Agricultural Water Management 97 (2010) 1876–1886

Contents lists available at ScienceDirect

Agricultural Water Management

journal homepage: www.elsevier.com/locate/agwat

Assessment of groundwater storage depletion by overexploitation using simple indicators in an irrigated closed aquifer basin in Iran

M. Van Camp[a,*], M. Radfar[a,b], K. Walraevens[a]

[a] Laboratory for Applied Geology and Hydrogeology, Ghent University, Krijgslaan 281 9000 Gent, Belgium
[b] Department of Water Engineering, Shahrekord University, Shahrekord, Iran

ARTICLE INFO

Article history:
Received 5 June 2009
Accepted 10 February 2010

Keywords:
Overexploitation
Indicator
Modeling
Irrigation

ABSTRACT

In arid and semi-arid regions irrigation is usually needed to provide enough water for crop growth in cultivated areas. As surface waters are scarce, especially in summertime when the water is needed, groundwater is heavily used to supply the water demand. Overexploitation of the aquifer in dry years causes depletion of the groundwater storage and systematical lowering of the piezometric levels. This is a particular problem in aquifers developed in closed basins where lateral inflow is nearly absent and replenishment is constrained by rainfall recharge. In this paper, simple indicators derived from meteorological data, abstraction rates and piezometric time series are compared with the groundwater storage depletion as obtained from a calibrated groundwater flow model. Application of the method to the overexploited Shahrekord basin in Iran shows that for the simulated period 1989–2003 an accumulative index of the difference of aquifer recharge, as calculated by a soil moisture balance method, and groundwater abstraction has a correlation coefficient of nearly one with model calculated storage. Indicators based on the filling index derived from piezometric time series or on the ratio of aquifer discharge to recharge have slightly lower correlations. The accumulated index indicator can be used to follow aquifer storage in the future without the need to run the full groundwater flow model. This simple approximation is restricted to aquifer systems with a limited lateral inflow and outflow.

1. Introduction

Aquifers are very convenient sources of water because they are natural underground reservoirs and can have an enormous storage capacity (Morris et al., 2003). As many aquifers contain high quality waters their application for drinking water production and human consumption is obvious and they are used worldwide as a source for irrigation water in agricultural regions. As their storage capacity is large, they can provide a continuous source of water, even in dry seasonal periods when rainfall is nearly absent and superficial water is fast depleted. To define how much groundwater can be abstracted from an aquifer, assuming that groundwater is a renewable resource. The safe-yield concept was introduced in the 1920s (Meinzer, 1920). It has been defined as the amount of water that can be withdrawn from the aquifer without producing an undesired result (Morris et al., 2003). Because of the somewhat diffuse definition, the concept is flawed and rather controversial (Bredehoeft et al., 1982; Bredehoeft, 1997, 2002; Sophocleous, 1997, 2000, 2001; Devlin and Sophocleous, 2005). Kalf and Woolley (2005) give a review of the definition and methodology of determining safe yield. More recently the concept of sustainability is very much alive, and is defined as the level of development of groundwater that meets the needs of the present generation without compromising the ability of future generations to meet their needs (Alley et al., 1999; Custodio, 2002; Morris et al., 2003; Alley and Leake, 2004; Maimone, 2004). But also sustainability is a manifold concept, approachable from many points of view (Llamas et al., 2006). When the total amount of abstraction from an aquifer is or will be close to, or greater than, the total recharge over several years, it is often said that there is overexploitation (Custodio, 2002). Basically, an aquifer is overexploited or at risk of being so when the quotient of pumping and recharge is greater than one (Vrbo and Lippanen, 2007). In practice, however, an aquifer is often considered as overexploited when some persistent negative results of aquifer development are felt or perceived, such as a continuous water-level drawdown, progressive water-quality deterioration, increase of abstraction cost, or ecological damage (Custodio, 2002). The term "intensive groundwater use" is applied when exploitation induces significant changes on natural aquifer dynamics (Llamas and Custodio, 2003; Custodio et al., 2005).

Although aquifer over-exploitation may be an ambiguous and controversial term, it is likely to become more important, especially in the arid and semi-arid regions of the world (where aquifer recharge is limited), as competing demands grow on a limited

* Corresponding author. Tel.: +32 92644655; fax: +32 92644988.
E-mail address: Marc.VanCamp@Ugent.be (M. Van Camp).

0378-3774/$ – see front matter
doi:10.1016/j.agwat.2010.02.006

resource (Morris et al., 2003). In many semi-arid to arid regions groundwater has been withdrawn at rates far in excess of recharge, leading to groundwater "mining", major water-level declines, increased pumping costs, and decreased well yields (Konikow, 2002). The term groundwater mining is used when conscious and planned abstraction rate greatly exceeds aquifer recharge (United Nations, 1992). Groundwater overdraft occurs when extraction exceeds both natural and induced aquifer recharge over long periods. Overdrafting aquifers is common and may be temporarily beneficial within a long-term water management strategy (Harou and Lund, 2008), but in arid and semi-arid regions, where natural aquifer recharge is limited, developing long-term sustainable exploitation strategies may be a difficult challenge. Despite the significance of groundwater for sustainable development, it has not always been properly managed, which often has resulted in depletion and degradation of the resource (IGES, 2008).

In the light of the risk of aquifers being overexploited or even mined, the assessment of groundwater storage depletion, caused by permanent or temporary groundwater overdraft, is crucial in a first step for evaluating groundwater imbalance. This is particularly so in arid and semi-arid region where there is a large need on irrigation water for supporting agriculture. Continuously declining piezometric levels are a first and easily recognizable sign of storage depletion, but a more quantitative approach is needed to sizing the real problem. In the past, some groundwater sustainability indicators have been proposed (Vrbo and Lippanen, 2007), some of which are related to storage depletion. The UNESCO proposed indicators have already been applied in some regions (Lavapuro et al., 2008). In this study, eight different indicators are evaluated for their ability to correlate with the groundwater storage in a closed aquifer basin, where lateral cross-border flow is small. Aquifer storage is computed with a distributed groundwater flow model that simulates the exploitation history (1989–2003) of the whole basin.

2. Materials and methods

2.1. Groundwater depletion indicators

As stated by UNESCO (Foster and Loucks, 2006) the concept of overexploitation is intended to indicate an imbalance within the groundwater budget of the aquifer. Storage change is often estimated as the residual of outflow and inflow, both of which are estimated with variable uncertainty. Storage change can also be estimated using a storage coefficient and water-level changes derived from measurements in wells. Poor definition of the hydrogeologic system and storage property distributions can result in large uncertainty in estimates of groundwater storage change derived from water-level change (Pool, 2002). Finally a hydrodynamic model using transient flow regimes can be used to simulate the evolution of groundwater storage.

In this paper, six different simple indicators for groundwater depletion are compared with groundwater storage (STOR) and changes in groundwater storage (DSTOR), as calculated with a distributed groundwater flow model. All indicators have been calculated as year averaged values, but are based on monthly computed numbers.

The first two indicators FI and AFI (Table 1) must be derived from piezometric time series. A filling index FI has been proposed by the UNESCO/IAEA/IAG Working Group on Groundwater (Vrbo and Lippanen, 2007) as an easy-applied indicator to represent changes in piezometric level for a given time period (e.g. a month) within the fluctuation range. It is defined as the filling level for a specific date within a historical period, which has been quantified as the quotient between the measured level with respect to the minimum historical level, and the difference between the maximum and minimum historical levels during the considered time period. The advantage of the filling index is that it is only based on measurements of piezometric levels and does not require discharge rates and/or quantifications of recharge or natural discharge. An overall filling index for the aquifer is calculated as the average value of filling indices derived for each observation point. The monthly filling indices can be averaged to obtain a mean annual value. The second indicator is the time accumulated filling index AFI.

Table 1
Definition of the groundwater depletion indicators.

Indicator	Explanation	Definition
FI	Filling index	$FI = \frac{PL_i - PL_{min}}{PL_{max} - PL_{min}}$
AFI	Accumulated Filling Index	$AFI = \sum_{i=1}^{m} FI_i$
RDR	Recharge–Discharge Ratio	$RDR = \frac{RECH_i}{DISCH_i}$
ARDR	Accumulated Recharge–Discharge Ratio	$ARDR = \sum_{i=1}^{m} \frac{RECH_i}{DISCH_i}$
RDD	Recharge–Discharge Difference	$RDD = RECH_i - DISCH_i$
ARDD	Accumulated Recharge–Discharge Difference	$ARDD = \sum_{i=1}^{m} (RECH_i - DISCH_i)$

The other four indicators are derived from recharge and discharge components of the considered aquifer or basin. Indicators based on a ratio require that the denominator is not zero. In arid and semi-arid regions aquifer recharge can be absent in some periods (months) or even years. Therefore the ratio of recharge over discharge is more recommended than the inverse definition. The RDR indicator is defined as the ratio of recharge to discharge and ARDR is its time accumulated form. The RDR and ARDR indicators have always positive values and smaller values indicate a larger depletion of groundwater storage. Differencing the two components in the RDD and ARDD indicators can give negative values. Here depletion is characterised by negative numbers. Definitions of the indicators are summarized in Table 1 where:

FI_i = filling index for month i
PL_i = piezometric level (m) in month i
PL_{min} = minimum piezometric level (m) for the considered time series
PL_{max} = maximum piezometric level (m) for the considered time series
$RECH_i$ = aquifer recharge (m^3/day) in month i
$DISCH_i$ = aquifer discharge (m^3/day) in month i

2.2. Application to the Shahrekord Plain aquifer in Iran

2.2.1. Location and physiography

The Shahrekord region is located in southwest Iran close to the city of Isfahan (Fig. 1). It is a 645 km^2 large plain at an altitude of 2000–2300 m above mean sea level and surrounded by mountainous terrain reaching an altitude well above 3000 m. It is an important agricultural region and has a significant socio-economic function. Of the total plain area, around 200 km^2 or 31% is cultivated land (as derived from Landsat TM satellite images). Cultures include wheat, barley, sugar beet, alfalfa, potato and trees like alabama and apple. The cultivated land parcels are intensively irrigated using water from wells located in between the cultures, from karizes and to a minor extent, from springs. Karizes (or qanats) are constructed as gently sloping subsurface galleries extending into the water table at their deepest part, which can be accessed from the surface through series of vertical shafts, and which are gravita-

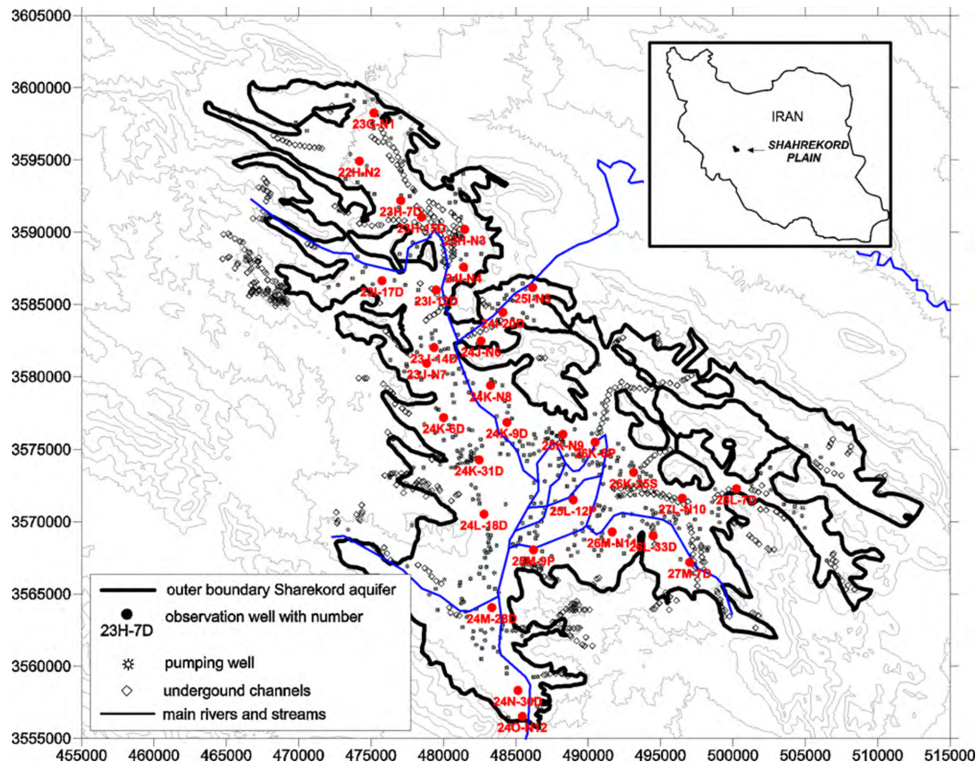

Fig. 1. Location and situation of the Shahrekord Plain basin (topography, hydrography, observation wells (with numbers), pumping wells (squares), karizes (diamonds), springs (stars)).

tionally draining the groundwater to the galleries outlet. They can deliver large quantities of water to the surface without need for pumping. Karizes allow water to be transported over long distances in hot dry climates without loosing a large proportion of the water to seepage and evaporation.

2.2.2. *Climate*

Following the Köppen classification, the climate of the area is semi-arid, with a mean annual rainfall of 321 mm and a mean annual air temperature of 11.8 °C. Two different seasons can be distinguished. Most of the precipitation falls in winter time between November and April (Table 2). The coldest month is January with an average temperature below zero (−1.5 °C). About 40% of the precipitation occurs as snow. The dry season starts in June and has nearly no precipitation at all till October. Average temperatures rise till 24 °C in July. Winter season has 50% of total precipitation, autumn 28% and spring 22% according to the observations of the Shahrekord station. Summer is completely dry.

Table 2
Long-term monthly averages of precipitation, temperature and humidity in Shahrekord station.

Month	Precipitation (mm)	Average T (°C)	Average humidity (%)
January	60.8	−1.5	67
February	47.9	1.2	67
March	60.3	6.0	62
April	37.5	11.3	55
May	14.1	15.9	48
June	0.9	20.7	42
July	1.9	24.0	33
August	0.4	23.1	31
September	0.0	18.8	32
October	6.9	13.2	42
November	32.2	7.4	54
December	58.6	2.2	63

2.2.3. *Geology and hydrogeology*

The Shahrekord aquifer system is formed by the sedimentary filling in a nearly closed basin eroded in a limestone basement, which is considered as the substratum of the aquifer system. It consists of mainly sandy and silty deposits which can reach a total thickness of up to more than 100 m. An isopach map was made based on the well descriptions available. The upper part of the sequence, below ground surface, is usually more silty and can be considered as an aquitard, with a thickness up to 10–20 m. Locally, in the deeper part of the sequence a second aquitard is found, but as it is discontinuous, its importance for the general groundwater flow system is rather limited. The main sandy deposits form the main single aquifer of the basin. The hydrostratigraphy is schematised to a single aquifer which is covered by a superficial aquitard. Locally, the water table can be located in the covering aquitard or deeper in the underlying sandy layers.

2.2.4. *Piezometric levels*

Since 1984 a monitoring network of observations wells was installed and measured monthly. Since 1984 fifteen wells are used for this purpose, another four were added in 1987. They are spread out over the basin and their location is indicated in Fig. 1. The piezometric series were plotted in time graphs to investigate trends and temporal behavior. Many of the series show a significant and systematic decease in levels, mainly in the last decade. Especially at the end of the nineties, levels decreased faster. For the 19 observation wells the piezometric level lowered with an average of about 9.5 m between 1992 and 2002. Most of the series also show strong seasonal cycli, although these can vary significantly in size from year to year. Two factors are inducing this seasonality: recharge from rainfall which causes increased levels in wintertime, and high groundwater discharge rates in summer for irrigation water supply which cause declining levels in summer time. In the centre of the basin where most of the exploitation wells are located, the water table is now at depths of 10–20 m below ground surface. Before groundwater was intensively used (at least thirty years ago) water table was shallow and the valley was locally described as a "mud plain". Nowadays, only limited outflow of surface water by streams occurs. Only in winter time after rainy periods, there is temporarily flow out of the basin. Sporadic but discontinuous measurements of flow rates have been done in the late seventies during these floods, and recorded peak rates were no more than a few cubic meters a second. These last no longer than a few days.

Trend analysis on the 19 observation well series shows most of them have a declining trend. By fitting a linear trend line it was found that in only three wells a downward trend was absent (small regression line slope and correlation coefficient < 0.1). But most of the other wells have a downward trend between 10 and 80 cm/year with correlation coefficients of more than 0.4.

2.2.5. Groundwater balance

2.2.5.1. Recharge components. As the aquifer system is a nearly closed basin, the main natural aquifer recharge originates from rainfall during the winter season. In arid and semi-arid areas, where potential evapotranspiration equals or surpasses average precipitation, recharge is difficult to estimate (Kinzelbach et al., 2002; Anuraga et al., 2006). Aquifer recharge is estimated here using a soil moisture balance approach based on the THORNTHWAITE and MATHER method (Thornthwaite, 1948; Thornthwaite and Mather, 1955, 1957). In this study the program WATBUG (Willmott, 1977) was used for this purpose, based on monthly data for precipitation and temperature. As calculated recharge rates are strongly dependent on the used value for the water holding capacity (WHC), a sensitivity analysis was made for this parameter. Calculations were done with WHC values of 50, 100, 150 and 200 mm. The results of these WATBUG runs show the calculated recharge amounts (in mm/year) as a function of the yearly precipitation (Fig. 2). For each WHC value a linear regression was calculated (Table 3). A comparison was made between the time series of monthly calculated recharge amounts for the different WHC values and the piezometric time series in the observation wells. Yearly recharge of the water table only occurs when WHC values are no more than around 100 mm. This value was finally selected and used in the calculations.

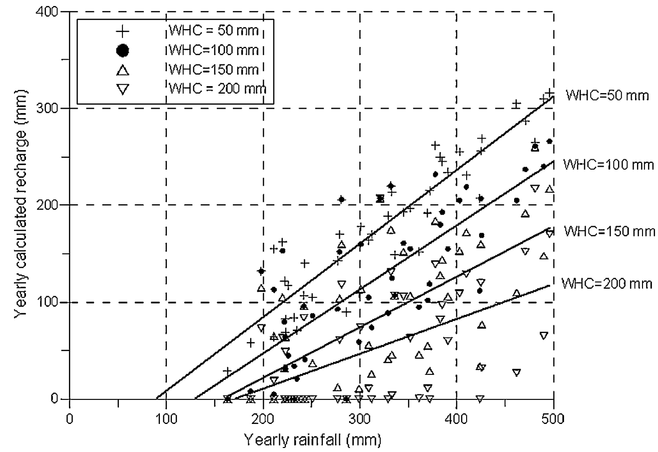

Fig. 2. Calculated yearly aquifer recharge according to the water holding capacity (WHC).

Table 3
Regression analysis between WATBUG calculated recharge rates (in mm/year) and yearly precipitation totals (in mm/year) for different WHC values.

WHC (mm)	A	B	R^2	*X*-intercept
50	0.7592	−67.24	0.83	88.56
100	0.6606	−84.90	0.61	128.51
150	0.5216	−82.26	0.43	157.70
200	0.3595	−61.06	0.28	168.84

During summertime, between 200 and 300 Mm^3 of water is used for irrigation. Because not all this water is evaporated or used for transpiration by the crops, a fraction will infiltrate and recharge the water table. Estimation of this return flow (Radfar, 2009) indicates that nearly 30% of the irrigation water can be viewed as return flow. As the total amount of groundwater discharge is large, exact

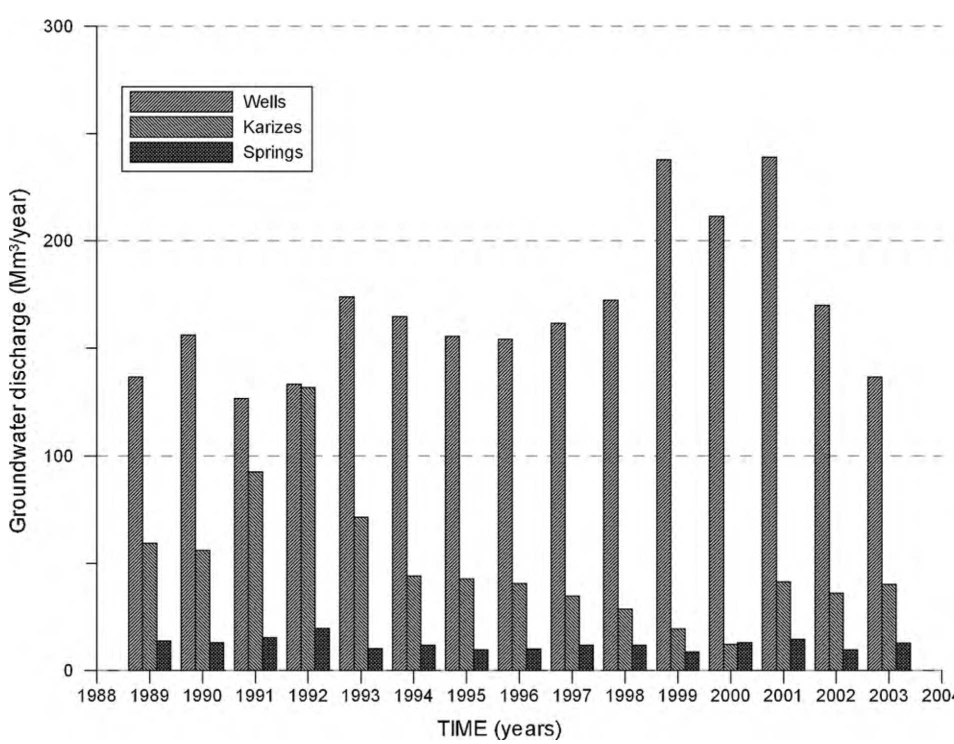

Fig. 3. Groundwater abstraction in the period 1989–2003.

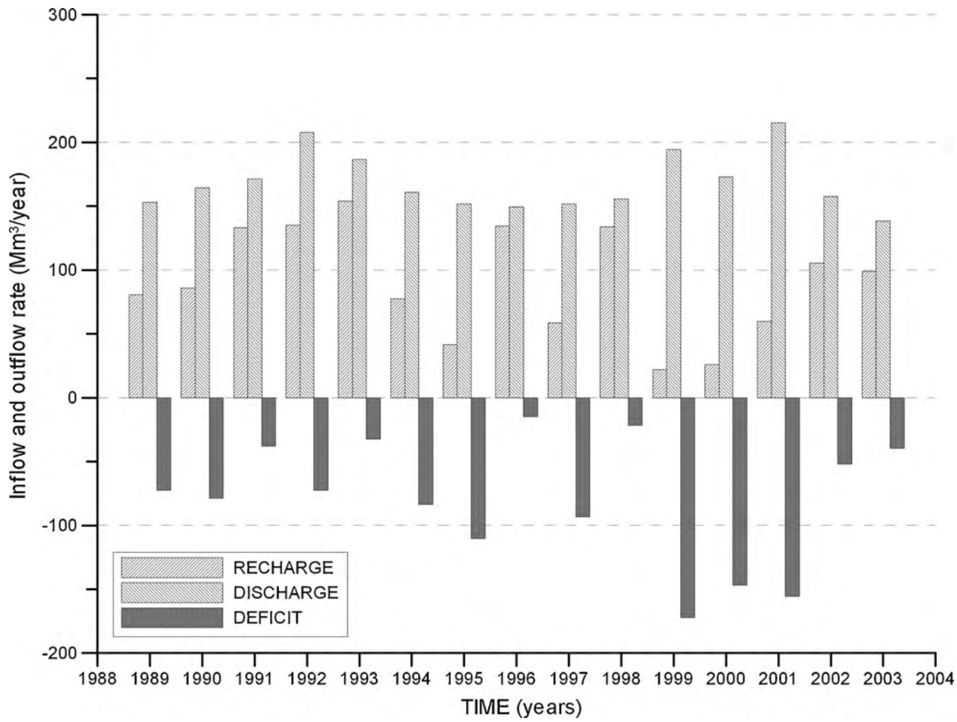

Fig. 4. Comparison of basin discharge and recharge between 1989 and 2003.

estimation of the return flow fraction is necessary for the overall water balance. Irrigation brings yearly at least 1 m of water on the cultivated land surface.

2.2.5.2. Discharge components. Water used for irrigation purposes is taken from three different sources. Spring water forms the smallest contribution. The last decade it contributes only to about 5.4% of the total irrigation amount. Karizes in use since old times capture water usually near the periphery of the plain and drain it to the plain centre and cultivated plots (see their location in Fig. 1). The last decade they contribute for around 21.9% of the total extracted irrigation water. The number of water producing wells has steadily increased in the last twenty to thirty years. Statistics exist of the number of wells since the fifties. Before 1979 only 116 wells are known. This increased to 320 in the mid eighties and recently in 2003 a total of 633 wells should exist. The location of these wells is recorded and indicated in Fig. 1. They are mainly located in or between cultivated plots and account for 72.6% of groundwater exploitation.

Streamflow out of the basin is neglectible as it only accidentically occurs after heavy rains in winter time. Therefore this discharge component is not considered in the water balance.

2.2.5.3. Groundwater balance for 1989–2003. From 1989 to 2003 detailed discharge rates from springs, karizes and wells are available (Fig. 3). For 1994 no data are available for wells extraction. A total water balance was constructed assuming 30% of the irrigation water being return flow (Radfar, 2009) and adding the calculated recharge components. Sporadic discharge by stream flow during winter floods is not considered as important, as is groundwater outflow through the basin outlet. These amounts are small compared to the other balance components. The total recharge and discharge components and the difference between them are presented in Fig. 4. At least from 1989 on (and probably even before 1989) there is a deficit which is usually between 50 and 100 Mm^3 a year. Smaller deficits occurred in 1993, 1996 and 1998. The most striking feature on the graph is the fact that the three largest deficits of the last 15 years occurred subsequently in 1999, 2000 and 2001. Yearly deficit was around 150 Mm^3.

2.2.5.4. Groundwater balance before 1989. No detailed information of the water balance components before 1989 is available. However, an approximative water balance has been reconstructed based on some assumptions. For groundwater extraction from wells, the total discharge is assumed proportional to the number of wells present in a specific year compared to the number of wells and total extraction rates for 1989. Discharge from springs and karizes, which are much longer in use, is assumed proportional to the yearly precipitation following a linear relationship derived from regression analysis of the data from 1989 to 2003.

2.2.6. Description of the groundwater flow model

The groundwater flow model uses MODFLOW (McDonald and Harbaugh, 1988) as the simulator. The outer limit of the modelled region is the border of the plain and was derived from elevation data and topographic maps. The model grid uses a uniformly spatial discretization of 200 m all over the model domain. The aquifer system is schematised in the model as a two-layer system, comprising the upper and lower half of the aquifer. The hydraulic conductivity and transmissivity of the aquifer system is derived from a series of pumping tests that have been done in different production wells (Radfar, 2009). Transmissivity values increase towards the center of the basin to more than around 1000 m^2/day. The model uses a transient flow regime with monthly stress periods and boundary conditions were defined on a monthly basis. Monthly recharge rates as calculated from the soil moisture water

Table 4
Errors on the calculated piezometric levels and drawdowns after the calibration runs.

Objective function	Average error (m)	Average absolute error (m)
Piezometric levels 1989	−1.41	7.03
Piezometric levels 2002	−1.35	10.05
Drawdowns 1989–2002	−0.06	4.17

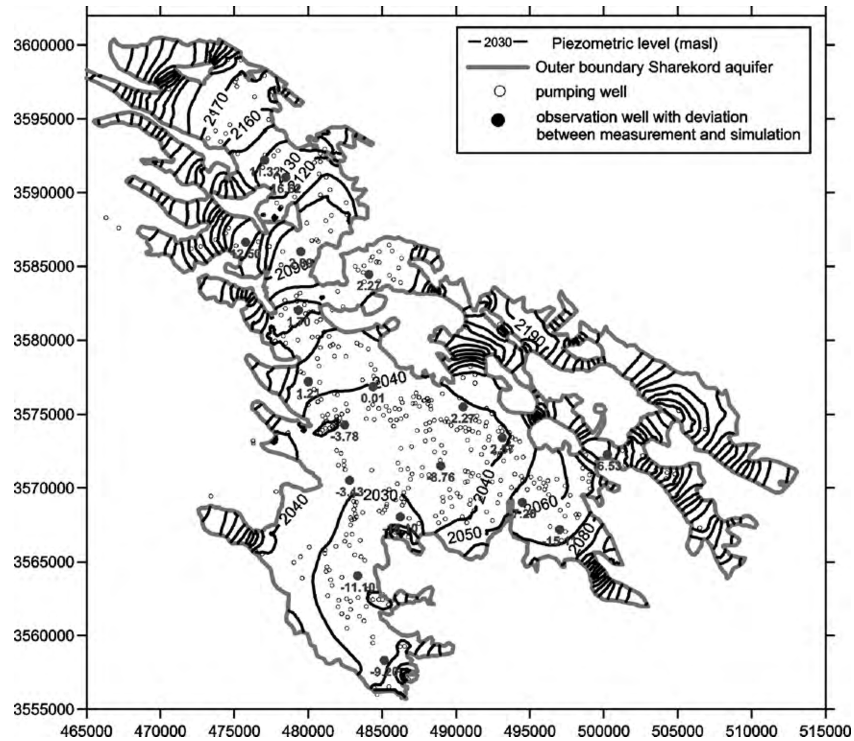

Fig. 5. Simulated piezometric levels in 2002 and differences between calculated and observed water levels in 19 piezometers.

balance were input into MODFLOWs recharge package. As irrigation is seasonal, total aquifer discharge from wells, karizes and springs was restricted to a six month period from April till September. The same applies for the irrigation return flow. The stream network is included into the model with the MODFLOW DRAIN module. River bottom levels were derived from a digital terrain model.

2.2.7. Calibration of the flow model

The specific yield of the upper aquifer layer is the only parameter adapted during the calibration runs. The model was calibrated based on piezometric levels of the years 1989 and 2002 and the drawdown between 1989 and 2002. Drawdowns were derived from the time series of the 19 observation wells. An average decline of about 9.5 m is found for the considered 13 year period. Compar-

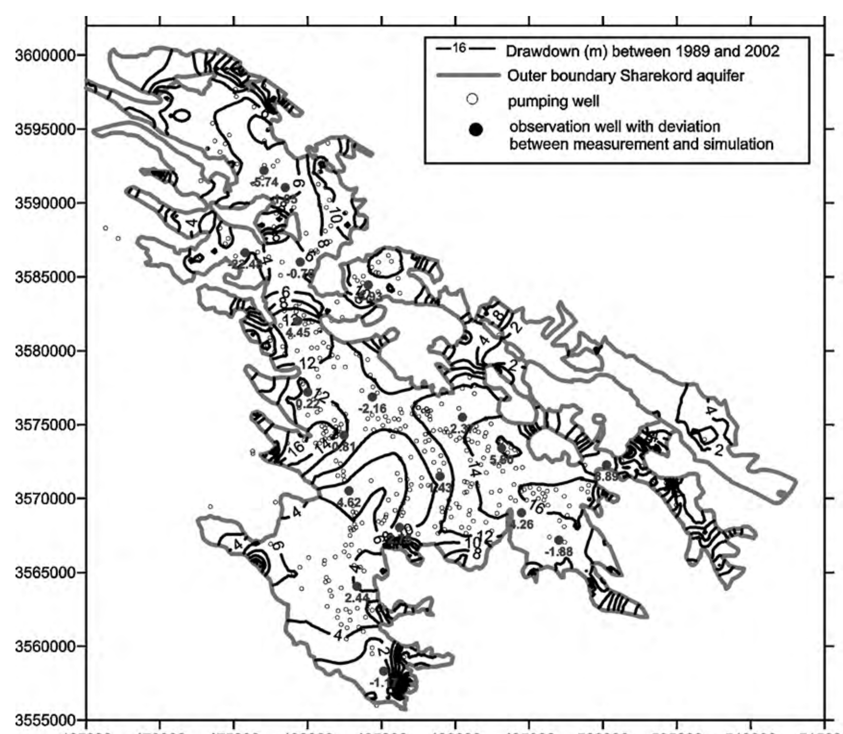

Fig. 6. Simulated drawdowns between 1989 and 2002 and differences between calculated and observed drawndowns in 19 piezometers.

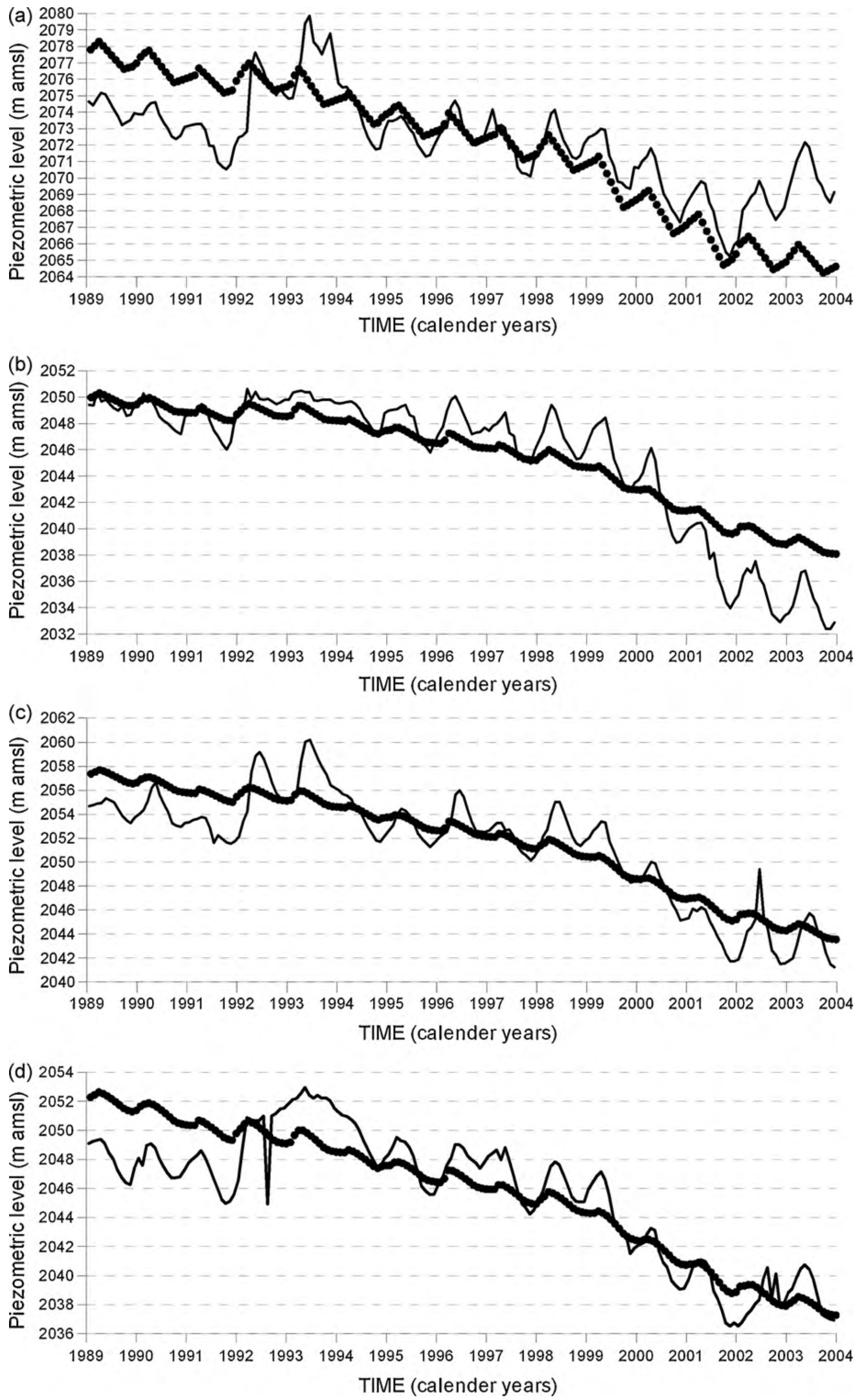

Fig. 7. Simulated and observed piezometric levels in selected observation wells (solid line = observed; dots = simulated): (a) 23J-14D, (b) 24K-9D, (c) 24K-6D, (d) 26K-6P.

ing runs with different values, a specific yield of 0.24 was retained. The average absolute error, defined as the arithmetic average of the absolute differences between calculated and observed levels, was around 7 m for the 1989 situation and around 10 m for 2002 (Table 4). These represent around 8 and 12% of the range in piezometric levels over the basin. Deviations are mainly caused by local impact of pumping wells in proximity of the observation wells. The average absolute error on drawdowns between 1989 and 2002 is only around 4 m. Average errors, defined as the arithmetic average of the difference between calculated and measured levels, are slightly negative for the piezometric levels, but close to zero for the drawdowns. The model can reproduce the observed depression cone and declining trend quite well.

3. Results and discussion

The model outputs monthly distributions of piezometric levels. These can be used to plot piezometric maps for selected years and months or construct time graphs for selected locations. The calculated piezometric levels for 2002 (averages of the twelve

Table 5
Model computed groundwater depletion rate for each year DSTOR (in Mm^3/year) and accumulated change in storage STOR (Mm^3 referenced to end 1988 situation) and the values of the various indicators.

Year	DSTOR (Mm^3)	STOR (Mm^3)	FI (−)	AFI (−)	RDR (−)	ARDR (−)	RDD (Mm^3)	ARDD (Mm^3)
1989	38.46	38.46	0.844	0.844	0.532	0.532	70.96	70.97
1990	69.81	108.27	0.799	1.643	0.515	1.047	79.39	150.35
1991	125.14	233.41	0.718	2.361	0.758	1.805	41.73	192.08
1992	7.42	240.83	0.64	3.001	0.637	2.442	76.16	268.24
1993	65.33	306.16	0.867	3.868	0.862	3.304	24.88	293.12
1994	129.47	435.63	0.995	4.863	0.498	3.801	77.27	370.12
1995	78.81	514.44	0.798	5.661	0.267	4.069	112.65	483.05
1996	43.88	558.32	0.677	6.338	0.901	4.970	15.33	498.39
1997	96.15	654.47	0.701	7.039	0.369	5.339	99.13	597.52
1998	30	684.47	0.579	7.618	0.870	6.208	19.84	617.36
1999	151.8	836.27	0.614	8.232	0.114	6.322	170.46	787.83
2000	165.54	1001.81	0.465	8.697	0.146	6.468	150.44	938.27
2001	167.15	1168.96	0.261	8.958	0.262	6.730	166.47	1104.75
2002	76.29	1245.25	0.158	9.116	0.613	7.343	65.62	1170.38
2003	68.34	1313.59	0.301	9.427	0.645	7.988	53.69	1224.07

monthly levels) are plotted on a piezometric map (Fig. 5). Differences between calculated and measured water levels in the 19 observation wells are also indicated, and the location of pumping wells is annotated with circles. A map of calculated drawdown between 1989 and 2002 is generated from the piezometric level grids (Fig. 6). The largest drawdowns (more than 10 m) are found in the central part of the plain where most of the cultivated land parcels and irrigation wells are located.

For four observation wells located in different parts of the plain, time graphs of observed and calculated levels are plotted (Fig. 7). Observation well 23J-14D is located in the northern half of the plain. On this location, levels are initially calculated a few meters too high before the wet seasons of 1992 and 1993, when aquifer recharge seems to be underestimated by the model. Maybe lithological conditions favour local infiltration and cause more intense recharge events or the local value for specific yield is lower than the uniform value of 0.24 used in the simulation. The model is however not sufficiently refined to include these small-scale local heterogeneities. But the declining trend after 1994 is well reproduced, although in the last two years levels seem to have risen on this location. Observation wells 24K-9D and 24K-6D are located 7 and 5 km to the south. The declining trend is reproduced although seasonal fluctuations are larger in the observation series. In well 24K-9D the drawdown is more than 15 m, much of which has developed in barely two years (2000 and 2001). The model does not simulate such large drawdown. Maybe local pumping rates have increased here more. Well 26K-6P is located in the central part of the plain and shows a drawdown of around 15 m but the decline was more spread over the last decade here. The model is simulating the observed trend well.

From the monthly water balances of the model, values for the monthly change in storage (DSTOR) were obtained and accumulated (STOR). Yearly accumulated values for DSTOR and STOR are included in Table 5. Monthly and yearly averaged values for DSTOR are plotted in Fig. 8. Over the considered 15 year's period, at least around 1300 Mm^3 have been extracted from storage, which means the yearly deficit is 80–90 Mm^3. Average yearly recharge for the whole basin (1989–2003) is around 88.8 Mm^3. Average groundwater discharge is thus around twice its recharge rate. Even if groundwater exploitation would be completely halted, it would take at least another 15 years for the basin to recover.

It can be expected that the yearly calculated indicators FI, RDR and RDD have a rather good correlation with the yearly change in storage DSTOR, while the accumulated indices AFI, ARDR and

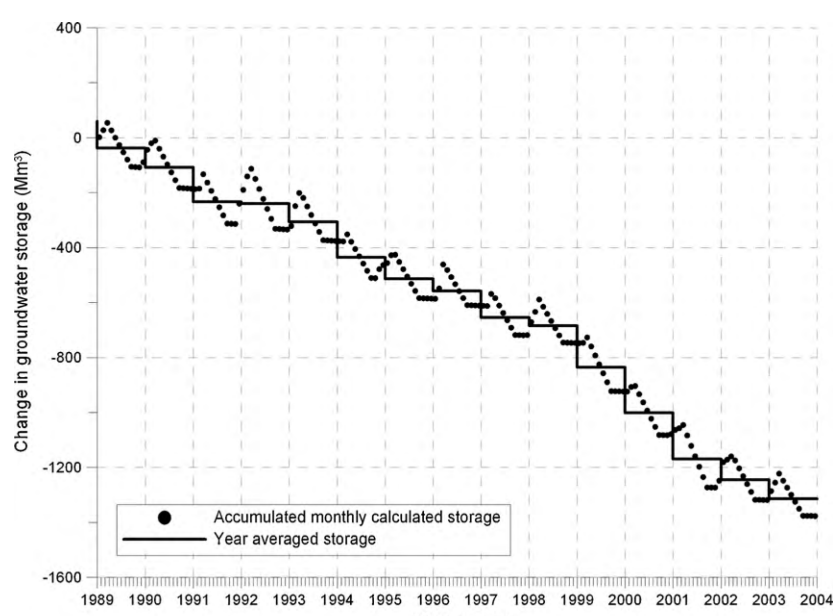

Fig. 8. Model calculated groundwater storage (STOR) depletion in 1989–2003 (dots are monthly values, solid line is year averaged).

Table 6
Linear correlation coefficients r between indicators and accumulated change in groundwater storage (STOR) and change in storage (DSTOR).

indicator	Definition	Storage STOR	Storage depletion rate DSTOR
FI	Filling Index	0.86	−0.19
AFI	FI-accumulative	0.95	0.40
RDR	RECH/DISCH-ratio	0.30	−0.69
ARDR	RDR-accumulative	0.96	0.34
RDD	RECH–DISCH	0.35	−0.72
ARDD	RDD-accumulative	1.00	0.40

ARDD are more related to the accumulated change in storage STOR, which gives an indication of aquifer storage depletion over the years (Table 6 and Fig. 9). The filling index based indicators FI and AFI show both a good correlation with the accumulated change in storage STOR, logically slightly higher for the AFI index (Fig. 9b). The relatively high correlation between FI and STOR (r = 0.86) can be explained by the increasing downward trend in piezometric levels at the end of the 1990s, related to the dry period 1999–2001, when basin storage was already largely depleted. Correlation between FI and the depletion rate (Fig. 9a) is rather small (−0.19), which may be caused by the local impact of pumping wells on nearby observation wells. This may be the weak point of the filling index method in regions with lots of pumping wells.

The indicators based on recharge and discharge RDR and RDD show a good (negative) correlation with the depletion rate (yearly change in storage) with correlation coefficients r of around −0.7 (Fig. 9c and e). Surprisingly, correlation with the accumulated change in storage (Table 6) is not very low with r values of around 0.30. This may be explained by the fact that 1999, 2000 and 2001 were very dry years with large yearly deficits on the water balance

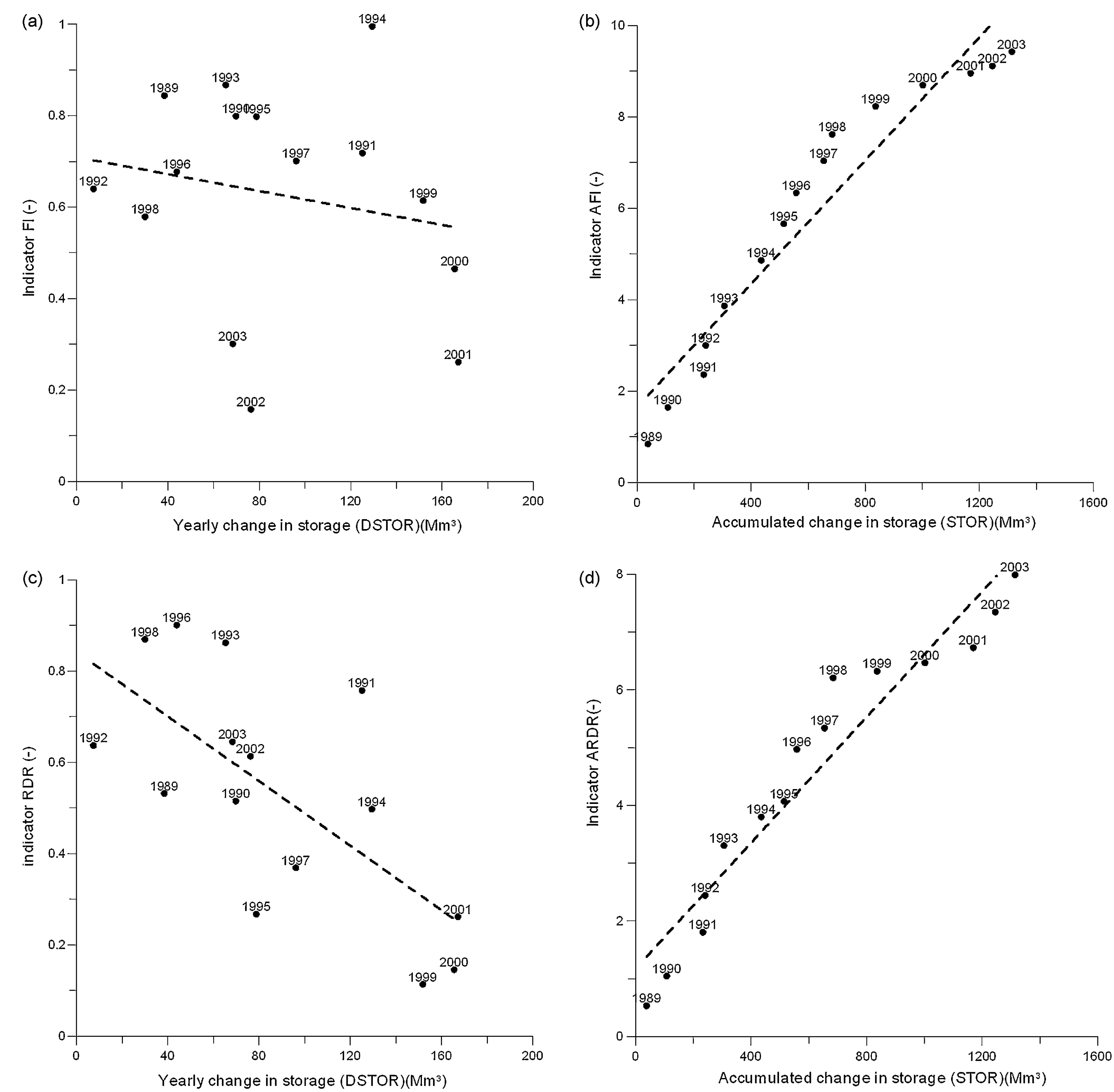

Fig. 9. Correlation between the indicators and the model calculated change in groundwater storage (DSTOR) and accumulated change in groundwater storage (STOR).

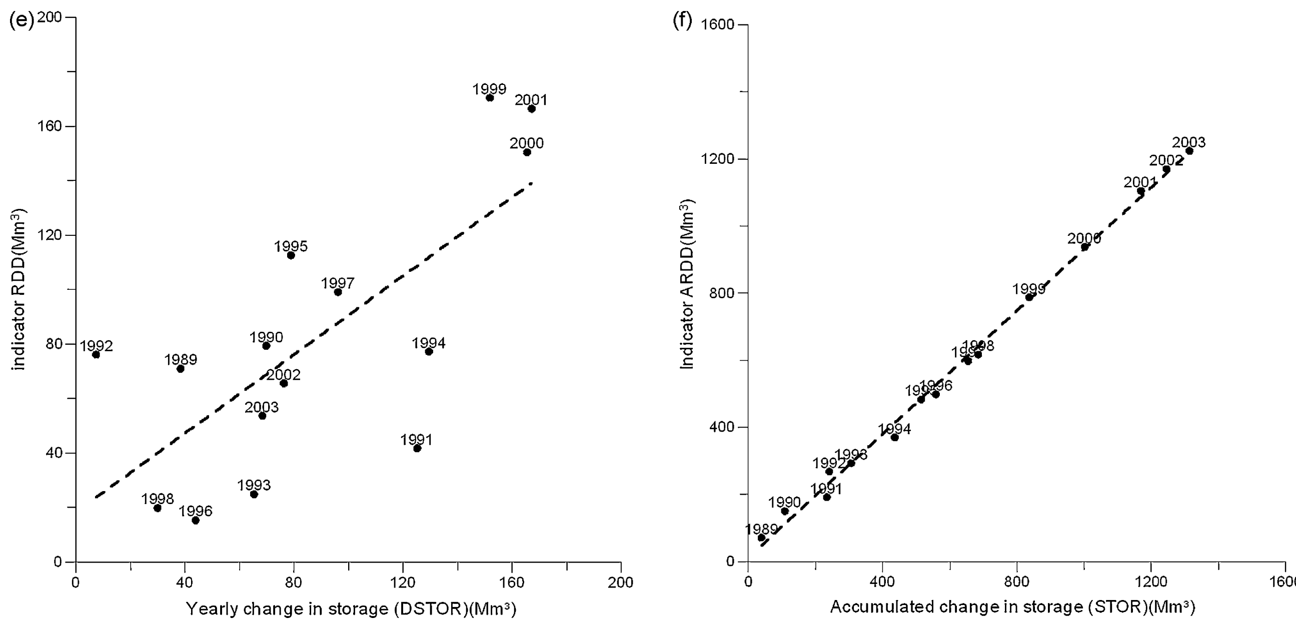

Fig. 9. (*Continued*).

of the basin and accordingly exceptionally low values for RDR and high values for RDD, while at the same time accumulated storage depletion was near its maximum. The correlation between STOR and both RDR and RDD may be overestimated by this fact. The accumulated indicators ARDR and ARDD show very good correlations ($r > 0.96$) with the accumulated change in storage (Fig. 9d and f). As the Shahrekord aquifer is located in a nearly closed basin with very limited lateral inflow and outflow of water, changes in storage are nearly exclusively controlled by balancing exploitation rates versus recharge from rainfall events.

The (very) high correlation degree of some indicators may, at least partly, be enhanced, by the following facts:

- The aquifer system has been continuously depleted during the last 15 years and an overall rather systematic declining trend in piezometric levels and hence groundwater storage, is observed.
- The model uses a single uniform value for specific yield (0.24).
- Accidently a succession of three dry years occurred near the end of the considered 15 year's period when storage depletion was already at its maximum. These three years had exceptional values for the recharge related indicators.
- In the Shahrekord aquifer lateral inflow and outflow is quite limited, making the water balance very sensitive to temporal variations in recharge.
- Both the model and the indicators RDR, ARDR, RDD and ARDD are based on the same dataset of recharge estimations and recorded discharge amounts.

4. Conclusions

In overexploited aquifer systems in closed basins with very limited lateral inflow and outflow, three different indicators based on an accumulated index show a good correlation ($r > 0.95$) with groundwater storage. The simple accumulated difference between estimated aquifer recharge and groundwater discharge gives the best correlation ($r > 0.99$) with aquifer storage, better than indicators based on the ratio of abstraction rate to aquifer recharge ($r = 0.96$), or indicators based on the filling index, derived from piezometric time series ($r = 0.95$). The yearly difference between recharge and discharge gives the best indication for the rate at which the groundwater storage is depleted. Although the accumulated filling index showed the smallest correlation ($r = 0.95$) with storage, it has the advantage of being solely based on simple water-level measurements and does not need complicated calculations using uncertain parameterisations. However, an extensive monitoring network is a necessity. This has been demonstrated for an overexploited basin in Iran where lateral inflow and outflow are small compared to recharge and discharge rates. The use of the proposed indicator(s) is only recommended in this type of aquifers.

Acknowledgements

The authors would like to thank the Ministry of Power of Iran and in particular appreciate the Departments of Water and Meteorology in Shahrekord for their support in data collection. The second author would like to thank the Ministry of Research, Science and Technology of Iran and Shahrekord University for their Financial support and Ghent University for accepting to guide him during his PhD research.

References

Alley, M., Leake, S.A., 2004. The journey from safe yield to sustainability. Ground Water 42 (1), 12–16.
Alley, W.M., Reilly, T.E., Franke, O.L., 1999. Sustainability of ground-water resources. U.S. Geological Survey Circular, 1186.
Anuraga, T.S.K., Ruiz, L., Mohan Kumar, M.S., Sekhar, M., Leijnse, A., 2006. Estimating groundwater recharge using land use and soil data: a case study in South India. Agricultural Water Management 84, 65–76.
Bredehoeft, J., 1997. Safe yield and the water budget myth. Ground Water 35 (6), 929.
Bredehoeft, J.D., 2002. The water budget myth revisited: why hydrogeologists model. Ground Water 40 (4), 340–345.
Bredehoeft, J.D., Papadopulus, S.S., Cooper Jr., H.H., 1982. Groundwater: The Water-budget Myth in Scientific Basis of Water Resource Management, Studies in Geophysics. National Academy Press, Washington, D.C, pp. 51–57.
Custodio, E., 2002. Aquifer overexploitation: what does it mean? Hydrogeology Journal 10 (2), 254–277.
Custodio, E., Kretsinger, V., Llamas, M.R., 2005. Intensive development of groundwater: concept, facts and suggestions. Water Policy 7, 151–162.
Devlin, J.F., Sophocleous, M., 2005. The persistence of the water budget myth and its relationship to sustainability. Hydrogeology Journal 13 (1), 549–554.
Foster, S., Loucks, D.P. (Eds.), 2006. Non-Renewable Groundwater Resources. A Guidebook on Socially-sustainable Management for Water Policy Makers. IHP-VI, Series on Groundwater No. 10; UNESCO 2006.

Harou, J.J., Lund, J.R., 2008. Ending groundwater overdraft in hydrologic-economic systems. Hydrogeology Journal 16 (6), 1039–1055.
IGES, 2008. Climate Change Policies in the Asia-Pacific: Re-Uniting Climate Change and Sustainable Development. Kanagawa, Japan.
Kalf, F.R., Woolley, D.R., 2005. Applicability and methodology of determining sustainable yield in groundwater systems. Hydrogeology Journal 13 (1), 295–312.
Kinzelbach, W., Aeschbach, W., Alberich, C., Goni, I.B., Beyerle, U., Brunner, P., Chiang, W.H., Rueedi, J., Zollmann, K., 2002. A survey of methods for groundwater recharge in arid and semi-arid regions. Early Warning and Assessment Report Series, UNEP/DEWA/RS.02-2. United Nations Environment Programme, Nairobi, Kenya.
Konikow, L.F., 2002. Groundwater depletion and over-exploitation: a global problem. 2002 Denver Annual Meeting, October 27–30, 2002, Geological Society of America.
Lavapuro, M., Lipponen, A., Artimo, A., Katko, T.S., 2008. Groundwater sustainability indicators: testing with Finnish data. Boreal Environment Research 13, 381–402.
Llamas, M.R., Custodio, E., 2003. Main Common Concepts, Relevant Factors, and Some Suggestions. In: Llamas and Custodio (Eds.). Intensive Use of Groundwater: Challenges and Opportunities. Balkema. Publishers. Dordrecht. Published also in IHP-VI, series on Groundwater, n(4, 18 p. UNESCO (2002)).
Llamas, M.R., Martinez-Santos, P.A., de la Hera, A., 2006. The Manifold Dimensions of Groundwater Sustainability: An Overview. In: Ragone, S., de la Hera, A, Hernandez-Mora, N (Eds.), The Global Importance of Groundwater in the 1st Century: Proceedings of the International Symposium on Groundwater Sustainability. National Ground Water Association Press, Ohio, USA. ISBN 1-56034-131-9, pp. 105–116.
Maimone, M., 2004. Defining and managing sustainable yield. Ground Water 2 (6), 809–814.
McDonald, M.G., Harbaugh, A.W., 1988. A Modular Three-Dimensional Finite-Di4fference Ground-Water Flow Model: U.S. Geological Survey Techniques of Water Resources Investigations, Book 6, Chapter A1, 586 pp.
Meinzer, O.E., 1920. Quantitative methods of estimating groundwater supplies. Geological Society of America Bulletin 31, 329–338.
Morris, B.L., Lawrence, A.R.L., Chilton, P.J.C., Adams, B., Calow, R.C., Klinck, B.A., 2003. Groundwater and its Susceptibility to Degradation: A Global Assessment of the Problem and Options for Management. Early Warning and Assessment Report Series, RS. 03-3. United Nations Environment Programme, Nairobi, Kenya.
Pool, D.R., 2002. Gravity methods for monitoring groundwater storage change in groundwater depletion and over-exploitation: a global problem. 2002 Denver Annual Meeting, October 27–30, 2002. Geological Society of America.
Radfar, M., 2009. Hydrogeological and Hydrochemical Characterisation and Modelling of the Tertiary-Quaternary Aquifer System in Shahrekord Plain–Iran. Ph.D. Dissertation. Ghent University.
Sophocleous, M., 1997. Managing water resources systems: why “safe yield” is not sustainable. Ground Water 35 (4), 561.
Sophocleous, M., 2000. From safe yield to sustainable development of water resources:the Kansas experience. Journal of Hydrology 235 (1–2), 27–43.
Sophocleous, M., 2001. Environmental implications of intensive groundwater use with special regard to streams and wetlands. In: Intensive Use of Groundwater. Challenges and Opportunities. A.A. Balkema Publishers. The Netherlands.
Thornthwaite, C.W., 1948. A new and improved classification of climates. Geographical Review 38 (1), 55–94.
Thornthwaite, C.W., Mather, J.R., 1955. The Water Balance. Publ. in Climatology, 8(1), C.W Thornthwaite & Associates, Centerton, New Jersey.
Thornthwaite, C.W., Mather, J.R., 1957. Instructions and Tables forComputing Potential Evapotranspiration and the Water Balance. Publ. in Climatology, 10(3), C.W. Thornthwaite & Associates, Centerton, New Jersey.
United Nations, 1992. Interregional workshop on groundwater overexploitation in developing countries. Gran Canaria, Canary Islands, Spain, United Nations, Department of Technical Cooperation for Development. Prepared by J. Dijon and E. Custodio. New York—Barcelona. U.N. INT/90/R43, New York.
Vrbo, J., Lippanen, A. (Eds.), 2007. Groundwater Resources Sustainability Indicators. UNESCO, Paris.
Willmott, C.J., 1977. Watbug: A FORTRAN IV Algorithm for Calculating the Climatic Water Budget. Publications in Climatology 30, C.W. Thornthwaite Assoc., Laboratory of Climatology, Elmer, NJ, p. 55.