# Environmental isotopic study on the recharge and residence time of groundwater in the Heihe River Basin, northwestern China

**Zongyu Chen · Zhenlong Nie · Guanghui Zhang · Li Wan · Jianmei Shen**

**Abstract** The recharge and origin of groundwater and its residence time were studied using environmental isotopic measurements in samples from the Heihe River Basin, China. $\delta^{18}O$ and $\delta D$ values of both river water and groundwater were within the same ranges as those found in the alluvial fan zone, and lay slightly above the local meteoric water line ($\delta D=6.87\delta^{18}O+3.54$). This finding indicated that mountain rivers substantially and rapidly contribute to the water resources in the southern and northern sub-basins. $\delta^{18}O$ and $\delta D$ values of groundwater in the unconfined aquifers of these sub-basins were close to each other. There was evidence of enrichment of heavy isotopes in groundwater due to evaporation. The most pronounced increase in the $\delta^{18}O$ value occurred in agricultural areas, reflecting the admixture of irrigation return flow. Tritium results in groundwater samples from the unconfined aquifers gave evidence for ongoing recharge, with mean residence times of: less than 36 years in the alluvial fan zone; about 12–16 years in agricultural areas; and about 26 years in the Ejina oasis. In contrast, groundwater in the confined aquifers had $^{14}C$ ages between 0 and 10 ka BP.

**Résumé** La recharge et le temps de résidence de l'eau souterraine ont été étudiés par des techniques isotopiques dans le bassin de la Rivière Heihe, Chine. Les valeurs de $\delta^{18}O$ et $\delta D$ de la rivière et des eaux souterraines se retrouvent dans le même ordre de grandeur que ceux de la zone alluviale, et sont légèrement au dessus de la Droite Météorique Locale ($\delta D = 6.87\delta^{18}O + 3.54$). Ceci indique que les rivières s'écoulant des zones montagneuses contribuent substantiellement et rapidement aux ressources en eau des sous-bassins du Sud et du Nord. Les valeurs en $\delta^{18}O$ et $\delta D$ des eaux souterraines dans la partie captive des aquifères de ces sous-bassins, sont proches les unes des autres. Un enrichissement des isotopes lourds par évaporation semble évident. L'augmentation en $\delta^{18}O$ la plus prononcée est observée au droit des zones agricoles, reflétant l'influence de l'irrigation. Le tritium dans les parties libres des aquifères, apporte la preuve d'une recharge récente, et un temps de résidence de 36 ans dans la zone alluviale, 12 à 16 dans la zone agricole, et 26 ans dans l'oasis d'Ejina. Le contraste avec les eaux des parties captives de l'aquifère est marqué par un âge $^{14}C$ compris entre 0 et 10 ka BP.

**Resumen** Se ha estudiado la recarga y origen del agua subterránea y su tiempo de residencia usando mediciones de isótopos ambientales en muestras de la Cuenca del Río Heihe, China. Los valores de $\delta^{18}O$ y $\delta D$ en muestras de agua de río y agua subterránea se encuentran dentro de los mismos rangos que los encontrados en la zona de abanico aluvial y ligeramente por encima de la Línea de Agua Meteórica Local ($\delta D = 6.87\delta^{18}O + 3.54$). Este resultado indica que los ríos de montaña contribuyen de manera considerable y rápida a los recursos hídricos en las subcuencas del norte y del sur. Los valores de $\delta^{18}O$ y $\delta D$ en agua subterránea en los acuíferos no confinados de estas subcuencas son bastante parecidos entre sí. Existe evidencia de enriquecimiento de isótopos pesados en agua subterránea ocasionado por evaporación. El incremento más pronunciado en el valor de $\delta^{18}O$ se presentó en áreas agrícolas lo cual refleja la mezcla adicional de flujo de retorno de riego. Los resultados de tritio en muestras de agua subterránea de los acuíferos no confinados proporcionaron evidencia de la recarga actual con tiempos de residencia promedio: menos de 36 años en la zona del abanico aluvial; cerca de 12–16 años en áreas agrícolas; y cerca de 26 años en el oasis Ejina. En contraste, el agua subterránea en los acuíferos confinados tuvo edades $^{14}C$ que variaron de 0 a 10 ka antes del presente.

**Keywords** Arid regions · Stable isotopes · Radioactive isotopes · Groundwater recharge · Groundwater age

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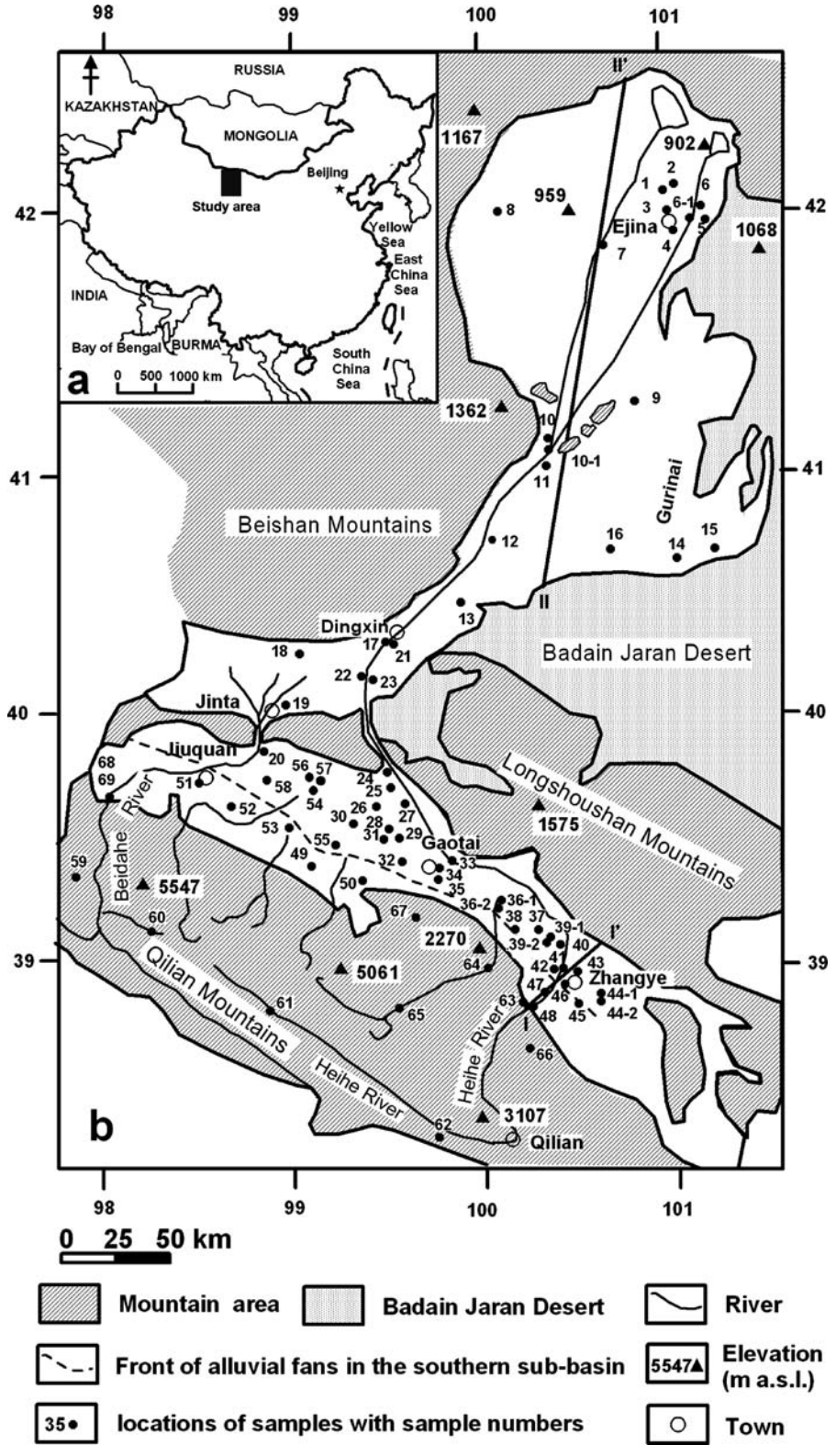


**Fig. 1** **a** Study area in northwestern China and **b** geographic features of the Heihe River Basin, locations of samples with sample numbers, and location of cross-sections I I′ and II II′ (shown in Fig. 2)

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## Introduction

The Heihe River runs in the second largest inland arid catchment of China. It is an eco-environmentally fragile area and scarce in water resources. Sustainable water management is vital for maintaining the ecologic system in spite of an increasing demand for water supply. The corresponding development and management of the groundwater resources in this area require detailed and reliable information on the origin and the natural recharge rate of groundwater. Two fundamental questions are related to the management of the groundwater resources:

1. What is the source of recharge? (Or where does recharge occur?)
2. How old is this groundwater? (Is there current replenishment?)

In the Heihe River Basin, Fan (1991) and Chen (1997) suggested that groundwater was recharged by rivers infiltration in the piedmont of the Qilian Mountains. However, the recharge process is not well understood yet and the groundwater residence time is unknown due to the hydrogeological complexity. Naturally occurring stable ($^{2}H$, $^{18}O$) and radiogenic ($^{3}H$, $^{14}C$) isotopes in water have been widely used over the last 50 years to address problems related to the recharge and the residence time of groundwater (e.g. Fontes 1980; Gonfiantini 1986; Clark and Fritz 1997; Nativ and Riggio 1989; Wood and Sanford 1995). The purpose of this study is to identify sources of recharge, to localize recharge areas and to determine the groundwater residence time by employing isotopic measurements.

## Study area

The arid Heihe River Basin is located at 98°00′–101°30′ E, 38°00′–42°30′ N and has a drainage area of 130,000 km$^2$ (Fig. 1). It consists of three parts, namely the upper mountainous area (the source of the Heihe River), the middle oasis area (incorporating towns such as Zhangye and Jiuquan), and the lower terminal arid area around Ejina. The elevations of these three reaches are 5,000–2,000 m above sea level (asl), 1,700–1,300 m asl, and 1,450–910 m asl, respectively.

The upper reach with an elevation of 2,000–5,000 m asl is the headwater area in the Qilian Mountains with a mean annual temperature of −3 to 4°C. Glaciers covering 421 km$^2$ are distributed above elevation 4,500 m asl (Chen 2002). At elevations above 4,000 m asl, vegetation is very sparse and dominated by cushion plants. Meadows

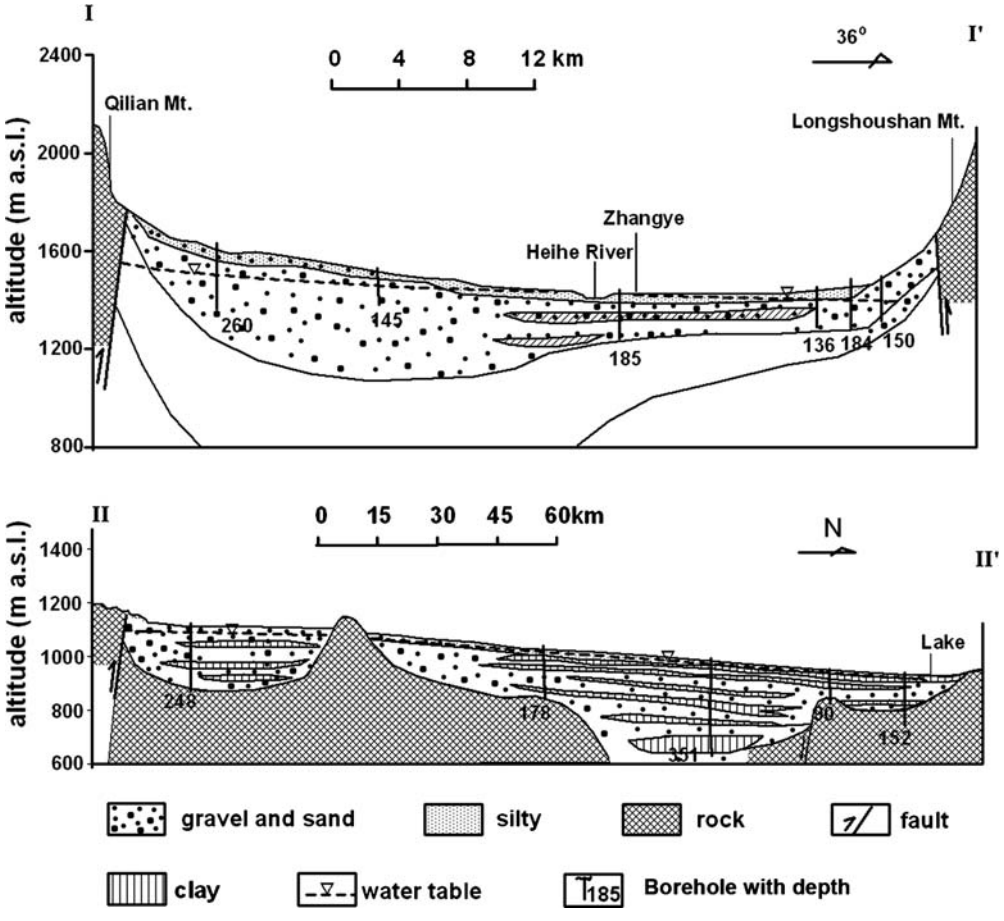


**Fig. 2** Hydrogeologic cross-sections along transect I-I′ and II-II′ in Fig. 1


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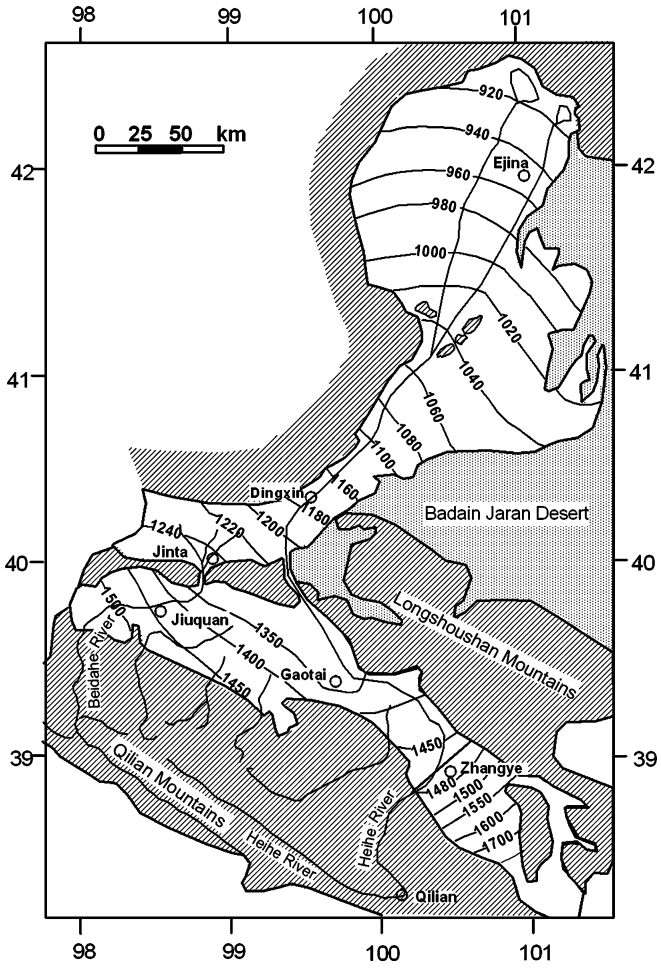


**Fig. 3** Contour map of shallow groundwater level (in meters above sea level). See Fig. 1 for key

and shrubs occur below 3,300 m asl. The mean annual precipitation ranges from 200 to 500 mm and may reach 700 mm in the peak areas (Wang and Cheng 1999). At elevations above 4,000 m asl, precipitation is dominated by snow. The maximum precipitation occurs between 2,400–3,000 m asl.

The middle and lower reaches of the Heihe River Basin are separated into the southern and northern sub-basins by the Longshoushan Mountains whose elevation ranges from 500 m asl to 2,500 m asl. The southern sub-basin, with an elevation of 1,300–1,700 m asl, is an alluvial plain which consists of alluvial fans and floodplain downgradient of the distal fan-front. The cultivated oasis in the middle reaches, dominated by irrigated farmland, is mainly located in the floodplain. The mean annual temperature in this sub-basin is about 3–7°C. The mean annual precipitation ranges from 50 mm to 150 mm and the mean potential evaporation rate is about 2,000–2,200 mm/year (Gao 1991). The relatively few rain events are not enough to generate runoff. The northern sub-basin consists of the Jinta-Dingxin area and Ejina area. Except for the Jinta-Dingxin agricultural area, the riverbank area and the Ejina oasis, most areas in this sub-basin are defined as Gobi Desert. The mean annual temperature is about 8°C. The mean annual precipitation is only 42 mm and the mean potential evaporation rate is 2,300–3,700 mm/year (Wang and Cheng 1998).

Seventeen rivers with a total runoff of $3.44\times10^9$ $m^3$/year, originate from the Qilian Mountains and flow towards the basin. The Heihe River is the largest perennial river. It flows through the whole basin into the terminal lakes in the Ejina oases with a total length of 812 km. Most tributaries are ephemeral rivers. They disappear in the piedmont after approaching the mouths of the mountain valleys. The Beidahe River, in the west part of the catchment, is the second largest river which also originates from the Qilian Mountains and flows through Jiuquan into the Jinta water reservoir.

Surface runoff from the upper mountain area by rain, melt water of snow and glaciers is the only source of water available in the basin. The increase of water demand in the middle oases area, mainly for irrigation of agricultural land, has resulted in the decrease of surface water supply for the lower arid area. Even in the middle oasis area (including Jinta-Dingxin area), over 80% of the total discharge has been diverted from main river courses to many irrigation canals, resulting in not only the rise of groundwater level inside the cultivated oases, but also the increase in area affected by soil salinization (Gao 1991).

Groundwater development is mainly concentrated in the southern sub-basin for agricultural and industrial purposes. A small amount of groundwater is used in the Dingxin-Jinta agricultural area of the northern sub-basin, because surface water is the main source for irrigation in this area. The hydrogeological setting of basins was studied by Chen (1997) and Fan (1991). There are two separate groundwater systems in the whole basin, the southern sub-basin system and the northern sub-basin system. Each system has independent hydrological conditions.

In the southern sub-basin, the regional Quaternary aquifers consist of alluvial fans in the piedmont plain and fluvial deposits in the lower part of the sub-basin. The aquifer system includes a thick unconfined zone consisting of coarse-grained gravel and sand with a thickness up to 1,000 m in the piedmont plain, and a confined part consisting of medium to fine and silty sand with a thickness of 50–200 m in the floodplain (Fig. 2). A thrust along the foot of the Qilian Mountains in a NW–SE direction, produces a steep hydraulic gradient across the fault. It is difficult for mountain groundwater to flow into the basin by lateral flow and therefore it moves to the surface as springs at the foot of the mountain (Fan 1991; Chen 1997). The rivers originating in the Qilian Mountains are the main recharge source for aquifers. Groundwater in this sub-basin flows generally from the piedmont area towards the centre of the basin (Fig. 3). Because the Longshoushan Mountains acts as a barrier for groundwater flow, groundwater discharges at the low part of this sub-basin by upward leakage and springs. The depths to the water table range from 50–200 m in the upper

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alluvial fan to 3–5 m in the lower part of the floodplain. The piezometric surface of the confined aquifer is 2–3 m higher than the water table of unconfined aquifer. Groundwater in the unconfined aquifer is fresh (total dissolved solids (TDS) <1,000 mg/L) and of Ca–$HCO_3$ type in the piedmont plain. Groundwater is brackish (TDS=1,000–5,000 mg/L) and of Ca–Mg–$SO_4$–$HCO_3$ or Na–Mg–$SO_4$–Cl type in the floodplain. The groundwater in the confined aquifer is fresh (TDS<1,000 mg/L) and of Ca–Mg–$HCO_3$–$SO_4$ type.

In the northern sub-basin, the regional Quaternary aquifers consist of eolian and lacustrine deposits. The aquifer system changes from a small unconfined gravel-pebble zone to a multi-layered zone which consists of sand and silt (Fig. 2). The Heihe River is the main recharge source for aquifers. The depth to the water table is 10–30 m in the south part of this sub-basin and gradually changes to 3 m below the surface in the Ejina oasis. Groundwater flows from the southwest to the northeast (Fig. 3) and discharges by evaporation. Shallow groundwater is of Ca–Mg–$HCO_3$–$SO_4$ or Ca–Mg–$SO_4$ type with a TDS of 1,000–3,000 mg/L. The confined aquifer is deeper than 70 m, and consists of sandy gravel separated by the discontinuous clay-dominant layers, and is distributed more towards the northeast of Ejina. Artesian wells are found occasionally in the north part of the basin. From south to north, the depth to the groundwater level for the confined aquifer ranges from 15 to 1 m. Total dissolved solids in this water is less than 1,000 mg/L. The groundwater is of Ca–Mg–$HCO_3$–$SO_4$ type.

## Materials and methods

Samples of water were collected for isotope analysis between May and August 2001 along the flow path from the headwater area to the discharge area (Fig. 1). Most samples were analyzed for tritium, oxygen-18, deuterium, carbon-14 and carbon-13. All groundwater samples were collected directly from the wellhead. Prior to sampling, the wells were pumped until at least three well volumes were drawn away. Samples for $\delta^{18}O$, $\delta D$ and tritium were collected in 1-L glass bottles with gas-tight caps. The samples were prepared for dissolved inorganic carbon and subsequent carbon isotope determination, by adding excess $BaCl_2$ to 120 L of water previously brought to pH ≥12 by addition of $CO_2$-free NaOH and collecting the precipitated $BaCO_3$.

Temperature, pH and electric conductivity (EC) were measured in the field with a WTW Multi 340i handheld meter which was calibrated before use. Alkalinity was measured by Gran titration method (Gran 1952). Isotopic analyses were done by the Central Laboratory of Hydrogeology, Ministry of Land and Resources, China.

Stable isotope ratios of waters were determined using the $CO_2$ equilibration method (Epstein and Mayeda 1953) and the zinc-reduction method for D/H ratios (Coleman et al. 1982), followed by analysis with a Finnigan MAT 251 mass spectrometer. The isotope compositions were reported in standard δ-notation representing per mil deviations from the V-SMOW standard (Vienna Standard Mean Ocean Water). Precisions for $\delta D$ and $\delta^{18}O$ are ±1.0 and ±0.2‰, respectively.

Tritium was determined on electrolytically enriched water samples by low-level proportional counting. The results were reported as tritium units with a typical error of ±1 TU (Eichinger 1980). The detection limit for tritium measurement in the laboratory is about 2–3 TU. The $^{14}C$ of dissolved inorganic carbon (DIC) was determined radiometrically by liquid scintillation counting after conversion to benzene. And the specific $^{14}C$ activity was reported in the unit of percent modern carbon (pMC). The detection limit was 0.7–1.0 pMC. The $\delta^{13}C$ analyses were mass spectrometrically determined and reported as δ-values related to the V-PDB (Vienna Pee Dee Belemnite) standard. The precision of $\delta^{13}C$ was ±0.5‰.

The stable isotopic compositions of groundwater were used to identify the sources and area of recharge. Tritium activity was measured to distinguish pre-bomb (pre-1952) water from younger water and estimate the mean residence time (MRT). The isotopes ($^{13}C$ and $^{14}C$) of carbon in DIC were measured to determine the groundwater age (Clark and Fritz 1997). The analytical results of samples are presented in Table 1.

## Results

### Isotopes in precipitation

Isotopes in precipitation have been measured for samples collected in Zhangye (38.93° N 100.43° E, altitude 1483 m asl) by the International Atomic Energy Agency (IAEA) since 1986. The weighted mean values of $\delta^{18}O$ and $\delta D$ were −6.0 and −42‰, respectively —Global Network of Isotopes in Precipitation Database (2004). The local meteoric water line (LMWL), $\delta D=6.87\delta^{18}O+3.54$, was obtained by the least squares fit (LSF) method. For 2001, the weighted mean tritium value was about 43 TU. On the basis of the data from seven representative Chinese GNIP (Global Network of Isotopes in Precipitation) stations during 1986–1998, Liu (2001) discussed the distribution of tritium values for precipitation in China and showed that Zhangye was the centre of a high tritium value area. He found that the tritium values in Zhangye precipitation were about 4 times higher than the values at the coastal station, Haikou (20.02° N 110.21° E, altitude 15 m asl), in southern China. He suggested that the reason for the high tritium value of Zhangye precipitation was due to nuclear tests and the latitudinal effect (concentration of tritium activity increasing with geographic latitude, going from the equator to the poles). Comparing tritium values of Zhangye precipitation with that of other Chinese GNIP stations, it appeared to be possible that this high tritium level was the result of the nuclear tests carried out in the desert region near the study area before the last above-ground nuclear blast in 1980.

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**Table 1** Analytical results for water samples in the Heihe River Basin

| Sample No. | Well name | Altitude (m asl) | Well depth (m) | Temp. (°C) | EC (μS/cm) | pH | $\delta^{18}O$ (‰) | δD (‰) | d-excess (‰) | $^{3}H$ (TU) | $\delta^{13}C$ (‰) | $^{14}C$ (pMC) |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| Runoff | | | | | | | | | | | | |
| 6–1 | | 916 | | | | | −4.9 | −50 | −11 | 28 | | |
| 10–1 | | 1,056 | | | | | −5.0 | −53 | −13 | 42 | | |
| 17 | | 1,176 | | 25.0 | 2,319 | 9.4 | −1.7 | −32 | −18 | 28 | | |
| 20 | | 1,315 | | 26.1 | 975 | 8.1 | −7.4 | −51 | 8 | 40 | | |
| 23 | | 1,240 | | | 2,437 | | 0.8 | −49 | −55 | 45 | | |
| 24 | | 1,270 | | | 1,721 | 8.8 | −6.8 | −53 | 1 | 25 | | |
| 33 | | 1,320 | | | 1,556 | | −7.2 | −56 | 2 | 23 | | |
| 41 | | 1,458 | | 22.5 | | 7.8 | −8.0 | −56 | 8 | 53 | | |
| 59 | | 3,285 | | 7.7 | 367 | 7.9 | −9.2 | −51 | 23 | 49 | | |
| 60 | | 3,855 | | 7.2 | 386 | 8.1 | −8.6 | −45 | 24 | 43 | | |
| 61 | | 3,796 | | 14.9 | 4,217 | 8.0 | −7.4 | −49 | 10 | 40 | | |
| 62 | | 3,027 | | 12.9 | 411 | 8.0 | −7.7 | −45 | 17 | 63 | | |
| 63 | | 1,659 | | 14.3 | 468 | 8.0 | −8.1 | −50 | 15 | 58 | | |
| 64 | | 1,700 | | 13.9 | 499 | 7.8 | −8.3 | −54 | 12 | 41 | | |
| 65 | | 1,900 | | 14.2 | 417 | 8.0 | −8.5 | −54 | 14 | 31 | | |
| 68 | | | | | | | −12.2 | −80 | 18 | 27 | | |
| 69 | | | | | | | −14.1 | −87 | 26 | 12 | | |
| Springs | | | | | | | | | | | | |
| 66 | | 2,267 | | 15.2 | | 7.0 | −10.0 | −56 | 24 | 12 | | |
| 67 | | 1,908 | | 20.0 | | 7.8 | −9.6 | −62 | 15 | 2 | | |
| Unconfined aquifer | | | | | | | | | | | | |
| 1 | HX1 | 920 | 9 | 11.7 | 2,347 | 7.6 | −6.9 | −46 | 9 | 33 | | |
| 3 | HX3 | 925 | 10 | 11.4 | 2,424 | 7.4 | −5.9 | −47 | 0 | 36 | | |
| 5 | HX5 | 923 | 8 | 12.2 | 1,363 | 7.9 | −6.2 | −48 | 2 | 32 | | |
| 10 | HX10 | 1,056 | 40 | 14.8 | 2,002 | 7.8 | −7.6 | −51 | 10 | 31 | | |
| 11 | HX11 | 1,063 | 90 | 9.0 | 1,154 | 7.7 | | | | 27 | | |
| 12 | HX12 | 1,103 | 50 | 13.5 | 1,164 | 7.6 | | | | 36 | | |
| 13 | HX13 | 1,144 | 60 | 15.5 | 1,376 | 7.9 | | | | 21 | | |
| 14 | HX14 | 1,050 | 10 | 13.0 | 2,813 | 7.7 | −1.7 | −45 | −31 | 41 | | |
| 15 | HX15 | 1,020 | 10 | 13.0 | 2,6583 | 8.0 | −3.5 | −56 | −28 | 25 | | |
| 16 | HX16 | 1,100 | 30 | 15.5 | 2,502 | 8.1 | −1.9 | −41 | −26 | 19 | −6.1 | 63.0 |
| 19 | HX19 | 1,217 | 60 | 11.8 | 944 | 7.4 | −8.4 | −54 | 13 | 39 | | |
| 21 | HX22 | 1,176 | 36 | 10.5 | 2,521 | 7.3 | −5.9 | −52 | −5 | 30 | | |
| 26 | ZY22 | 1,352 | 90 | | 646 | 7.2 | −9.7 | −61 | 17 | 3 | | |
| 27 | HX31 | 1,310 | 12 | 14.5 | 1,656 | 7.9 | −8.0 | −53 | 11 | 9 | | |
| 34 | HX26 | 1,352 | 11 | 13.9 | 982 | 7.6 | −7.8 | −52 | 10 | 21 | −9.4 | 87.8 |
| 36–2 | HX42–2 | 1,388 | 81 | 11.3 | 1,999 | 8.0 | −8.6 | −56 | 13 | 78 | | |
| 37 | HX47–1 | 1,453 | 51 | 11.3 | 769 | 7.8 | −8.0 | −54 | 10 | 52 | −9.9 | 94.3 |
| 38 | HX41–1 | 1,456 | 14 | 12.0 | 6,781 | 7.7 | −8.0 | −52 | 12 | 94 | −8.4 | 78.3 |
| 39–2 | HX46–2 | 1,441 | 89 | 11.7 | 518 | 7.8 | −8.1 | −53 | 12 | 164 | | |
| 40 | ZY5 | 1,441 | 60 | | 1,083 | 7.4 | −7.7 | −49 | 13 | 136 | | |
| 42 | HX45 | 1,458 | 6 | 11.8 | 464 | 7.6 | −7.9 | −52 | 11 | 52 | | |
| 43 | HX37 | 1,458 | 9 | 11.5 | 950 | 7.7 | −8.7 | −57 | 13 | 16 | −8.5 | 72.1 |
| 44–2 | HX36–2 | 1,516 | 70 | 10.1 | 1,516 | 7.4 | −8.3 | −54 | 12 | 46 | −10.9 | 112.0 |
| 46 | ZY8 | 1,618 | 60 | | 709 | 7.4 | −8.7 | −56 | 14 | 38 | | |
| 47 | HX59 | 1,538 | 100 | 10.1 | 806 | 7.4 | −7.9 | −53 | 10 | 59 | −8.9 | 111.2 |
| 48 | HX29 | 1,708 | 280 | 11.3 | 487 | 7.7 | −8.3 | −54 | 12 | 66 | −4.4 | 90.5 |
| 49 | HX49 | 1,553 | 124 | 13.0 | 1481 | 8.0 | −9.5 | −56 | 20 | 25 | −5.4 | 81.3 |
| 50 | HX48 | 1,753 | 280 | 14.3 | 439 | 7.8 | −9.4 | −55 | 20 | 27 | | |

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| | | | | | | | | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| 51 | ZY21 | 1,314 | 80 | 13.8 | 411 | 7.9 | −8.6 | −55 | 14 | 33 | −2.0 | 93.4 |
| 52 | HX50 | 1,464 | 270 | 11.3 | 985 | 7.9 | −9.0 | −56 | 16 | 105 | −9.4 | 110.0 |
| 53 | HX76 | 1,390 | 100 | 11.0 | 1,366 | 7.6 | −9.1 | −57 | 16 | 54 | | |
| 56 | HX75 | 1,386 | 20 | 14.1 | 706 | 7.5 | −9.7 | −58 | 20 | 2 | | |
| Confined aquifer | | | | | | | | | | | | |
| 2 | HX2 | 922 | 90 | 13.1 | 1,487 | 8.0 | −6.1 | −53 | −4 | 3 | −5.5 | 18.3 |
| 4 | HX4 | 939 | 140 | 14.6 | 1,294 | 7.9 | −7.2 | −54 | 4 | 4 | −5.9 | 26.6 |
| 6 | HX6 | 916 | 100 | 13.1 | 2,253 | 7.7 | −6.3 | −57 | −7 | 8 | −7.0 | 38.7 |
| 7 | HX7 | 924 | 130 | 11.0 | 1,275 | 7.8 | −10.3 | −66 | 16 | 8 | −4.7 | 27.4 |
| 8 | HX8 | 939 | 102 | 13.8 | 1,556 | 8.1 | −7.5 | −55 | 5 | 5 | −5.7 | 22.5 |
| 9 | HX9 | 1,002 | 70 | 14.0 | 656 | 7.6 | −8.5 | −57 | 11 | 2 | −5.2 | 48.6 |
| 18 | HX18 | 1,223 | 70 | 11.8 | 1,100 | 7.7 | −10.0 | −57 | 23 | 10 | −4.8 | 55.7 |
| 22 | HX23 | 1,240 | 70 | 10.5 | 2,135 | 7.3 | −8.3 | −48 | 18 | 8 | | |
| 25 | ZY23 | 1,318 | 130 | 14.5 | | 7.0 | −11.1 | −72 | 17 | 3 | −7.7 | 36.8 |
| 28 | ZY25 | 1,378 | 168 | 14.0 | | 8.4 | −10.0 | −65 | 15 | 3 | −4.4 | 18.4 |
| 29 | ZY26 | 1,378 | 94 | 13.0 | | 8.0 | −9.8 | −63 | 15 | 3 | −5.0 | 25.6 |
| 30 | ZY18 | 1,418 | 150 | 13.0 | 552 | 8.5 | −9.7 | −61 | 17 | 3 | −3.8 | 34.9 |
| 31 | HX79 | 1,431 | 150 | 13.0 | 433 | 8.4 | −9.6 | −62 | 15 | 5 | −4.7 | 26.3 |
| 32 | HX80 | 1,431 | 150 | 13.0 | 395 | 8.0 | −9.6 | −60 | 17 | 10 | | |
| 35 | HX27 | 1,421 | 200 | 16.0 | 919 | 7.8 | −9.1 | −51 | 22 | 2 | −4.4 | 41.4 |
| 36–1 | HX42–1 | 1,388 | 157 | 11.3 | 278 | 8.2 | −9.6 | −59 | 2 | 76 | | |
| 39–1 | HX46–1 | 1,441 | 151 | 11.9 | 222 | 8.1 | −9.8 | −56 | 22 | 44 | | |
| 44–1 | HX36–1 | 1,516 | 130 | 10.6 | 643 | 7.6 | −9.3 | −55 | 19 | 22 | | |
| 45 | HX35–1 | 1,483 | 185 | 11.7 | 256 | 9.4 | −10.0 | −58 | 22 | 2 | −6.7 | 67.8 |
| 54 | ZY16 | 1,702 | 300 | 11.0 | 386 | 7.8 | −9.6 | −61 | 16 | 3 | −5.2 | 13.6 |
| 55 | ZY17 | 1,700 | 293 | | | | −9.2 | −63 | 11 | 3 | | |
| 57 | HX77 | 1,387 | 120 | 16.3 | 483 | 8.4 | −9.9 | −63 | 16 | 2 | −5.5 | 39.2 |
| 58 | HX78 | 1,381 | 100 | 12.4 | 1,478 | 8.1 | −10.0 | −62 | 18 | 3 | −5.5 | 48.7 |

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**Table 2** Summary of isotopic compositions in basins

| Area | *n* | Mean $\delta^{18}O$ (‰) | Mean δD (‰) | Mean d-excess (‰) | Samples for statistics |
|---|---|---|---|---|---|
| Surface water in mountain area | 7 | −8.3±0.2 | −50±1 | 16±2 | 59, 60, 61, 62, 63, 64, 65 |
| Surface water in basins | 8 | −4.4±1.1 | −50±3 | −9.8±7.3 | 6–1, 10–1, 17, 23, 20, 24, 33, 41 |
| Alluvial fan in southern sub-basin | 7 | −8.8±0.2 | −55±1 | 15±2 | 47, 48, 49, 50, 51, 52, 53 |
| Unconfined aquifer in southern sub-basin | 13 | −8.4±0.2 | −54±1 | 13±1 | 26, 27, 34, 36–2, 37, 38, 39–2, 40, 42, 43, 44–2, 46, 56 |
| Confined aquifer in southern sub-basin | 15 | −9.8±0.1 | −61±1 | 17±1 | 25, 28, 29, 30, 31, 32, 35, 36–1, 39–1, 44–1, 45, 54, 55, 57, 58 |
| Unconfined aquifer in northern sub-basin | 12 | −5.3±0.8 | −49±1.6 | −6±5 | 1, 3, 5, 10, 14, 15, 16, 21 |
| Confined aquifer in northern sub-basin | 8 | −8.0±0.6 | −56±2 | 8±4 | 2, 4, 6, 7, 8, 9, 18, 22 |

## *Isotopes in surface water*

For the stable isotope composition of the Heihe River, Zhang et al. (2005) found that the mean $\delta^{18}O$ values increase along the river course, based on the monthly samples collected at six gauge stations (S1, S2, S3, S4, S5 and S6) from May 2001 to May 2002. The locations of S1, S2, S3, S4, S5, and S6 were at the same locations as those of surface water Nos. 63, 41, 33, 24, 17 and 6-1, respectively (Fig. 1). The mean $\delta^{18}O$ value at S1, which was located at the outlet of the mountain valleys, was −8.2±0.3‰ (Zhang et al. 2005). Seven samples of the current study (sample Nos. 59, 60, 61, 62, 63, 64 and 65) were collected from rivers at an altitude of 2,000–4,000 m asl in the headwater area in May 2001. These samples had tritium values of 31–63 TU which were similar to those of recent rainwater in Zhangye. The $\delta^{18}O$ and δD values for these samples ranged from −9.2 to −7.4‰ and −54 to −45‰, respectively (Tables 1 and 2) and with the mean

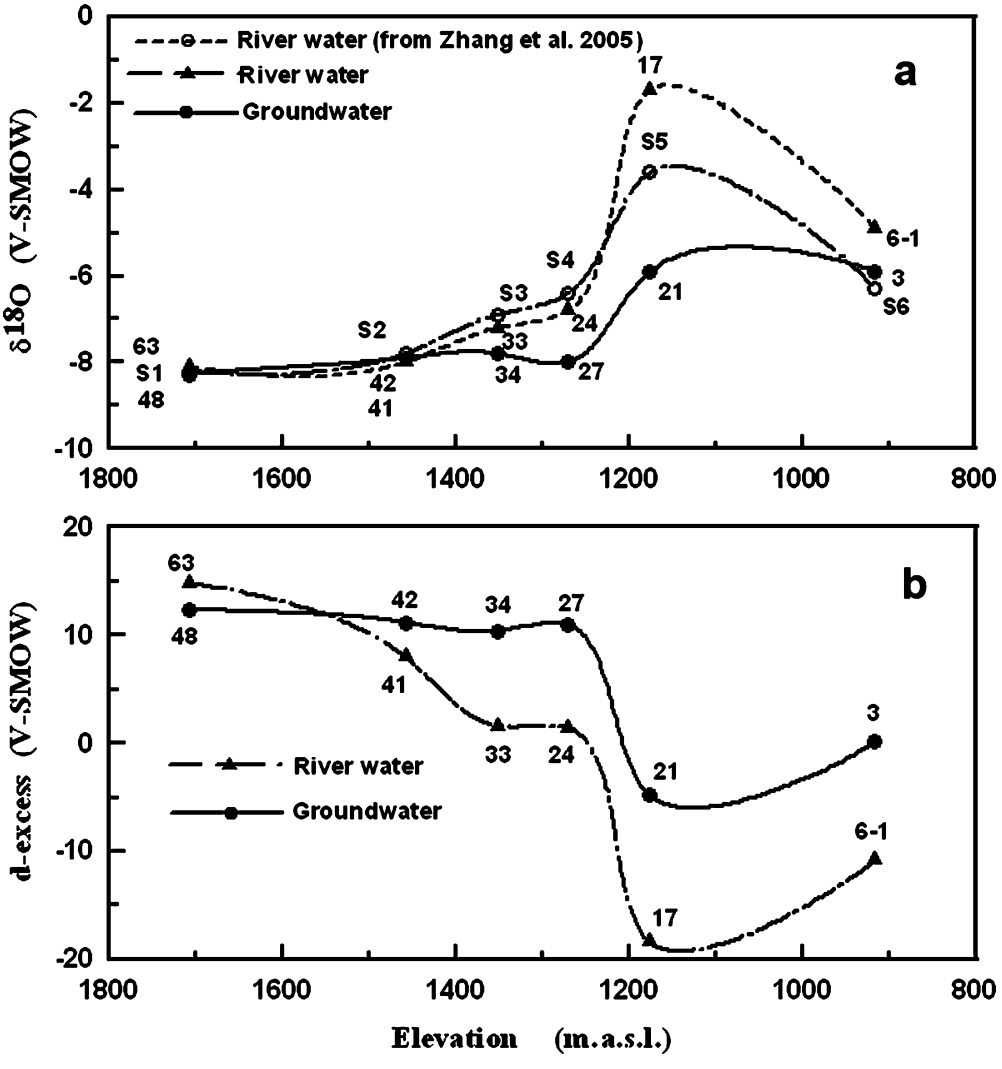


**Fig. 4** **a** $\delta^{18}O$ and **b** d-excess values at surface water and groundwater sites along the Heihe River course

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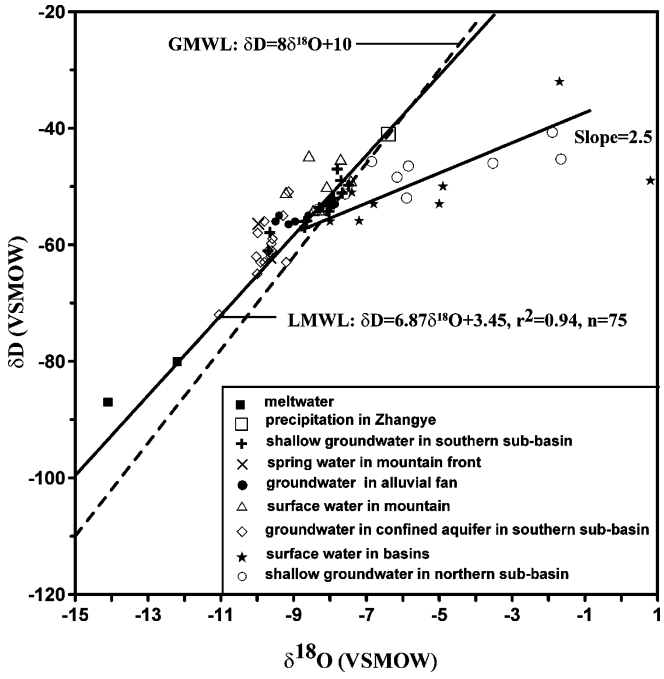


**Fig. 5** Plot of δD vs. $\delta^{18}O$ for waters in the Heihe River Basin

$\delta^{18}O$ value of 8.3±0.3‰ ($n$=7) which was close to the mean value at S1. This mean value was more negative than the mean value of precipitation in Zhangye. The recharge altitude of this water was calculated as 2,300–4,300 m asl according to the mean altitude effect of about −0.25‰/100 m. It is the representative for the precipitation in the Qilian Mountain area. Two samples (sample Nos. 68 and 69) thought to be melt water from mountains were collected in the Beidahe River during March and April, 2002. They have tritium values of 12 TU and 27 TU (Table 1). Their $\delta^{18}O$ values are −12.2 and −14.1‰ respectively; their δD values are −87 and −80‰, respectively.

The mean $\delta^{18}O$ values of monthly samples collected by Zhang et al. (2005) from the Heihe River were −7.8±0.4‰ at S2, −6.9±1.5‰ at S3, −6.4±1.5‰ at S4, −3.6±2.2‰ at S5 and −6.3±1.5‰ at S6, and showed more positive rising along the river course. The $\delta^{18}O$ values of samples collected along the Heihe River as part of this study had a similar trend with that of the mean values from S1 to S6 (Fig. 4a). The $\delta^{18}O$ (δD) values increased from −8.1‰ (−58‰ for δD) at the outlet of the mountain (sample No. 63) to −4.9‰ (−50‰ for δD) at the end of the river in Ejina (sample No. 6-1). The most positive $\delta^{18}O$ and δD values were found in Dingxin agricultural area (sample Nos. 17 and 23).

The "deuterium excess" values $d$ ($d$=δD−8$\delta^{18}O$, Dansgarrd 1964) for the 17 samples ranged from +26 to −55‰. A significant difference in the behaviour of the deuterium excess at mountain areas and basins has been observed (Table 1 and Fig. 4b). The deuterium excess values for samples collected in the mountains area ranged between +10 and +26‰. The river has very high deuterium excess in the mountains area, suggesting that the stream water has been recycled through the atmosphere in arid conditions. In contradiction to these findings, the deuterium excess from the river samples in the basin ranged between +2 and −55‰. This indicated that evaporation is probably occurring in ponds, reservoirs and swamps in the basin. The relationship between δD and $\delta^{18}O$ of water samples in Table 1 was plotted with the local meteoric water line (LMWL) and global meteoric water line (GMWL) in Fig. 5. The samples for mountain areas were above the LMWL. These samples may be from recent local precipitation and are less subject to evaporation than other samples in the basin. The river samples collected in the basin lay to the right of the LMWL and show that the river is typically subjected to evaporation, being enriched with $\delta^{18}O$ and δD and resulting in a lower slope value. The mean values of δD and $\delta^{18}O$ were about 4.4±1.1 and 50±3‰ ($n$=8, sample Nos. 6-1, 10-1,17, 20, 23, 24, 33 and 41) with a mean d-excess of 9.8±7.3‰.

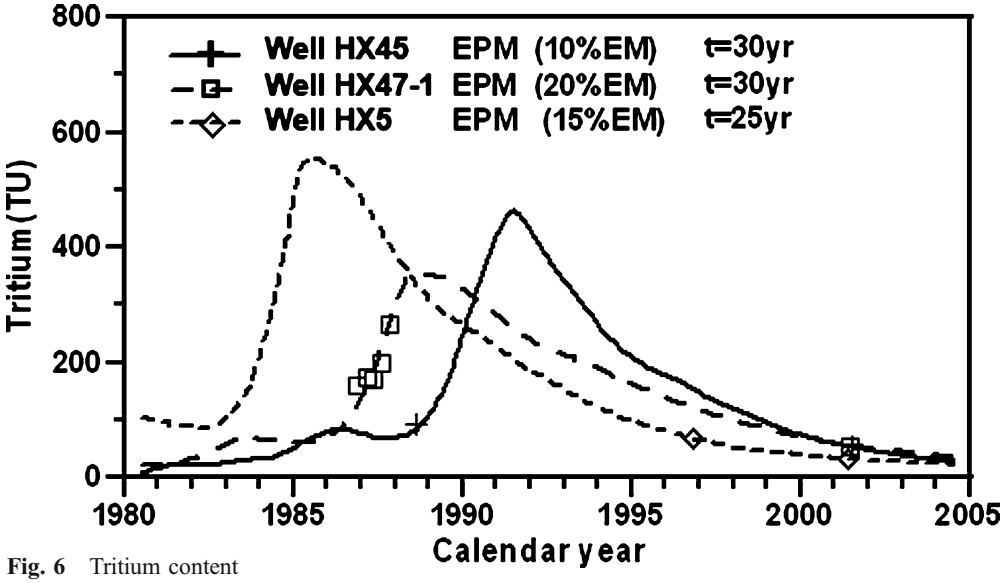


**Fig. 6** Tritium content vs. time, showing the fit of exponential piston flow model (*EPM*) variables for selected wells. *EM*=fraction of exponential flow

**Fig. 6** Tritium content

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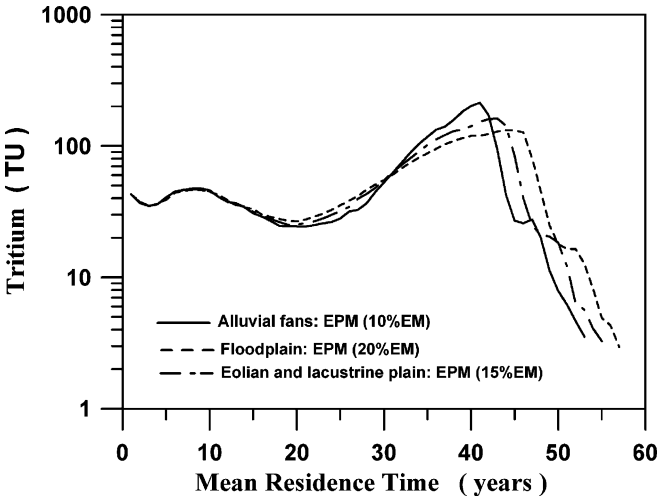


**Fig. 7** Output concentration of tritium as a function of the mean residence time for groundwater in different areas, using the exponential piston model (*EPM*) variables

## Isotopes in groundwater

Similar to river water, $\delta^{18}O$ and δD values for groundwater in the unconfined aquifer became positive along the flow path from the southern sub-basin to the northern sub-basin. Groundwater from the alluvial fan aquifer (sample Nos. 47, 48, 49, 50, 51, 52 and 53) were characterized by low $\delta^{18}O$ values ($-8.8<\delta^{18}O<-7.9‰$) and low δD values ($-57<\delta D<-53‰$). The samples from the wells drilled into the unconfined aquifers in the floodplain of the southern sub-basin (sample Nos. 26, 27, 34, 36-2, 37, 38, 39-2, 40, 42, 43, 44-2, 46 and 56) showed similar stable isotope compositions to those of the alluvial fan aquifer. They had a range of $\delta^{18}O$ values between −9.7 and −7.7‰ and δD values between −61 and −49‰. The recharge altitudes of these waters were calculated to be 2,300–4,300 m asl, which represented the altitude of the Qilian Mountains. The stable isotope compositions of the samples from the northern sub-basin (sample Nos. 1, 3, 5, 10, 14, 15, 16, and 21) deviated even more from the LMWL with a slope of 2.5 and may reflect enrichment due to evaporation processes. They had a wide range of $\delta^{18}O$ ($-7.6‰<\delta^{18}O<-1.7‰$) and δD ($-56‰<\delta D<-41‰$) values. The relatively high deuterium excess was observed in the southern sub-basin ($d$=10 to 20‰), and the low deuterium excess was found in the northern sub-basin ($d$=−31 to 10‰) (Fig. 4b). The low deuterium excess in the northern sub-basin probably indicated enrichment by evaporation.

The tritium content of groundwater in the basin ranged from 2 TU up to 164 TU. The high tritium contents (>60 TU) were found for the groundwaters emerging along the front of the alluvial fan zone

**Table 3** Tritium concentrations and MRT interpretation of tritium using an exponential piston flow model (EPM)

| Sample No. | Well name | $^3H$ (TU) | Mixing fraction (%) | Mean ages (in years) based on the EPM model | Interpreted age (years) |
|---|---|---|---|---|---|
| Unconfined aquifer | | | | | |
| 1 | HX1 | 33 | 15 | 15, 25, 46 | 25 |
| 3 | HX3 | 36 | 15 | 3, 13, 26, 46 | 26 |
| 5 | HX5 | 32 | 15 | 15, 25, 46 | 25 |
| 10 | HX10 | 31 | 15 | 16, 26, 46 | 26 |
| 11 | HX11 | 27 | 15 | 18, 22, 47 | 22 |
| 12 | HX12 | 36 | 15 | 3, 13, 26, 46 | 26 |
| 13 | HX13 | 21 | 15 | 48 | 48 |
| 14 | HX14 | 41 | 15 | 1, 5, 11, 27, 46 | 27 |
| 15 | HX15 | 25 | 15 | 20, 47 | 20 |
| 16 | HX16 | 19 | 15 | 50 | 50 |
| 19 | HX19 | 39 | 10 | 2, 5, 12, 27, 46 | 12 |
| 21 | HX22 | 30 | 10 | 16, 24, 46 | 16 |
| 26 | ZY22 | 3 | 10 | 58 | 58 |
| 27 | HX31 | 9 | 10 | 54 | 54 |
| 34 | HX26 | 21 | 10 | 50 | 50 |
| 36–2 | HX42–2 | 78 | 10 | 34, 47 | 34 |
| 37 | HX47–1 | 52 | 10 | 29, 48 | 29 |
| 38 | HX41–1 | 94 | 10 | 36, 47 | 36 |
| 39–2 | HX46–2 | 164 | 10 | 44 | 44 |
| 40 | ZY5 | 136 | 10 | 44 | 44 |
| 42 | HX45 | 52 | 10 | 29, 48 | 29 |
| 43 | HX37 | 16 | 10 | 52 | 52 |
| 44–2 | HX36–2 | 46 | 10 | 8, 28, 48 | 28 |
| 46 | ZY8 | 38 | 10 | 2, 5, 12, 26, 48 | 26 |
| 47 | HX59 | 59 | 20 | 31, 43.5 | 31 |
| 48 | HX29 | 66 | 20 | 31, 43.5 | 31 |
| 49 | HX49 | 25 | 20 | 18, 22, 47 | 22 |
| 50 | HX48 | 27 | 20 | 17, 24, 45 | 24 |
| 51 | ZY21 | 33 | 20 | 14, 27, 44 | 27 |
| 52 | HX50 | 105 | 20 | 34, 43 | 34 |
| 53 | HX76 | 54 | 20 | 30, 44 | 30 |
| 56 | HX75 | 2 | 10 | 59 | 59 |

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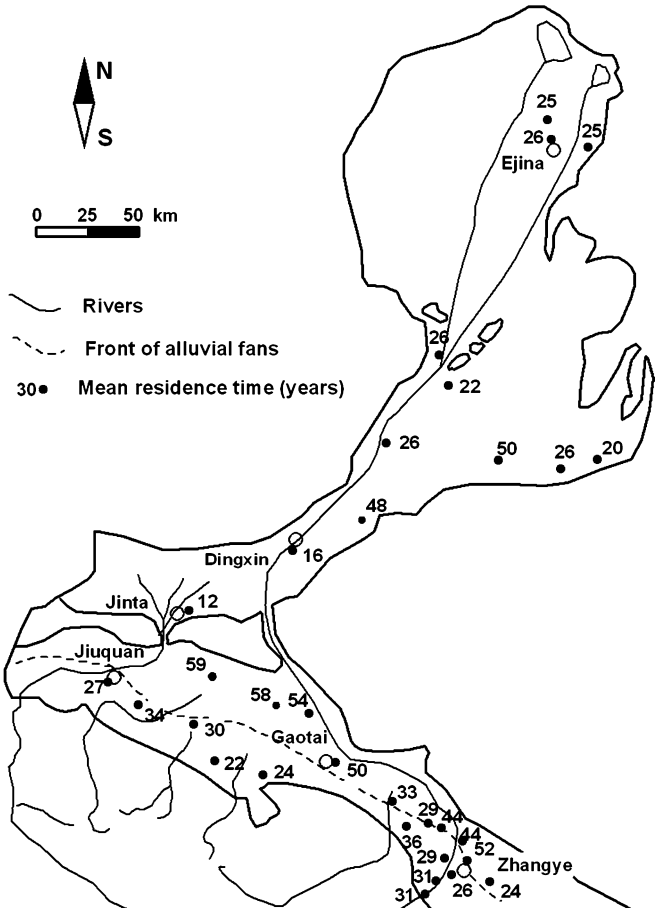


**Fig. 8** Mean residence time for shallow groundwater in the Heihe River Basin

(sample Nos. 36-1, 36-2, 38, 39-2, 40, 47, 48 and 52), while the low (<3 TU) tritium contents (sample Nos. 2, 9, 25, 26, 28, 29, 30, 35, 45, 54, 56, 57 and 58) were found for the groundwaters from the central part of the basin, where the artesian conditions start to become evident in wells. The relatively high tritium content (30–60 TU) in wells located at the northern sub-basin near the river was attributed to admixture of surface water with high tritium content.

Similar to the tritium content trend, the $^{14}C$ activity exceeds 70 pMC in groundwater around the piedmont of the southern sub-basin, and $^{14}C$ activity less than 70 pMC is found in the centre of the southern sub-basin and Ejina area in the northern sub-basin.

## Discussion

### *Mean residence time (MRT) and groundwater age*

Tritium is a reliable tracer to distinguish groundwater recharged during the pre-bomb time from younger water (Clark and Fritz 1997). Several models, using different transit time distributions, have been suggested for the use of tritium in the estimation of mean residence time of

**Table 4** The calculated $P_{CO2}$, saturation index of calcite ($SI_{CaCO3}$) and $\delta^{13}C_{DIC}$ from samples of confined aquifer in southern sub-basin. *CAL* calculated; *MEAS* measured

| Sample No. | Well name | $K^+$ (mg/L) | $Na^+$ (mg/L) | $Ca^{2+}$ (mg/L) | $Mg^{2+}$ (mg/L) | $Cl^-$ (mg/L) | $SO_4^{2-}$ (mg/L) | $HCO_3^-$ (mg/L) | $^{14}C$ (pMC) | Temp. (°C) | pH | LOG $PCO_2$ | LOG $SICaCO_3$ | $\delta^{13}C_{DIC-CAL}$ (‰) | $\delta^{13}C_{DIC-MEAS.}$ (‰) |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| 2 | HX2 | 4.8 | 204.5 | 50.1 | 51.0 | 211.3 | 302.6 | 208.7 | 18.3 | 13.1 | 8.0 | −2.8 | 0.2 | −5.8 | −5.5 |
| 4 | HX4 | 5.7 | 112.8 | 50.3 | 74.5 | 122.0 | 331.4 | 156.2 | 26.6 | 14.6 | 7.9 | −2.8 | 0.0 | −5.5 | −5.9 |
| 6 | HX6 | 6.5 | 223.9 | 42.4 | 99.7 | 271.6 | 560 | 239.2 | 38.7 | 13.1 | 7.7 | −2.4 | −0.2 | −6.7 | −7.0 |
| 7 | HX7 | 6.5 | 191.6 | 45.7 | 36.0 | 111.3 | 307.4 | 229.4 | 27.4 | 11.0 | 7.8 | −2.5 | 0.0 | −6.7 | −4.7 |
| 8 | HX8 | 4.6 | 256.2 | 47.9 | 33.5 | 245.3 | 293.5 | 209.9 | 22.5 | 13.8 | 8.1 | −2.9 | 0.3 | −5.9 | −5.7 |
| 9 | HX9 | 6.9 | 63.9 | 37.3 | 30.6 | 59.6 | 116.2 | 183.1 | 48.6 | 14.0 | 7.6 | −2.5 | −0.2 | −7.3 | −5.2 |
| 25 | ZY23 | 11.3 | 206.0 | 41.7 | 91.9 | 187.9 | 317 | 385.9 | 36.8 | 14.5 | 7.0 | −1.5 | −0.6 | −6.6 | −7.7 |
| 28 | ZY25 | 8.2 | 32.0 | 50.5 | 44.5 | 17.0 | 129.7 | 331.9 | 18.3 | 14.0 | 8.0 | −2.6 | 0.5 | −5.6 | −4.4 |
| 29 | ZY26 | 2.8 | 33.1 | 51.0 | 52.9 | 21.3 | 100.9 | 331.9 | 25.6 | 13.0 | 8.0 | −2.6 | 0.5 | −5.7 | −5.0 |
| 30 | ZY18 | 2.2 | 48.2 | 18.0 | 17.3 | 39.0 | 42.3 | 145.2 | 34.9 | 13.0 | 8.5 | −3.4 | 0.3 | −12.0 | −3.8 |
| 31 | HX79 | 2.4 | 22.5 | 20.0 | 39.4 | 13.5 | 52.8 | 233.1 | 26.3 | 13.0 | 8.4 | −3.1 | 0.4 | −11.3 | −4.7 |
| 35 | HX27 | 4.1 | 103.0 | 53.7 | 32.7 | 149.6 | 146.0 | 163.5 | 41.4 | 16.0 | 8.1 | −2.9 | 0.3 | −5.0 | −4.4 |
| 45 | HX35–1 | 2.8 | 28.2 | 12.0 | 16.5 | 17.7 | 8.6 | 112.3 | 67.8 | 11.7 | 9.4 | −4.4 | 0.9 | −14.7 | −6.7 |
| 54 | ZY16 | 7.9 | 61.9 | 67.3 | 110.1 | 64.5 | 436.1 | 253.8 | 13. 6 | 11.0 | 7.8 | −2.5 | 0.2 | −4.7 | −5.2 |
| 57 | HX77 | 5.3 | 48.9 | 22.0 | 24.5 | 26.2 | 87.4 | 173.3 | 39.2 | 16.3 | 8.4 | −3.3 | 0.4 | −10.0 | −5.5 |
| 58 | HX78 | 7.5 | 97.8 | 60.2 | 102.9 | 56.0 | 483.2 | 183.1 | 48.7 | 12.4 | 8.1 | −2.9 | 0.3 | −5.0 | −5.5 |


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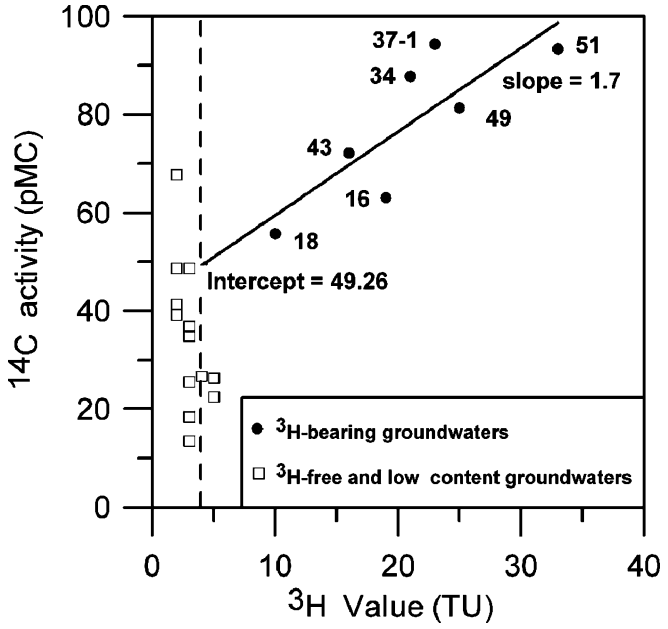


**Fig. 9** Plot of $^{3}H$ vs. $^{14}C$ for selected groundwater samples which are thought not to contain bomb-tritium. The slope for higher tritium content samples intercepts with the detection limit (*dashed line* at 3 TU), to provide an estimate of the initial $^{14}C$ activity of groundwater

groundwater (Maloszewski and Zuber 1982). In this study, the exponential piston flow model (EPM) was used to estimate the mean residence times of groundwater. This model has been found to produce satisfactory fit for groundwater samples collected from northwestern China (Gao 1995). The tritium input function required for this model was adopted from the available records of the station at Zhangye. However, systematic measurements of tritium in Zhangye precipitation did not begin until 1986. Thus, the tritium values in precipitation during 1953–1959 were approximated from the correlation with the data of Ottawa, Canada, while the values during the period of 1960 to 1986 were estimated from the Doney model (Doney et al. 1992). The parameters for EPM were obtained by matching the model output to the measured tritium data. The published tritium data available from well HX45 (located at the distal fan), well HX47-1 (located in an irrigation area) and well HX5 (located at Ejina) since 1980 (Zhao 2002), have been used together with this study's tritium data to estimate the model parameters. Exponential piston flow models with 10, 20, 15% exponential flow, compared with the total flow, were identified from matches with tritium data for the sites of piedmont, floodplain and desert area (Ejina), respectively (Fig. 6). The tritium output curves for different values of parameters in the year 2001 are shown in Fig. 7. And the values of MRT calculated for samples are listed in Table 3.

In the southern sub-basin, groundwater samples (sample Nos. 26, 27, 39-2, 40 and 56) from the lower part of the floodplain had bomb and pre-bomb tritium levels that gave unambiguous mean residence times of 44–59 years. Groundwater samples downgradient of the front of distal fans (sample Nos. 34, 36-2, 37, 38, 42, 43, 44-2, and 46) and from the alluvial fans themselves (sample Nos. 47, 48, 49, 50, 51, 52 and 53) were mostly in the ambiguous tritium range and had ages between 20 years and 50 years. Ambiguity could be resolved with hydrogeologic information. Because of the high permeability of coarse-grained gravel and sand, these waters ages would possibly range from 22 years to 36 years. The spatial distribution of MRT (Fig. 8) showed an increase from the piedmont towards the central zone of the sub-basin. Based on this spatial distribution, two main directions of groundwater flow were defined: from the SE towards the NW and from the SW towards the NE. These two flow paths are in agreement with the water table contours (Fig. 3) of the south sub-basin. Thus, the groundwater recharge was identified in the piedmont of Qilian Mountains.

In the northern sub-basin, groundwater samples (sample Nos. 1, 3, 5, 10, 11, 12, 14, 15, 19 and 21) in the Dingxin agricultural area and Ejina oasis gave ambiguous

**Table 5** Groundwater ages determined by application of $^{14}C$-DIC models

| Sample No. | Well name | Uncorrected age (year BP) | $^{3}H$ corrected age (year BP) | Tamers model age (year BP) | $\delta^{13}C$ mix model age (year BP) | IAEA model age (year BP) |
|---|---|---|---|---|---|---|
| 2 | HX2 | 14021 | 7802 | 8520 | 7437 | 1011 |
| 4 | HX4 | 10938 | 4719 | 5480 | 4696 | −1569 |
| 6 | HX6 | 7840 | 1621 | 2578 | 2696 | −3565 |
| 7 | HX7 | 10706 | 4487 | 5346 | 3146 | −3367 |
| 8 | HX8 | 12343 | 6124 | 6786 | 5988 | −414 |
| 9 | HX9 | 5962 | −257 | 712 | −1209 | −7384 |
| 25 | ZY23 | 8260 | 2041 | 4117 | 2918 | −2450 |
| 28 | ZY25 | 14044 | 7825 | 8529 | 6002 | −353 |
| 29 | ZY26 | 11268 | 5049 | 5757 | 4084 | −2357 |
| 30 | ZY18 | 8703 | 2484 | 3042 | 2581 | −3963 |
| 31 | HX79 | 11032 | 4813 | 5394 | 3546 | −2983 |
| 35 | HX27 | 7299 | 1080 | 1750 | −839 | −7043 |
| 45 | HX35−1 | 3219 | −3000 | −2502 | −1817 | −8523 |
| 54 | ZY16 | 16512 | 10299 | 11127 | 9621 | 3086 |
| 57 | HX77 | 7748 | 1529 | 2097 | 1073 | −5170 |
| 58 | HX78 | 5951 | −268 | 420 | −569 | −7078 |

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**Table 6** The $\delta^{18}O$ and $\delta D$ values for selected samples and correction for evaporation using evaporation line with a slope (S) of 2.5

| Sample No. | Raw $\delta^{18}O$ (‰) | Raw $\delta D$ (‰) | Corrected $\delta^{18}O$ ($S$=2.5) (‰) | Corrected $\delta D$ ($S$=2.5) (‰) |
|---|---|---|---|---|
| Surface water in basins | | | | |
| 6–1 | −4.9 | −50 | −9.4 | −61 |
| 10–1 | −5 | −53 | −10.0 | −66 |
| 17 | −1.7 | −32 | −7.2 | −46 |
| 20 | −7.4 | −51 | −8.2 | −53 |
| 24 | −6.8 | −53 | −9.0 | −59 |
| 33 | −7.2 | −56 | −9.5 | −62 |
| 41 | −8 | −56 | −9.0 | −59 |
| Mean value | −5.9±0.8 | −50±3 | −8.9±0.4 | −58±2.5 |
| Unconfined aquifer (north basin) | | | | |
| 1 | −6.9 | −46 | −7.3 | −47 |
| 3 | −5.9 | −47 | −8.1 | −52 |
| 5 | −6.2 | −48 | −8.4 | −54 |
| 10 | −7.6 | −51 | −8.2 | −52 |
| 14 | −1.7 | −45 | −10.2 | −67 |
| 15 | −3.5 | −56 | −9.3 | −60 |
| 16 | −1.9 | −41 | −9.0 | −59 |
| 19 | −8.4 | −54 | −8.4 | −54 |
| 21 | −5.9 | −52 | −9.3 | −61 |
| Mean value | −5.3±08 | −49±1.6 | −8.7±03 | −56±2.0 |

ages between 12 and 47 years. The water tables, both at Dingxin and in the Ejina area, were very shallow (3 m) and therefore the water age could be young. In the Dingxin area, river water was the main source for irrigation. Thus, the groundwater MRT should be less than 20 years. In the Ejina area, Zhang (2005) suggested that groundwater MRT in this area was about 10 years. Since the river water and lake were the main recharge source for groundwater in Ejina, and the lake dried up during 1992–2000, groundwater MRT should be greater than 10 years. The range of MRT in this area was adopted as between 20 and 30 years. For the groundwater samples in the Gobi Desert (sample Nos. 10, 11, 12, 13, 14, 15 and 16), the MRT was selected as greater than 20 years, which referred to the result from the study by Zhang (2005). The groundwater MRT was thus resolved, and is listed in Table 3 and shown in Fig. 8. The spatial distribution of MRT suggested that groundwater recharge both in the Dingxin and Ejina oases were from the Heihe River. Groundwater flowed from Dingxin towards the NE and N. This was also in an agreement with the water table contours (Fig. 3).

Groundwaters in the confined aquifer with relatively low tritium content (sample Nos. 2, 4, 6, 7, 8, 9, 22, 25, 28, 29, 31, 30, 35, 45, 54, 55, 57 and 58) were supposed to be recharged before 1952. Their ages were determined by $^{14}C$ dating methods. The samples from confined aquifers, found to contain considerable tritium (sample Nos. 18, 32, 36–1, 39-1 and 44-1), were modern water or a mixture of modern water and old water. They can not be dated by $^{14}C$ dating methods. For $^{14}C$ dating of groundwater, the initial $^{14}C$ activity ($A_0$) at the time of recharge was required as a starting point. Much of the carbon in groundwater is derived from gaseous $CO_2$ in the vadose zone. However, this carbon, containing high levels of $^{14}C$, was diluted by low $^{14}C$ activity carbon dissolved from minerals during groundwater recharge. For this reason, the estimation of the initial $^{14}C$ activity ($A_0$) was difficult. There were several methods of correcting the initial $^{14}C$ activity for dead-C dilution (Vogel 1970; Tamers 1975; Pearson 1965; Mook 1976; Fontes and Garnier 1979). These methods considered either only chemical mixing between the carbon compounds, or isotopic fractionation. Geyh (2000) gave a summary overview of the current state-of-art in the radiocarbon dating of groundwater.

Generally, the concentration of dissolved inorganic carbon (DIC) was controlled initially by variations in dissolved $CO_2$ present in the soil zone, which was then taken up by reactions within the aquifers (Fontes 1983; Kalin 1999). The $\delta^{13}C_{DIC}$ values of samples had a range of −7.7 to −4.4‰. The possible sources of $\delta^{13}C_{DIC}$ were dissolved $CO_2$ and the dissolution of calcite. The extent of calcite dissolution depended on the openness of the system. The combination of $P_{CO2}$ (partial pressure of carbon dioxide), mDIC (the molar concentration of DIC) and $\delta^{13}C_{DIC}$ can provide an indication of recharge conditions (Clark and Fritz 1997). Non-measured parameters were assigned the following values: $\delta^{13}C$ value of mineral carbonate 1.5‰ and $\delta^{13}C$ value of the soil gas, −23‰. The calculated $P_{CO2}$ (Table 4) was less than that in soil ($P_{CO2}$ about $10^{-1.8}$–$10^{-2.0}$). Thus, calcite dissolution was taking place under closed system conditions and the initial carbonic acid from soil $CO_2$ was consumed by calcite dissolution. The saturation index (SI) of calcite calculated from the ion activity product and the solubility constant indicated that groundwater had dissolved calcite to the saturation point. Most theoretical $\delta^{13}C$ values of DIC calculated for the samples were close to the measured values and supported the closed system dissolution of carbonate. However, four groundwater samples (sample Nos. 30, 31, 45 and 57) have big deviations from the measured $\delta^{13}C$ value. In view of the fact that the pH values (8.4–9.4) of these samples were higher than groundwater samples in the recharge area, the calculated $\delta^{13}C_{DIC}$ should be lower than the measured values. This suggests that the carbonate system has

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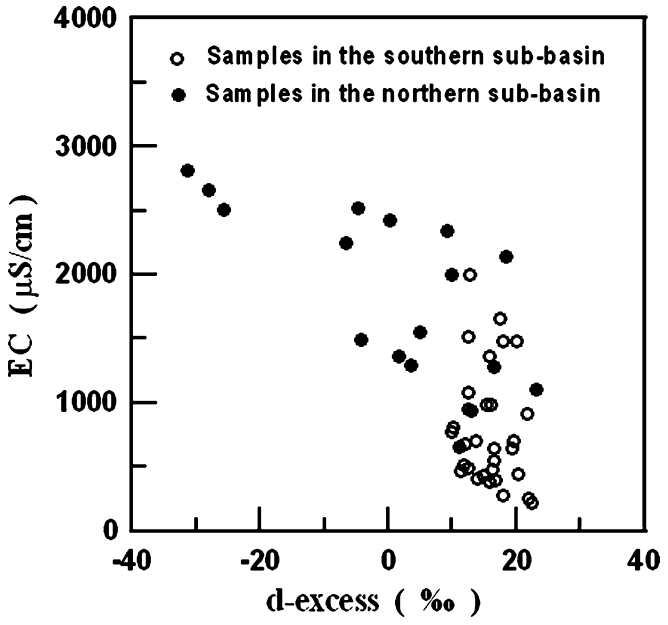


**Fig. 10** Relationship between d-excess and EC for groundwater samples in unconfined aquifers

evolved during the residence time in the aquifer. The difference may reflect reaction in the aquifer such as matrix exchange, etc. (Clark and Fritz 1997).

To calculate the ages of groundwater recharged prior to 1952, an assumption was made in the first approximation that calcite was dissolved under fully closed system conditions, where there was no exchange with soil $CO_2$ during calcite dissolution. In this case, more than 50% of the carbon of DIC was derived from dissolving carbonate minerals (Tamers 1975). A second, alternative approach was to apply the $\delta^{13}C$ mixing model (Pearson 1965). It considered the $\delta^{13}C$ enrichment during dissolution of $CO_2$ and the evolution of $\delta^{13}C$ of DIC (Clark and Fritz 1997). Finally, the IAEA (International Atomic Energy Agency) model was used (Salem et al. 1980) to consider the mixing and isotope exchange. Both the Tamers model and the $\delta^{13}C$ mixing model gave a similar mean $A_0$, 51.9± 2.45 pMC and 44.98±5.21 pMC, respectively. However, the IAEA model gave a low mean $A_0$ (20.92±2.75) and actually overcorrected to produce future ages.

Another reliable approach to the estimation of $A_0$ is to plot the $^3H$ value vs. the $^{14}C$ activity (Buckau et al. 2000; Geyh 2000). In Fig. 9, the $^3H$ value for selected groundwater samples which were believed to not contain bomb $^3H$, were plotted against the $^{14}C$ activity. The data points can be divided into two groups. One is groundwaters with tritium content near or less than the detection limit. The other group contains those groundwaters that have a linear correlation between the tritium value and $^{14}C$ activity (but thought to be free of bomb $^{14}C$). The initial $^{14}C$ activity, $A_0$=49.26 pMC, was estimated by extrapolating the fit line for these samples to the detection limit of $^3H$ (3 TU). This $A_0$ value was similar to the values calculated with the Tamers model and the $\delta^{13}C$ mixing model. The groundwater age calculated by the Tamers model and $\delta^{13}C$ mixing models are shown together with the other models in Table 5. The IAEA model age was unlikely to be representative of the groundwater since the model overcorrected the ages. Both the Tamers model ages and the $\delta^{13}C$ mixing model ages were not far from ages calculated with the $^3H$ method and likely to be the most representative.

In view of $\delta^{13}C$ mixing model ages, groundwater in the southern sub-basin falls between the modern age and 9621 years BP. Modern groundwater was found along the front of alluvial fans near the Heihe River and Beidahe River, while more than 1,000-year-old water was found in the central sub-basin of the southern sub-basin. The water age increased from SE to NW starting from 0 year at Gaotai (well HX27) and rising to 9,000 years BP at the discharge zones (well ZY16). The ages in the northern sub-basin increased towards the north and west. Ground-

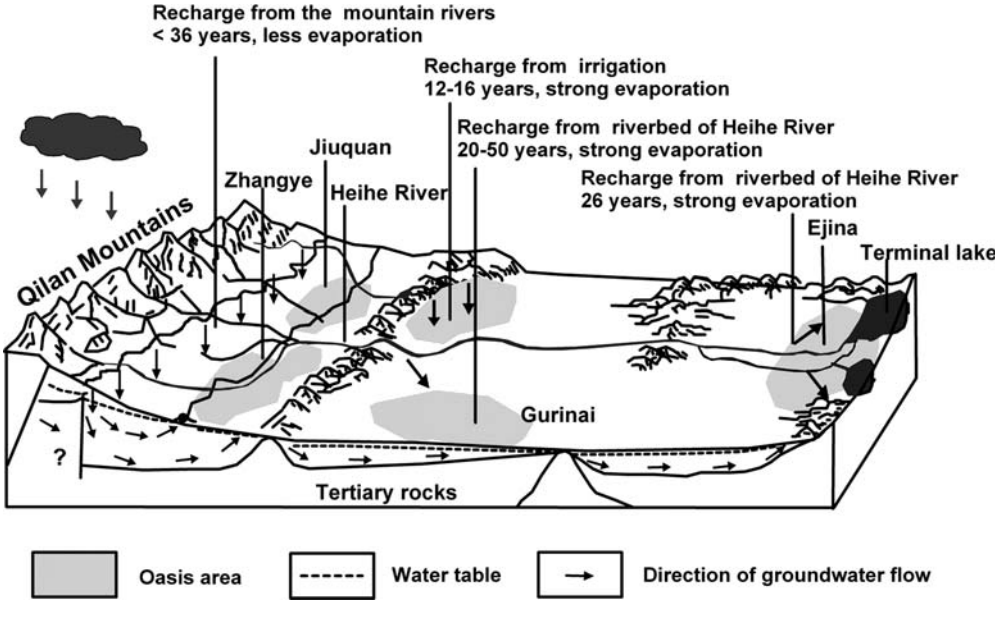


**Fig. 11** The significant modern recharge mechanisms in the arid Heihe River Basin (diagram of conceptual landscape revised from Liu 1998)

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water ages for the Ejina oasis (Wells: HX2, HX4, HX6 and HX7) rose from 2,696 years BP to 7,437 years B.P. Well HX8 (sample No. 8), located in the Gobi Desert, west of the Ejina oasis had an age of 5,988 years BP. Another well HX9 (sample No. 9), located in the Gobi Desert area, south of the oasis showed a modern age which suggested that groundwater near this well was recharged by river water before 1952. It seemed that groundwater flowed from the south towards the north and from the east to the west. The wells in Ejina oasis (wells: HX4, HX6 and HX7) and well HX8 in the Gobi Desert all had the same potentiometric level of about 940 m asl, while the potentiometric level of well HX9 is about 1,000 m asl. Thus, the age difference between the former wells (HX8 and wells in Ejina) and well HX9 reflected the different flow velocity from well HX9 to well HX8 and wells in Ejina.

## Origin and recharge areas of groundwater

The average environmental isotope contents of the water samples collected at different sites were reported in Table 2 and showed that recharge source for groundwater was from mountain rivers. The $\delta^{18}O$ and $\delta D$ values of water samples collected from the alluvial fans at wells with sample Nos. 47–53 ($\delta^{18}O$ −8.8±0.2‰; $\delta D$ −55±1‰; $n$=7) and the unconfined aquifer ($\delta^{18}O$ −8.4±0.2‰; $\delta D$ −54±1‰; $n$=13; sample Nos. 26, 27, 34, 36–2, 37, 38, 39-2, 40, 42, 43, 44-2, 46 and 56) in the southern sub-basin were within the same range as those collected from the mountain rivers ($\delta^{18}O$ −8.3±0.2‰, $\delta D$ −50±1‰; $n$=7, sample Nos. 59–65). These waters were isotopically depleted compared with the precipitation in Zhangye. Thus, the precipitation may not be the dominant mechanism for groundwater recharge. The closeness of the groundwater and river composition and similar d-excess indicated that the mountain river was a major contributor to recharge in the southern basin. The calculated recharge altitudes indicated recharge water came from the Qilian Mountains.

Groundwater samples collected in the southern sub-basin and surface water samples collected in the mountain areas had relatively low $\delta^{18}O$ and $\delta D$ values. These depleted samples were above the LMWL showing a deuterium excess (Fig. 5). It was rare to find surface water or groundwater that plotted above the line. However, in low-humidity regions, re-evaporation of precipitation from local surface waters created vapour masses with isotopic contents that plotted above the local meteoric water line (Clark and Fritz 1997). The shift above the LMWL may suggest that the vapour evaporated from the hydrologically closed desert basin. This also indicated that the recharge waters infiltrated rapidly through the highly permeable zone after precipitation or snow melting before significant loss by evaporation. The Qilian Mountains had the greatest precipitation and logically provide the most recharge to the basin. The high tritium contents in the alluvial fan may be due to short turnover time of groundwater or to direct inflow from the mountain rivers. The d-excess values for the samples from the mountain rivers and aquifers in the southern sub-basin ranged from +10 to 26‰. It also suggested that the source of recharge for groundwater was the mountain river and the recharge waters infiltrated rapidly, with no significant loss by evaporation.

The ranges of $\delta^{18}O$ and $\delta D$ values in groundwaters (sample Nos. 1, 3, 5, 10, 14, 15, 16 and 21) collected from unconfined aquifers in the northern sub-basin are −8.9 to −1.7‰ and −56 to −41‰, respectively. The ranges were close to those of the samples collected from the river in the southern sub-basin (−8.0 to 0.8‰ and −56 to −32‰; respectively). The relationship between $\delta^{18}O$ and $\delta D$ in river samples and groundwater samples in the northern sub-basin suggested that kinetic fractionation via evaporation from the soil or water surface had influenced the isotope values. Groundwater samples from the northern basin lay to the right of the LMWL (Fig. 5). The scattering of samples can be attributed to evaporation. The original isotope values of the water prior to evaporation can be inferred by determining the slope of the change in $\delta^{18}O$ and $\delta D$ as the isotope values deviate from the LMWL (Clark and Fritz 1997). The original values of $\delta^{18}O$ and $\delta D$ were determined by calculating the intersection with the LMWL of an evaporation line (slope=2.5). Each corrected value was listed in Table 6. The corrected values showed that $\delta^{18}O$ values ranged from −10.2 to −7.3‰ with a mean value of −8.7‰±0.3 for groundwater, and −10.0 to −7.2‰ with a mean value of −8.9‰±0.4 for river waters. The corrected values showed that $\delta D$ values ranged from −67 to −47‰ with a mean value of −56‰±2.0 for groundwater, −46 to −66‰ with a mean value of −58‰± 2.5 for river waters. These corrected values were in the ranges of the mountain river waters. This finding indicated that the recharge source was the Heihe River. Groundwater samples in the northern sub-basin had d-excess values between −31 and 10‰. The low deuterium excess in the northern sub-basin probably indicated enrichment by evaporation. One groundwater sample (sample No. 21) was collected at Dingxin agricultural area with d-excess of −5‰. Two river samples (sample Nos. 17 and 23) were collected at Dingxin agricultural area. They had a d-excess value of −18 and −55‰, respectively. This lowest d-excess value may have resulted from re-evaporation of irrigation return flow. Three groundwater samples (sample Nos. 14, 15 and 16) with d-excess values between −31 and −26‰, were collected from shallow wells near the Gurinai grassland. This was in agreement with the observations of Geyh and Gu (1999). These results suggested that this extremely low deuterium excess was caused by the physical processes occurring in the unsaturated zone which modified the isotopic composition of the pore water in arid regions. The plot of the EC vs. the d-excess values clearly indicates two groups of groundwater (Fig. 10). One is the groundwater in the southern sub-basin whose d-excess values (+10 to +20‰) do not vary with increasing EC. This indicates that evaporation does not appear to influence salinization. However, the other groundwater group, in the northern sub-basin, clearly has

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a tendency for decreasing d-excess with increasing EC values. This reflects the effect of strong evaporation.

Deep groundwater samples collected from the south sub-basin (sample Nos. 25, 28, 29, 30, 31, 32, 35, 36-1, 39-1, 44-1, 45, 54, 55, 57 and 58) had depleted levels of heavy isotopes (−9.8±0.1‰ for $\delta^{18}O$; −61±1‰ for δD; $n$=15) compared with those collected from wells in alluvial fans and shallow wells. They also lay above the LMWL showing a deuterium excess between 10 and 22‰. The $\delta^{18}O$ and δD values of two springs (−9.6 and −62‰, respectively, for spring sample No. 66; −10.0 and −56‰ respectively, for spring No. 67), collected from the mountain valley, were close to the values of the confined aquifer in the southern sub-basin. The $^{14}C$ ages of these samples were up to about 9,000 years BP. Thus, the recharge water for the confined aquifer may be derived from deep sources. However, the tritium in some of these groundwater samples near the river suggested that they were recharged during the past 40 years. Two recharge mechanisms for deep groundwater possibly occurred in the south sub-basin. One involved recharge via infiltration from the rivers that drain the mountains, and the other involved lateral recharge from mountains through fractures.

The d-excess of deep groundwater in Ejina was significantly lower than that of the meteoric water. The $^{14}C$ data showed that groundwater in the deep confined aquifer may date, in part, to the climate optimum in the Holocene. It is possible that recharge took place during the warmer climate in this period. However, the high tritium content in some samples may be caused by the mixture of shallow groundwaters with paleowaters.

In summary, modern recharge to the aquifers in the Heihe River Basin occurred via three main processes (Fig. 11):

1. Recharge occurred directly by rapid infiltration from a series of ephemeral or perennial mountain streams. This recharge has taken place in the alluvial fan zone where the water table was 50–200 m below ground level. Less evaporation occurred during recharge. The mean residence time of groundwater was less than 36 years.
2. Irrigation return flow mainly occurred in the Jinta-Dingxin agricultural area, where water from the Heihe River provided the major source of irrigation water for land users. Strong evaporation occurred prior to recharge and the mean residence time of groundwater was 12–16 years.
3. Infiltration from the Heihe River was along the riverside. Strong evaporation occurred prior to recharge and the mean residence time of groundwater was 26 years in the Ejina oasis.

## Conclusions

Groundwater in the Heihe River Basin mainly recharges from the drainage of mountain rivers approaching from the Qilian Mountains. Recharge to the aquifer occurs in the highly permeable zone of the piedmont and along the riverside in the Ejina area, by infiltration from the Heihe River and irrigation return flow in agricultural areas. The mean residence time of groundwater is less than 36 years in the alluvial fan zone and about 44–59 years down-gradient towards the flood plain. The mean residence time of groundwater is 12–16 years in Dingxin agricultural areas while it is about 25 years in the Ejina area. However, groundwater of confined aquifers is old with ages between 0 and 10,000 years BP. Exploitation of this groundwater is considered to be “mining” if more water is withdrawn than is replenished.

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