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GROUND-WATER STORAGE CALCULATION IN KARST AQUIFERS WITH ALLUVIUM OR NO-FLOW BOUNDARIES

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Abstract: The determination of water-budget parameters, such as change in storage and subsurface inflow and outflow, is costly and unreliable due to heterogeneities of karst aquifers. Some karst aquifers may have one or a combination of boundaries such as impermeable formations, alluvial aquifers, and known ground-water divides. Karst water only discharges through springs or flows to the adjacent alluvium. A new procedure is proposed to estimate volume of storage in region during the dry season in these settings. The subsurface inflow and outflow can be measured in the adjacent alluvium using equipotential and flow lines, cross-sectional area, and transmissivity of the alluvial aquifer. The dry season makes it possible to calculate the karst spring recession coefficient and karst aquifer dynamic volume at the beginning and end of the hydrological year. The change of storage is the difference between the dynamic volumes of the karst aquifer at the beginning and end of the hydrological year. The volume of water which flows to the adjacent alluvium or spring is measured by plotting the discharge as a function of time and estimating the recession coefficient at the beginning (or end) of the hydrological year. Known equations are used to calculate the dynamic volume of springs. A general equation is proposed to calculate the dynamic volume of a karst aquifer when there is a combination of springs, and subsurface inflow and outflow from the karst aquifer. The proposed method is applicable to the Zagros Folded Zone in Iran.

Introduction

Karst aquifers are classified on a scale that ranges from diffuse, mixed, to conduit (White, 1969). Two extremes can be recognized: a diffuse aquifer which is poorly integrated and has little influence on karst ground-water circulation (Shuster and White, 1971; Atkinson, 1977) and a conduit aquifer, in which ground-water flow is dominated by the conduit system (White, 1988). Most karst aquifers contain both diffuse and conduit regimes in which water flows in conduits as well as fractures and pores. These characteristics of karst aquifers result in extensive variations of hydraulic parameters and water-surface elevations, even over short distances.

The determination of some water-budget parameters, such as storage volume and subsurface inflow and outflow, require the measurement of water levels in boreholes and hydraulic parameters in exploration wells. It is very difficult to obtain reliable ground-water parameters and water-table configurations in a heterogeneous karst system.

The objective of this study is to propose equations for estimating the change of storage in a semi-arid karst system with no-flow boundaries (impermeable or known ground-water divide) in which the karst water discharges only from a karst spring. This method is applied in the Sheshpeer aquifer of Iran. A general equation is proposed to calculate the dynamic volume and change of storage of a karst aquifer when there is a combination of springs, and subsurface inflow and outflow from the karst aquifer to the adjacent alluvial aquifers. The karst subsurface inflow and outflow can be measured in the adjacent alluvium.

Water Budget

A hydrologic budget is a quantitative evaluation of inflow, outflow and the change in storage over a specified time interval, usually a hydrological year. A hydrological year starts and ends when storage is at its minimum. The mass-balance equation for a karstic high mountain aquifer can be simplified as follows

$$I_a + I_s + I_{ss} = O_a + D + E + \Delta V \qquad (1)$$

where I_a is the subsurface inflow from the adjacent aquifer, I_s is the effective recharge, I_{ss} is surface-water seepage which reaches the phreatic ground water, O_a is subsurface outflow to an adjacent aquifer, D is the discharge by pumping wells, springs and qanats (a qanat is a sloped longitudinal channel which is dug below the ground-water surface to intercept ground water and at the end of the qanat, water discharges by gravity into an irrigation channel), E is evaporation from the water table, and ΔV is the change in ground-water storage between the beginning and the end of the hydrological year and may be positive or negative. If the karst aquifer is in direct contact with an adjacent aquifer, the subsurface inflow and outflow can be measured in the adjacent aquifer. This will later be explained in the discussion section.

If the run-off leaves a karst formation, joining the main drainage on a non-karst formation, effective recharge (I_s) is estimated in several typical surface sub-basins by the following equation

$$I_s = P - (R + E_t) \tag{2}$$

where P is precipitation, R is run-off and E_t is actual evapotranspiration or sublimation. If the karst area is covered by snow, then sublimation should be measured instead of evapotranspiration. Runoff (R) is estimated by measuring the flow rates at the outlet of the typical karstic sub-basin. Evapotranspiration (E_t) can be estimated by direct field measurement (soil-water content between two consecutive rainfalls or lysimeters) or by climatological and crop data (Jensen, 1981). Sublimation is measured by lysimeter or calibrated equations.

If a permanent river or channel flows on the surface of a karst aquifer, its seepage (I_{ss}) can be estimated by measuring the discharges at the beginning and end of the reach which traverses the karst aquifer.

Change in Storage

The change in storage can be calculated for a simple karst system if the following conditions are met:

1. The karst aquifer is bounded by impermeable layers (Fig. 1a) and/or a known ground-water divide. The impermeable layers and ground-water divide are no-flow boundaries.
2. The karst water discharges only from a karst spring.
3. There is no precipitation during the dry season, such that the spring recession coefficient can be calculated at the beginning and end of the hydrological year.

The volume of stored water in the saturated zone above the level of the outflow spring is termed the dynamic volume of the spring (Mangin, 1975; Ford and Williams, 1994). If there is no precipitation and base-flow conditions prevail, the dynamic volume gradually discharges from the spring until the spring is completely dry. If the spring discharge is plotted as a function of time, the dynamic volume at any time t is equal to the area under the curve bounded between the time t and the time when discharge reaches zero. On a semi-logarithmic scale, this curve becomes a line with slope $-\alpha$, called the recession coefficient (Fig. 2a). Maillet (1905) proposed the following simple exponential equation for this line

$$Q = Q_0 e^{-\alpha t} \tag{3}$$

where Q is the discharge ($m^3\ s^{-1}$) at time t, Q_o is the discharge at time zero, t is the time elapsed (day) between Q and Q_o, and α is the recession coefficient (day^{-1}). Thus the spring dynamic volume (V), in m^3, is (Milanović, 1981; Ford and Williams, 1994)

$$V = C \int_t^\infty Q_0 e^{-\alpha t}\, dt = C \frac{Q}{\alpha} \tag{4}$$

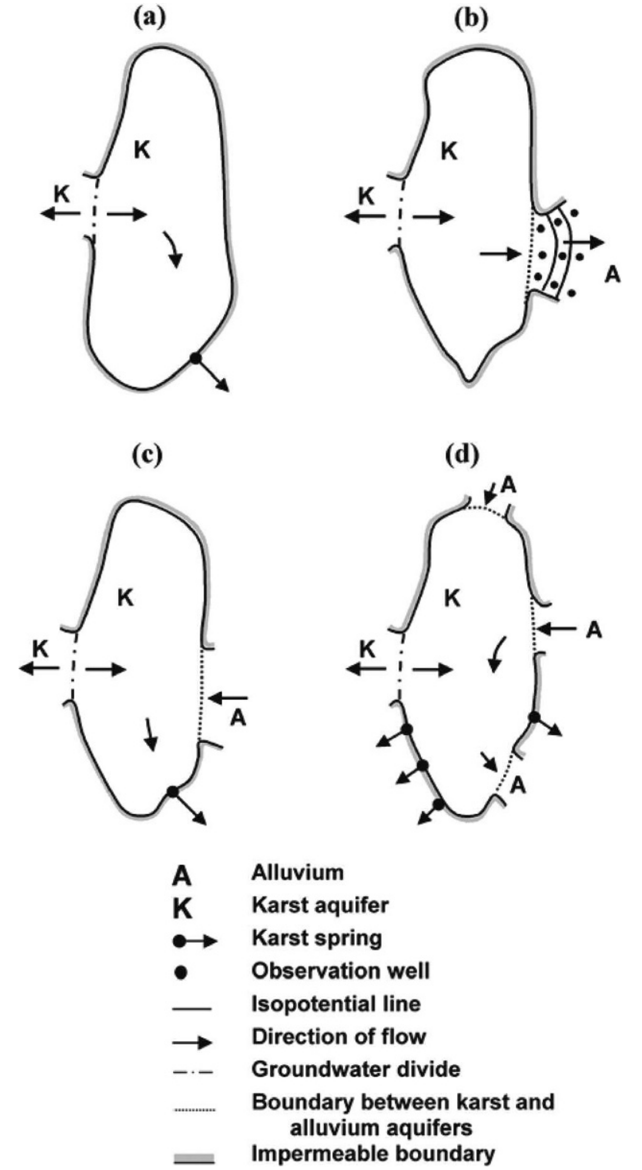

Figure 1. Schematic diagram of the boundary conditions of karst aquifers.

The constant C, equal to 86400, is the unit conversion factor (days to seconds).

The change of storage in the karst aquifer (ΔV) is estimated as the difference between the dynamic volume at the beginning (V_{S1}) and end (V_{S2}) of the hydrological year.

$$\Delta V_S = V_{S1} - V_{S2} \tag{5}$$

The subscript s denotes spring. If there is no change in the recession coefficient after the beginning or end of the hydrologic year, the change of storage may be estimated by a simple combination of Equations (4) and (5).

$$\Delta V_S = C \left[\frac{Q_{S1}}{\alpha_{S1}} - \frac{Q_{S2}}{\alpha_{S2}} \right] \tag{6}$$

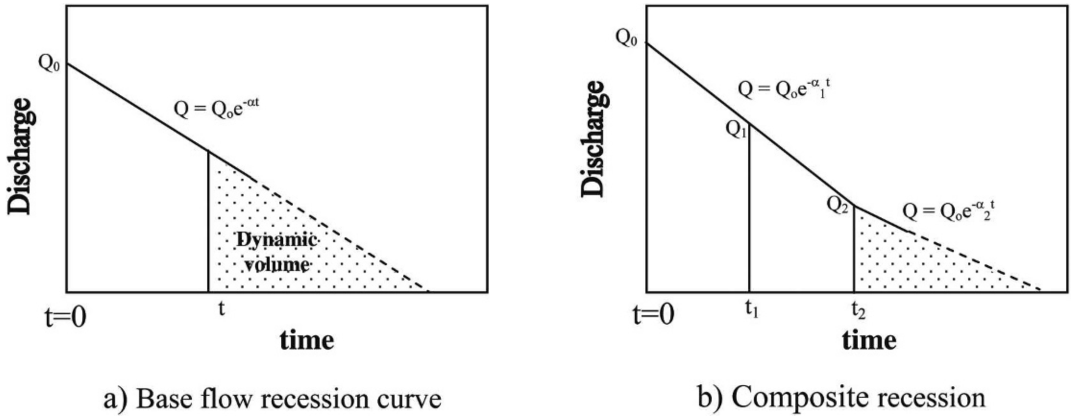

Figure 2. Schematic diagram of recession curves: a) base flow recession curve; b) composite recession curve.

The recession coefficient may change during the recession process (Forkasiewicz and Paloc, 1967). Therefore, if there is a change of recession coefficient (Fig. 2b), the dynamic volume is calculated as follows:

$$V = C\int_{t_1}^{t_2} Q_0 e^{-\alpha_1 t}\, dt + C\int_{t_2}^{\infty} Q_0 e^{-\alpha_2 t}\, dt = C\left(\frac{Q_1 - Q_2}{\alpha_1} + \frac{Q_2}{\alpha_2}\right) \qquad (7)$$

in which α_1 and α_2 are the recession coefficients (day^{-1}) of the two slopes respectively, Q_1 is the discharge (m^3 s^{-1}) at the time that dynamic volume is being estimated, Q_2 is the discharge (m^3 s^{-1}) at the point where the slope changes, t_1 is the time (day) when the dynamic volume is being estimated, and t_2 is the time (day) when the slope changes. If several recession coefficients are observed after the time when dynamic volume is being estimated, a similar approach results in the following general equation

$$V = C\left[\frac{Q_1 - Q_2}{\alpha_1} + \frac{Q_2 - Q_3}{\alpha_2} + \frac{Q_3 - Q_4}{\alpha_3} + \cdots + \frac{Q_{n-1} - Q_n}{\alpha_{n-1}} + \frac{Q_n}{\alpha_n}\right] \qquad (8)$$

where n is the number of recession coefficients. To estimate the dynamic volume more accurately, it is suggested that discharge be measured very accurately and at least weekly since the dynamic volume is very sensitive to the value of α and the discharge.

SIMPLE CASE STUDY

In the simple case study, the karst water only discharges through a karst spring and there is no inflow from the adjacent alluvium. The study area is located to the west of Shiraz, Iran. Extensive geological, geomorphological, hydrogeological (including dye tracing), hydrochemical and hydrological studies have been carried out in this study area (Pezeshkpoor, 1991; Porhemat, 1993; Raeisi et al., 1993; Raeisi and Karami, 1996; Raeisi and Karami, 1997; Raeisi et al., 1999). The Barm-Firooz and Gar (Mor and Gar Mountains) anticlines extend in the general direction of the Zagros Mountain Range (Fig. 3). The exposed cores of the anticlines dominantly consist of the calcareous Sarvak Formation (Albian-Turonian), underlain and overlain by impermeable shales of the Kazhdumi (Aptian-Cenemonian) and Pabdeh-Gurpi (Cretaceous-Tertiary) Formations, respectively. The tectonic features are a main thrust, and normal and strike-slip faults. The strike-slip fault has produced suitable conditions for extensive karstification. The most important exokarst feature is the presence of 160 sinkholes on the northern flank of Gar Mountain, and 99 sinkholes in Barm-Firooz Mountain (Fig. 3). The Sarvak Formation is present at the highest points of the study area, with a maximum elevation of 3714 m asl. and a minimum of 2110 m asl.

Out of twelve springs emerging from the Sarvak Formation, only Sheshpeer Spring occurs on the northern flank, with a mean annual discharge of 3247 L s^{-1}. One of the springs, Berghan, has a mean annual discharge of 632 L s^{-1}. The mean annual discharge of all the other springs ranges from 1.41 to 68.34 L s^{-1} (Raeisi and Karami, 1997). Precipitation occurs during late fall, winter and early spring. There is no precipitation during the rest of the year. The precipitation in winter is mainly in the form of snow, most of which melts by mid-spring. The average annual precipitation at Berghan station (elevation 2110 m) is 750 mm. Using the regional relationship between elevation and rainfall, the average annual precipitation of the Sheshpeer catchment area is calculated to be 1350 mm (Raeisi et al., 1993 and Porhemat, 1993). Soil cover overlies 35% of the Sarvak Formation. About 95 percent of the soils belong to the regosol and lithosol categories, and the remaining 5 percent are related to the brown soils (Raeisi and Karami, 1996). The study area is a natural pasture.

The 81 km^2 catchment area of the Sheshpeer Spring (Fig. 3) consists primarily of the northern flank of Barm-Firooz and Gar Mountains and a small portion of the

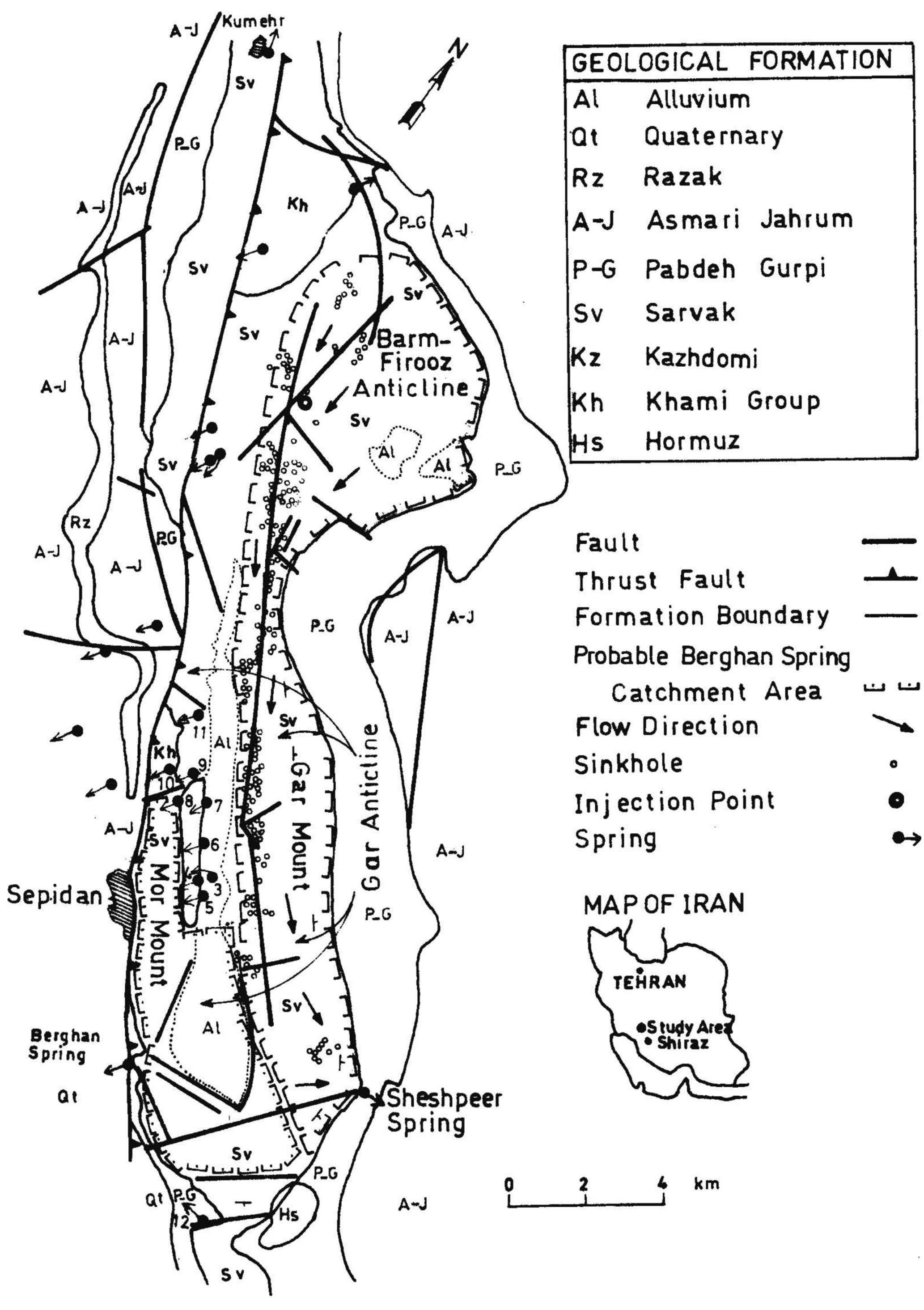

Figure 3. Hydrogeological map of the study area (after Raeisi and Karami, 1996).

southern flank of Barm-Firooz Mountain (Pezeshkpoor, 1991). The indicated boundary for the catchment area is based on:

1. The northern flank of the Gar and Barm-Firooz anticlines have been brought up by tectonic stress, such that the aquifers of the northern and southern flanks have been disconnected and the underlying impermeable Khazdumi Formation is outcropping in a part of the anticline core (Fig. 3). Dye-tracing tests confirm this hydrogeological disconnection between the southern and northern flanks (Raeisi et al., 1999).
2. The northwest and northeast sides of the anticlines are bounded by the impermeable Pabdeh-Gurpi Forma-

tions (Fig. 3 and 4). Water flow under the Pabdeh-Gurpi Formations is not possible because (a) karstification usually does not occur under an 800-meter thick layer of Pabdeh-Gurpi Formations, (b) there are no outcrops of Sarvak Formation in parallel adjacent anticlines, and (c) no dye was detected in the springs of the adjacent anticlines.

3. The sinkholes are only located in the catchment area of Sheshpeer Spring. Sodium fluorescein injected in a sinkhole 18 km away from the Sheshpeer spring (Fig. 3) appeared only in this spring (Raeisi et al., 1999).

It may be concluded that the basin divide of Sheshpeer Spring is a no-flow boundary and all the recharged water emerges only from the Sheshpeer spring.

The hydrographs of Sheshpeer Spring for 1991 and 1992 are presented in Figure 5. The discharge data in the summer of 1991 were measured monthly. Although this reduces the precision of the recession coefficient, it is still accurate enough to describe the proposed method. The recession curves of Sheshpeer Spring for the beginning and end of the hydrological year are presented in Figure 6. The reference time is May 1, so the beginning of the hydrological year (Sep. 21, 1991) is day 144 and the end of the hydrological year is day 509. By fitting the data to Equation (3), the following relation is obtained:

$$Q = 2.9002e^{-0021t} \tag{9}$$

Using Equation (9) for the beginning of the hydrological year (t =144), the discharge is 2.143 $m^3\ s^{-1}$. From Equation (4), the dynamic volume at the beginning of the hydrological year is 88.2 $\times 10^6$ m^3.

At the end of the hydrological year (September 21, 1992), the recession equation is as follows (Fig. 6):

$$Q = 19.937e^{-0.0038t} \tag{10}$$

At the end of the hydrological year (t = 509 day) the discharge (Q_1) is 2.88 $m^3\ s^{-1}$. In 1992, discharge was only measured up to the end of November (Fig. 5), therefore it is not known if the recession curve continues with the same slope or shows a change of slope after this time. In 1991, the recession coefficient changed to 0.0021 at a discharge of 2.5 $m^3\ s^{-1}$. Because the recession coefficient is dependent on discharge, a similar coefficient is also expected in 1992 at discharges lower than 2.5 $m^3\ s^{-1}$. Taking this into account, the recession curve of 1991 is transferred to 1992 starting at a discharge (Q_2) equal to 2.5 $m^3\ s^{-1}$ (dotted line on Fig. 6). The dynamic volume at the end of the hydrological year is calculated to be 111.5 $\times$ 10^6 m^3 using Equation (5). The change of storage using Equation (7) is 23.3 $\times 10^6$ m^3.

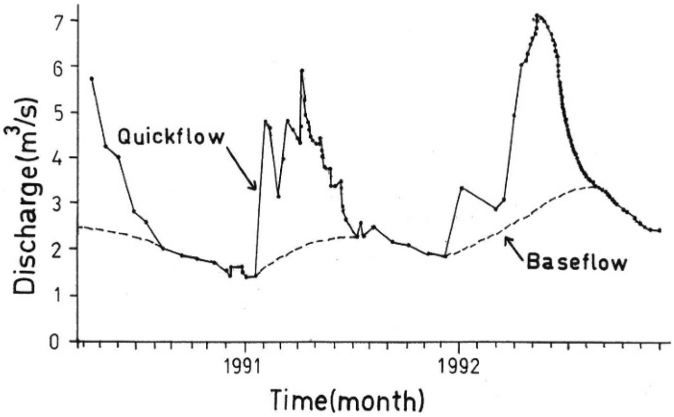

Figure 5. Hydrograph of Sheshpeer spring.

The subsurface inflow (I_a) and subsurface outflow (O_a), and evaporation from the water table (E) in Equation 1 are zero in the study area. There are neither wells nor qanats in the study area, so the total volume of Sheshpeer spring discharge (D) was 112.2 $\times$ 10^6 m^3 during the hydrological year. Applying the above data to Equation (1), a value of 134.7 $\times$ 10^6 m^3 is obtained for effective recharge from precipitation and surface water seepage (I_S). Because there are neither rivers nor channels in the catchment area, the volume of precipitation reaching ground water (I_S) is 134.7 $\times$ 10^6 m^3. The volume of rainfall was about 146.5 $\times$ 10^6 m^3 during the hydrological year. This shows that the infiltration coefficient is 91.9%.

Porhemat (1993) measured runoff and sublimation directly, and estimated snow melt and sublimation by the

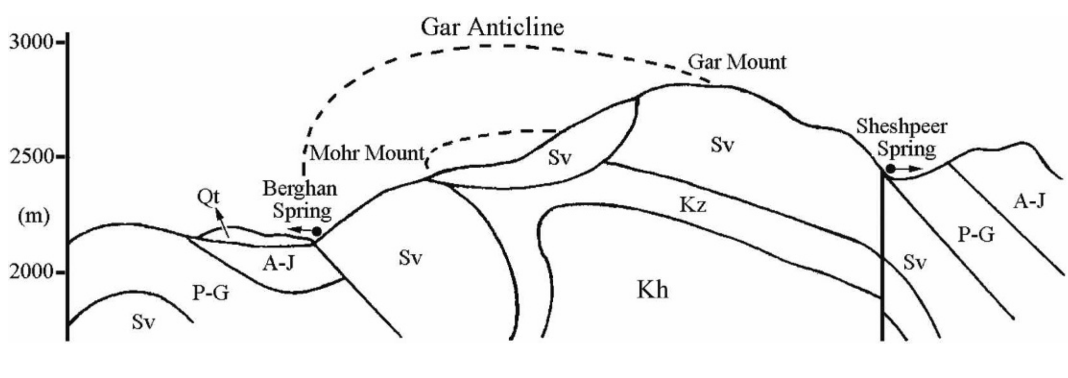

Figure 4. Geological cross section between Berghan and Sheshpeer Springs, The legend is shown on Figure 3 (after Raeisi et al., 1993).

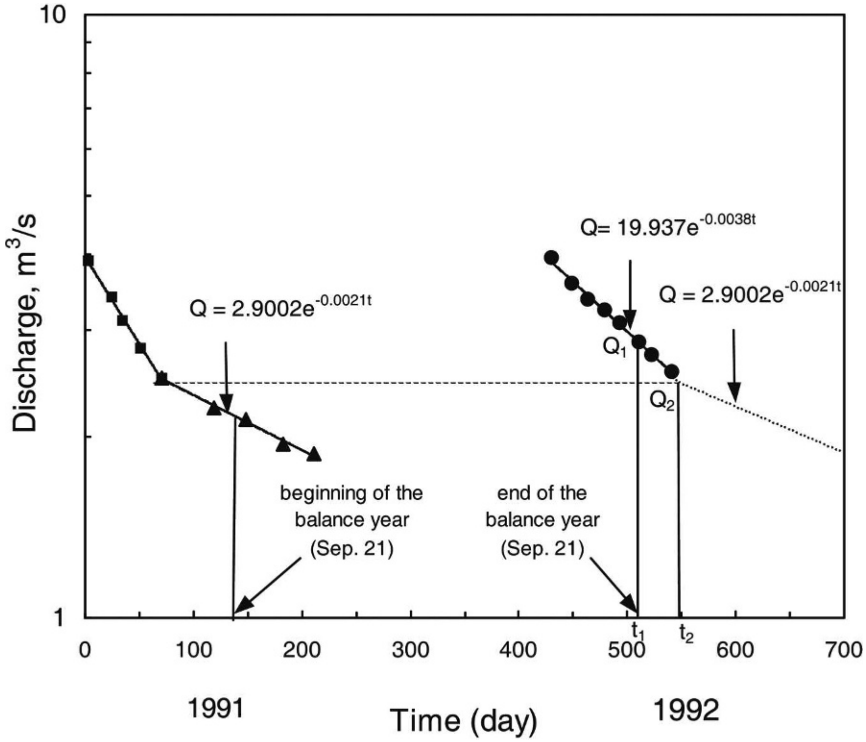

Figure 6. Recession curves of Sheshpeer spring around the beginning and end of the hydrological year.

regionally calibrated equations in three small catchment areas in the Barm-Firooz anticline. The average percentage of sublimation was 5.2%, runoff was 25.3%, and infiltration from melting snow was 69.5%. The catchment of the study area is covered partly by sinkholes; therefore, parts of the runoff flow into the sinkholes, justifying the value of 91.9% for the infiltration coefficient estimated from water-budget equations. Discharge was measured monthly instead of weekly, reducing the accuracy of recession coefficient. This in turn affects the estimated value of change of storage.

Discussion

Boreholes and exploration wells are very small features on the scale of the heterogeneities of a karst aquifer. Values of hydrodynamic parameters obtained from pump tests vary widely over short distances depending on exactly where the wells are drilled (White, 1988). A well that taps a connection with the conduit system can produce very large yields of water with negligible drawdown. A well drilled a few meters away in an unfractured or unkarstified block of limestone may have a very small yield of water. Therefore, it is very difficult to obtain reliable groundwater parameters and water-table configurations in such a heterogeneous system. In addition, the construction of observation and pumping wells is very expensive in the highland karst zone of Iran due to deep water tables and lack of accessible roads.

If the karst aquifer is in direct contact with an adjacent alluvium aquifer overlying an impermeable layer, the subsurface outflow from the karst aquifer can be measured in the adjacent alluvium aquifer, thus avoiding the heterogeneity of the karst aquifer. The outflow from the karst aquifer is equal to subsurface inflow into the adjacent alluvial aquifer. A piezometric network is constructed in the adjacent alluvium. The subsurface outflow can be measured in the adjacent alluvium using the flow net formed by equipotential and flow lines, cross-sectional area, and transmissivity (Todd and Mays, 2005). This is a commonly used method in alluvial aquifers. Three rows of piezometers are constructed in the adjacent alluvium, perpendicular to the direction of flow (Fig. 1b). All the piezometers must reach the top of the impermeable layer beneath the alluvium. The outflow may be measured monthly during the dry season and more frequently during the wet season; therefore, the total annual volume of outflow can be determined. The advantage of this method is that one is no longer confronted with the problems of the high expenses of constructing piezometric networks and pumping wells in a highland karst aquifer, the hydraulic-parameter heterogeneity, and complexity of water-surface configuration.

To overcome the problems of karst heterogeneity, several methods have been proposed in order to avoid the direct measurement of storage coefficients. Milanović (1981) suggested estimating the storage coefficient using the recession coefficient of a spring hydrograph. The assumptions of homogenous medium, one dimensional and diffuse flow, constant horizontal cross-section of the aquifer and average gradient of the energy potential are not characteristics of karst aquifers. Spatial variations of storage coefficients in an aquifer cannot be detected with this method because the discharge can only be measured at one point. White (1988) suggested estimating the water budget over a sufficiently long time such that the change of storage approaches zero. Unfortunately, long time data are rarely available for many karst regions. In addition, the water budget cannot be determined for a wet, a dry, or a specific year. Milanović (1981) suggested choosing a balance year with negligible change of storage. In this method, the project should be postponed until the occurrence of a year with negligible change of storage.

The simplified case of a karst aquifer with no-flow boundaries having its water discharged only through springs was explained earlier. In addition, the change in storage may be determined while the karst aquifer is partially or totally in direct contact with the alluvium. Two different karst aquifers with simplified boundary conditions are considered for estimating the change in storage, and finally, the general case will be discussed:

Case 1

The karst aquifer is in direct contact with the impermeable layer, known ground-water divide, and/or alluvial aquifer (Fig. 1b). The adjacent alluvial aquifer is located over an impermeable formation, such that the karst subsurface water flows only to the adjacent alluvium. There is no inflow into the karst aquifer from the adjacent alluvial aquifer. The alluvium dynamic volume is defined as the volume of water stored in a karst aquifer above the impermeable basement which is located beneath the adjacent alluvium. This water is gradually discharged to the adjacent alluvium and decreases to zero under the assumption of no precipitation.

The subsurface outflow from the karst aquifer can be measured in the adjacent alluvium using the method explained in the previous paragraphs. This outflow must be measured at least once a week in the dry season, starting from two to three months before the beginning (and end) of the hydrological year until the first rainfall. The alluvium recession coefficient is then estimated as the slope of the semi-logarithmic plot of outflow versus time using the Maillet (1905) equation. Equations (3) and (4) are used to estimate the alluvium dynamic volume. In this case, the change of storage (ΔV_O) is the difference between the alluvium dynamic volume at the beginning of the hydrological year (V_{O1}) and the alluvium dynamic volume at the end of the hydrological year (V_{O2}):

$$\Delta V_O = V_{O1} - V_{O2} \tag{11}$$

where the subscript o denotes outflow from the karst aquifer. If there is no change in the recession coefficient after the beginning or end of the hydrological year, the change of storage may be estimated by a simple combination of Equations (4) and (11)

$$\Delta V_O = C\left[\frac{Q_{O1}}{\alpha_{O1}} - \frac{Q_{O2}}{\alpha_{O2}}\right] \tag{12}$$

Case 2

The karst aquifer boundary is similar to Case 1, but there is inflow from the adjacent alluvium into the karst aquifer. Karst water discharges only from a karst spring (Fig. 1c). In this case, water discharging from the spring is a combination of water stored in the karst aquifer and the incoming water from the adjacent alluvium

$$V_S = V_a + V_E \tag{13}$$

where V_S is the total volume of water discharging from the karst spring and can be measured directly, and V_E is the dynamic volume of the adjacent alluvium from which water enters the karst aquifer. V_E is estimated by the same procedure mentioned for the flow of karst water to an adjacent aquifer. V_a is the dynamic volume of the karst aquifer and may be estimated using equation 13. In this case the change of storage is:

$$\Delta V = V_{a1} - V_{a2} = (V_{S1} - V_{E1}) - (V_{S2} - V_{E2}) \tag{14}$$

or

$$\Delta V = (V_{S1} - V_{S2}) - (V_{E1} - V_{E2}) \tag{15}$$

The subscripts 1 and 2 denote the beginning and end of the hydrological year respectively.

General Case

A karst aquifer has various springs and inflow and outflow sections (Fig 1d). The general equation is a combination of all the above cases

$$\Delta V = \sum_{j=1}^{n} (V_{S1j} - V_{S2j}) + \sum_{i=1}^{m} (V_{O1i} - V_{O2i}) - \sum_{k}^{p} (V_{E1k} - V_{E2k}) \tag{16}$$

in which V_{S1} and V_{S2} are the spring dynamic volumes, V_{O1} and V_{O2} are the outflow dynamic volumes of the adjacent alluvium, and V_{E1} and V_{E2} are the inflow dynamic volumes of the adjacent alluvium. The subscripts 1 and 2 denote the beginning and end of the hydrological year respectively and n, m, and p are the number of springs, and outflow and inflow sections of the adjacent alluvium respectively. All the above dynamic volumes can be measured using Equations (4), (5) or (6). If there is no change in the recession coefficient after the beginning or end of the hydrological year, the aquifer change of storage may be estimated by a simple combination of Equations (4) and (16)

$$\Delta V = \sum_{j=1}^{n} \left(\frac{Q_{S1j}}{\alpha_{S1j}} - \frac{Q_{S2j}}{\alpha_{S2j}} \right) + \sum_{i=1}^{m} \left(\frac{Q_{O1i}}{\alpha_{O1i}} - \frac{Q_{O2i}}{\alpha_{O2i}} \right) - \sum_{k}^{p} \left(\frac{Q_{E1k}}{\alpha_{E1k}} - \frac{Q_{E2k}}{\alpha_{E2k}} \right) \quad (17)$$

The changes of storage are calculated in all cases without the necessity of measuring the storage coefficients in exploitation wells, which are not representative of karst aquifers. This increases the accuracy of the proposed method in a complex heterogeneous karst system.

Characteristics of the Zagros Karstic Folded Zone

The proposed method is applicable in a karst aquifer with the special boundary conditions. This type of aquifer is typical in the Zagros Orogenic System, especially in the Folded Zone. The Zagros Highland occupies the borderlands of Iran, from eastern Turkey to the Oman Sea. The Zagros Orogenic System may be divided longitudinally into the folded zone, imbricated zone, and thrust zone. The folded zone, the southwestern half of the orogenic system, is a zone of strong folding produced for the most part by late Pliocene orogeny. This zone is characterized by a repetition of long and regular anticline and syncline folds. The anticlines are normally mountain ridges, mostly composed of limestone, and the synclines are valleys and plains. James and Wynd (1965) and Falcon (1974) described the stratigraphic and structural characteristics of the Zagros sedimentary sequence in detail. Ashjari and Raeisi (2006) showed that most of the aquifers in the Zagros Folded Zone form a broad highland aquifer with the following characteristics:

1. The karstic aquifers are sandwiched between two thick impermeable formations such that the hydrogeological connections between them are disconnected, except in the rare occasions that a major fault juxtaposes the two karstic formations. A cross section at the Kermansh province, west of Iran, shows that the impermeable marly and shaly Pabdeh-Gurpi Formations disconnect the hydrogeological relationship between the karstic Asmari and Illam Formations (Fig. 7). The underlying impermeable Pubdeh-Gurpi Formations inhibit loss of water from the outcropped karstic Asmari Formation to the underlying buried karstic Illam Formation.
2. The outcropped karstic units are present in the high mountains in the Simple Folded Zone. The impermeable overlying formations are eroded from the high elevation parts of the anticlines, and they are mostly exposed at the foot of the anticlines or buried under a thin alluvium (Fig. 7). The impermeable formation surrounds the karstic formation such that each anticline seems to be an independent aquifer unless a major fault changed the normal anticline setting. When the contact of karstic formations with the overlying impermeable formations acts as a local base of erosion, karst water discharges as a spring at the contact zone (Fig. 7). When there is direct connection between the karstic formation and the alluvium, karst water flows to the adjacent alluvial aquifer above the contact zone of karstic formation and the overlying impermeable layer (Fig. 7).
3. Flow from one karstic anticline to the parallel adjacent karstic anticlines is unlikely because a) the karstic formations in synclines are normally buried under very thick overburden. Karstification is usually not significant at greater depths. b) The elevations of the underlying impermeable formations under the crest of parallel anticlines are higher than the ground-water level in adjacent alluvial aquifer, such that the alluvium water cannot flow toward the adjacent anticlines.

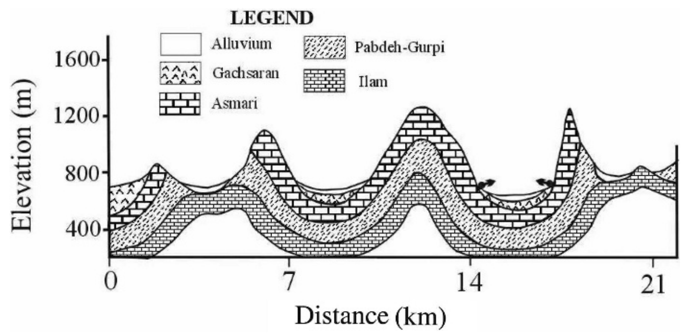

Figure 7. Alvand Basin cross section (after Karimi, 2003).

These characteristics imply that karstic formations in the Zagros anticlines are mostly independent aquifers. The aquifer boundary is limited to the overlying and underlying impermeable formations and/or adjacent alluvium. The general direction of flow is mostly toward the local base of erosion, parallel to the strike. Karst water discharges at springs or flows into the adjacent alluvial aquifer where a direct connection exists between the alluvium and the karst formation.

Examples, such as Case 1, 2, or other complex cases, require a special arrangement of piezometers in the adjacent alluvium (Fig. 1b) to enable the preparation of a reliable flow net. It was not possible to find a piezometric network similar to that of Figure 1b in Iran. In most cases, only one piezometer was found, and the piezometers were often shallow or no hydraulic conductivity data were available. Satisfactory research requires tremendous budgets; it is not possible to construct the piezometers and exploration wells with the limited research budgets of universities. It is our hope that the publication of this paper will encourage research budget support from government offices.

CONCLUSIONS

A new method is proposed to calculate the ground-water storage in karst aquifers with specific boundaries in a semi-arid karst region. These aquifers are delineated by one or more boundaries such as impermeable formations, known ground-water divides or alluvial aquifers. Karst water either discharges as a spring at the foot of the anticlines or flows into adjacent alluvial aquifers. The dry season makes it possible to calculate the recession coefficient and dynamic volume of a karst aquifer. For such conditions, several equations are proposed to estimate the budget parameters of subsurface inflow and outflow of the aquifer, and the change of storage. Subsurface inflow and outflow from the karst aquifer can be measured in the adjacent alluvium instead of the karstic aquifer itself, avoiding the heterogeneities of complex karst systems and the high expenses of constructing accessible roads, deep piezometers, and pumping wells in highlands. The change of storage is estimated as the difference between the dynamic volumes at the beginning and end of the hydrological year. The dynamic volumes of springs are calculated by the known method of plotting discharge as a function of time and using the related equations. The dynamic volume of a karst aquifer which flows to the adjacent alluvium is similarly estimated by plotting the discharge versus time, where the discharge is measured in the adjacent alluvium instead of the karst aquifer. A general equation is presented for estimating dynamic volume for the case where there is a combination of springs and inflow and outflow from adjacent aquifers. The value of the dynamic volume is very sensitive to the values of the recession coefficients, therefore it is highly recommended that the discharge be measured accurately on a weekly basis. The theory was applied for a special case where the karst water discharges only from a spring. It is recommended that the study be extended for a case with the adjacent alluvium boundary.

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REFERENCES

Ashjari, J., and Raeisi, E., 2006, Anticlinal structure influences on regional flow, Zagros, Iran: Journal of Cave and Karst Studies, v. 68, no. 3, p. 119–127.

Atkinson, T.C., 1977, Diffuse flow in limestone terrain in the Mendip Hills, Somerset (Great Britain): Journal of Hydrology, v. 35, p. 93–100.

Falcon, N.L., 1974, Southern Iran: Zagros Mountains in Mesozoic-Cenozoic orogenic belts: Geological Society of London, Special Publication, v. 4, p. 199–211.

Forkasiewicz, J., and Paloc, H., 1967, Le regime de tarissement de la Foux-de-la-Vis. Etude preliminaire: Chronique d'Hydrogeologie, BRGM, v. 3, no. 10, p. 61–73.

Ford, D.C., and Williams, P.W., 1994, Karst geomorphology and hydrology: Winchester, Massachusetts, Unwin Hyman, 601 p.

Jensen, M.E., 1981, Design and operation of farm irrigation systems: The American Society of Agricultural Engineering, Michigan, 829 p.

James, G.A., and Wynd, J.G., 1965, Stratigraphic nomenclature of Iranian Oil Consortium Agreement area: Bulletin of American Association of Petroleum Geologists, v. 49, no. 12, p. 2182–2245.

Karimi, H., 2003, Hydrogeological behavior of Alvand karst aquifers, Kermanshah, [Ph.D. thesis]: Shiraz, Shiraz University, Iran, 282 p.

Mangin, A., 1975, Contribution à l̀etude hydrodynamique des aquiferes karstiques. DES Thesis: Dijon University, France,

Maillet, E., 1905, Essais d̀hydraulique souterraine et fluviale, Hermann, Paris,

Milanović, P.T., 1981, Karst Hydrogeology: Fort Collins, Colorado, Water Resources Publications, 434 p.

Pezeshkpoor, P., 1991, Hydrogeological and hydrochemical evaluation of Kuh-e Gar and Barm-Firooz springs [M. Sc. Thesis]: Shiraz, Shiraz University, Iran, 282 p.

Porhemat, J., 1993, Evaluation of hydrological balance parameters in karstic highland catchment area. [M. Sc. Thesis]: Shiraz, Shiraz University, Iran, 427 p.

Raeisi, E., Pezeshkpoor, P., and Moor, F., 1993, Characteristics of karst aquifer as indicated by temporal changes of the springs physico-chemical parameters: Iranian Journal of Science and Technology, v. 17, p. 17–28.

Raeisi, E., and Karami, G., 1996, The governing factors of the physical and hydrochemical characteristics of karst springs: Carbonates and Evaporites, v. 11, no. 2, p. 162–168.

Raeisi, E., and Karami, G., 1997, Hydrodynamics of Berghan karst spring as indicators of aquifer characteristics: Journal of Cave and Karst Studies, v. 59, no. 3, p. 112–118.

Raeisi, E., Zare, M., and Eftekhari, P., 1999, Application of dye tracing for determining characteristics of Sheshpeer karst spring, Iran: Theoretical and Applied Karstology, v. 11–12, p. 109–118.

Shuster, E.T., and White, W.B., 1971, Seasonal fluctuations in the chemistry of limestone springs: A possible means for characterizing aquifers: Journal of Hydrology, v. 14, p. 93–128.

Todd, D.H., and Mays, L.W., 2005, Groundwater Hydrology: John Wiley and Sons, Inc., 636 p.

White, W.B., 1969, Conceptual models for carbonate aquifer: Ground Water, v. 7, no. 3, p. 15–21.

White, W.B., 1988, Geomorphology and hydrology of karst terrains: New York, Oxford University Press, 464 p.