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Use of a hydrogeochemical approach in determining hydraulic connection between porous heat reservoirs in Kaifeng area, Henan, China

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Abstract

In this paper a case study of hydraulic connectivity in a 300–1600 m deep, low temperature, sedimentary geothermal system in Kaifeng area, Henan province, China is presented. Based on lithologic data from 52 geothermal wells and chemical data on geothermal water (GW) from six depth-specific and representative wells, the system was chemically grouped into two main hot reservoirs (300–1300 m and 1300–1600 m deep), which were in turn, divided into six sub-reservoirs (SRs). Data on stable isotope (^{2}H and ^{18}O) ratios, radioactive isotope (^{14}C) radiation in conjunction with computation of mineral–fluid chemical equilibria were used to establish the recharge source (a mountainous region in the southwestern part of Zhengzhou, 60 km away); evaluate groundwater age which varied with well depth from 15630 ± 310 a to 24970 ± 330 a; and assess the chemical equilibrium state within the system. The results of different analysis did not suggest an obvious hydraulic connection between the two main hot reservoirs. The location of the recharge zone and the geohydrologic characteristics of the study area demonstrate that the GW utilized from the system is mainly derived from confined waters of meteoric origin.

1. Introduction

The city of Kaifeng lies in the mid-southern part of Huabei (North China) plain on the southern bank of the Yellow river within latitudes 34°10′6″ and 34°55′57″ N and longitudes 114°4′36″ and 114°56′32″ E. The study area is 120 km^2 (Fig. 1). Since 1980, a geothermal water (GW) resource in Kaifeng has been used for recreation, bathing, swimming, heat supply, health care, drink processing and Tung wood block treatment. The geothermal system in Kaifeng is considered to be a macro-geothermal field with possible daily water withdrawal amounting to 45,771.64 m^3 and total thermal energy averaging 1.221×10^{15} kcal equivalent to 5.112×10^{12} MJ or 7.099×10^8 tonnes of standard coal (Lin et al., 1999). The utilization of GW can generate appreciable economic and social

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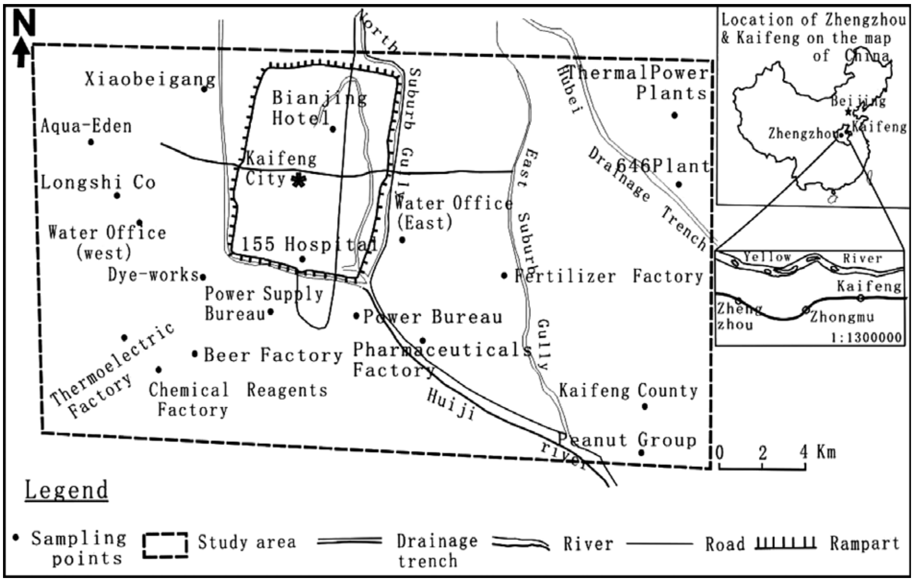

Fig. 1. Location map of the study area and sampling points.

returns. However, this utilization is being carried out in conditions which do not allow rationalized management. Particularly, the location of producing wells lacks scientific basis, and this affects the output and sustainability of the GW resource in the region. So a systematic characterization of the GW bearing formations to elucidate hydraulic connectivity within the system is appropriate. Conventional methods for the classification of hot reservoirs are mostly based on lithologic borehole data analysis. Moreover, the estimation of hydraulic connectivity between strata in a geothermal system depends on dynamic water level data. On the other hand, classification of a hot reservoir in a deeply buried geothermal system lacking dynamic water level data, poses severe problems in characterizing the properties of the system. The likely lithologic discontinuity between thermal water bearing formations, here referred to as sub-reservoirs (SRs) in the study area, has raised concern over heat transmissibility between the aquifer layers, origin and age of the GWs, geothermal gradient and ground temperature distribution, and water circulation conditions. However, enough data and information on the water chemistry of the geothermal system in Kaifeng is available, hence providing the scientific basis for the use of hydrogeochemistry to estimate hydraulic connectivity in the system. Subsequently, along with stable and radioactive isotope methods, a mineral–fluid chemical equilibrium approach was used to estimate any hydraulic connections between the SRs. The chalcedony geothermometer was used to evaluate the reservoir temperatures. At present, there are 52 geothermal wells with depth ranging from 440 to 1580 m, and water temperatures between 25 and 65 °C in the study area. Sixteen of these wells were sampled for the determination of chemical constituents in water; 9 wells were sampled for stable and radioactive isotopes (Lin et al., 2004). In order to ascertain accurate analysis, water chemical constituents from six representative wells only were used in this study. Previous isotope data on meteoric and surface waters in the Zhengzhou–Kaifeng area were also used (courtesy of the IAEA station in Zhengzhou, Henan Province).

The aim of this study was to develop a procedure that would further allow the determination of a hydraulic connection between the different SRs of the porous geothermal field in the Kaifeng area, so as to better plan investigation, evaluation and management of the regional GW resource.

2. Geological setting and geothermal background

The study area is in the southeastern corner of the Jiyuan–Kaifeng hollow geologic structure unit. Topographically, Kaifeng is characterized by very low relief from west to east. Alluvial and eolian landscapes are the most common geomorphic features, which were shaped during numerous flooding events and diversion of the Yellow River. The natural ground surface elevation ranges from 69 to 74 m above sea level. Based on hydrogeologic data, the system was divided into 6 SRs primarily made up of sedimentary deposits (silt, sandstone, clay and mudstone) from the Pleistocene (Q1) and Neogene (N) epochs, which are the main formations exploited. The bedrock is made up of metamorphic formations of Paleozoic origin, where fault-block structures and fractures dominate the tectonics. The 300–1300 m formations belong to the Minhuazhen group (Nm) while the 1300–1600 m formations are of the Guantao group (Ng) origin. Lithologic data show interbedded clay formations of 20–40 m between different SRs which may affect the hydraulic connectivity between different SRs, thus lowering efficiency of producing wells, in terms of unit water discharge (Table 1 and Fig. 2 showing a partial stratigraphical cross-section of the Zhengzhou–Zhongmu–Kaifeng area). The Nm and Ng groups are different in their water circulation properties; they have different storage temperatures, which increase with depth as established by previous studies (Lin et al., 2004). To the west of the study area there are mountains with limestone of Ordovician origin, where the recharge may enter with respect to the general regional flow direction (W–E). The Kaifeng region is characterized by a succession of west-eastwards horizontally stretched confined aquifers, with a minimum and maximum water yield of 18.01 m^3/h and 123.74 m^3/h, respectively (Table 1). The main discharge mode is the extraction of static water from geothermal wells.

The Kaifeng area is a low temperature geothermal system with an average geothermal gradient of 3.460 °C/100 m and an average heat flow of 59.80 mW/m^2. GW is quite abundant in the region. Its formation and distribution characteristics in the mid-eastern plain area of Henan province are, to a certain degree, representative of the North China Basin (Chen et al., 1994). The geothermal system in the Kaifeng area is a result of both thermal conduction and convection phenomena. However, it should be noted that thermal conduction is much more dominant. Convective transport occurs even in static groundwater, and appears to take place in the study area. It is controlled by the thermal conductivity of the geologic formations and the contained pore water (Freeze and Cherry, 1979). In Kaifeng, the geothermal system is characterized by an unevenly distributed ground temperature, which is mainly controlled by the north-eastwards geologic structures. The bedrock geologic structure (fractures and fissures) influences ground temperature distribution, which manifests higher temperature with increasing depth, this in turn, affects the deep groundwater activity. The groundwater activity is characterized by strong reactions between thermal water and the surrounding rocks; hence there are more dissolved chemicals with more complex constituents within the hottest aquifer. In the North China Basin, most of the geo-temperature anomalies can be related to the pre-Cenozoic basement uplifts (Wang et al., 1993). The more developed

Table 1
Lithologic and hydrogeologic characteristics of the Sub-reservoirs

SRs	300–450	450–600	600–800	800–1000	1000–1300	1300–1600
Geologic period	Q_1, Nm	Nm	Nm	Nm	Nm	Ng
Lithology	Silt, medium sand and clay	Medium, fine sand and clay	Fine, medium sand and clay	Fine to medium silty and sandy clay	Medium to fine silty sand and clay	Limestone, marl and mudstone
Number and thickness of layers in aquifer	9 layers, 50 m	7–9 layers, 50–65 m	7–10 layers, 50–70 m	8 layers, 70–90 m	10–13 layers, 140 m	15 layers, 140 m
Confining units between the SRs	30 m of clay	40 m of clay	30 m of clay	20–40 m of clay	>20 m of clay	20 m of Mudstone
Number of wells	13	14	9	2	13	1
Water temperature (°C)	25–30	30–38	35.5–45	40.5–50	49–55	65
Depth of static water level (m)	6.77–17.0	20.18–39.85	14.09–27.5	16.80–18.95	17.00–20.4	24.57
Depth of dynamic water level (m)	20.18–39.85	24.62–41.20	22.29–44.15	44.00–56.74	25.30–74.92	51.80
Water yield (m^3/h)	46.57– 123.74	25.0–29.3	40.0–60.0	31.0–47.92	18.01–55.0	52.0
Specific capacity (m^3/h)	3.97–7.71	4.31–7.55	2.71–6.19	1.14–1.30	0.75–2.81	2.52

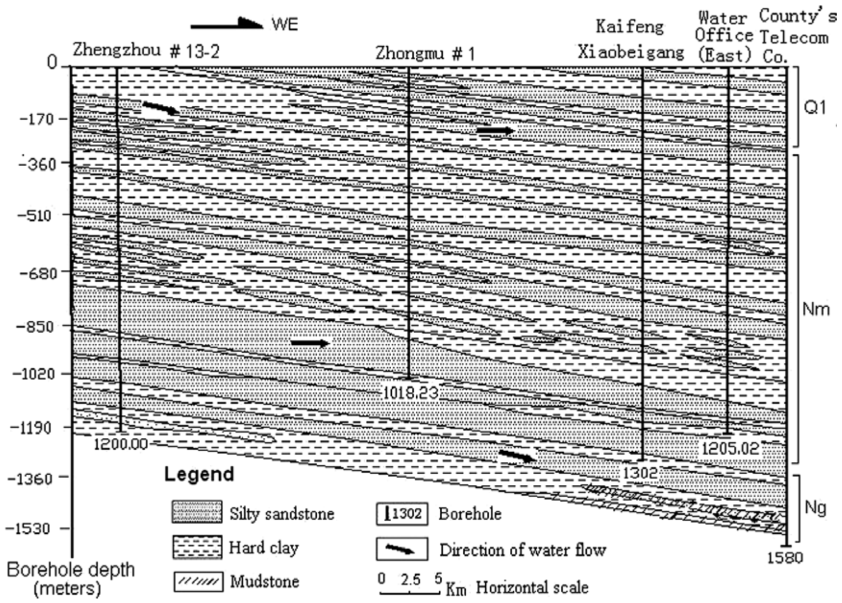

Fig. 2. Stratigraphical cross-section of the Zhengzhou–Kaifeng area showing the lithologic composition of the geothermal system.

the bedrock geologic structures (in terms of abundance of openings and their interconnectivity), the more intense the deep hot water activity, and thus the ground temperatures become higher. This finding is consistent with earlier studies by Förster et al. (1997), who observed that temperature anomalies of the order of 5–7 °C might reflect fluid movement along faults and openings, rather than causing changes in thermal conductivity resulting from different pore fluids within a system controlled by geologic features.

3. Sampling and analysis

3.1. Hot reservoir grouping

To determine the hydrogeologic characteristics of the study area, lithologic samples were collected from 52 geothermal boreholes and hydrogeological parameters from the main SRs recorded and used (Table 1). Using the geothermal borehole lithologic data, in conjunction with hydrogeologic parameters, the 300–1600 m hot reservoir depth was established and vertically divided into 6 SRs groups. Analysis of data from Table 1 shows an average thickness of clay deposits between each set of SRs to be more than 20 m. These clay deposits form the aquifuge layers of the system. From a hydrogeological point of view, none or a very poor hydraulic connection exists between different SRs mainly due to the clay layers, with the exception of the basement system in the southwestern area in Zhengzhou, with well developed faults and where the recharge comes from.

In order to chemically characterize the SRs and to verify or invalidate the lithologic and hydrogeologic divisions of the system, conventional techniques were used to analyze for the main chemical constituents of the GW from six representative wells. These wells were chosen and sampled so that they are depth-specific with respect to the 6 SRs. The samples were sent for analysis at the Station of Environmental Hydrogeology of the Mining Office in Henan. The results of the analyses are presented in Table 2.

3.2. Determination of the age and origin of the geothermal water

To establish the geochemical behavior of the geothermal system and the hydraulic relationship between the different SRs, stable and radioactive isotope measurements and analyses were carried out on nine water samples collected from representative geothermal wells (with respect to the 6 SRs). A cold ground water sample from the Hospital 155 well in Kaifeng, and a sample from Jiangang reservoir near Zhengzhou were also collected. Previous results of stable isotope analysis of four rainwater samples from Zhengzhou, Xinmi city, Weishi County and Kaifeng city, respectively, were used as well (data obtained from the IAEA station in Henan). All the samples were submitted to the

Table 2
Chemical composition of geothermal water at representative sampling locations in the study area (mg/L)

Sampling locations	Beer factory	Power bureau #2	Peanut group Co.	Longshi Co.	Water office (East)	County's telecom
SRs (m)	300–450	450–600	600–800	800–1000	1000–1300	1300–1600
Well depth (m)	450.00	550.70	660.00	860.00	1205.00	1586.00
T (°C)	29.0	30.0	38.0	40.5	53.0	65.0
pH	8.4	8.2	8.1	8.1	8.2	7.2
SiO_2	17.42	19.00	18.00	20.00	27.31	42.00
Na	185.48	184.00	210.0 0	222.30	232.55	1600.00
K	1.67	1.90	2.50	1.80	3.37	25.00
Ca	12.26	8.02	6.01	4.01	3.66	101.80
Mg	10.30	5.47	2.43	2.43	1.73	27.36
HCO_3	374.30	359.41	445.45	452.77	530.00	236.76
SO_4	84.50	85.97	50.91	66.76	35.60	44.67
Cl	50.88	43.60	33.68	40.06	28.30	2578.99
Sr	0.28	0.21	0.21	0.24	0.24	7.65
Li	0.02	0.021	0.012	0.02	0.01	0.03
Fe	0.04	0.20	0.04	0.14	0.57	0.20
Al	$<0.01^{+}$	$<0.01^{+}$	0.02	$<0.01^{+}$	0.05	$<0.01^{+}$
F	1.30	0.80	0.50	0.96	1.41	0.74
Br	$<0.01^{+}$	0.30	$<0.01^{+}$	0.20	0.14	3.00
I	$<0.05^{\#}$	$<0.05^{\#}$	0.05	0.15	0.07	1.00
Hardness $CaCO_3$	73.00	42.50	25.00	20.00	16.20	366.50
TDS	721.10	690.00	751.80	791.84	837.70	4627.20

$^{+,\#}$ Water constituents for which the lower detection limits of the analytical apparatus are 0.01 and 0.05, respectively (here, when Al < 0.01 mg/L, the value 0.01 is used by default to achieve computing).

Institute of Hydrogeology and Environmental Geology at the Chinese Academy of Geology in Shijiazhuang city for analysis. The stable isotopes were determined using a mass spectrometer. ^{2}H and ^{18}O were determined by equilibration with CO_2 and U reduction, respectively. Analytical results are expressed as δ^2H (‰) and δ^{18}O (‰) from VSMOW (Vienna standard mean ocean water) with analytical precision of ±0.20‰ and ±1.5‰, respectively. The results are shown in Table 3.

3.2.1. Radioactive isotope measurements

Previous studies by Lin et al. (1999) and the results of the present determinations show that the concentration of ^{3}H in the GW in the study area is low: $T < 0.50$ TU (see Table 3); hence ^{3}H could not be reliably used to estimate the age of the regional thermal water. The ^{14}C technique was the only alternative radioactive isotope method available to estimate the relative age of the thermal water in the region, though controversies over correcting the ^{14}C and the applicability of corrected ^{14}C for groundwater age dating still remain. The use of ^{14}C measurements and the piston-flow model (Przewlocki and Yurtsever, 1974; Zuber, 1986) allow estimation of GW age, and were actually adopted for evaluating the degree of hydraulic connectivity between the 6 SRs.

3.2.2. Stable isotope measurements

To determine the origin of the GW in Kaifeng, the previous results of the ^{2}H and ^{18}O ratios from some 57 meteoric water samples collected 3 km south of Zhengzhou at latitude 34.72°N, longitude 113.63°E, from September 1985 to July 1992 (data from the IAEA station in Henan), were used in this study. Data from the isotope analyses was used to plot the meteoric water line of the SW zone of Zhengzhou (1), which manifests a very strong affinity with the meteoric water line of mainland China: $\delta D = 7.81\delta^{18}O + 8.16$. In the Kaifeng area, nine GW samples were analyzed for δD and δ^{18}O ratios; and then a correlation equation between these two ratios established (2).

$$\delta D = 8.01\delta^{18}O + 8.28 \quad (1)$$

$$\delta D = 6.41\delta^{18}O - 10.31 \quad (2)$$

3.3. Mineral–fluid chemical equilibrium state assessment

3.3.1. Assessment method

In conjunction with the use of stable and radioactive isotope methods to determine some characteristics of the geothermal system in Kaifeng, the mineral–fluid chemical equilibrium approach was used to further characterize the equilibrium state

Table 3
Results of isotope measurements in the study area

Sampling points	Type of sample	Well depth (m)	δD (‰)	$\delta^{18}O$ (‰)	T (TU)	^{14}C (%)	Age (a)
Zhengzhou	Rain	N/A	−38.90	−6.03	37.09 ± 3.05	Not applicable (N/A)	
Xinmi city	Rain	N/A	−54.64	−7.35	17.97 ± 3.11	N/A	
Weishi county	Rain	N/A	−96.08	−11.91	–	N/A	
Kaifeng city	Rain	N/A	−58.85	−7.99	15.47 ± 3.23	N/A	
Zhengzhou Jiangang reservoir	Surface water	N/A	−57.91	−7.74	19.75 ± 2.99	N/A	
155 hospital	Cold water	280	−72.78	−10.18	<0.50	N/A	
Fertilizer factory	Hot water	440	−75.52	−10.54	–	15.093 ± 0.57	15,630 ± 310
Water office (West)	Hot water	595	−72.22	−10.00	<0.50	12.102 ± 0.34	17,460 ± 230
Peanut group Co.	Hot water	660	−71.29	−9.37	<0.50	7.456 ± 0.30	21,460 ± 330
Power bureau #2	Hot water	803	−72.00	−9.59	–	7.755 ± 0.40	21,140 ± 430
Longshi Co.	Hot water	860	−71.00	−9.85	<0.50	10.154 ± 0.33[a]	18,910 ± 270[a]
Bianjing hotel	Hot water	1200	−75.27	−9.57	<0.50	6.162 ± 0.21	23,040 ± 280
Water office (East)	Hot water	1205	−73.14	−9.41	–	5.404 ± 0.28	24,120 ± 430
Water office deep well (West)	Hot water	1231	−78.26	−10.50	–	6.913 ± 0.22	22,090 ± 260
County's telecom	Hot water	1580	−68.80	−8.62	<0.50	4.876 ± 0.19	24,970 ± 330

[a] Subjective ^{14}C content and age for water sample from Longshi Co. well.

of minerals and water contained in the 6 SRs of the geothermal system in an effort to confirm the assumption established by lithologic data. Determination of chemical equilibrium between minerals and GW in a geothermal system (hot reservoir) can be accomplished by using a mineral–fluid chemical equilibrium approach (the saturation index: log (Q/K)) (Reed, 1982) as follows:

$$\log\ (Q/K)_j = \log \prod_{i=1}^{n} a_{i,j}^{V_{i,j}} - \log K_j \tag{3}$$

where, Q and K are the respective actual and theoretical activity of surrounding mineral j.

The actual activity Q_j formula is written as:

$$Q_j = \prod_{i=1}^{n} a_{i,j}^{V_{i,j}} \tag{4}$$

where, a_{ij} is the activity of the ith component of mineral j; V_{ij} is the reaction coefficient of the ith component of mineral j, and $\prod_{i=1}^{n}$ expresses the product of n numbers.

The saturation index is used to determine whether a chemical equilibrium exists between a mineral j and water at temperature T (°C). When log $(Q/K)_j$ is equal to zero, a chemical equilibrium exists between the mineral j and the water; when it is greater than zero, the water is oversaturated with respect to the mineral; when it is smaller than zero, the water is undersaturated with respect to the mineral.

Using data on analyzed water samples (Table 2), the saturation index of each mineral in the surrounding rock within the SRs (600–1600 m) was calculated at a given temperature using the computer program WATCH (Bjarnason, 1994). Curves for the corresponding relationship Log (Q/K) vs. T were plotted (Fig. 5) to estimate the mineral–water equilibrium state. SRs at 300–600 m depth were not considered because of their relatively low water temperature (25–38 °C) and lack of sufficient chemical data at that level.

3.4. Geothermometer temperature estimation

Up to the present, numerous geothermometers have been used to estimate geothermal reservoir temperatures. Each of them is applicable at different temperature ranges, pH and cation concentrations. The use of SiO_2, Na–K, and K–Mg geothermometers are among the most common in

Table 4
Concentration of chemical geothermometers in thermal waters

SR (m)	Sampling locations	Sampled water temperature (°C)	SiO_2 (mg/L)	Na (mg/L)	K (mg/L)	Mg (mg/L)
300–450	Beer factory	29.0	17.42	185.48	1.67	10.30
450–600	Power bureau #2	30.0	19.00	184.0	1.90	5.47
600–800	646 Plant	35.0	18.00	212.50	1.50	4.68
	Peanut group	38.0	18.00	210.00	2.50	2.43
800–1000	Power bureau #1	39.0	20.00	214.00	2.20	3.04
	Longshi Co.	40.5	20.00	222.50	1.80	2.43
1000–1300	Bianjing hotel	50.5	29.48	365.00	3.50	3.04
	Water office (East)	53.0	27.31	232.55	3.37	1.34
	Dongjing hotel	54.0	29.00	280.00	2.30	2.43
	County water supply	50.0	27.31	300.00	3.00	2.31
	155 Hospital	52.0	28.00	230.00	2.00	2.31
	Water bureau (East suburb)	53.5	27.54	231.00	2.44	2.41
1300–1600	County's telecom	65.0	42.00	1600	25.00	27.36

practice. The SiO_2 (quartz) geothermometer is generally applicable to geothermal systems at pH < 10, and temperature >150 °C with or without steam loss. The Na–K geothermometer is generally applicable to geothermal systems containing waters with pH > 7, and temperature >150 °C. However, Fournier (1977) and Arnorsson et al. (1983) have successfully established empirical temperature functions for a chalcedony geothermometer and Arnorsson et al. (1983) for a low-albite and microcline Na–K geothermometer in the ranges 25–180 °C and 25–250 °C, respectively. The K–Mg geothermometer is applicable to low temperature (<100 °C) geothermal systems with pH > 7, but equilibrium is attained very fast and it often approaches sampling temperature.

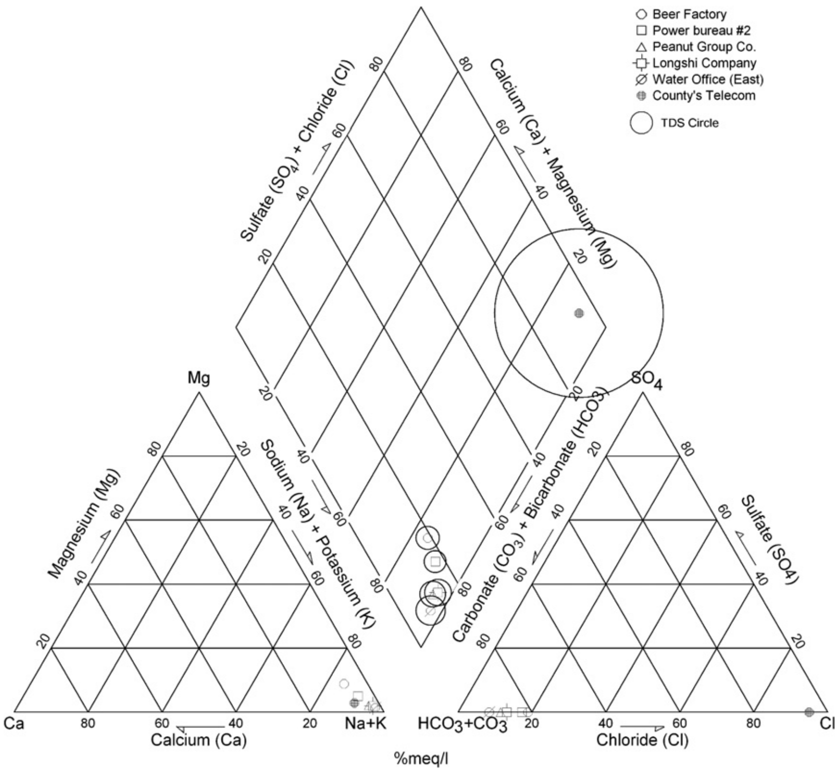

Fig. 3. Piper diagram showing the two main types of geothermal water in Kaifeng.

Based on the chemical analysis of 13 GW samples in Kaifeng (Table 4), the applicability of these three chemical geothermometers to this present study was simultaneously investigated through plotting of SiO_2–T, Na/K–T, and $K/(Mg)^{1/2}$–T scatter plots, which were respectively compared to the standard curve of each suggested approach (chalcedony geothermometer with no steam loss (Arnorsson et al., 1983; Fournier, 1977), Na–K geothermometer (Arnorsson et al., 1983), and K–Mg geothermometer (Giggenbach, 1986)). Finally, the chalcedony geothermometer suggested by Arnorsson et al. (1983) was adopted to estimate the temperatures of the hot water reservoirs in the system.

4. Results and discussion

According to data on water quality analysis (Table 2); the geothermal system in Kaifeng was chemically classified into two main groups: the 300–1300 m and 1300–1600 m range groups. The 300–1300 m depth range group (Nm + Ng) contains fresh and soft GW (with total dissolved solids - TDS < 1 g/L and hardness <100 mg/L) of Na–HCO_3 type; with pH between 8.1 and 8.4. The concentration of K^+ and Na^+ ions increase linearly with depth while those of Ca^{2+} and Mg^{2+} decrease linearly with depth. The concentration of HCO_3^- ion increases linearly with depth while those of Cl^- and SO_4^{2-} ions decrease slightly with increasing depth. These changes in the 300–1300 m group appear to be controlled mainly by geo-temperature. Samples collected from this group plot closely in the Piper diagram shown in Fig. 3.

The 1300–1600 m depth range group (Ng) contains a semi-mineralized to hard GW (with TDS of 4.66 g/L and a hardness of 366.5 mg/L) of Na–Cl type with a pH of 7.2. The major ion concentrations in this group are far greater than those in the first group, except for HCO_3^-, due to the rise of salinity with increasing depth, which indicates the maturation of the Na–Cl type of water in this

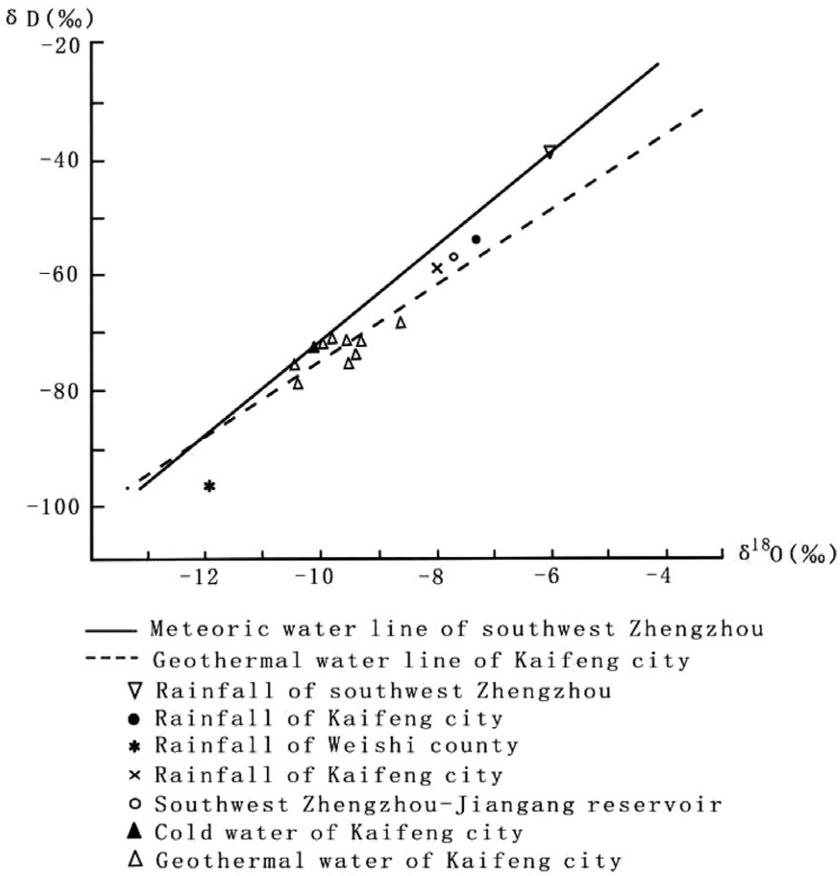

Fig. 4. Isotopic relationship of water resources in Kaifeng to the meteoric water in Zhengzhou.

formation. Higher concentration of trace elements, such as Li, Sr, Br and I are found on average in the Ng formations. From the Piper diagram shown in Fig. 3, the GW sample collected from this group (Count's telecom well) clearly appears distinctive compared to the samples from the first group in terms of salinity and water type. This finding allows the claim that this sample belongs to a confined aquifer not hydraulically connected to the previous group. The chemical characteristics of the water samples from these two groups are clearly different. In sum, no obvious hydraulic connection between the SRs within the system is clearly demonstrated; otherwise a mixture of the two types would have resulted.

The results of the analysis of the radioactive measurements in Table 3 shows that: (1) The age of GW increases with depth in closely distributed geothermal wells. For instance, geothermal wells at the Water office (west), Longshi company and the deep Water office well (west) respectively illustrate such a trend. The age differences suggest that different SRs may not have a direct hydraulic connection; otherwise water of the same age would have been noted. (2) It was noted that the ^{14}C content of GW decreases with depth while the GW age increases with depth with the exception of water sampled from the Longshi Company well. Such a phenomenon may arise from sampling and/or measurement errors. (3) In the same SR, the age of GW increases

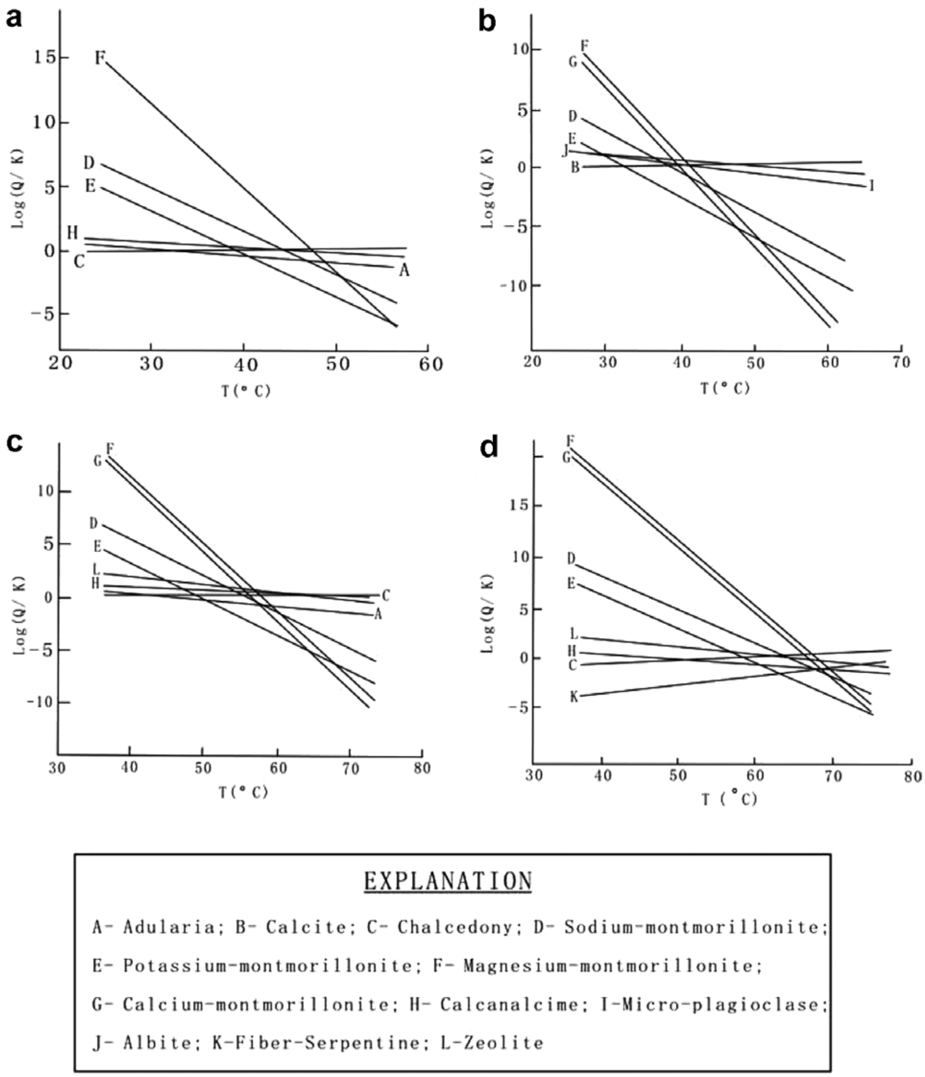

Fig. 5. log (Q/K)–T relationship curves in the 600–800 m SR (a), 800–1000 m SR (b), 1000–1300 m SR (c), and 1300–1600 m SR (d) showing chemical equilibria between minerals and water at different temperature intervals.

from west to east, e.g. in the 1000–1300 m SR, GW age in the deep-well at the Water office (west) and the well at Water office (east) increases from 22090 ± 260 a to 24120 ± 430 a. Within a 6 km horizontal stretch, the aquifer hydraulic conductivity was noted to be 8.1 mm/d (Lin et al., 1999) validating that GW indeed flows very slowly from the west towards the east. (4) With the average GW age in the study area being 20440 ± 320 a, which is representative of GW age in the North China basin; despite the controversies over groundwater age dating using corrected ^{14}C; and accounting for the water flow speed of 8.1 mm/d, thus the GW recharge zone for the 300–1300 m depth range group aquifer is estimated to be located beyond 60 km at the nearest. From Fig. 4, it can be seen that the GW sample from the County's telecom plots away from the others; moreover the ^{14}C content of that sample is smaller than that of the others, supporting the prediction of no obvious connection between the two groups of aquifers. The origin and residence time of the GW in the 1300–1600 m group need to be further investigated, because of the likely intense water-rock interaction at a greater depth in the system, where temperature is much higher.

Comparison of the local meteoric water line for Zhengzhou (1) with the correlation line (2) for δD and $\delta\ ^{18}O$ of the GW in Kaifeng (Fig. 4) shows mostly strong similarities. However, isolated points outside the local meteoric water line of Zhengzhou suggest some abnormalities. These points are located to the right of the local meteoric water line (1) and the line obtained by Eq. (2). They represent samples collected from the deep GW Zone (especially at 1200 m, 1205 m and 1231 m depths). This phenomenon may be the result of changes in ^{18}O ratios in the groundwater due to chemical reactions with the host rock even though the water temperature is <100 °C. Thus, accounting for the overall regional water flow direction and the results of the above studies, it can be stated that the GW utilized from the geothermal system in Kaifeng (with a mean value of $\delta D = -73.06‰$, $\delta^{18}O = -9.72‰$) is mainly derived from the confined static water of meteoric origin, originating in the mountainous region in the southwestern part of Zhengzhou (with a mean value of $\delta D = -38.90‰$, $\delta^{18}O = -6.03‰$), which is more than 60 km away.

The main formation minerals of the hot reservoir region are: Na-montmorillonite, K-montmorillonite,

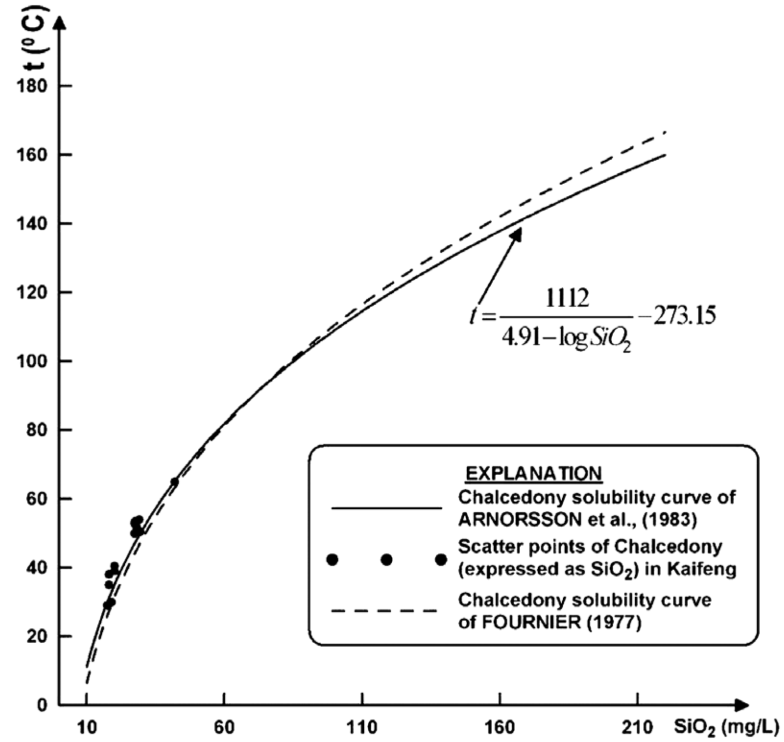

Fig. 6. Temperature dependence of chalcedony concentrations (expressed as SiO_2) in waters from 13 geothermal wells in Kaifeng.

Mg-montmorillonite, Ca-montmorillonite, chalcedony, adularia, calcite, calcanalcime, microplagioclase, albite, fiber-serpentine and zeolite. These minerals are, however, not all present in a significant amount in each SR. A study of the log (Q/K) vs. T curves for these minerals at depth intervals of 600–800 m, 800–1000 m, 1000–1300 m and 1300–1600 m reveals that minerals and GW reach chemical equilibria in the temperature intervals 29–45 °C, 33–47 °C, 45–60 °C, and 58–72 °C (Fig. 5a–d), respectively. The attainment of chemical equilibria between the minerals and GW in the present study is indeed consistent with Giggenbach (1981) and Arnorsson et al. (1983) who claimed that it is now generally accepted that mineral–water equilibrium has been attained in geothermal waters for all major components except mobile elements like Cl when temperatures are above 100 °C and in some cases at temperatures as low as 50 °C.

However, because Al values are only available for two samples out of six in this study, concerns over the attainment of equilibrium between Al – bearing minerals and GW can be raised. Fortunately for such a subjective Al data case, the study of Pang and Reed (1998) attempted to resolve the problem. They pioneered the FixAl approach to forcing water to equilibrium with Al-bearing minerals such as microcline, with very good results. Even though, the state of equilibrium is beyond the foremost objectives of this study, the Al data problem was not addressed in this paper.

From Fig. 6, it can be seen that the scatter plots of the chalcedony geothermometer fit best the temperature function curve of Arnorsson et al. (1983), and secondly that of Fournier (1977). The changes of temperature from downhole to surface are exhibited from the deepest well (County's telecom) with the highest measured temperature (65 °C) to the shallowest well (Beer factory) with the lowest measured temperature (29 °C). It can be said that the chalcedony geothermometer responds quite fast to temperature changes in Kaifeng. From the temperature function curve of Arnorsson et al. (1983), it can be estimated that the geothermometer temperatures in Kaifeng range from 29.93 to 65.17 °C within the depth range 450–1600 m. This temperature range is indeed consistent with the temperature range of waters sampled for this present study. The results obtained using the Na–K geothermometer suggested by Arnorsson et al. (1983) were less satisfactory. The majority of the data points for waters sampled show scatter around the Na/K standard

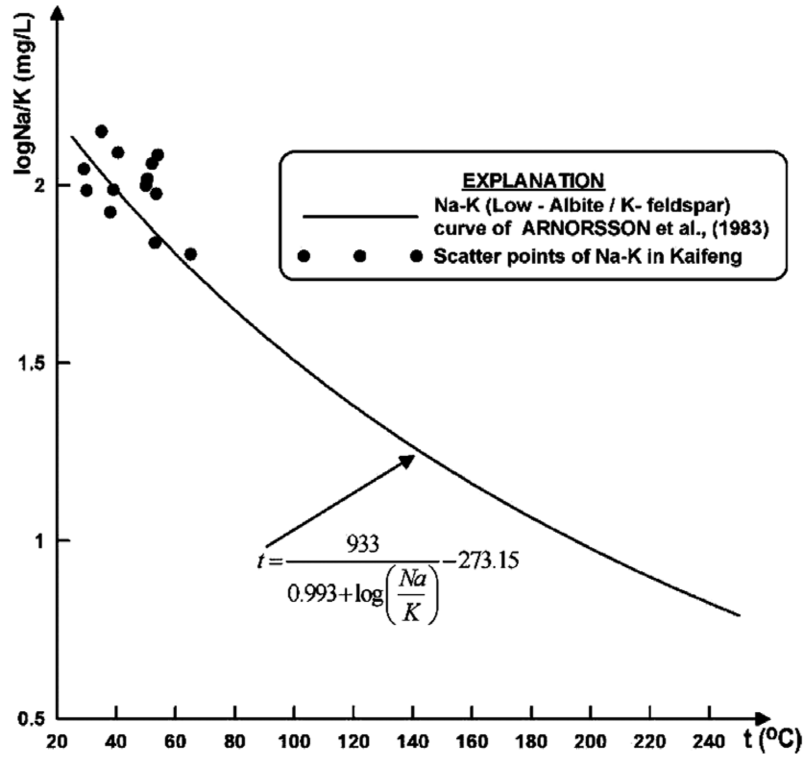

Fig. 7. Temperature dependence of Na/K concentrations in waters from 13 geothermal wells in Kaifeng.

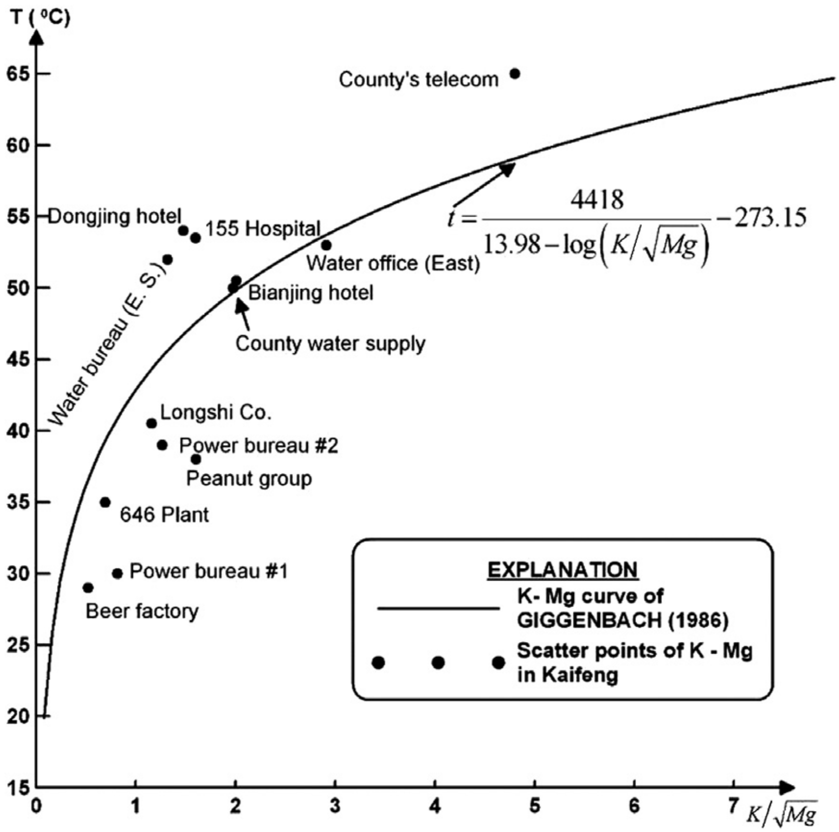

Fig. 8. Temperature dependence of $K/(Mg)^{1/2}$ concentrations in waters from 13 geothermal wells in Kaifeng.

curve (Fig. 7). The calculated geothermometer temperatures range for the above mentioned depth range is 33.45–59.70 °C. As for the results derived from the use of the K–Mg geothermometer of Giggenbach (1986), the K–Mg scatter points do not plot close to the standard curve (Fig. 8). The calculated geothermometer temperatures range from 36.67 to 59.03 °C. The Na–K and K–Mg geothermometer temperatures are not nearer to revealing the sampling temperature. Thus, it is assumed that the chalcedony geothermometer is the most appropriate for the investigation of the geothermal system in Kaifeng.

5. Conclusions

The results of the study are summarized as follows:

(1) According to the overall stratigraphy and the depth of the geothermal system in Kaifeng, along with the water chemical properties, isotope measurements and mineral–fluid chemical equilibrium computations, the GW recharge is not from precipitation in the Kaifeng area, but rather from meteoric water infiltration in the southwestern mountainous region of Zhengzhou, which is more than 60 km away.

(2) No obvious hydraulic connection exists between the two main hot reservoirs in the study area. The approaches used in this paper have validated the assumptions previously derived from the lithologic analysis by Lin et al. (1999).

(3) Starting from the 600–800 m SR and extending into the 1300–1600 m SR, the chalcedony geothermometer responds faster to temperature changes, thus it can be used to estimate the reservoir temperature in the study area.

(4) Finally, drawing on the results of different studies, the conceptual model of the regional heat storage was estimated to be a recharge/discharge type with no obvious upward vertical leakage. The horizontal runoff is guaranteed by lateral water flow in the main confined aquifers.

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