoirs (Dreybrodt et al. 2002; Hiller et al. 2011; Romanov et al. 2007). The contamination of abstraction wells can be minimized by keeping a sufficient distance from the zone of infiltration (Masciopinto et al. 2012) while being aware that karst aquifers reveal a wide range of groundwater flow velocities.

In order to make use of the limited knowledge about the local hydraulic characteristics of karst aquifers, derived from hydraulic tests, geophysical measurements and water-level observations (Daher et al. 2011), additional hydrochemical data need to be obtained, for example, to identify possible interactions of different water sources. Major ions in the water can indicate chemical processes (Lakshmanan et al. 2003), whereas the comprehension of the geochemical history and the hydrological conditions of the water can be obtained from isotopic data. Oxygen-18 ($\delta^{18}O$) and deuterium ($\delta^{2}H$) create a relationship between groundwater and its place of origin, whereas tritium ($\delta^{3}H$) and carbon-14 allow for determination of the mean residence time of young groundwater (<60 years; Gonfiantini et al. 1998) and old groundwater, respectively (Plummer 2005). This information is suitable, for example, to identify and avoid the exploitation of fossil groundwater. Additionally, both applications are useful for documenting the mixing of different waters (Etcheverry and Vennemann 2009).

The Wala reservoir is one of the larger MAR test sites in carbonate aquifers, with a mean annual recharge of around 6.7 million cubic meters (MCM; 2002–2012). Water is recharged mainly by natural seepage from the reservoir and small amounts via recharge wells (since 2011) and then recovered at Hidan wellfield. The wellfield serves as an important drinking-water supplier of the cities of Amman, Madaba and nearby communities.

This study presents the evaluation and interpretation of hydrochemical and isotopic data, recharge rates from the reservoir, abstraction rates and water level records from the aquifer, which have been obtained from the database of the Ministry of Water and Irrigation (MWI), Jordan. It is intended to understand and assess the dynamic system of natural recharge and groundwater flow as well as the spatiotemporal impact of artificial recharge and abstraction on the karst aquifer. The central research questions are concerned with (1) the proportion of reservoir infiltration in the abstracted groundwater and the associated groundwater level fluctuations, (2) the influence of sedimentation on the reservoir infiltration, (3) the impact of recharge on the groundwater salinity, and (4) an understanding of the extent of rock dissolution. The results of this study serve as the basis for a conceptual groundwater flow model of the Wala area and as a reference example for future MAR operations in karst aquifers in semiarid regions.

Site description

Geographical setting

Jordan's landscape is mainly characterized by the Jordan Valley, the adjacent mountain range and the mountain plateau extending to the east of it. The Wala reservoir is located in the mountain range around 40 km south of Jordan's capital, Amman, at an elevation of 485 m above sea level (asl) (Fig. 1a). Its surface catchment is a part of the Mujib drainage basin and the Dead Sea groundwater basin (Al-Assa'd and Abdulla 2010) and extends east of the reservoir (Fig. 1b) with a total area of around 1,770 km^2 and an elevation of around 1,000 m asl in the northeast to around 960 m asl in the southeast.

The climate in Jordan is semiarid with rainfall mainly during winter. Rainfall data in Wala catchment reveal an increasing trend from north to south with mean annual values of 500 mm (north) and 100 mm (south), whereas the rainfall maximum can reach 800 mm (north) and 250 mm (south). Dry years supply less than 200 mm (north) to 50 mm (south). It can be inferred that most of the runoff is generated in the northern part of the catchment. Mean annual potential evaporation ranges from 2,300 mm (north) to 3,000 mm (south; Margane et al. 2009). The evaporation rate of the Mujib basin ranges from 85 to 93 % and the infiltration rate from 3 to 11 % of the total precipitation (El-Naqa 1993). The remaining rainfall generates surface runoff, which ranges from 0.5 to 10 % of the total rainfall, and flows via numerous tributaries into Wadi Wala. On its way downwards, water passes into Wadi Hidan (variant spelling: Haidan, Heidan, Heedan) and finally discharges into the Dead Sea. Groundwater flow generally follows the topographic gradient from east to west (Fig. 1b).

The Hidan wellfield is located around 7 km west of Wala dam at an elevation of around 350 m asl and comprises 16 active abstraction wells with depths between 100 and 200 m. Wala abstraction wells are spread along the wadi and in the valley between the adjacent hills, between the Hidan wellfield and the dam, with depths between 100 and 400 m (Fig. 1c).

Geology and hydrogeology

The Wala reservoir and its catchment are embedded into a sequence of Upper Cretaceous to Eocene sedimentary rocks (Fig. 2), predominantly limestone, dolomite and chalk, and partly covered with Quaternary deposits. The structural setup of the area is within the tectonic framework of the Jordan Rift Valley, which has been tectonically active mainly in the past 12 million years (Bayer et al. 1988).

Marl and limestone sequences of the A5/6 formation (Fig. 1d) underlie the regional aquifer system of the Wadi

DOI 10.1007/s10040-015-1233-6

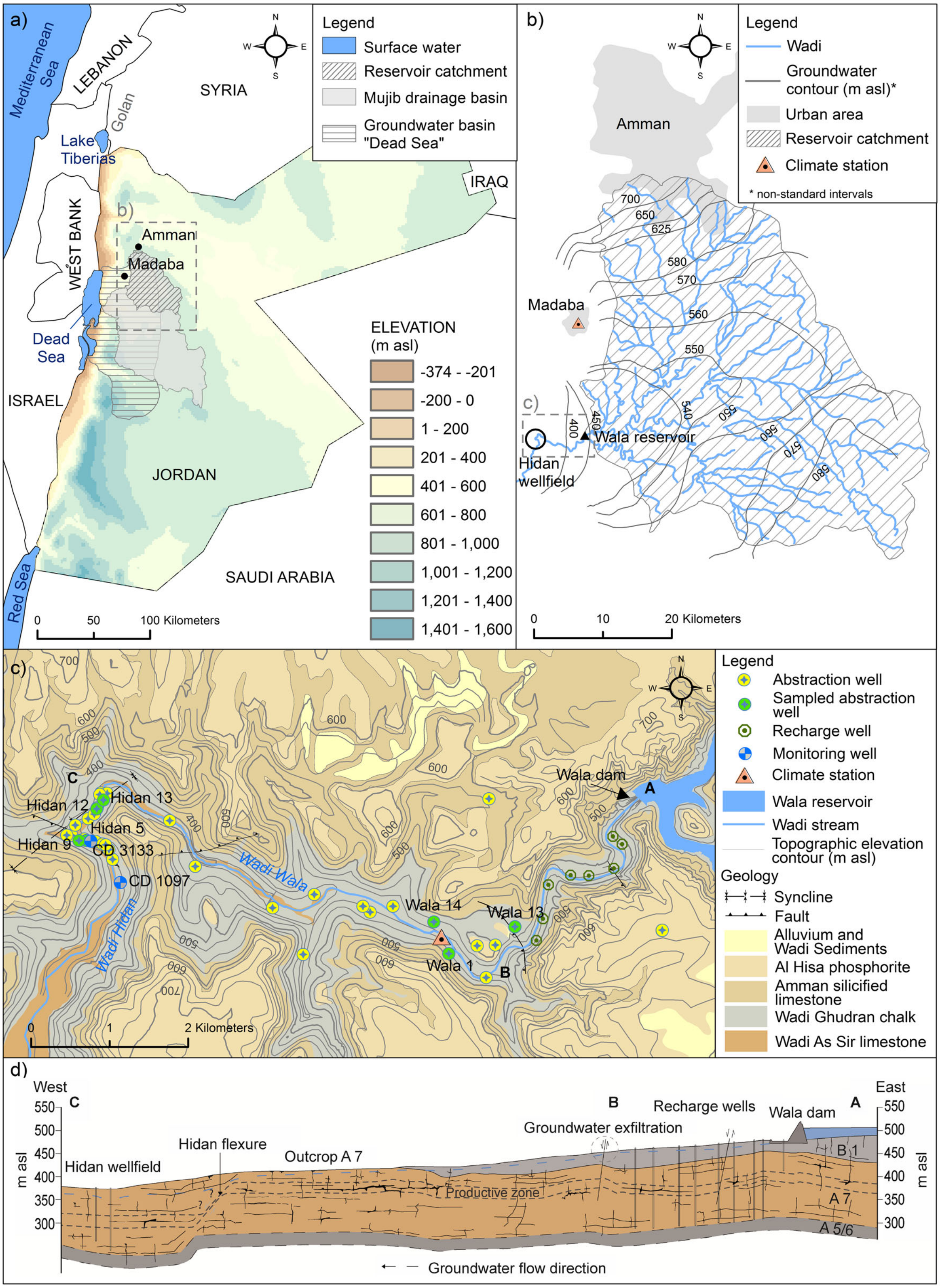

Fig. 1 **a** Geographical setting of Jordan and location of the Dead Sea groundwater basin, the Mujib drainage basin and the surface catchment of Wala reservoir. **b** The Wala reservoir catchment with groundwater flow direction and the location of the reservoir and Hidan wellfield. **c** The geology of the Wadi Wala area, recharge, abstraction and observation wells (modified after Al Hunjul 1993). **d** Schematic geological profile along Wadi Wala with the location of the recharge wells and abstraction wells (modified after Humphreys 1991, stratigraphy as Fig. 2)

As Sir (A7) formation. This consists of a well-bedded hard and dense limestone, where, in some studies, thin beds of evaporates such as gypsum, anhydrite and halite are observed (Powell and Moh'd 2011; El-Naqa 2004). The

 DOI 10.1007/s10040-015-1233-6

Period	Epoch	Group	Formation: rock type	Symbol	Inferred thickness (m)	Aquifer type	
Tertiary	Paleocene	Belqa	Umm Rijam: limestone, chalk, chert	B4	120-150	Aquitard	
	Maastrichtian/Paleocene		Muwaqqar: chalk, marl, limestone, chert	B3	~100	Aquitard	
Upper Cretaceous	Maastrichtian		Al Hisa: phosphatic limestone, chert	B2b	40-65	Aquifer	
	Campanian		Amman: limestone, chert	B2a	75	Aquifer	
	Santonian		Wadi Ghudran: chalk; Member: Dhiban	B1c	12-15	Aquitard	Reservoir
			Member: Tafilah	B1b	~ 50	Poor aquifer	
			Member: Mujib	B1a	12-15	Aquitard	
	Turonian	Ajlun	Wadi As Sir: limestone	A7c	~120-150	Aquifer	Recharge
				A7b		Aquifer	
				A7a		Aquifer	
	Cenomanian		Shueib: marl, limestone	A5/6	100-165	Aquitard	

Fig. 2 Stratigraphy of formations at the Wala reservoir test site (after Humphreys 1991; Margane et al. 2002)

limestone outcrops along Wadi Wala around 3 km east of Hidan wellfield, constituting the main regional aquifer system, and can be subdivided into three hydraulic zones with less permeable parts on the top and the bottom and a productive zone in between. Groundwater flow takes place primarily horizontally along the bedding planes, which are enlarged due to rock dissolution. Vertical flow takes place mainly along joints and fractures (Humphreys 1991). Numerous swallow holes are present along the wadi, filled and traversed by the surface water, which is discharged into the wadi from temporal springs around 2 km downstream of the Wala reservoir (Fig. 1d). Higher karstification along the wadi course and thus higher permeability values for the A7 outcrops along the wadi can be assumed.

The Ghudran formation builds the basement of Wala reservoir and is divided into three members: Dhiban (B1c), Tafilah (B1b) and Mujib (B1a). The first and third members consist mainly of chalk and act as aquitards. The second consists of chalk and limestone beds and is considered to be a poor aquifer. The basement of the limestone and chert sequences of the Amman (B2) formation starts at the top of Wala dam construction and reveals subhorizontal bedding with several minor fold structures and faults. Outcrops of the B3 and B4 formation are found in the Wala catchment and along the wadi and consist of chalk, limestone and chert with thin beds of gypsum and halite (B3) and of limestone, chalk and chert (B4; Humphreys 1991; Bender 1968; Margane et al. 2002).

Around 5 km west of the dam, a flexure crosses the wadi from E to WSW with a dipping of around 40–60° to the north. There, the Ghudran formation is constrained and thinned out between the A7 and B2 formation and might act as a hydraulic barrier for horizontal groundwater flow but promotes the vertical flow along numerous fractures. The axis of a syncline follows the direction of the wadi (NE–SW) at Hidan wellfield and crosses another small flexure (SE–NW) orthogonally (Fig. 1c).

Methodology

Available data and methods

Hydrological, hydrogeological and hydrochemical data were obtained from the database of the Ministry of Water and Irrigation (MWI). The available data and the applied methods for analyzing the chemical constituents and physical parameters are listed in the following:

- Electrical conductivity (EC) and pH were analyzed in situ with field equipment.
- Bicarbonate was determined by titration in the laboratory, 2–4 h after sampling.
- Major anions and cations were measured using an ion chromatography method.
- Oxygen-18 was analyzed with the CO_2 equilibrium/Delta Plus XP isotope ratio mass spectrometer (hydrogen deuterium oxygen, HDO), using a platinum catalyst method. Precision is±0.15 ‰, according to Bajjali and Abu-Jaber (2001) and Bajjali (2006).
- Deuterium was analyzed using the Delta Plus XP isotope ratio mass spectrometer (hydrogen deuterium oxygen, HDO), using the platinum catalyst method. Precision is±1 ‰, according to Bajjali and Abu-Jaber (2001) and Bajjali (2006).
- Tritium was determined using the electrolytic tritium enrichment method. Precision is±1 TU, according to Bajjali and Abu-Jaber (2001) and Bajjali (2006).
- ^{14}C was measured with a benzene synthesis line and liquid scintillation counter.
- ^{13}C was measured by mass spectrometry.

Time series of hydrogeological data:

- Monthly water level from observation wells CD 3133 and CD 1097.
- Daily water level from Wala reservoir. Geometric data of Wala reservoir include the water surface area and reservoir volume for each centimeter of water level.

DOI 10.1007/s10040-015-1233-6

- Daily precipitation (rainfall collector) and evaporation (measured with a class A pan) from Wadi Wala station.
- Monthly abstraction rates from Hidan and Wala wells.

Isotope data exist from 1994 to 2010, whereas major ions, temperature and EC are available from 2001 to 2011. Data from 1994–2010 for major ions are not available, but additional EC data exist from 1982 to 2000 from 342 samples. Groundwater level data are available from 1994 to 2012. Data gaps in the time series exist due to irregular operation of the wells. Most of the abstraction wells are equipped with taps for manual control, which allow permanent or individual sampling at any time. All data are listed in the electronic supplementary material (ESM).

Major ions

Data from 51 water samples are taken from the Wala reservoir (2003–2006), two Hidan wells (2001–2011) and two Wala wells (2003–2009). Their locations are displayed in Fig. 1c. Wells with only one sample were not considered. The differentiation of the water types was done on the basis of the major ions Na^+, K^+, Mg^{2+}, Ca^{2+}, Cl^-, SO_4^{2-} and HCO_3^-. To check the reliability of the results, the error of the charge balance of the ions was calculated to be less than 5 %, which excluded around 20 % of the available data. Total dissolved solid (TDS) values were deduced from the EC (TDS≈0.65 × EC). In order to calculate the saturation index (SI), samples with missing temperature measurements were complemented by the average temperature from other samples of the same well.

Isotopes

Existing tritium (^3H), deuterium (δ^2H) and oxygen-18 ($\delta^{18}O$) data are evaluated for 3 Wala wells, 14 Hidan wells, 6 recharge wells, precipitation from Madaba climate station (Fig. 1b) and the Wala reservoir. The timeframes of the 133 tritium and 163 deuterium and oxygen-18 samples differs greatly, and for several wells only one record exists. A few data are available from 1990 at Hidan wellfield; all other data are from 2002 to 2010. Rainfall data already exist from 1987 until 2000, while data for the reservoir exist from 2002 to 2010.

The δ^2H and $\delta^{18}O$ data have been evaluated to determine the origin of groundwater and to identify possible mixing of water with different isotopic signatures. Therefore, the values are put in relation to the global meteoric water line (GMWL) (Craig 1961). The rainfall of regional origin is represented by the Mediterranean meteoric water line (MMWL; Gat et al. 1969).

Additional information about mixing ratios has been obtained from tritium data. Here, mean residence time of groundwater can be determined for timeframes of less than 60 years, due to its half-life of around 12.43 years (Etcheverry and Vennemann 2009). Values of 0.8–4 tritium units (TU) in groundwater point to the mixing of old and recent water, whereas values between 5 to 10 TU come from recent recharge (after Clark and Fritz 1997, in Ofterdinger et al. 2004).

Carbon-14 and carbon-13 data of one Hidan and two Wala wells are available from 2002 to 2006, to determine the groundwater age of 8 samples. The method is based on the fact that ^{14}C cuts off from atmospheric exchange when it infiltrates underground. From that point, it can be used for age dating of groundwater bodies, due to its radioactive decay and half-life of around 5,730 years (Gonfiantini and Zuppi 2003; Etcheverry and Vennemann 2009). The formula for age dating is (Plummer and Glynn 2013):

$$T = \frac{5730}{\ln 2} \ln \left(\frac{\delta^{14}C_{\mathrm{Initial}}}{\delta^{14}C_{\mathrm{Measured}}} \right) \quad (1)$$

where T is the time and 5730 is the modern half-life of ^{14}C.

In order to solve the formula for T, the initial ^{14}C activity is required, which is always reported in percent modern carbon (pmc). For calibration, the oxalic acid standard is used, where 95 % of the standard corresponds to 100 % pmc (Gonfiantini and Zuppi 2003; Meredith 2009). It is known from previous studies that dissolution and precipitation of $CaCO_3$ influences the ^{14}C content in groundwater. Therefore, to estimate the source of the different carbon quantities, the ^{13}C content of the total dissolved inorganic carbon (DIC) can be consulted (Kattan 1995). The ^{13}C enrichment and ^{14}C depletion in DIC is caused by the isotope exchange between DIC and carbonate rock (Pearson and Hanshaw 1970; Geyh 1970). ^{13}C activities are related to the Vienna Pee Dee Belemnite (VPDB) standard. Several models have been developed in the past decades to estimate the initial ^{14}C activity (Vogel 1968; Tamers 1967; Eichinger 1983; Salem et al. 1980; Evans et al. 1979; Pearson and White 1967; Fontes and Garnier 1979).

The Fontes and Garnier (1979) equation has been chosen for this study, because it considers most of the chemical processes in carbonate aquifers such as the carbonate, gypsum and soil gas CO_2 dissolution, and the exchange of CO_2 in the gas and aqueous phase, as well as the calcite-bicarbonate exchange (Plummer and Glynn 2013). For comparison, the differing results from the IAEA model (Plummer and Glynn 2013) and Evans model (1979) are also applied. The IAEA model is a further development of the Pearson model, which is based on the isotopic mass balance relation of ^{13}C and ^{14}C, and considers the equilibrium of bicarbonate with the gas phase and the subsequent dissolution of solid carbonate. The Evans (1979) model considers the isotopic fractionation, due to the dissolution and precipitation process of calcium carbonate in groundwater, when equilibrium between calcite and dissolved bicarbonate is supposed (Plummer and Glynn 2013). For the calculations, the following basic assumptions are made according to studies from Kattan (1995) and Kattan (2001) in Syria: initial ^{14}C

DOI 10.1007/s10040-015-1233-6

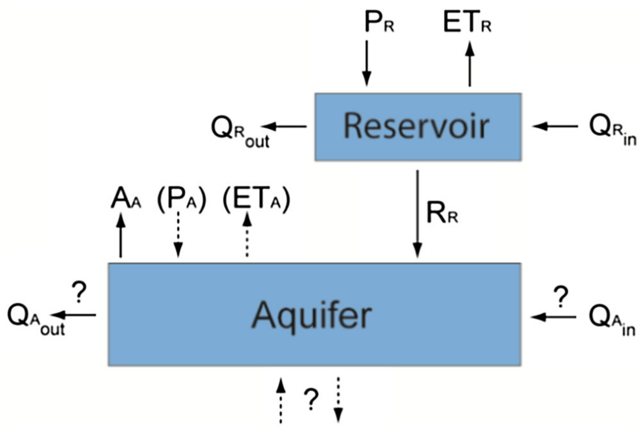

Fig. 3 Chart of the water balance of the Wala reservoir and the aquifer

activities in the rock are 0 and 1 ‰ for ^{13}C. Activities in soil CO_2 are 100 ‰ for ^{14}C and −21 ‰ for ^{13}C.

Water balance

The amount of input and output from the reservoir is reflected in the daily water level changes and calculated with the related reservoir volume increase (input) or decrease (output). The surface water inflow (Q_{Rin}) is calculated by subtracting the amount of precipitated water (P_R) on the lake surface from the total water volume increase (input). The aquifer recharge (R_R) is calculated by subtracting the daily amount of evaporated (ET_R) water from the total water volume decrease (output). The height of precipitation and evaporation is calculated by multiplying the daily values with the water surface area from the corresponding day (see Fig. 3).

A simultaneous determination of Q_{Rin} and R_R cannot be made because both factors depend on water level increase and water level decrease respectively. Therefore, calculation errors occur on days with inflow into the reservoir. The amount of surface outflow (Q_{Rout}) is measured and calculated by the Ministry of Water and Irrigation (MWI 2012). The aquifer output is determined using the monthly abstraction rates from the wells (A_A). The provided amount from natural groundwater flow is unknown (Q_{Ain} and Q_{Aout}). Direct recharge along the wadi between the reservoir and wellfield by precipitation (P_A), and by loss through evapotranspiration (ET_A) are neglected, as well as possible water exchange with a deeper aquifer.

Results and discussion

Water balance

From 2002 to 2012, the reservoir received around 136 MCM of floodwater during the winter seasons, while around 52 MCM were lost via the spillway and around 84 MCM were stored. Approximately 74.1 MCM of the stored water infiltrated into the ground naturally and around 7.8 MCM were lost by evaporation. The difference of 2.1 MCM is composed of the roughly 1.5 MCM of remaining water in the reservoir and a loss that is not covered by the calculations .

The annual amount of infiltration ranges from around 1 MCM in 2008 to around 13 MCM in 2003 with an average of around 6.7 MCM. A clear decrease in infiltration can be observed (Fig. 4) with time, due to ongoing sedimentation in the reservoir. Thus, the initial storage capacity of around 9.3 MCM is reduced to approximately 7.7 MCM in 2012, which was caused from the increasing sediment in the reservoir. The total infiltration from 2002 to 2012 facing complete abstraction of around 129 MCM from Hidan and Wala wells, meaning roughly 57 % of the abstracted water is provided by the reservoir. The proportion of infiltrated water on the abstracted water was about 71 % in the period from 2002 to 2007 and decreased to around 42 % from 2008 to 2012. Yearly abstraction rates range from around 10 MCM in 2002 to more than 13.7 MCM in 2008 with an average of 11.7 MCM, where around 96 % is used for drinking-water supply and around 4 % for irrigation. Overexploitation, accompanied by the decreasing infiltration rate from the reservoir, causes water level drop in Hidan wellfield, as observed in 2008/2009 and 2011/2012 (Fig. 4). The yearly amount of groundwater flow to the

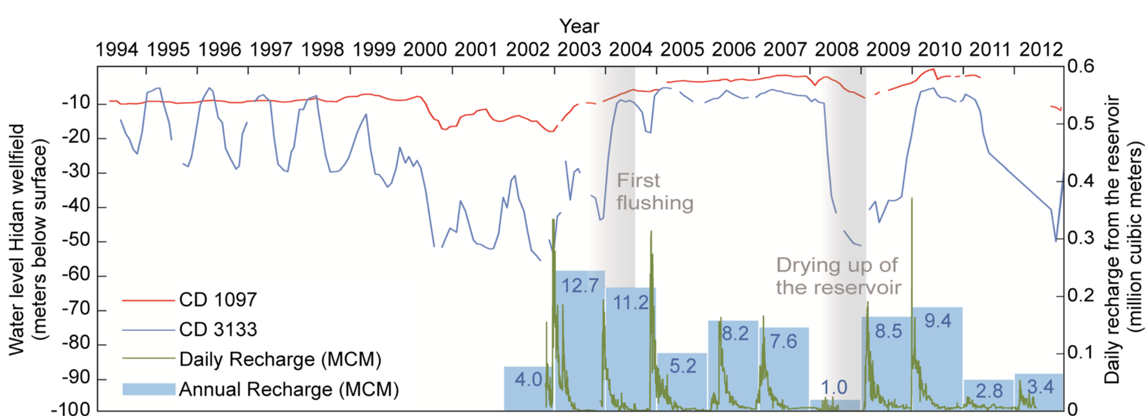

Fig. 4 Infiltration from the aquifer and water level fluctuations in observation wells *CD 1097* and *CD 3133* of Hidan wellfield (Fig. 1c)

 DOI 10.1007/s10040-015-1233-6

Hidan wellfield is difficult to determine due to variations in natural recharge and withdrawals in the catchment. Assuming the mean annual abstraction rate of around 11.7 MCM and artificial recharge of around 7.3 MCM keep the water level constant, the natural groundwater flow provides a minimum of 4.4 MCM on average per year to Hidan wellfield

General hydrogeochemistry

Table 1 presents a summary and statistical analysis of the available data from 2001 to 2011. The mean pH of the reservoir is 7.9, and for the groundwater samples is between 6.9 and 7.7. Temperature data are rarely documented and range between 23.5 and 30.5 °C in the groundwater and 23.0 °C in the reservoir. Water temperature in the reservoir is taken from the surface and does not represent the entire lake. Electrical conductivity (EC) data show higher values in Wala wells than in Hidan wells and indicate a better connection of Hidan wells to the conduit and fracture network, where flow velocities are generally higher and rock–water interaction is less. The significant differences in the average ion concentration concern mainly sulfate, chloride and bicarbonate. However, differences occur also among Hidan wells. Here, the impact of less-mineralized surface water from wadi infiltration at the outcrops of A7 aquifer (Fig. 1d) might be marginal. EC values of reservoir water are generally lower and vary due to occasional filling with low-mineralized floodwater or evaporation processes during summer. The average EC of around 470 μS/cm reveals that surface water has already dissolved rock during its flow through the soil and epikarst of the catchment.

The Piper diagram (Fig. 5) reveals a dominance of calcium and bicarbonate in groundwater and surface water, as is expected from limestone and dolomite. The anion triangle shows that the percentage of bicarbonate in surface water is much higher than in groundwater, whereas cations are much more clustered. The order of the abundance of the ions in both waters is Ca^{2+}>Na^{+}>Mg^{2+}>K^{+} and HCO_3^{-}>Cl^{-}>SO_4^{2-}.

Mean nitrate values of 4.2 mg/L for the reservoir are lower than expected from the influence of agricultural and farming activities in the catchment as well as the sewage from nearby communities. The low values can be explained by nitrate consumption by cyanobacteria and algae at the water surface. Mean concentrations in Wala and Hidan wells range from 11.2 to 27.5 mg/L and indicate return flow from irrigation areas around the wells carrying ammonia (NH_3), which is turned into nitrate by nitrification (Schmoll 2006). Potassium in reservoir water samples ranges between 7 and 15.6 mg/L, whereas groundwater values range from around 3 to 12 mg/L. Here, the influence of fertilizers is likely.

The excess of Cl^{-} over Na^{+} (Fig. 6b) in the groundwater points to additional sources of chloride, e.g. $CaCl_2$, $MgCl_2$ or KCl, which explains the slight excess of Ca^{2+} and Mg^{2+} over SO_4^{2+} and HCO_3^{-} in some groundwater samples. Higher Na^{+} values in the reservoir

Table 1 Physico–chemical parameters of 51 Wala reservoir and groundwater samples from 2001 to 2011

SampleID	Statistic	pH	Temp (°C)	EC (μS/cm)	TDS (mg/L)	Na (mg/L)	K (mg/L)	Mg (mg/L)	Ca (mg/L)	Cl (mg/L)	SO_4 (mg/L)	HCO_3 (mg/L)	NO_3 (mg/L)	SI Calcite	SI Aragonite	SI Dolomite	SI Gypsum	SI Anhydrite	pCO2 in vol % of [atm]
Reservoir	Min	7.5	23.3	231	192	18.2	5.9	4.1	23.7	16.0	13.0	99.4	0.8	0.1	−0.1	−0.3	−2.7	−2.9	0.1
(No. =7)	Max	8.3	23.3	633	876	95.2	15.6	53.4	84.0	147.7	146.9	372.1	9.5	1.2	1.1	2.5	−1.3	−1.5	1.1
	Ave	8.0	23.3	462	438	37.2	9.6	19.8	51.1	47.3	39.0	230.1	4.2	0.6	0.4	0.9	−2.2	−2.4	0.3
	SD	0.2	0.0	151	218	24.7	3.1	15.8	19.9	42.7	48.5	96.7	2.6	0.4	0.4	0.9	0.4	0.4	0.3
Wala 1	Min	7.2	27.5	1092	710	75.7	3.1	34.1	95.2	104.0	74.9	309.3	13.9	0.2	0.1	0.4	−1.5	−1.7	1.1
(No. = 8)	Max	7.5	27.5	1501	1046	123.9	12.1	61.4	109.8	182.8	186.2	410.5	42.2	0.6	0.5	1.3	−1.1	−1.3	2.5
	Ave	7.3	27.5	1341	916	98.0	8.7	46.9	104.5	154.7	153.8	357.4	27.5	0.4	0.3	0.8	−1.2	−1.4	1.8
	SD	0.1	0.0	139	121	13.7	2.9	8.4	4.4	22.8	35.7	29.6	9.1	0.1	0.1	0.3	0.1	0.1	0.4
Wala 14	Min	6.6	30.2	1278	831	84.0	3.9	45.5	100.2	131.0	138.7	343.4	6.8	−0.2	−0.3	−0.3	−1.2	−1.4	0.6
(No. = 10)	Max	7.9	32.2	1430	1024	97.8	7.0	63.4	123.9	170.4	191.5	425.2	15.6	1.1	1.0	2.4	−1.0	−1.2	12.3
	Ave	7.4	31.0	1327	942	88.9	5.4	50.1	110.5	143.8	169.5	386.9	11.2	0.6	0.4	1.2	−1.1	−1.3	2.6
	SD	0.3	0.9	45	63	4.1	1.0	4.9	8.1	11.2	15.3	22.2	2.8	0.3	0.3	0.7	0.0	0.0	3.3
Hidan 9	Min	7.1	23.5	767	546	36.8	3.1	27.2	78.8	59.3	65.3	265.4	17.8	0.1	−0.1	0.1	−1.6	−1.8	0.2
(No. = 13)	Max	8.2	28.2	1240	772	70.4	5.9	47.1	91.8	127.8	101.7	323.9	33.7	1.1	1.0	2.1	−1.4	−1.6	2.9
	Ave	7.8	26.2	931	641	56.2	4.4	35.0	85.4	93.7	79.1	300.1	23.3	0.7	0.6	1.4	−1.5	−1.7	0.7
	SD	0.3	2.0	105	68	7.5	1.0	5.9	3.8	15.0	9.5	18.6	3.8	0.3	0.3	0.5	0.0	0.0	0.7
Hidan 13	Min	7.0	25.5	928	603	62.8	3.1	32.7	88.0	106.5	80.2	305.0	17.3	0.1	0.0	0.2	−1.5	−1.7	0.3
(No. = 13)	Max	8.1	29.3	1271	830	80.5	6.7	48.0	110.0	147.3	135.4	361.7	28.3	1.1	0.9	2.1	−1.2	−1.5	3.5
	Ave	7.6	27.2	1090	758	73.1	5.2	38.0	98.3	123.0	102.5	337.4	21.9	0.6	0.5	1.2	−1.4	−1.6	1.1
	SD	0.3	1.6	86	70	5.9	0.9	4.0	5.5	9.2	13.5	15.8	3.2	0.3	0.3	0.6	0.1	0.1	0.8

Min minimum, *Max* maximum, *Ave* average, *SD*standard deviation, *N*number of samples. Temperature data are rarely measured and are completed by adding the average value of the sampling point

DOI 10.1007/s10040-015-1233-6

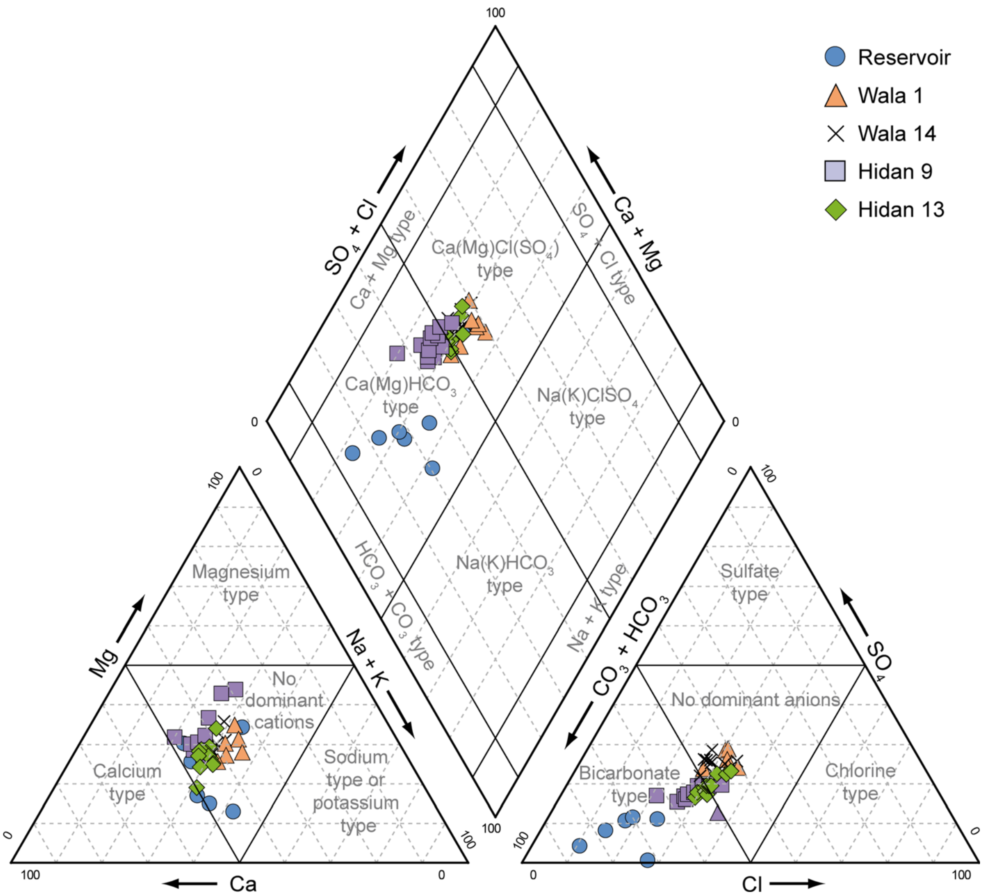

Fig. 5 Piper diagram displaying water samples from Wala *reservoir*, Wala wells *1* and *14*, and Hidan wells *9* and *13*. All samples are slightly dominated by calcium and bicarbonate; higher proportions of bicarbonate are present in the reservoir

water might come from cation exchange, where Ca^{2+} is replacing the Na^{+} on clay minerals. However, Na^{+} and K^{+} are also the result of silicate dissolution (Bakalowicz 1994), whereas high Cl^{-} and SO_4^{2-} content is often associated with wastewater.

Furthermore, the groundwater samples reveal that Cl^{-} is contributing more to the salinity than SO_4^{2-} (Fig. 6c). The same applies for Ca^{2+} over Mg^{2+}, which is explained by the presence of Ca^{2+} in more minerals than Mg^{2+}. Only a few samples contain the same amount of Mg^{2+} and Ca^{2+} (Fig. 6d).

Almost all water samples show slight supersaturation with respect to calcite, dolomite and aragonite. The fact that $SI_{Calcite}$ values over 0.5 are always associated with a CO_2 partial pressure of less than 1 % of the atmospheric pressure indicates degassing of CO_2 from almost half of the water samples. Consequently, SI values are computed to be too high and only a slight supersaturation can be assumed. Only one sample reveals a CO_2 partial pressure above 10 %. Such high values are typical for active tectonic areas (Bakalowicz 1994) and point to CO_2 originating from a deeper aquifer. Saturation of gypsum, anhydrite and sodium chloride is not reached due to their high solubility.

The difference of the average ion concentration in the surface (reservoir) and groundwater (Hidan well 9+13) and the total amount of infiltrated water from 2002 to 2012 (74 MCM) is taken for a rough calculation of the total volume of dissolved rock (only calcite, gypsum and halite are considered), which resulted in a value of around 6,500 m^3. Around 1,550 m^3 of it is attributed to calcite, 2,400 m^3 to halite and 2,550 m^3 to gypsum. The total volume of dissolved rock is less than 0.004 % of the volume of the productive aquifer section along the wadi corridor (≈1 km). However, the hydraulic impact of this dissolution cannot be evaluated by simple means, as it strongly depends on the spatial distribution pattern of the dissolution phenomena.

Investigations of Goode et al. (2013) on salinity trends in Jordan ascribe its increase mainly to the lower parts of basins and areas of discharge and withdrawals, influenced by the natural recharge and the geological setting. However, water-table declines do not necessarily correlate to salinity increases. The EC values of 342 samples, taken from 18 Hidan and 12 Wala wells, indicate a significant increase of the salinity from 1980 to 2011 (Mann Kendall trend test, p-value for Hidan wells 3.3E-3; for Wala wells <1E-4), visualized in Fig. 7a with average values of all wells and samples from the 1980s, 1990s and 2000s. Similar observations for salinity increase were done by Schmidt et al. (2013) in a comparable aquifer system at the western margin of the Lower Jordan Valley and

 DOI 10.1007/s10040-015-1233-6

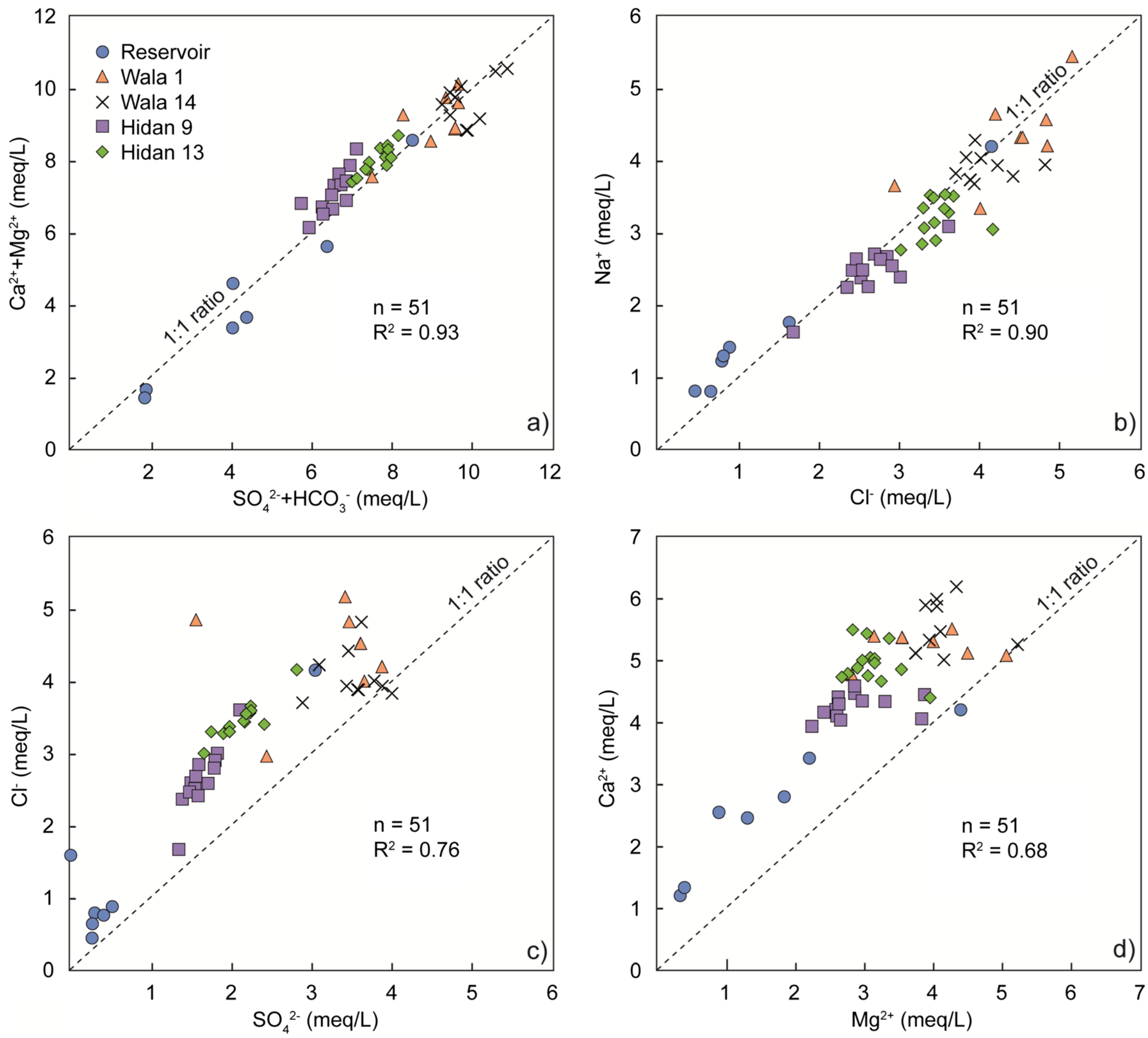

Fig. 6 Binary diagrams from waters of Wala reservoir, Wala wells and Hidan wells **a** $Ca^{2+}+Mg^{2+}$ vs. $SO_4^{2-}+HCO_3^-$, **b** Na^+ vs. Cl^-, **c** Cl^- vs. SO_4^{2-} and **d** Ca^{2+} vs. Mg^{2+}

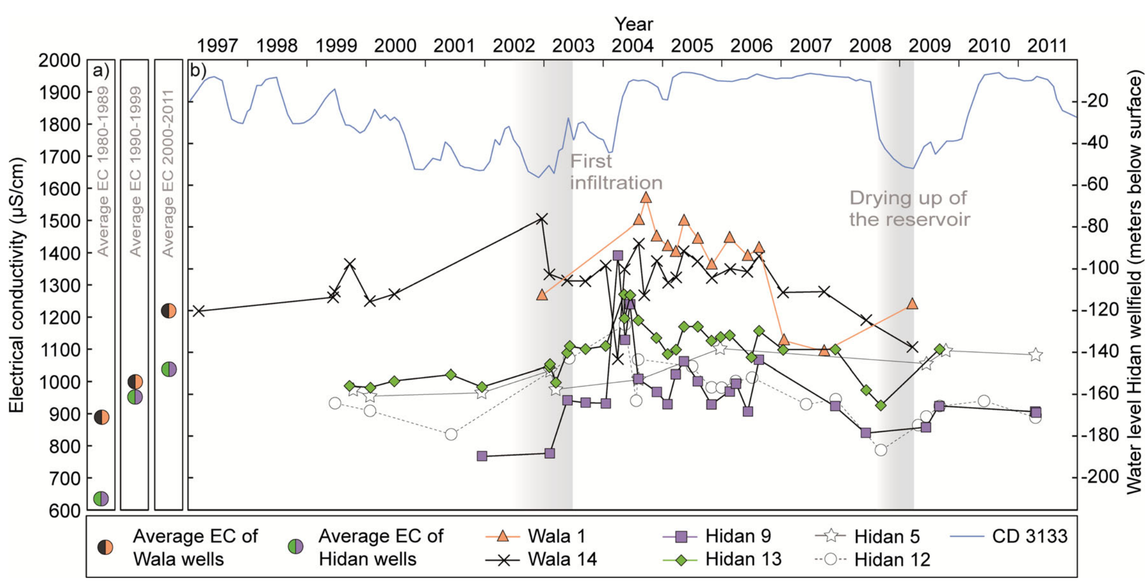

Fig. 7 **a** Electrical conductivity changes at Hidan and Wala wells, and **b** water levels in individual wells. During the first flushing of the aquifer the salinity increased in all wells (between 2003 and 2005). The infiltration from the Wala reservoir water decreased from 2007 and the reservoir dried up in 2008, which caused a decrease in the EC of water in the Hidan wells. Long-term increase of average EC is observed from 1980 to 2011

 DOI 10.1007/s10040-015-1233-6

attributed to progressive impact of wastewater. A final conclusion about the reason for long-term salinity increase in the study area is not possible due to missing a detailed survey of groundwater chemistry and water level. Otherwise, changes could be attributed to a specific water constituent, or temporal EC variations could be identified and correlated with groundwater levels and abstraction rates. Therefore, the impact of wastewater, along with declining water tables in the entire catchment, is probable. Substantial amounts of additional rock dissolution can be excluded.

EC data for four Hidan wells are taken as an example for the wellfield from 1999 to 2011 and compared to water level records of well CD 3133 (Fig. 7b). Water level correlations with the two Wala wells are not possible. Hidan wells show almost constant EC values from 1999 to 2002, but temporal fluctuations cannot be excluded, partly because only one annual record exists. The EC peak in 2003/2004 is interpreted as a reaction to the hydraulic impulses from reservoir infiltration. It is assumed that (1) higher-mineralized groundwater from the rock matrix is pushed into the fracture and conduit network, (2) there is mobilization of fine material, e.g. the leaching of precipitated salt from irrigation return flow in the unsaturated zone, (3) the impact of wastewater from the adjacent area is too minor to raise salinity constantly, and (4) rock dissolution is excluded because this process would not cause a single peak but a general long-term increase of salinity. EC increase in Hidan wells can be attributed mainly to chloride and sulfate, which supports the assumption of mixing with water from areas of higher salinity (Wala wells). It is doubtful, however, whether the amount of salt from irrigation return flow is sufficient to constantly increase the salinity, but it is plausible that it could cause a temporal effect. Salt residues associated with the approximately 1–2 MCM of annual irrigation applied to the unsaturated zone are certainly some tens of tons. The sharp decrease of the EC in Hidan wells in 2007 and 2008 can be related to the drying up of the reservoir in 2008, when infiltration had almost stopped and the groundwater level dropped. At the same time, EC values decreased almost to corresponding values before 2001, and natural hydraulic conditions appeared again. The results show that the system strongly reacts to combined changes of groundwater level, abstraction rates and infiltration.

Isotopes

Deuterium and oxygen-18

The $\delta^{18}O$ values of rainfall range from −8.33 to −0.05 ‰ with corresponding $\delta^{2}H$ values of −42.5 to −0.9 ‰. These are mainly scattered along the MMWL, meaning Mediterranean origin, and a few samples are close to the GMWL (Fig. 8). The wide range of values for reservoir water can be attributed to the large catchment area, where precipitation with different isotopic signatures occurs and is influenced by the condensation process. Additionally, evaporation processes in the open reservoir lead to a high variation. The input signals of reservoir water are scattered along the MMWL and the GMWL and show $\delta^{18}O$ values

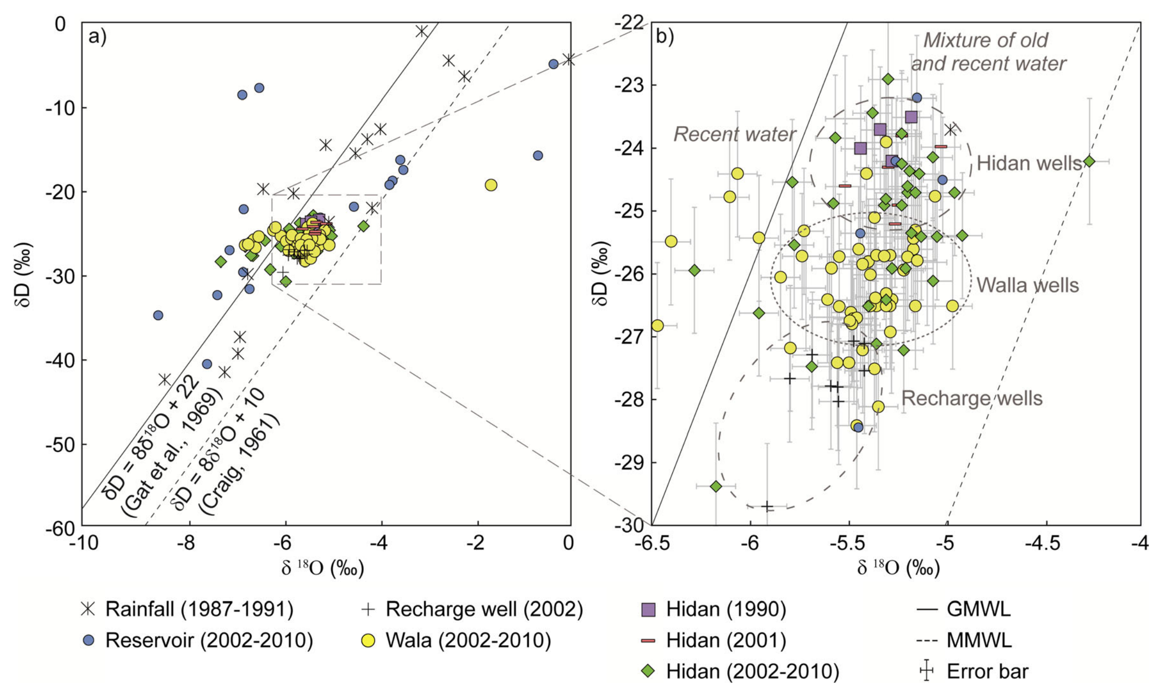

Fig. 8 Distribution of the $\delta^{2}H$ (δD) and $\delta^{18}O$ for samples of **a** rainfall, the reservoir, the recharge well and Wala groundwater, and **b** Hidan groundwater. Samples of Wala, recharge and Hidan wells are clustered along the *MMWL* and the *GMWL*, whereas data from precipitation and the reservoir are more scattered

 DOI 10.1007/s10040-015-1233-6

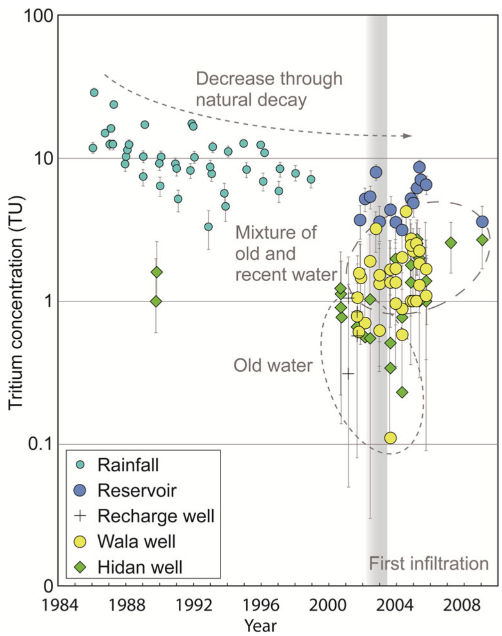

Fig. 9 Tritium (^{3}H) content of precipitation and Wala reservoir shows the decrease over time as a consequence of natural decay. The concentration in groundwater increases after 2002, caused by the mixing with recent water from Wala reservoir infiltration

ranging from −8.46 to −0.4 ‰ and δ^2H values from −40.7 to −21.6 ‰.

Samples of Wala, recharge and Hidan wells are more clustered and reveal values in the range of −7.2 to −1.6 ‰ for δ^{18}O and from −30.8 to −19.3 ‰ for δ^2H. Most of the samples are located between the MMWL and the GMWL. Samples above the MMWL were collected after 2002. Missing annual time series of single wells does not allow the detection of seasonal cycles or variations attributed to, for example, the impact of the reservoir.

Water samples from Wala wells and recharge wells are generally more depleted in δ^2H as compared to Hidan wells (Fig. 8), but have only a slight difference in δ^{18}O. This also applies to the time before the first infiltration, because some samples were collected before November 2002. Given the fact that evaporation creates a flatter line, it is obvious that water with a different isotopic signature rises towards Hidan wellfield, pointing to the mixing of water of different ages.

Tritium

The tritium values of rainfall, reservoir and groundwater samples are displayed in Fig. 9. Values for the rainfall samples in the Wala catchment confirm the decrease through natural radioactive decay from 1987 to 2000. Initial values range between around 11–30 TU and drop down to less than 10 TU.

Tritium values of reservoir water are similar to those of local rainfall and range between 3.1 and 8.7 TU with an average of 5.2 TU. It is obvious that ^{3}H values from rainfall and, hence, from the reservoir water vary strongly due to evaporation and condensation in the atmosphere. Initial values at Hidan and Wala wells in 2001–2002 of around 0.8 TU rise to a mean value of 2.3 TU in the years 2008–2010. This confirms the increasing contribution of reservoir water to the abstraction at Hidan wellfield. The ^{3}H values before 2002, of less than 1 TU in Hidan and Wala wells, indicate that the groundwater originates from a mixture of submodern (recharge before 1952) and recent water and it is evident that natural groundwater has a mean residence time of more than 60 years.

Mixing calculations of the median concentrations of ^{3}H in the reservoir and groundwater, recently and before the construction of the dam, show the proportion of the reservoir water in the groundwater is approximately 66 % (until 2007). This almost corresponds to the calculated 71 % (2002–2007) and 57 % (2002–2012) from the water balance.

Carbon-13 and carbon-14

The mean value of δ^{14}C activity of all wells is 13±1.09 pmc with a maximum of 29.11±1.44 pmc and a minimum of 4.97±0.60 pmc. Corresponding δ^{13}C values range from −10.60 to −13.41 ‰ with an average of −12.32 ‰. Table 2 shows the results of the calculated mean residence times of the groundwater for Fontes and Garnier (1979), the IAEA model (Plummer and Glynn 2013) and Evans model (1979).

Values of Wala well 13 and 14 range between 9,700 and 24,900 years and of the Hidan well 9 between 4,000 and 13,000 years. It is expected that there will be similar values along the flow path or at least a

Table 2 Calculated mean residence times of groundwater after Fontes and Garnier (1979); Plummer and Glynn (2013) and Evans (1979)

Sampling point	Date	^{14}C (pmc)	δ^{13}C (‰)	F and G (years)	IAEA (years)	Evans (years)
Wala 14[a]	14.10.2002	8.60±1.11	−11.76	24,741	19,174	19,174
Wala 14	17.03.2003	11.63±0.70	−13.39	19,344	17,643	17,643
Wala 14	09.07.2003	7.14±1.27	−12.60	26,835	21,227	21,227
Wala 14	05.04.2006	18.56±1.44	−13.10	14,781	13,628	13,628
Wala 13[a]	14.10.2002	5.50±0.76	−13.10	28,517	23,695	23,695
Wala 13	17.03.2003	4.97±0.60	−13.41	27,654	24,880	24,880
Wala 13	09.07.2003	13.62±0.90	−12.99	21,872	16,235	16,235
Hidan 9	02.12.2002	19.44±1.13	−11.24	17,136	12,090	12,090
Hidan 9	10.07.2003	17.13±1.55	−11.02	16,250	13,091	13,091
Hidan 9	05.04.2006	29.11±1.44	−10.60	8,711	8,306	8,306

[a] Before first reservoir infiltration (31.10.2002)

 DOI 10.1007/s10040-015-1233-6

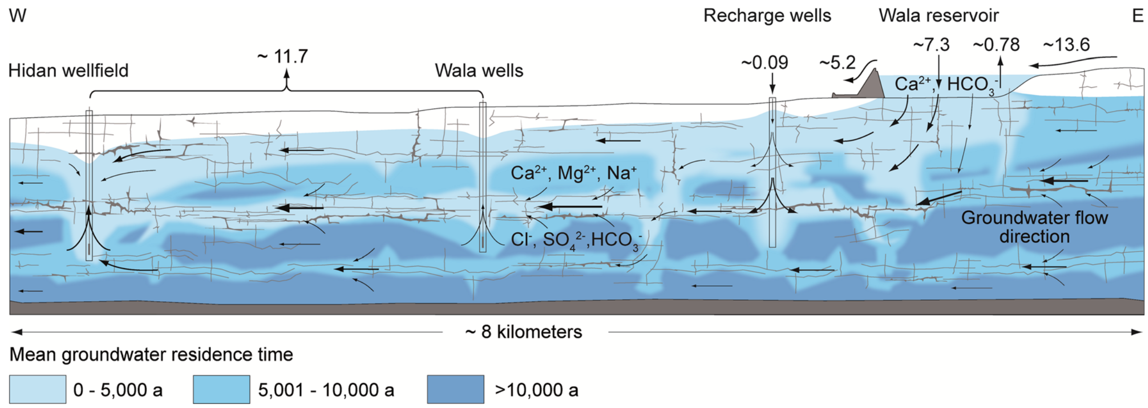

Fig. 10 Schematic profile along the wadi, showing mixture of young and old groundwater as a consequence of different flow regimes in the fracture and conduit network of the karst aquifer. Drawdown cones are seen at Wala and Hidan wells and infiltration mounds at the recharge well. Water balance values are displayed in million cubic meters as annual average (2002–2012)

slight increase. The fact that water becomes younger along the flow path points either to mixing with surface water, which is excluded by the tritium results, or mixing with groundwater of different ages, which agrees with the δ^2H and δ^{18}O results. The decrease of mean groundwater residence time between 2002 and 2006 is related to the increasing proportion of reservoir water infiltration. It can be derived from the calculated ages that water samples from 2002 originate from the late Pleistocene (older than 11,700 years) and some originate from the early to middle Holocene (11,700 to recent).

Conceptual groundwater flow model

The Hidan and Wala wells are fed by natural groundwater with mean residence times of several thousand years, originating from local precipitation. Generally, a higher productivity is attributed to the middle part of the Wadi As Sir aquifer between the Wala reservoir and Hidan wellfield. Likewise, an enhanced groundwater flow takes place along the wadi corridor, approved by the high recovery of infiltrated water at Hidan wellfield and the significant water-level increase after the first infiltration. This confirms the assumption made by Green et al. (2014) about increased karstification along wadis in karst terraines of semi-arid areas. Karst-specific flow patterns through fine matrix pores, fractures and conduits are confirmed by the occurrence of different EC values in the wells. Furthermore, isotope data indicate mixing of water from zones with different degrees of karstification and fracturing at the Wala and Hidan wells caused by the different drawdowns from pumping activities and the associated vertical flow components (Fig. 10). Such differences in mean groundwater residence times in production wells are often associated with the interaction of the aquifer and the well (Jurgens et al. 2014; Eberts et al. 2012; Zinn and Konikow 2007; Małoszewski and Zuber 1982). Upwelling of substantial amounts of water from a deeper aquifer can be neglected because of the underlying B5/6 aquitard. Also, lateral groundwater flow to the Hidan and Wala wells from the north and south, which could provide younger water, is unlikely because of the <8 km aquifer extension in these directions and the >300 m thickness of the unsaturated zone that receives only little rainfall.

Conclusion

The Wala reservoir is an example of a successfully managed aquifer recharge into a karst aquifer in a semi-arid region and demonstrates the promising potential that comes along with this technique regarding water management in Jordan.

Dry soil cover in the catchment leads to high sedimentation in the reservoir, which continuously decreases the infiltration rate and, thus, the proportion of artificially recharged water in the abstracted groundwater. The result will be a progressively dropping water level in the aquifer, when abstraction rates remain constant. Dry years will magnify this effect. Natural groundwater flow to Hidan wellfield is strongly dependent on the abstraction taking place in Wala catchment, upstream of the reservoir. Additional abstraction wells along Wadi Wala will directly affect Hidan wellfield as the water levels drop. Furthermore, lower abstraction yields are expected due to smaller degrees in fracturation and karstification compared to Hidan wellfield. Salinity fluctuations in the wellfield are expected to be on the same order of magnitude as from 2002 to 2012 of around 200 µS/cm.

Acknowledgements The authors thank the Ministry of Water (MWI), the Jordan Valley Authority (JVA) and the Water Authority of Jordan (WAJ) for their support and provision of the data. Furthermore, the German Federal Ministry of Education and Research (BMBF) are acknowledged for funding the SMART Project (Sustainable Management of Available Water Resources with Innovative Technologies) (FKZ 02WM1079-1086 and

DOI 10.1007/s10040-015-1233-6

FKZ02WM1211-1212). Special thanks go out to Prof. Michel Bakalowicz for fruitful discussion, Marian Bechtel for language editing of the manuscript, and the editor Dr. Vincent Post and an anonymous reviewer for their helpful suggestions.

References

Al-Assa'd TA, Abdulla FA (2010) Artificial groundwater recharge to a semi-arid basin: case study of Mujib aquifer, Jordan. Environ Earth Sci 60(4):845–859

Al-Hunjul NG (1993) Geological map of Madaba. Madaba 3153 II, scale 1:50 000. Geological Mapping Division, National Mapping Project, Amman, Jordan

Bajjali W (2006) Recharge mechanism and hydrochemistry evaluation of groundwater in the Nuaimeh area, Jordan, using environmental isotope techniques. Hydrogeol J 14(1–2):180–191

Bajjali W, Abu-Jaber N (2001) Climatological signals of the paleogroundwater in Jordan. J Hydrol 243(1):133–147

Bakalowicz M (1994) Water geochemistry: water quality and dynamics. Groundw Ecol 1:97–127

Bakalowicz M (2005) Karst groundwater: a challenge for new resources. Hydrogeol J 13(1):148–160

Bayer HJ, Hötzl H, Jado AR, Röscher B, Voggenreiter W (1988) Sedimentary and structural evolution of the northwest Arabian Red Sea margin. Tectonophysics 153(1):137–151

Bender F (1968) Geologie von Jordanien [Geology of Jordan]. 230 pp, Borntraeger, Berlin

Bouwer H (2000) Integrated water management: emerging issues and challenges. Agric Water Manag 45(3):217–228

Bouwer H (2002) Artificial recharge of groundwater: hydrogeology and engineering. Hydrogeol J 10(1):121–142

Clark ID, Fritz P (1997) Environmental isotopes in hydrogeology. Lewis, Boca Raton, FL

Craig H (1961) Isotopic variations in meteoric waters. Science 133:1702–1703

Daher W, Pistre S, Kneppers A, Bakalowicz M, Najem W (2011) Karst and artificial recharge: theoretical and practical problems—a preliminary approach to artificial recharge assessment. J Hydrol 408(3):189–202

Dillon P (2005) Future management of aquifer recharge. Hydrogeol J 13(1):313–316

Dillon P, Pavelic P, Page D, Beringen H, Ward J (2009) Managed aquifer recharge: an introduction. Waterlines Report Series no. 13, February 2009, National Water Commisssion, Canberra, Australia, 77 pp

Dreybrodt W, Romanov D, Gabrovsek F (2002) Karstification below dam sites: a model of increasing leakage from reservoirs. Environ Geol 42(5):518–524

Eberts SM, Böhlke JK, Kauffman LJ, Jurgens BC (2012) Comparison of particle-tracking and lumped-parameter age-distribution models for evaluating vulnerability of production wells to contamination. Hydrogeol J 20(2):263–282

Eichinger L (1983) A contribution to the interpretation of ^{14}C groundwater ages considering the example of a partially confined sandstone aquifer. In: Stuiver M, Kra RS (eds) Proceedings of the 11th International 14C Conference. Radiocarbon 25(2): 347–356

Einsiedl F (2005) Flow system dynamics and water storage of a fissured-porous karst aquifer characterized by artificial and environmental tracers. J Hydrol 312(1):312–321

El-Naqa A (1993) Hydrological and hydrogeological characteristics of Wadi el Mujib catchment area, Jordan. Environ Geol 22(3):257–271

El-Naqa A, Al Kuisi M (2004) Hydrogeochemical modeling of the water seepages through Tannur Dam, southern Jordan. Environ Geol 45(8):1087–1100

El-Naqa A, Al-Shayeb A (2009) Groundwater protection and management strategy in Jordan. Water Resour Manag 23(12):2379–2394

Etcheverry D, Vennemann T (2009) Isotope im Grundwasser: Methoden zur Anwendung in der hydrogeologischen Praxis. Umwelt-Wissen no. 0930 [Isotopes in groundwater: methods for use in hydrogeological practice. Environmental studies no. 0930]. Bundesamt für Umwelt, Bern, Switzerland, 121 pp

Evans GV, Otlet RL, Downing A, Monkhouse RA, Rae G (1979) Some problems in the interpretation of isotope measurements in United Kingdom aquifers. In: Isotope hydrology II. IAEA, Vienna, pp 679–708

Fontes JC, Garnier JM (1979) Determination of the initial ^{14}C activity of the total dissolved carbon: a review of the existing models and a new approach. Water Resour Res 15(2):399–413

Gat JR, Mazor E, Tzur Y (1969) The stable isotope composition of mineral waters in the Jordan Rift Valley, Israel. J Hydrol 76:334–352

Geyh MA (1970) Carbon-14 concentration of lime in soils and aspects of the carbon-14 dating of groundwater. In: Isotope hydrology. IAEA, Vienna, pp 215–223

Goldscheider N, Drew D (2007) Methods in karst hydrogeology. International Contributions to Hydrogeology, 26, CRC, Boca Raton, FL

Gonfiantini R, Zuppi GM (2003) Carbon isotope exchange rate of DIC in karst groundwater. Chem Geol 197(1):319–336

Gonfiantini R, Fröhlich K, Araguas-Araguas L, Rozanski K (1998) Isotopes in groundwater hydrology. In: Kendall C, McDonnell JJ (eds) Isotope tracers in catchment hydrology. Elsevier, Amsterdam, pp 203–246

Goode DJ, Senior LA, Subah A, Jaber A (2013) Groundwater-level trends and forecasts, and salinity trends, in the Azraq, Dead Sea, Hammad, Jordan Side Valleys, Yarmouk, and Zarqa groundwater basins, Jordan. US Geol Surv Open File Rep 2013-1061

Green RT, Bertetti FP, Miller MS (2014) Focused groundwater flow in a carbonate aquifer in a semi-arid environment. J Hydrol 517:284–297

Hiller T, Kaufmann G, Romanov D (2011) Karstification beneath dam-sites: from conceptual models to realistic scenarios. J Hydrol 398(3):202–211

Humphreys H, Partners (1991) Dams on Wadi Wala and Wadi Mujib, part IV: site investigation. Internal report of the Jordan Valley Authority of Jordan, Amman

Jurgens BC, Bexfield LM, Eberts SM (2014) A Ternary age-mixing model to explain contaminant occurrence in a deep supply well. Groundwater Suppl 1:25–39

Kattan Z (1995) Chemical and environmental isotope study of the fissured basaltic aquifer system of the Yarmouk basin (Syrian Arab Republic). In: Proceedings of a symposium on isotopes in water resources management. IAEA-SM-336/28, vol. 2, IAEA, Vienna

Kattan Z (2001) Use of hydrochemistry and environmental isotopes for evaluation of groundwater in the Paleogene limestone aquifer of the Ras Al-Ain area (Syrian Jezireh). Environ Geol 41(1–2):128–144

Lakshmanan E, Kannan R, Senthil Kumar M (2003) Major ion chemistry and identification of hydrogeochemical processes of groundwater in a part of Kancheepuram district Nadu, India. Environ Geosci 10(4):157–166

Małoszewski P, Zuber A (1982) Determining the turnover time of groundwater systems with the aid of environmental tracers: 1. models and their applicability. J Hydrol 57(3):207–231

Margane A, Hobler M, Almomani M, Subah A (2002) Contributions to the hydrogeology of northern and central Jordan. Geologisches Jahrbuch, Reihe C 68:52

Margane A, Hamdan I, Hajali Z (2009) Delineation of surface water protection zones for the Wala Dam. Technical report no. 12, MWI, Amman, Jordan

Masciopinto C, La Mantia R, Pollice A, Laera G (2012) Managed aquifer recharge of a karstic aquifer in Nardó, Italy. Reclaim Water: Advances in Water Reclamation Technologies for Safe Managed Aquifer Recharge, 47, European Commission, Brussels

DOI 10.1007/s10040-015-1233-6

Mays L (ed) (2009) Integrated urban water management: arid and semi-arid regions: UNESCO-IHP, vol 3. UNESCO, Paris

Meredith K (2009) Radiocarbon age dating groundwaters of the West Canning Basin, Western Australia. Dept. of Water, Perth, Australia

MWI (2004) National Water Master Plan (NWMP). Ministry for Water and Irrigation, Amman, Jordan

MWI (2012) Database. Ministry for Water and Irrigation, Amman, Jordan

Ofterdinger US, Balderer W, Loew S, Renard P (2004) Environmental isotopes as indicators for ground water recharge to fractured granite. Ground Water 42(6):868–879

Pearson FJ, Hanshaw BB (1970) Sources of dissolved carbonate species in groundwater and their effects on carbon-14 dating. Isot Hydrol 1970:271–286

Pearson FJ, White DE (1967) Carbon 14 ages and flow rates of water in Carrizo Sand, Atascosa County, Texas. Water Resour Res 3(1):251–261

Plummer LN (2005) Dating of young groundwater. In: Isotopes in the water cycle. Springer, Dordrecht, The Netherlands, pp 193–218

Plummer LN, Glynn PD (2013) Radiocarbon dating in groundwater systems, chap 4. In: Isotopes methods for dating old groundwater. IAEA, Vienna

Powell JH, Moh'd BK (2011) Evolution of Cretaceous to Eocene alluvial and carbonate platform sequences in central and south Jordan. GeoArabia-Middle East Petrol Geosci 16(4):29–82

Romanov D, Gabrovsek F, Dreybrodt W (2007) Leakage below dam sites in limestone terrains by enhanced karstification: a modeling approach. Environ Geol 51(5):775–779

Salem O, Visser JH, Dray M, Gonfiantini R (1980) Groundwater flow patterns in the western Lybian Arab Jamahiriaya. In: Arid-zone hydrology: investigations with isotope techniques. IAEA, Vienna, pp 165–179

Schmidt S, Geyer T, Marei A, Guttman J, Sauter M (2013) Quantification of long-term wastewater impacts on karst groundwater resources in a semi-arid environment by chloride mass balance methods. J Hydrol 502:177–190

Schmoll, Oliver (2006) Protecting groundwater for health: managing the quality of drinking-water sources. World Health Organization, Geneva

Tamers MA (1967) Radiocarbon ages of groundwater in an arid zone unconfined aquifer. Geophysical Monograph Series, vol 11, AGU, Washington, DC, pp 143–152

Vanderzalm JL, Page DW, Barry KE, Dillon PJ (2010) A comparison of the geochemical response to different managed aquifer recharge operations for injection of urban stormwater in a carbonate aquifer. Appl Geochem 25(9):1350–1360

Vogel JC (1968) Investigation of groundwater flow with radiocarbon. In: Isotopes in hydrology. International Atomic Energy Agency, Vienna, pp 355–369

Wolf L, Werz H, Hoetzl H, Ghanem M (2007) Exploring the potential of managed aquifer recharge to mitigate water scarcity in the Lower Jordan River Basin within an IWRM approach. In: Proceedings of the 6th International Symposium on Managed Artificial Recharge of Groundwater, ISMAR6, Phoenix, AZ, 28 October–2 November 2007

World Health Organization (ed) (2004) Guidelines for drinking-water quality: recommendations, vol 1. WHO, Geneva

Zinn BA, Konikow LF (2007) Potential effects of regional pumpage on groundwater age distribution. Water Resour Res 43(6)

DOI 10.1007/s10040-015-1233-6