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ORIGINAL ARTICLE

# Tracing infiltration and recharge using stable isotope in Taihang Mt., North China

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**Abstract** The groundwater in headwater region is an important recharge source for the adjacent mountain-front plain. In order to reveal the relationship among precipitation, soil water and groundwater, from June to September in 2004, stable isotopes (deuterium and oxygen-18) in precipitation and soil waters at the depths of 10, 20, 30, 50, 70, 90, and 110 cm were analyzed at two sites covered by black locust (*Robinia Pseudoacia* L.) (Site A) and grass predominated by Themeda triandra (T. *japonica* (Willd.) Tanaka) and Bothriochloa ischaemum (B. *ischaemum* (L.) Keng) (Site B) in an experimental catchment at Taihang Mt., North China, respectively. The $\delta^{18}$O of precipitation in daily rain events shows large variations (–13.3 to –4.3‰) with a mean of 8.1‰. The $\delta^{18}$O and $\delta$ D of soil waters along profiles in two sites suggest that the influence of canopy cover was just up to 10 cm in top soil water. The soil water moved over the zero flux plane at 70 cm in-depth is expected to escape the evaporative effect at the end of September in both sites. The results show that the stable isotope, instead of tritium as tradition, can be used to trace the soil water behaviors based on the movement of isotopic peak along the vertical profiles in this semi-arid and semi-humid mountainous region. The infiltration depths of soil water in Taihang Mt. are 12 and 10 mm/day from June to September in 2004 in Site A and Site B, respectively. Tracing by stable isotope, recharge fluxes of soil water to local groundwater are of 3.8 and 3.2 mm/day in Site A and Site B, respectively. The results provide desirable information for assessment of local groundwater resources.

**Keywords** Stable isotopes · Precipitation · Soil water · Groundwater · Infiltration · Recharge · Taihang Mt.

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## Introduction

The recharge process to groundwater is very important to assess regional water resources in mountains. It is rarely measured in the field in semi-arid and semi-humid mountainous region. Taihang Mt. region is the most important recharge source for the adjacent North China Plain (NCP), an important crop production region in China, where more and more serious water problems such as the water table declines continuously and worsening of water quality occurs (Liu and He 1996; Liu et al. 2001; Yang et al. 2002; Tang et al. 2004). The proper simulation of this recharge requires detailed knowledge of hydrological processes integrating precipitation, soil water along the profile and groundwater.

Understanding the influence of vegetation on soil water movement is important for water resource management. The main recharge period is the rainy season in Taihang

Mt. Thus, the recharge processes in this period on the variation of soil water along the profile is very important. However, these processes are poorly understood in semi-arid and semi-humid environments.

Stable isotopes (oxygen–18 and deuterium) were used to estimate regional recharge and evaporation (Allison and Barnes 1983; Saxena and Dressie 1984) and water vapor transport in soils and interaction of water vapor between soil and air (Barnes and Walker 1989; Knowlton et al. 1992). Most of these studies were carried out in arid region (Mathieu and Bariac 1996), and few studies were reported in semi-arid and semi-humid, especially in mountainous region.

Stable isotopic compositions of soil water can reveal information about soil water fluxes (e.g., evaporation, transpiration and downward infiltration) that is difficult to extract by other techniques. Zimmermann (1967b) reported the effect of soil water evaporation on the stable isotopic compositions. After that, the variation of deuterium and Oxygen composition in soil water has been determined in many later studies (Barnes and Allison 1983, 1984; Allison et al. 1984; Bengtsson et al. 1987; Walker et al. 1988; Barnes and Walker 1989; Midwood et al. 1993; Liu et al. 1995; Gehrels et al. 1998; Gazis and Feng 2004). The experiments on vegetated soil water were also rarely carried out. Newman et al. (1997) compared stable isotope profiles in soils under two different plant communities in semi-arid piñon-juniper woodland and ponderosa pine forest. Gehrels et al. (1998) tracked the behavior of soil water by monitoring the soil water fluctuations and changes in the environmental tracer oxygen–18 in vertical profiles at test sites with grassland, heathland and forest. Brodersen et al. (2000) reported the effect of vegetation structure on the stable isotope oxygen–18 composition of rainwater and soil water in the unsaturated zone in southern Germany. However, no references were reported in the stable isotope composition along the soil profiles in semi-humid and semi-arid mountain region in North of China.

Infiltration of precipitation and subsequent downward percolation, affected by soil texture, structure, wetness and water potential, is a complex process. It can be described as piston flow, in which water from more recent precipitation events (the newer water) pushes the older soil water to flow downward (Zimmermann et al. 1966, 1967a, b). Different infiltration profiles of precipitation with individual isotopic composition result in different isotopic profiles. An abrupt ''isotopic front'' within the soil can be observed after an isotopically distinct rainfall event. Therefore, by measuring isotopic compositions of precipitation, soil water, and groundwater, it is possible to identify the infiltration of precipitation and the recharge depth to groundwater by calculating the movement of isotopic peak along the soil profiles.

Tritium has been widely used to trace temporal variation of soil water movement, retention time and recharge to subsurface groundwater, especially in the unsaturated soil and arid environment (Allison and Hughes 1972; Bengtsson et al. 1987; Lin and Wei 2006). However, tritium cannot be taken as a tracer in humid environment because it is hard to observe its peaks along the soil profile in which the tritium changes frequently with the infiltration of precipitation.

The objective of this study is trying to apply stable isotope (oxygen–18 and deuterium) to investigate the recharge processes including precipitation, soil water and groundwater in slope covered by representative vegetation, and to use stable isotope as tracers for calculating the recharge of soil water to groundwater in semi-arid and semi-humid Taihang Mt. region, North China.

## Site description

The experiment was conducted in Niujiazhuang Catchment (area of 9.3 $km^2$) in the middle part of Taihang Mt., situated around 50 km southwest of Shijiazhuang, capital city of Hebei Province, China (Fig. 1). In the catchment, there is the Hilly Ecosystem Experimental Station in Taihang Mt. (114°15′E, 37°52′N, altitude 350 m above sea level), Chinese Academy of Sciences. The study area has a temperate climate with the average annual air temperature of about 13°C, and the maximum and minimum monthly temperature are usually observed in July and January with mean values of 26.3°C and –1.6°C, respectively. The annual pan evaporation is 1,934 mm/year. In the study area, it can be divided visibly into dry (from October to June) and rainy (from July to September) season (Fig. 2). Annual precipitation ranges from 390 to 750 mm, with 67.8% of that available between June and September. The vegetation coverage in the catchment is a mosaic of grassland, deciduous and coniferous mountainous forests, plantation and agricultural crops. The dominant stand species are xerophyte and semi-xerophyte grass (Themeda triandra, Bothriochloa ischaemum), shrubs (black locust, Ziziphus jujube, vitex) and planted vegetation such as persimmon (D. *Kaki* L.f.), apricot (P. *armeniaca* var. *ansu* Maxim), Chinese jujube (Z. *jujube Mill.*). Soils in this area generally belong to the mountainous cinnamon soil. Soil profiles, with depth of 20–120 cm predominantly, are generally poorly stratified with thin O horizon (0–2 cm in thickness) and poorly developed A (3–10 cm) and B horizons (20–70 cm), and contain abundant rock chips. No runoff was observed during the experimental period in the small catchment located at the west of the Hilly Ecosystem Experimental Station (see Fig. 1).

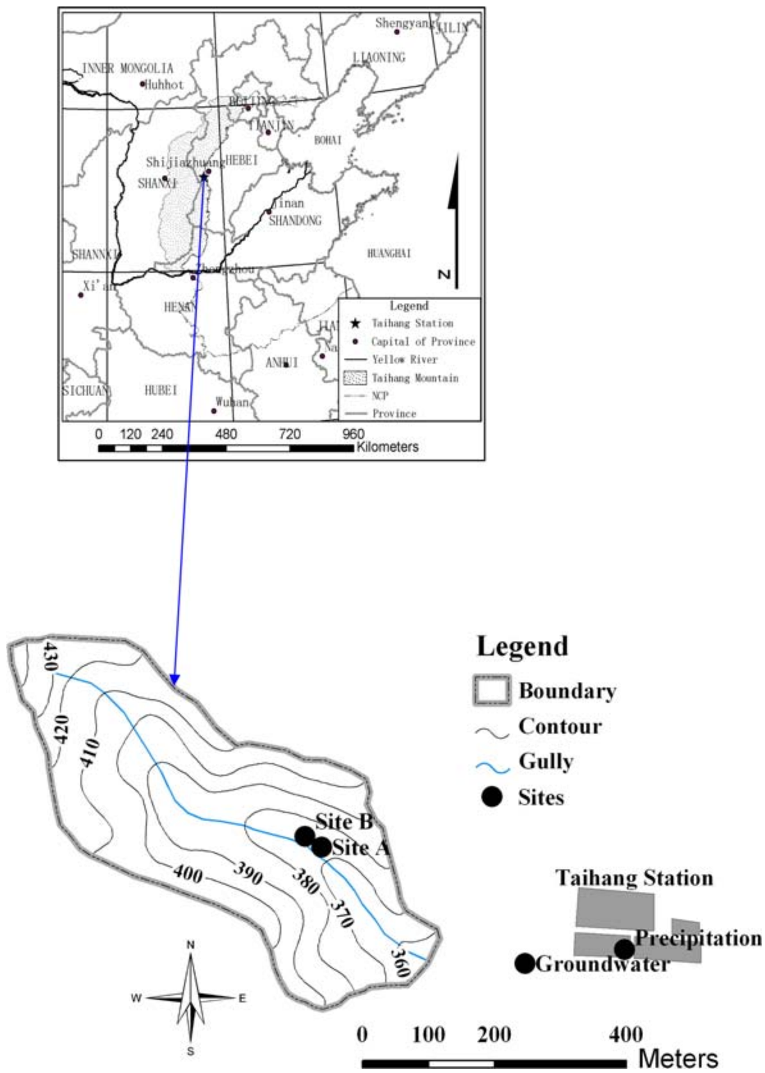


**Fig. 1** Location of the Spring catchment at the west of Taihang Station

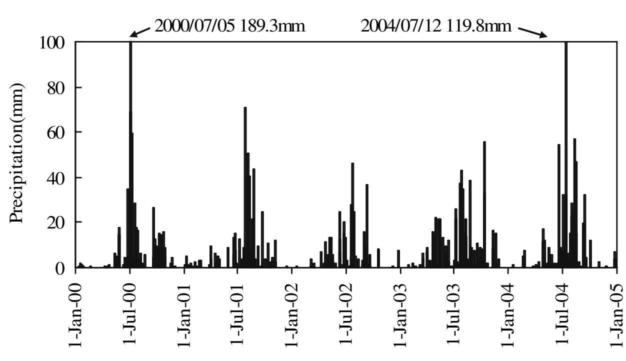


**Fig. 2** Daily precipitation in Taihang Mt. from 2000 to 2004. It is obvious that the rainy season is from July to September and the over 20 mm rainfall events are limited

## Methods

The experimental sites, Site A covered by robinia (*Robinia Pseudoacia* L.), and Site B grass-predominated by Themeda triandra (T. *japonica* (Willd.) Tanaka) and Bothriochloa ischaemum (B. *ischaemum* (L.) Keng), were in the small catchment located at the west of station (Fig. 1). The precipitation samples were collected by a 500 ml vial through a 200 mm diameter funnel in which a pingpong ball was set to avoid evaporation. The rainfall samples were gathered into two polyethylene air-tight vials of 100 ml as soon as possible after a rain event. After sampling, all vials were cleaned with distilled water and reset for next event.

The precipitation was measured by rain gauge (Rain Collector II, Davis Instruments Corp., USA, with accuracy of 0.25 mm) and recorded by event logger (HOBO, Onset Comp. Corp., USA).

At the experimental sites, suction lysimeters were at 10, 20, 30, 50, 70, 90, and 110 cm of depths. The suction lysimeters were constructed of Teflon pipe with a porous ceramic cup and installed at the bottom of 5-cm diameter vertical holes excavated with a hand-operated bucket auger and backfilled with the excavated material. Soil water was collected by applying a vacuum of about –0.8 MPa to the suction lysimeter. By this method, 10–500 ml of water samples can be taken with little effects on Oxygen–18 and deuterium composition (Swistock et al. 1989).

Soil water potential was measured by tensiometer set at the depths as same as suction lysimeters. The data were recorded manually every day with precision of 1 mmHg.

The groundwater was sampled every 5 days. It was taken additionally if the precipitation was more than 20 mm and lower than 50 mm. When it was more than 50 mm, groundwater would be taken everyday in the following 4 days. All samples were stored in 100 ml bottles and sealed with adhesive tape to prevent evaporation.

The oxygen–18 and deuterium compositions were measured with a mass spectrometer Finnigan Mat 253 (Finnigan, USA) and are expressed as $\delta$-values in parts per million relative to the international standard, VSMOW. All the stable isotope measurements were conducted in TC/EA method at the Stable Isotope Laboratory of the IGSNRR, CAS. The reproducibility of the $\delta$D values and $\delta^{18}$O values are ±2 and ±0.1‰, respectively.

## Results

General characteristics of isotopic composition in precipitation, soil water and groundwater are shown in Table 1. They are described as following:

### Isotopic composition in precipitation

The relationship of deuterium composition ($\delta$D) and oxygen–18 composition ($\delta^{18}$O) based on 22 samples of rain is shown in Fig. 3. On this plot, the data define a clear linear trend that represents the local meteoric water line for Taihang Mt. (LMWL) region during the rainy season (from June to September). The slope (7.6) and intercept (4.5) of the LMWL are both slightly lower than that of the global meteoric water line (GMWL, 8 and 10; Craig 1961). At Taihang station, the $\delta^{18}$O value of precipitation ranged from –4.3 to –13.3‰, with arithmetic mean value of –8.1‰ and standard deviation of 2.3‰, the deuterium ranged from –26 to –99‰, –57‰ and standard deviation of 18‰, respectively. The seasonal variation of $\delta^{18}$O and $\delta$D in precipitation is shown in Fig. 4. It can be found that the isotopic values are lower on 19 June and after 14 September, whereas they are higher from July to August in general. The less negative isotopic values were also observed in heavy rain event (13 July) and after a long rainy duration (23 July) when the precipitation became isotopically depleted.

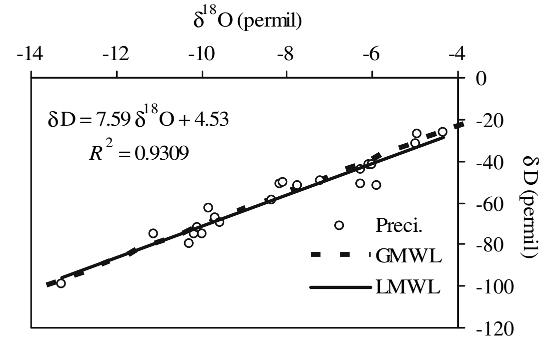


**Fig. 3** $\delta$D–$\delta^{18}$O relationship from Taihang station precipitation samples with regression lines (*solid line*) and regression equation. The global meteoric water line (GMWL; *dashed line*) is also indicated for reference

### Isotopic composition in soil water

Figure 5 shows the stable isotope relationship based on 125 samples from 14 rainfall events. The results indicate that all isotope dots of soil water fall on or along the right-side of the LMWL. Most of oxygen-18 compositions were higher than –8‰ at 10 cm, whereas those at the depths

**Table 1** General characteristics of isotopic composition in precipitation, soil water and groundwater at Taihang Mt.

| Types | Depth | Sample amount | $\delta^{18}$O (‰l) | | | $\delta$D (‰) | | |
|---|---|---|---|---|---|---|---|---|
| | | | Max. | Min. | Mean | Max. | Min. | Mean |
| Precipitation | | 22 | –4.3 | –13.3 | –8.1 | –26 | –99 | –57 |
| Soil water | 10 | 25 | –5.4 | –10.6 | –7.6 | –49 | –84 | –60 |
| | 20 | 27 | –3.6 | –10.3 | –8.1 | –33 | –89 | –64 |
| | 30 | 25 | –7.3 | –10.6 | –8.9 | –48 | –83 | –71 |
| | 50 | 19 | –7.1 | –10.4 | –8.9 | –52 | –84 | –70 |
| | 70 | 28 | –7.4 | –9.8 | –8.3 | –57 | –73 | –65 |
| | 90 | 11 | –6.4 | –7.7 | –7.3 | –51 | –60 | –58 |
| | 110 | 26 | –6.9 | –8.2 | –7.7 | –54 | –74 | –61 |
| Groundwater | | 24 | –8.1 | –13.4 | –9.3 | –57 | –87 | –67 |

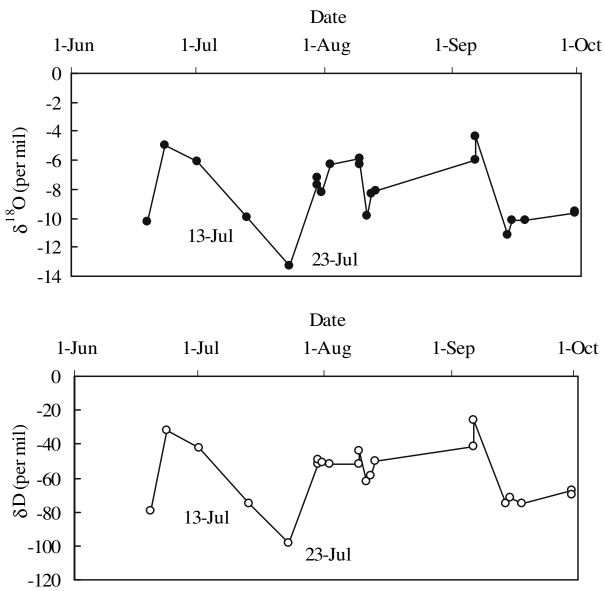


**Fig. 4** Seasonal variation of isotopic composition in precipitation

from 30 to 70 cm were lighter than –7.5‰. Generally, the values of $\delta^{18}$O and $\delta$D in soil water decrease with depth at the top layer above 30 cm at both sites. These trends were also reported in other arid and semiarid regions (e.g., Zimmermann et al. 1967a; e.g., Barnes and Allison 1983; Shurbaji and Phillips 1995), because the vapor-dominant transport is near the surface controlling the isotopic composition. In contrast, the trends below 30 cm in depth isotopically enrich with the depth. The variations of isotopic values at depths of 90 cm in Site B and 110 cm in Site A were small. The isotopic compositions at both sites showed that standard deviations declined in the vertical profiles from surface (Fig. 6). It was considered as more effects of evaporation on the topsoil, that was similar at low rainfall sites in Hawaii (Hsieh et al. 1998). The change range of $\delta^{18}$O decreased from 4.9 and 6.5‰ at 10 and 20 cm to 1.3‰ at 110 cm in Site A. Those data in Site B declined from 4.6 and 3.4‰ to 1.3‰ at 90 cm. The $\delta$D values had the similar trend as $\delta^{18}$O in the two sites.

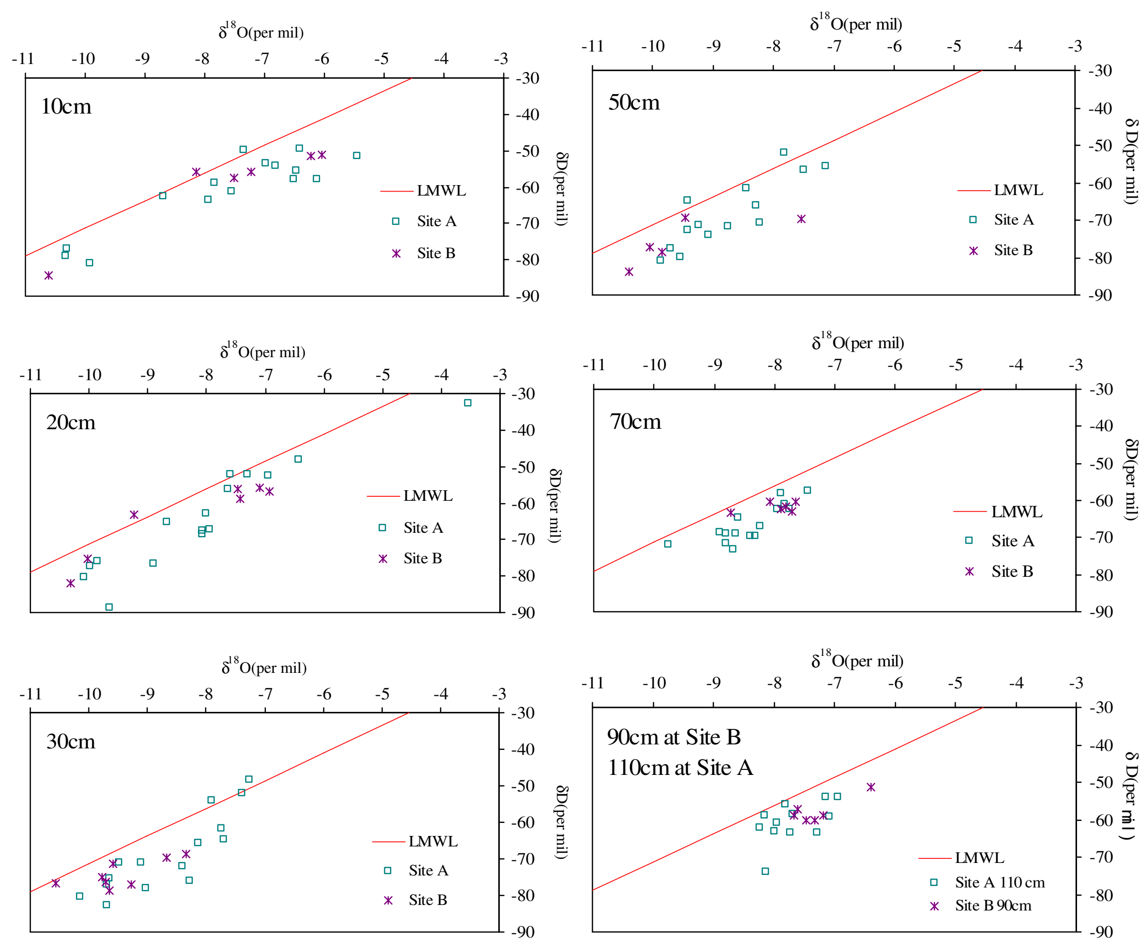


**Fig. 5** $\delta$D–$\delta^{18}$O relationship of soil water at different depths of Site A and Site B

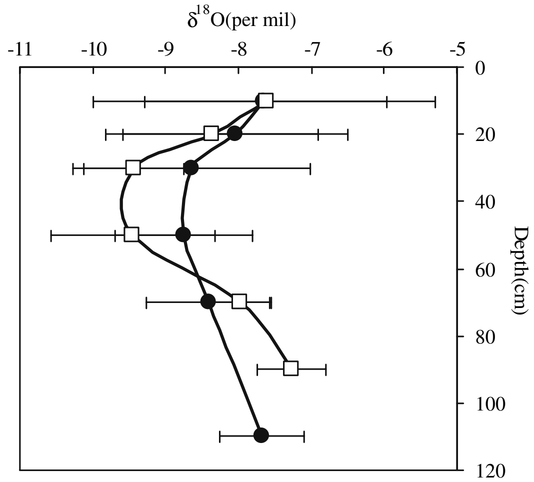


**Fig. 6** Variation of $\delta^{18}$O along soil profile at Site A (*square*) and Site B (*solid-circle*). One standard deviation is represented by the bars

Isotopic composition in groundwater

In order to explore the relationship between soil water and groundwater, the groundwater was sampled after rain stop. The data were plot in Fig. 7. All values of $\delta^{18}$O were lighter than –8‰. Its $\delta^{18}$O and $\delta$D values fluctuated slightly with a mean of –9.3 and –67‰, standard deviation of 1.3 and 6.7, and variation range of 5.3 and 30‰, respectively.

## Discussion

Evaporation and isotope in soil water

Precipitation is the only source for soil water in Taihang Mt. region. Evaporation is the major factor for enrichment of isotopes at topsoil. Moreover, the isotope composition is also influenced by its composition in precipitation. By comparing soil water isotope composition at 10 and at 20 cm in Fig. 5, the trend of isotope values suggests that the soil water originated from precipitation and the fractionation controlled by evaporation during the infiltration caused the $\delta^{18}$O and $\delta$D departing to the right-side of LMWL. In the vertical profile of isotopes in soil water, the most enriched values before the rainy season were found at the depths of 10 and 20 cm. However, the most $^{18}$O-depleted water also occurs at the top layer (Fig. 5) after isotope-depleted precipitation during the rainy season. It is further evidence that the infiltration of precipitation can be traced by changes of stable isotope values with depth. At 10 cm of the Site B, the evaporation trend line ($\delta$D = 2.4 $\delta^{18}$O–36.9, $R^2$ = 0.84) can be plotted clearly by the data from 31 July to 6 August when a series of rain events with lighter and lighter amount. The similar evaporation trend line cannot be found distinctly in Site A. Many previous researches indicated that the surface cover of soil could have dramatic effects on the soil physical environment near the soil surface, especially in depths from top to 40 cm (Chung and Horton 1987; Bristow and Horton 1996). Under cover condition, the net radiation is expected decline significantly and soil surface temperatures decrease. Consequently, cover suppresses the soil water evaporation to a large extent comparing to bare soil (Chung and Horton 1987). Therefore, the difference of evaporation trend line between two sites with distinct canopy covers is likely explained by the influence of cover on stable isotopic composition in top soil water.

Theoretically, the soil water's isotope signal of every rainfall event recorded on the soil layer will not be changed until the soil layer accepted the ''new'' water that had different isotopic composition (Zimmermann et al. 1967b). Figures 5 and 6 show that, along the vertical profile, the change range of $\delta^{18}$O and $\delta$D in soil become narrower and narrower. This phenomenon can be explained by the piston model that the soil water with higher isotopic composition came from accumulation of the upper layer soil water during the heavy rainfall events. In Taihang Mt. region, the amount effect of $\delta^{18}$O, i.e., the $\delta^{18}$O becomes lighter as the amount of rainfall increases, can be observed if the continuous rain events are excluded (Fig. 8). In the continuous rain events, rain-out of heavy isotopes in precipitation lead to the isotopic depletion, and usually, the amount of rainfall becomes lesser and lesser.

Figures 3 and 5 show that the isotopic compositions of soil water at 10 and 20 cm responded well to the precipi-

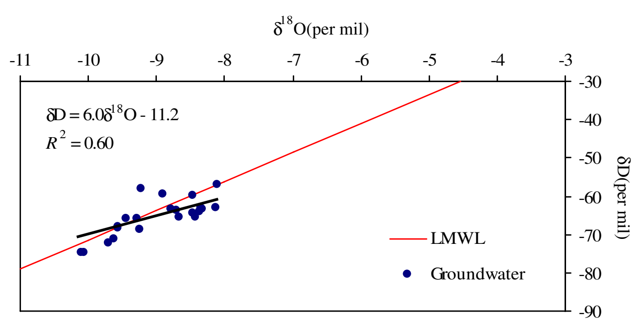


**Fig. 7** $\delta$D–$\delta^{18}$O relationship in groundwater

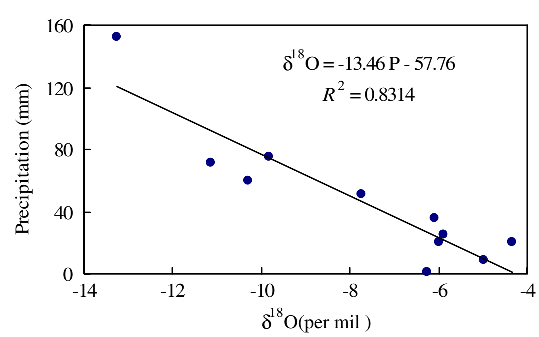


**Fig. 8** Amount effect of $\delta^{18}$O in precipitation

tation. From 30 to 110 cm in depth, all isotope values increased along the profiles (Fig. 5) in two sites. It indicates that the fractionation of soil water caused by evaporate effect declines up to 90 cm in Site B and 110 cm in Site A. It is different from the results reported by Newman et al. (1997). They indicated the evaporation effect was mainly to the upper 10 cm of forest soil in north New Mexico. Here, the results show that the vegetation has a weak effect on soil water isotope. It is the same as the previous results reported by Brodersen et al. (2000). It also infers that the impact of vegetation is limited to decrease the soil water evaporation below 30 cm. The data suggest that the evaporation should be paid more attention in assessment of mountainous region water resources.

According to the piston model (Zimmermann et al. 1967b), the soil water extracted at the beginning of rainy season recording the isotope information of precipitation on current event and the isotope signal in previous event, whose isotopic composition was highly fractionated by evaporate effect during the whole winter with a slight rainfall and snow. For instance, the $\delta$D and $\delta^{18}$O in soil water at depth of 10 cm were very light (–80 and –9.9‰) and were similar to the precipitation values of –79 and –10.3‰ on June 23. However, the values were –3.5 and –33‰ for $\delta^{18}$O and $\delta$D in 20 cm depths, respectively. On July 14 and 16, $\delta^{18}$O and $\delta$D values in top soil were controlled by the isotopic composition of precipitation, and they declined with the rainfall event with light fluctuation and slight increase at the end of September in Site A. Those of Site B at 70 cm and at 90 cm had the same trend as Site A. The changes at upper layer (0–30 cm) were determined by their current isotopic composition and evaporate fractionation with positive variation.

## Movement of soil water traced by stable isotope

July to September is the rainy season with few events more than 20 mm (Fig. 2), during which the daily maximum temperature can be higher than 35°C in Taihang Mt. but the annual mean temperature is just 13.5°C. Plants grow fast during this period. These cause the high evapotranspirative demand. The topsoil water is expected to decrease to lower than 7 wt% if no rain is available in 1 week.

Consequently, the vertical distributions of soil water potential changed greatly (Fig. 9). Similarly, the $\delta^{18}$O and $\delta$D values of soil water also greatly varied in the same period. Figure 9 shows that, from July to September, soil water potential changed from positive values to very negative values along the soil profiles. Furthermore, the most positive soil water potential corresponded with the most negative $\delta^{18}$O values, and vice versa. The most negative soil water potential and the less negative $\delta^{18}$O values were found on September 20, 2004. Total amount of 120 mm rainfall, the heaviest rainfall event in the year, was recorded

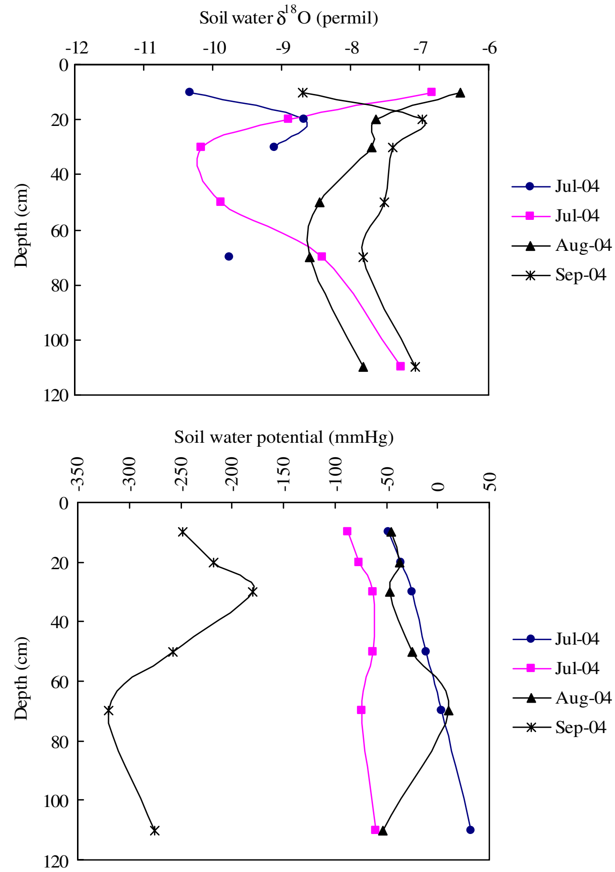


**Fig. 9** Soil water $\delta^{18}$O and potential profiles at Site A in 2004

on 12 July, 2004, with its isotopic composition of –10.0‰ for $\delta^{18}$O and –75‰ for $\delta$D, respectively. Additionally, the soil water at the depth of 10 cm was isotopically depleted as –10.3‰ for $\delta^{18}$O, lighter than that of precipitation. The soil water potential increased with increasing depth this day. Figure 9 shows the isotope profile has similar indication as the soil water potential. However, the soil water potential cannot sustain long time due to the extreme evapotranspirative demand for vegetation. The most negative soil water potential profiles were observed at the end of September. Although the soil water potential on September 20 indicated there was a positive gradient from 110 to 70 cm in depths, it was impossible for soil water to move upward. Correspondingly, no isotopic fractionation would occur below 70 cm. After October, the temperature was lower than 0°C and the soil water below depths of 50 cm tended to constant about 5 wt% in this region (Zhang et al. 2005). Hence, it is reasonable assumption that the isotopic composition will not change below 70 cm in depth after September.

## Infiltration and residue time of soil water

The soil water movement in the dry soils is controlled by both vapor-dominant transport near the surface and liquiddominant transport below the evaporation front (Walker et al. 1988) indicated by the $\delta^{18}O$ and $\delta D$ maximum bulge or peak. Using this maximum peak as an indicator, many researchers represented the soil water movement in dry region (Barnes and Allison 1983; Barnes and Walker 1989; Mathieu and Bariac 1996). In Taihang Mt., it has a long dry season from October to May in next year during the period, the rainfall and snow are small and the wind is strong. It causes the soil water at topsoil to decrease near to 5 wt% and its isotopic compositions enrich. Therefore, the heavy isotopic soil water will form at top soil during dry season, which gives an index to identify the soil water movement in the study area. Assuming that the effective rain in the period from June 23 and September 20, 2004 infiltrated into the soil can be described as a piston flow, the soil water reside time and recharge depth could be calculated approximately by considering the peak of $\delta^{18}O$.

Movement of isotope peak in soil vertical profile for each event during the rainy season was showed in Fig. 10. The $\delta^{18}O$ and $\delta D$ values in the first heavy precipitation with amount of 59 mm on June 19 were of –10.3 and –75‰, respectively. Consequently, the $\delta^{18}O$ values in soil water sampled on June 23 was found as –9.9 and –3.5‰ at depths of 10 and 20 cm, respectively. Apparently, soil water at depths of 10 cm recharged by precipitation with slight isotopic enrichment. However, the isotopic value at 20 cm did not deplete with that of precipitation. It suggests that the soil water in top layer with strongly isotopical enrichment caused by extreme evapotranspiration from last October to May was pushed downward this layer by piston-flow type. As the second heavy rain was in July 12 with precipitation of 120 mm, the $\delta^{18}O$ vertical profiles of soil water did not move downward because the sampled time was before the heavy rain stopped. Therefore, the isotopic peak still kept at depth of 20 cm. Variations of soil water

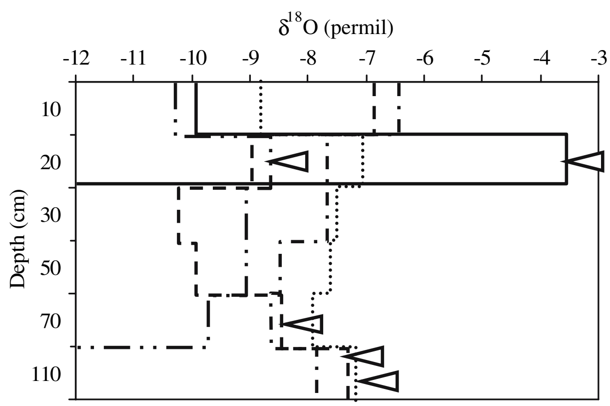


**Fig. 10** Movement of isotope peak along soil profile at Site A (–Jun. 23, -···Jul. 12, - - - Jul. 31, -·-Aug. 10, ......Sep. 20, their peaks located at 20, 20, 70, 110, and 110 cm, respectively, and the *triangle* indicates the peak position)

and evaporation in July and August were large. The heavy rains from July 13 to 25 account for the isotopic peak moving downward to 70 cm on July 31. The data determined on August 10 and on September 20 indicate that the isotopic peak arrived at a soil of 110 cm in depth, where (lower than 90 cm) exhibited a trend to constant for isotopes from the previous discussion. However, the isotopic variation in upper layer was still large because of evaporation during the shiny days and isotopic differences for each rain event. The $\delta^{18}O$ of soil water at a depth of 110 cm on September 20 was –7.1, 0.7‰ heavier than that of August 10.

Inasmuch as it is dry season from October to May with a few rains, and the evapotranspiration is extremely strong, the soil water potential and its isotope compositions corresponds to precipitation, and there are no driven forces strong enough to cause the soil water moving downward, then the incidental isotopic composition to enrich in soil below 100 cm in depth. Therefore, soil water movement in Taihang Mt. at both Site A and Site B was approximately 110 cm in 2004 in terms of piston-flow model and tracer evaluation technique described by Bengtsson et al. (1987). The soil water residence times from top to 90 cm and 110 cm in depth for Site B and Site A in 2004 were 9 months since the precipitation in September last year was similar to that of 2004. In rainy season (from June 19 with first heavy rain of 59 mm to September 20 with last heavy rain more than 30 mm), the mean depths of soil water movement are 12 and 10 mm/day for the water infiltrated into Site A and Site B, respectively.

## Recharge of soil water into groundwater

The isotopic compositions of groundwater and soil water in Site A and Site B strongly suggest that there are near relationship between the local groundwater system and soil water at 110 cm depth for Site A and 90 cm for Site B (Fig. 11). The Figs. 5 and 7 show that the isotopic compositions of groundwater are close to those of soil water at 50 cm in depth. The results indicate that those parts of water infiltrate into the depth of 50 cm will recharge possibly into the local groundwater system. However, the amount of recharge is decided by soil saturation status and water balance of soil mass between 50 cm to zero flux plane (ZFP) where the soil water will escape the evaporation effects and eventually reach to the groundwater table. No recharge will yield in unsaturated soil mass. The isotope differences between groundwater and soil water at 90 cm of Site B and 110 cm of Site A indicate that the evaporate effects in these depths becomes weak. The narrow range of $\delta D$ and $\delta^{18}O$ and the ZFP located at 70 cm in depth on September 20 supported the assumption that the soil water reaching the soil below the depths of 90 cm in

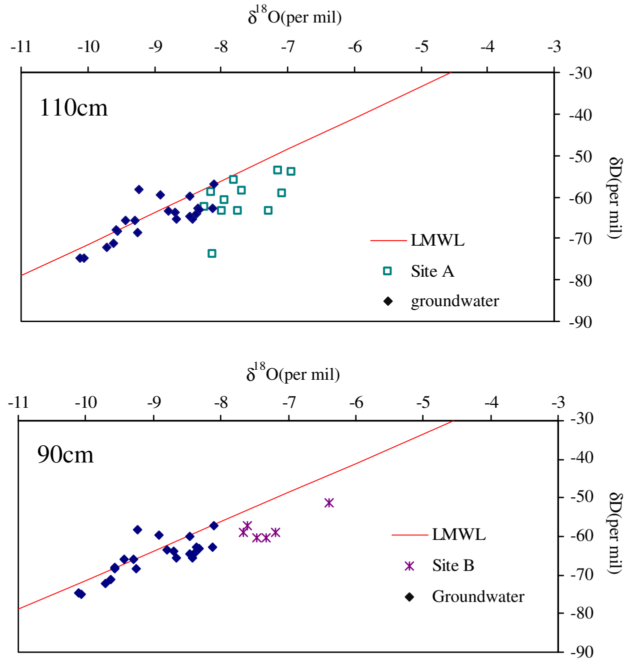


**Fig. 11** Relationship between deep soil water and groundwater

Site B and 110 cm in Site A can recharge local groundwater. Below ZFP, the isotope fractionation becomes very weak and can be ignored. Therefore, it is reasonable to assume that the soil water passed through the soil layer at depths of 90 cm in Site B and 110 cm in Site A will recharge to groundwater. The antecedent soil water contents from top to 110 cm in Site A and 90 cm in Site B were determined by 36.5 wt% in saturated status and those at the end of dry season was 5 wt%. Thus, the annual average recharge flux to groundwater was 3.8 and 3.2 mm/day from June to September in 2004 in Site A and Site B, respectively.

## Conclusions

This paper gives the local meteoric water line in Taihang Mt. region. The slope and intercept of the LMWL are both lower than that of GMWL indicates that the evaporation is stronger than the mean value in the global.

Comparison of $\delta^{18}$O and $\delta$D profiles for Site A covered by black locust and Site B predominated by Themeda triandra and Bothriochloa ischaemum in semi-arid and semi-humid Taihang Mt. region shows that the $\delta^{18}$O and $\delta$D in top soil water is strongly dependent on those values of precipitation and isotopically enriched by evaporation effect. The different canopy covers between two sites just influence the stable isotopic composition in top soil water in 10 cm.

A long dry season from last October to May with slight rainfall, stronger wind and heavy evaporation makes it possible to trace the temporal vertical movement of soil water by stable isotope oxygen-18, instead of traditional tritium. Consequently, the depths of soil water movement in 2004 are 110 cm in Site A and 90 cm in Site B with a moving ratio of 12 and 10 mm/day, respectively, during the rainy season from June to September in Taihang Mt. The soil water moved over the zero flux plane will escape the evaporative effect below 90 cm in Site B and 110 cm in Site A and eventually recharge into local groundwater system with recharge flux of 3.2 and 3.8 mm/day, respectively.

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## References

Allison GB, Barnes CJ (1983) Estimation of evaporation from non-vegetated surfaces using natural deuterium. Nature 301:143–145

Allison GB, Hughes MW (1972) Comparison of recharge to groundwater under pasture and forest using environmental tritium. J Hydrol 17:81–95

Allison GB, Barnes CJ, Hughes MW, Leaney FWJ (1984) The effect of climate and vegetation on oxygen-18 and deuterium profiles in soils Isotope Hydrology 1983. In: Proceedings of the Symposium IAEA, Vienna, pp 105–123

Barnes CJ, Allison GB (1983) The distribution of deuterium and $^{18}$O in dry soils: 1. Theory. J Hydrol 60:141–156

Barnes CJ, Allison GB (1984) The distribution of deuterium and $^{18}$O in dry soils: 3. Theory for non-isothermal water movement. J Hydrol 74:119–135

Barnes CJ, Walker GR (1989) The distribution of deuterium and oxygen-18 during unsteady evaporation from a dry soil. J Hydrol 112:55–67

Bengtsson L, Saxena RK, Dressie Z (1987) Soil water movement estimated from isotope tracers. Hydrol Sci J 32:497–520

Bristow KL, Horton R (1996) Modeling the impact of partial surface mulch on soil heat and water flow. Theor Appl Climatol 54:85–98

Brodersen C, Pohl S, Lindenlaub M, Leibundgut C, Wilpert KV (2000) Influence of vegetation structure on isotope content of throughfall and soil water. Hydrol Processes 14:1439–1448

Chung S-O, Horton R (1987) Soil heat and water flow with a partial surface mulch. Water Resour Res 23:2175–2186

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

Gazis C, Feng X (2004) A stable isotope study of soil water: evidence for mixing and preferential flow paths. Geoderma 119:97–111

Gehrels JC, Peeters JEM, De Vries JJ, Dekkers M (1998) Mechanism of soil water movement as inferred from 18O stable isotope studies. Hydrol Sci J 43:579–594

Hsieh JCC, Chadwick OA, Kelly EF, Savin SM (1998) Oxygen isotopic composition of soil water: quantifying evaporation and transpiration. Geoderma 82:269–293

Knowlton RG Jr, Parsons AM, Gaither KN (1992) Techniques for quantifying the recharge rate through unsaturated soils ASTM Special Technical Publication. ASTM, Philadelphia, pp 111–123

Lin R, Wei K (2006) Tritium profiles of pore water in the Chinese loess unsaturated zone: implications for estimation of groundwater recharge. J Hydrol 328:192–199

Liu CM, He XW (1996) Strategies for water problems in the 21st century in China. Scientific Press, Beijing

Liu B, Phillips F, Hoines S, Campbell AR, Sharma P (1995) Water movement in desert soil traced by hydrogen and oxygen isotopes, chloride, and chlorine-36, southern Arizona. J Hydrol 168:91–110

Liu C, Yu J, Kendy E (2001) Groundwater exploitation and its impact on the environment in the North China Plain. Water Int 26:265–272

Mathieu R, Bariac T (1996) An isotopic study ($^2$H and $^{18}$O) of water movements in clayey soils under a semiarid climate. Water Resour Res 32:779–789

Midwood AJ, Boutton TW, Watts SE, Archer SR (1993) Natural abundance of $^2$H and $^{18}$O in rainfall, soil moisture and plants in a subtropical thorn woodland ecosystem: implications for plant water use. International Atomic Energy Agency, Proceedings Series. IAEA, Vienna, pp 419–431

Newman BD, Campbell AR, Wilcox BP (1997) Tracer-based studies of soil water movement in semi-arid forests of New Mexico. J Hydrol 196:251–270

Saxena RK, Dressie Z (1984) Estimation of groundwater recharge and moisture movement in sandy formations by tracing natural oxygen-18 and injected tritium profiles in the unsaturated zone. In: Proceedings of the symposium on isotope hydrology in water resource development, IAEA, Vienna, pp 139–150

Shurbaji ARM, Phillips FM (1995) A numerical model for the movement of $H_2O$, $H_2^{18}O$, and $^2HHO$ in the unsaturated zone. J Hydrol 171:125–142

Swistock BR, Dewalle DR, Sharpe WE (1989) Sources of acidic storm flow in an Appalachian headwater stream. Water Resour Res 25:2139–2147

Tang C, Chen J, Shindo S, Sakura Y, Zhang W, Shen Y (2004) Assessment of groundwater contamination by nitrates associated with wastewater irrigation: a case study in Shijiazhuang region, China. Hydrol Processes 18:2303–2312

Walker GR, Hughes MW, Allison GB, Barnes CJ (1988) The movement of isotopes of water during evaporation from a bare soil surface. J Hydrol 97:181–197

Yang Y, Watanabe M, Sakura Y, Tang C, Hayashi S (2002) Groundwater-table and recharge changes in the Piedmont region of Taihang Mountain in Gaocheng City and its relation to agricultural water use. Water SA 28:171–178

Zhang Q, Li F, Ou G, Song X, Zhang W (2005) Response of soil moisture variation to precipitation and mulching measures. J Beijing For Univ 27:37–41

Zimmermann U, Münnich KO, Roether W (1966) Tracers determine movement of soil moisture and evapotranspiration. Science 152:346–347

Zimmermann U, Ehhalt DH, Münnich KO (1967a) Soil water movement and evapotranspiration: changes in the isotopic composition of the water. In: Proceedings of the IAEA symposium on the use of isotopes in hydrology. IAEA, Vienna, 1966, pp 567–584

Zimmermann U, Münnich KO, Roether W (1967b) Downward movement of soil moisture traced by means of hydrogen isotopes. Isotope techniques in the hydrologic cycle. Am Geophys Union Geophys Monogr Ser 11:28–36