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ORIGINAL ARTICLE

Hsueh-Yu Lu
Tsung-Ren Peng
Tsung-Kwei Liu
Chung-Ho Wang
Chih-Chao Huang

Study of stable isotopes for highly deformed aquifers in the Hsinchu-Miaoli area, Taiwan

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H.-Y. Lu (✉)
Department of Earth and Environmental Sciences,
National Chung Cheng University,
168, University Road,
Min-Hsiung Chia-yi, Taiwan
E-mail: seishei@eq.ccu.edu.tw
Tel.: +886-5-2720411
Fax: +886-5-2720807

T.-R. Peng
Department of Soil and Environmental Sciences, National ChungHsing University,
Taichung, Taiwan

T.-K. Liu
Department of Geosciences,
National Taiwan University,
Taipei, Taiwan

C.-H. Wang
Institute of Earth Sciences,
Academia Sinica, Taipei, Taiwan

C.-C. Huang
Central Geological Survey,
Ministry of Economic Affairs,
Taipei, Taiwan

Abstract This study was based on the analysis of isotopic compositions of hydrogen and oxygen in samples from precipitation, groundwater and stream water. In addition, parts of groundwater samples were dated by carbon-14 and tritium. These data are integrated to provide other views of the hydrologic cycle in the Hsinchu-Miaoli groundwater district. The groundwater district is principally composed of Pleistocene and Holocene aquifers. The Pleistocene aquifers are highly deformed by folding and faults into small subdistricts with areas of only tens of square kilometers. These aquifers are exclusively recharged by local precipitation. The Holocene aquifers cover narrow creek valleys, only tens of meters in thickness. The local meteoric water line (LMWL), constructed from rainfall samples in the Hsinchu Science Park, is described by the equation $\delta D = 8.02\delta^{18}O + 10.16$, which agrees with the global meteoric water line. In addition, the precipitation isotopic compositions can be categorized into two distinct end members: typhoon type and monsoon type. The groundwater isotopic compositions are perfectly located on an LMWL and can be considered a mixture of precipitations. Based on the mass balance of isotopic compositions of oxygen and hydrogen, infiltration is more active in the rainy season with depleted isotopic compositions. The amount of infiltration during May–September is roughly estimated to comprise at least 55% of the whole year's recharge. The isotopic compositions of stream water are expressed by a regression equation: $\delta D = 7.61\delta^{18}O + 9.62$, which is similar to the LMWL. Although precipitation isotopic compositions are depleted during summer time, the isotopic compositions contrarily show an enriched trend in the upstream area. This is explained by the opposite altitude effect on isotopic compositions for typhoon-related precipitations.

Keywords Stable isotopes · Groundwater recharge · Pleistocene aquifer · Mid-northern Taiwan

Introduction

Taiwan is an island situated in the Western Pacific Ocean with an area of about 36,000 km^2. Mountain ranges above 1,000 m a.s.l. cover an area of 31% and plains below 100 m a.s.l. have an equal proportion. Therefore, most of the river systems are short and narrow. Under this situation, reservoir construction is the only solution to maintain stable surface water supply. This strategy worked for more than 40 years until the climatic pattern changed dramatically. The area has suffered water shortages during three out of the last five summers. The utilization of groundwater resources was particularly appealing during those times and much interest was shown in investigating and monitoring groundwater resources. This work has been proceeding systematically for years in Taiwan under the national Groundwater Monitoring Network Plan. This current study is a part of that plan.

Oxygen and hydrogen isotopic compositions of water have served for decades as a natural tracer all over the world to characterize the provenance of water mass, including groundwater and surface water. In Taiwan, stable isotope analyses of water mass were performed in the early 1980s (e.g. Shieh et al. 1983; Peng 1995; Wang et al. 2001; Wang and Peng 2001). These studies solved some key points in the hydrological cycle, such as the identification of groundwater recharge areas (e.g. Peng et al. 2002) and seawater intrusion signals (e.g. Kuo and Wang 2001); and this information was determinately used for groundwater resource management and boundary conditions of simulation models.

Stable isotopic compositions of groundwater are principally controlled by the compositions of precipitation, evaporation, mixing of water mass and isotope fractionation with phases in the aquifer (Fontes 1980; Gat 1980, 1981, 1996; Rozanski et al. 1997). These mechanisms are usually of seasonal variation due to monsoon-controlled precipitation and the temperature effect (Gat 1980). Generally, the isotopic ratios of hydrogen and oxygen closely follow those in the local mean annual precipitation (Friedman et al. 1964; Fritz et al. 1987; Ingraham and Taylor 1991). In this study, stable isotopic compositions of groundwater, stream water and precipitation from different seasons are analyzed to discuss the infiltration process in detail.

Study area

The Hsinchu-Miaoli Groundwater District (HMGD) is in mid-northern Taiwan, which is the western flank of the Central Mountain Range uplifted by the collision between the Eurasia Plate and the Philippine Sea Plate. Elevations range from sea level to 500 m within a distance of 12 km and local slopes of up to 40% at the southeast corner. The geological structures orient along a NE–SW direction in the northern part of the study area and smoothly rotate to a NNE–SSW direction in the southern part (Chou 1977). This feature makes the narrowing central zone of HMGD (Fig. 1).

There are two kinds of aquifers in the HMGD: Holocene aquifer and Pleistocene aquifer (Fig. 1). The Holocene aquifers are made up of coarse alluvial deposits with higher hydraulic conductivity ($\sim 10^{-2}$–10^{-3} cm/s) and only cover the river valleys. The sediments are usually tens of meters in width and thinner than 50 m. The Pleistocene aquifers are composed of the Toukoshan Formation, i.e., a semi-consolidated conglomerate, lithic graywacke, subgraywacke and shale with much lower hydraulic conductivity ($\sim 10^{-4}$–10^{-5} cm/s). The Pleistocene aquifers are highly deformed due to strong neo-tectonism and their thickness varies from 100 to 1,000 m. The major basal aquitard in this area is the Pliocene consolidated shale formation that conformably underlies the Pleistocene aquifer and forms the eastern boundary of the HMGD.

The Pleistocene aquifers are not only highly deformed by folding, but are also cut off by faults, mostly strike-slip and thrust as shown in the conceptual model constructed using the section around the Hsinchu Science Park (Fig. 2). Accordingly, the HMGD is divided into small sub-districts with areas of only tens of square kilometers. These groundwater sub-districts are usually composed of folded Pleistocene aquifers in the central part and are unconformably underlain by Holocene aquifers on the boundaries. Since the Pleistocene aquifers are significantly higher in altitude, groundwater preferably infiltrates from the Pleistocene aquifers to the Holocene aquifers and is finally drained by the river system.

There are three major Holocene aquifers in order from north to south. (1) Toucian-Fongshan Alluvial Plain with a thickness of at least 50 m, (2) Jhonggang Alluvial Plain with a thickness of less than 30 m and (3) Houlong Alluvial Plain with a thickness of less than 10 m. Although the volume of Holocene aquifers may not be significant enough to contain considerable amounts of groundwater, the enormous Pleistocene aquifers provide adequate groundwater even though they have lower hydraulic conductivity.

Analytical method

Water samples were collected from stream, precipitation and monitoring wells in the HMGD during 2001–2003 (Fig. 3). The stream samples were taken at 41 locations along 11 creeks in both February 2001 and July 2001, representing dry and wet seasons, respectively. The groundwater samples were pumped from 33 newly constructed monitoring wells just after well completion and performance tests. This guarantees that the isotopic compositions are intact and they would not be altered by any chemical or biological reactions after well construction. The precipitations were continuously collected during January–June 2001 at the Hsinchu Science Park. All samples, including 84 stream samples, 33 groundwater samples and 35 precipitation samples, were analyzed for stable isotopic compositions of oxygen and hydrogen; and 27 groundwater samples were also analyzed for tritium and carbon-14 (C-14) ages.

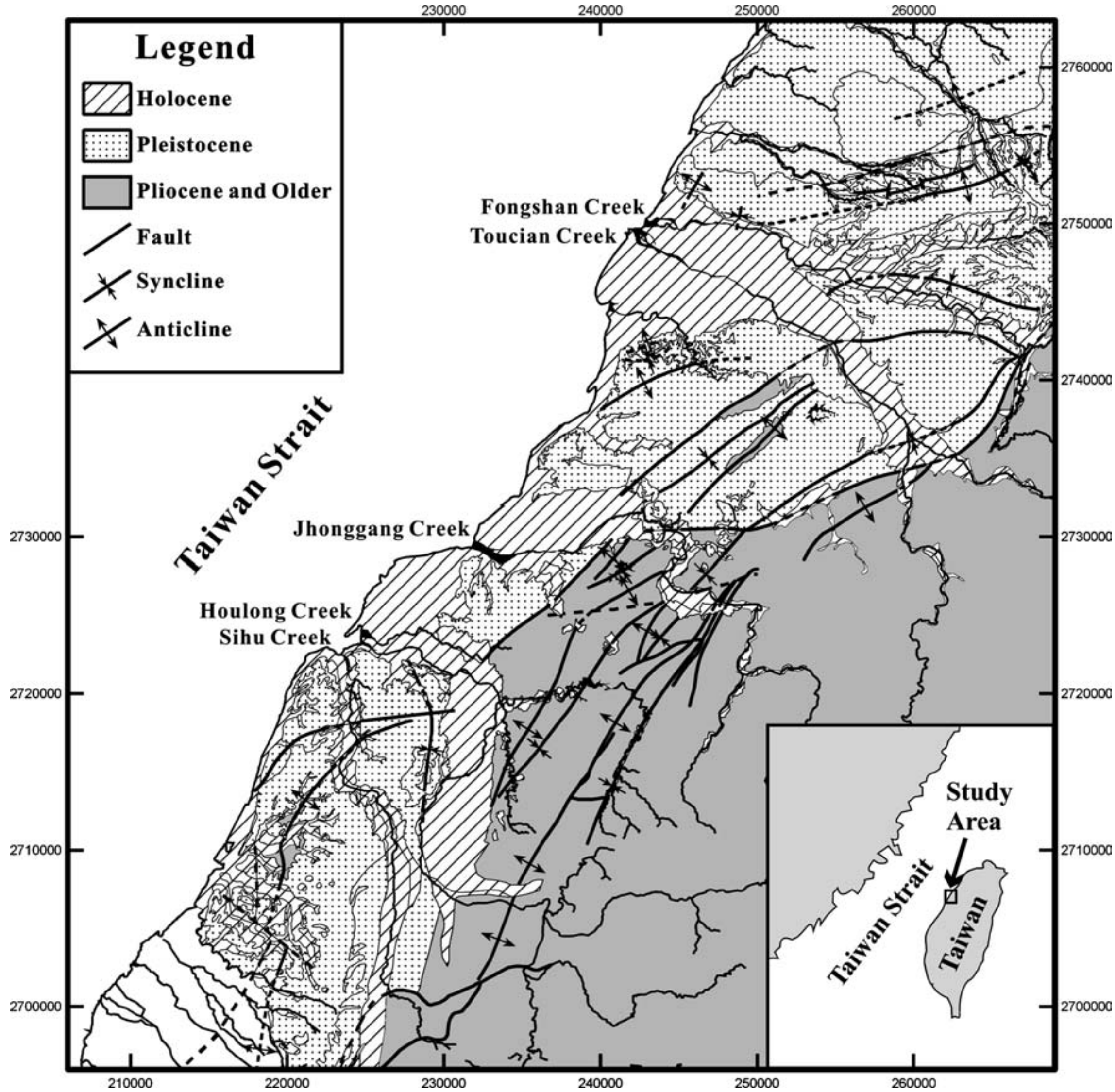

Fig. 1 Hydrogeological map of HMGD showing the distribution of Holocene aquifer, Pleistocene aquifer and basal aquitard. The aquifers are highly deformed by enormous geological structures

These samples were collected in clean, dry, 100-ml pyrex glass bottles and sealed with a Polyseal cap to prevent leakage and evaporation from altering isotopic compositions. Oxygen isotope ratios were analyzed using CO_2–H_2O equilibration (Epstein and Mayeda 1953). The equilibrated CO_2 gas was measured by a VG SIRA 10 isotope ratio mass spectrometer. The isotopic compositions of hydrogen were determined after reduction of water to H_2 using zinc metal supplied by the Biogeochemical Laboratory of Indiana University (Coleman et al. 1982) and then H_2 was measured by a VG MM602D isotope ratio mass spectrometer. All isotopic ratio results are reported using the δ-notation as a per mill (‰) relative to an international Vienna standard mean ocean water standard and normalized on scales such that the $\delta^{18}O$ and δD of standard light Antarctic precipitation are −55.5 and −428‰, respectively. In this study, the analytical errors (1σ) were 0.1‰ for $\delta^{18}O$ and 1.5‰ for δD.

Results and discussion

Isotopic compositions of local precipitation

Figure 4 presents the local meteoric water line (LMWL) $\delta D = 8.02\delta^{18}O + 10.16$ with the square of the correlation coefficient (r^2) of 0.98 for 35 samples collected at the Hsinchu Science Park (Appendix 1). This regression line is quite the same as the global meteoric water line (GMWL) $\delta D = 8\delta^{18}O + 10$ first proposed by Craig (1961) and means that water vapors are all from the

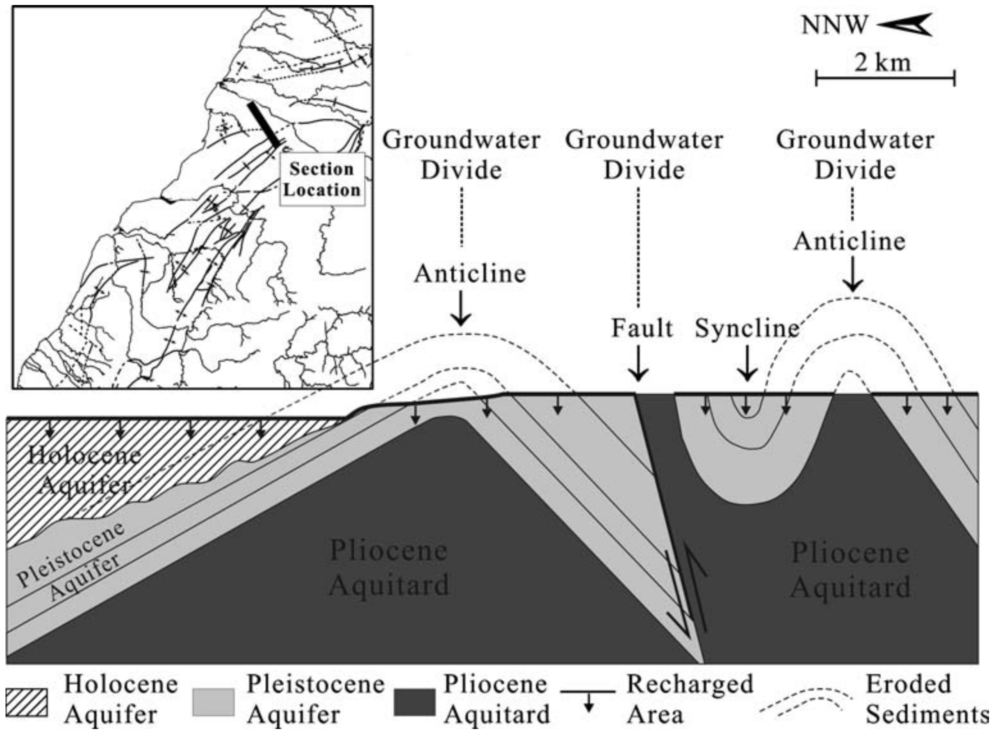

Fig. 2 The section around Hsinchu Science Park represents the conceptual model of groundwater system in HMGD. The section is only roughly on scale in horizontal but not on scale in vertical

open sea and there is no effective evaporation during precipitation (Craig et al. 1963). The stable isotopic compositions show two separated end members, depleted and enriched. The depleted isotopes were apparently precipitated by typhoon events or typhoon-related storms with higher rainfall intensity, whereas the enriched ones represent monsoon precipitations with lower rainfall intensity. There is only one exception for precipitation with the daily rainfall depth of 31 mm, which has −45 and −6.6‰ for δD and δ^{18}O. In Taiwan, a distinct typhoon season occurs from June to October, sharply peaking from August through September. An average of three to four typhoons pound Taiwan each year.

Precipitations from July to December 2003 were not collected in this study. In HMGD, June–August are the warmest months while January–February are the coldest. Although the warm season sample (July–August) was absent, its weather conditions are very similar to June, as are the isotopic compositions. Accordingly, the sampling period is sufficient to conclude that the isotopic compositions are separated into depleted typhoon and enriched monsoon types.

Isotopic compositions of groundwater

As shown in Fig. 4, most of the stable isotopic compositions of groundwater are nearly right on the LMWL (Appendix 2). Evaporation did not make considerable fractionation between δ^{18}O and δD during infiltration. Otherwise, the isotopic compositions would evolve off the LMWL to another line with gentle slope and not just locate on the LMWL. Obviously, the groundwaters suffer evaporation during infiltration, as the mean annual precipitation and evaporation are 1,620 and 740 mm (measured by 120 cm evaporation pan), respectively, during 2001–2004. Under these constraints, the groundwaters should be involved with evaporation processes under high humidity, and the δ^{18}O and δD values would evolve to the heavier side almost along the GMWL during infiltration. The original isotopic ratios may have to be more depleted before this profound evaporation occurs.

In addition, 27 groundwater samples were analyzed with tritium and C-14 by Liu (2001). The tritium measurements were made by liquid scintillation counting with an estimated detection limit of about 0.1 TU. To speak conservatively, groundwater having tritium concentrations greater than 1.0 TU is interpreted as water mixed with modern water or recharged after 1952 when tritium was released into the atmosphere as a result of the atmospheric testing of nuclear weapons. Therefore, tritium concentration is an excellent marker of groundwater age. If the precipitation and/or dissolution of some minerals in the aquifer were involved in changing stable isotopic compositions of groundwater, it may roughly show some relationship between isotopic ratios and the tritium concentration. This conclusion is also applicable to the C-14 age of groundwater. It should be

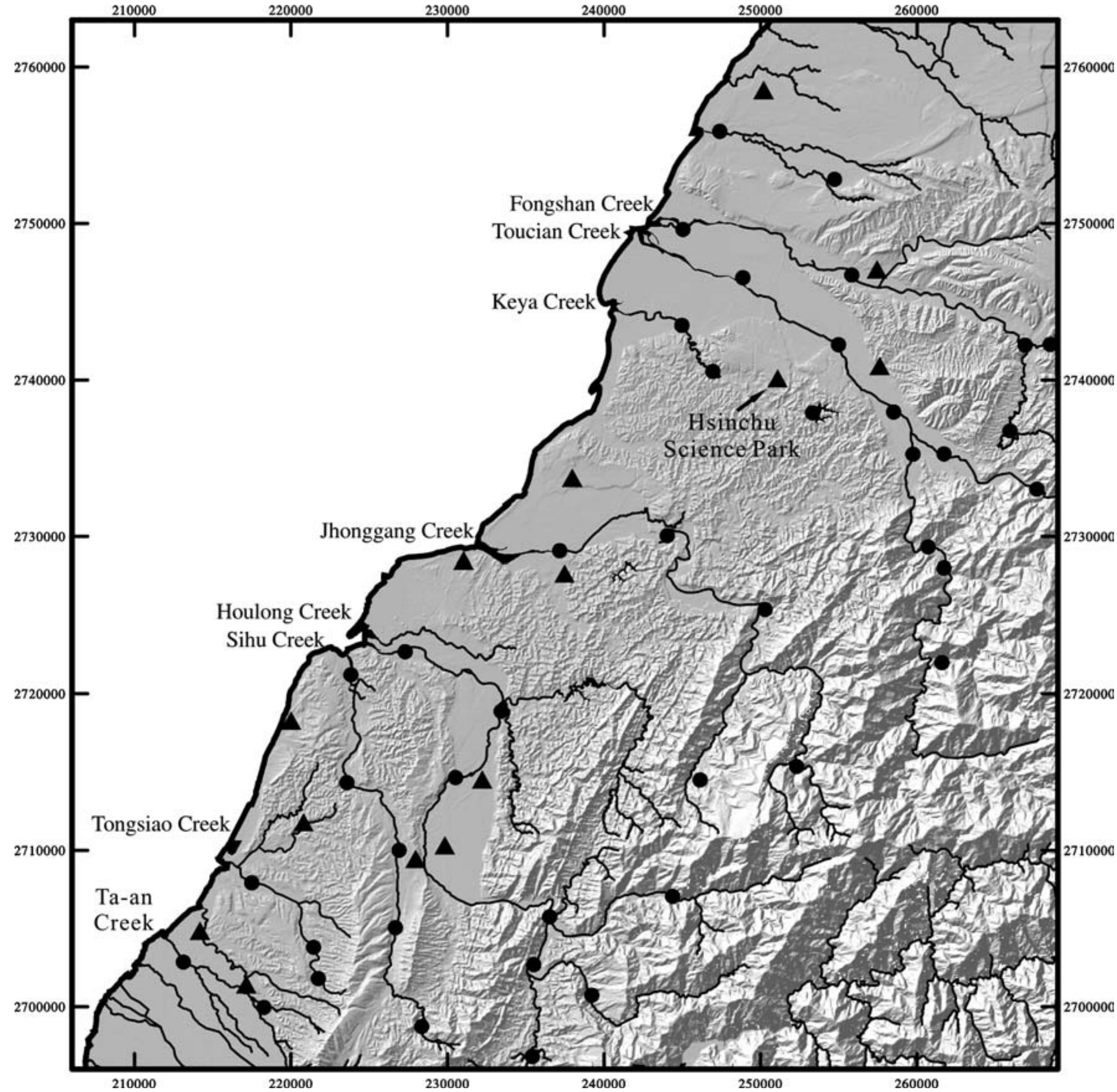

Fig. 3 Sampling sites of groundwater (*triangle*) and stream water (*circle*) samples. Groundwater samples from different monitoring depths were also collected from each site

noted that the C-14 ages reported in this study were not corrected using $\delta^{13}C$. It is usually used to overcome the contamination of carbonate in C-14 age dating because the carbonate can simultaneously provide dead carbon and different stable isotopic compositions. Figure 5 shows that the isotopic compositions of groundwater do not have a positive relationship with tritium and C-14 age, and confirms that the changes of δD and $\delta^{18}O$ from precipitation and dissolution of sand-and-gravel aquifer minerals can be ignored in the HMGD.

Table 1 shows the statistic information of the isotopic compositions for groundwater samples. The standard deviations of δD and $\delta^{18}O$ are 3.87 and 0.44‰, respectively, which are equivalent to 9.5 and 6.7% of the mean values for 33 samples. These data are distributed in a narrow range which locates nearly on the LMWL and between two precipitation types in the δD–$\delta^{18}O$ diagram (Fig. 4). In addition, according to the conceptual model of the groundwater system in HMGD (Fig. 2), the groundwater from higher elevations is not able to recharge the downstream aquifer where it is isolated by geological structures. The stream flow does not recharge the aquifer through riverbanks because the groundwater table is much higher than the adjacent river level. Consequently, it is plausible to use a two-end-member mixing model to describe the relationship between precipitation and groundwater. The groundwater isotopic compositions are very close to the long-term weighted average of the local meteoric water within the range of 5×5 km^2. This phenomenon was also reported by Darling and Bath (1988) as attenuation of the seasonal variations for isotopes while groundwater

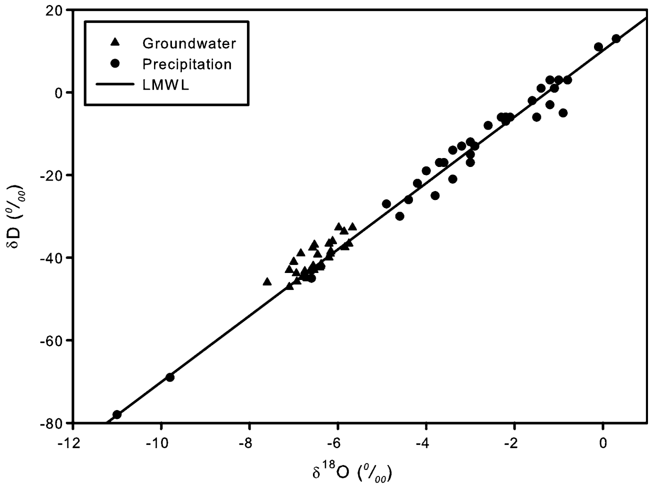

Fig. 4 Plot of δD versus δ^{18}O of precipitation samples (*solid circle*) and groundwater samples (*solid triangle*). The local meteoric water line is determined as $\delta D = 8.02\delta^{18}O + 10.16$ ($r^2 = 0.98$) from 35 samples collected from the Hsinchu Science Park

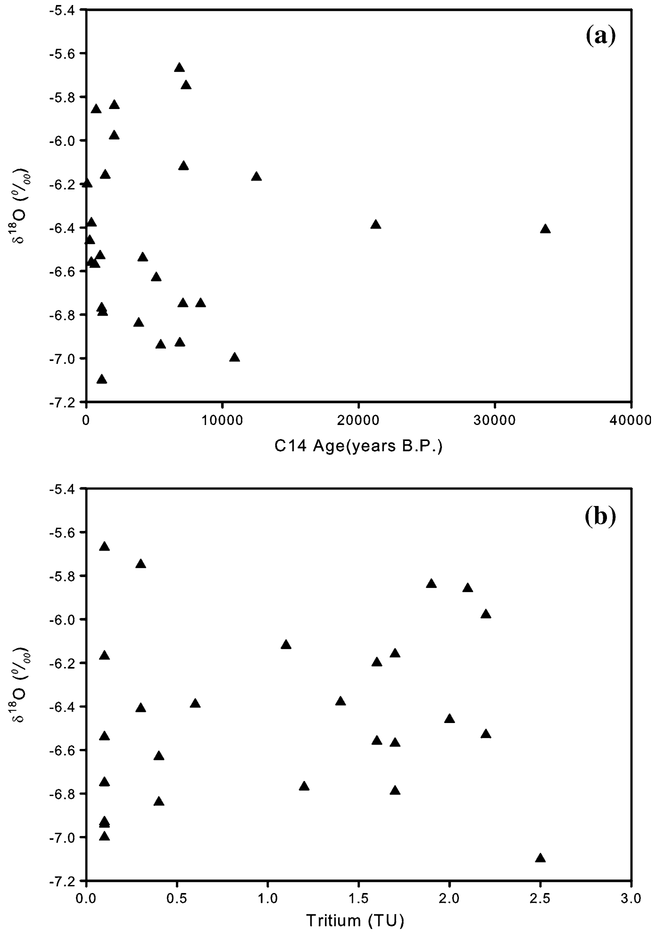

Fig. 5 Plots of δ^{18}O versus C-14 age (**a**) and the tritium concentration (**b**) for groundwater samples of HMGD. The plots of δD versus C-14 age and tritium concentration are not shown here due to the highly correlated linear relationship between δD and δ^{18}O

infiltrates through the unsaturated zone. The mixing of different recharge waters during infiltration through the vadose zone is affected by the heterogeneity of soil and residence time. In the Pleistocene aquifers composed of semi-consolidated soil with lower hydraulic conductivity, groundwater is expected to have a longer residence time and infiltrate through both fracture (fast path) and pore (slow path) to obtain a convergence of isotopic compositions of groundwater samples from distributed locations and depths.

Semi-quantitative analysis of groundwater recharge

According to the previous discussion, the groundwater in HMGD represents the long-term weighted average of local meteoric water. Under this exclusive condition, it is possible to have quantitative estimation for the recharge based on the mass balance of isotopic compositions. However, some difficulties do exist. (1) The amount of infiltration is complicated. It is principally influenced by the permeability and moisture content of the soil, the presence of vegetation and the volume and intensity of precipitation. The amount of infiltration is critical for calculating storm runoff. In hydrology, it is usually solved by the empirical formula, e.g. SCS curve-number method described in Patra (2001). (2) The travel time of water in the vadose zone is difficult to determine. (3) The precipitation samples in this study are not from a whole hydrological year.

Based on these reasons, only a semi-quantitative approach to groundwater recharge is allowed. In this study, the concept of the hydrologic φ-index method is employed to manage complications of the infiltration process. The φ-index assumes that the loss is uniformly distributed across the rainfall event and represents an average loss function whose value results in a volume of direct runoff equal to that measured (Viessman et al. 1989). If the φ-index is larger than the rainfall intensity during an interval of time, then the computed direct

Table 1 Statistical information for groundwater isotopic compositions in HMGD

Isotopes	Sample number	Average (‰)	Mode (‰)	Standard deviation (‰)	95% Confidence interval (‰)
δD	33	−40.7	−42.0	3.87	1.37
δ^{18}O	33	−6.5	−6.2	0.44	0.16

runoff during that period is zero. Although the concept is the same, the assumptions of the φ-index differ slightly in this study. (1) The modified φ-index, named as isotopic φ-index, is defined by the mass balance between the measured groundwater isotopic compositions and the weighted isotopic compositions below the φ-index in the hyetograph. Accordingly, the volume of rainfall below the isotopic φ-index represents the amount of infiltration but not the total losses for the traditional φ-index. (2) The isotopic φ-index should be treated as a non-uniform value. The hydrologic φ-index, the traditional one, is usually applied for an individual storm with an hourly hyetograph. Nevertheless, only isotopic compositions for the daily hyetograph are available, and those for July–December are absent. To approach this problem under the concept of a weighted average of isotopic compositions, an isotopic φ-index is defined based on a mean monthly hyetograph. Under this scenario, the isotopic φ-index would not remain constant through a year.

According to the above assumptions, the derivation of an isotopic φ-index needs a mean monthly hyetograph and the corresponding mean isotopic compositions. Firstly, the mean monthly hyetograph needs to be established with daily precipitation records for a selected duration. Theoretically, the duration should approximately agree with the travel time in the vadose zone. Although it is difficult to estimate travel time as mentioned above, the duration generally lasts for years. Even though the travel time is not an even number of hydrological years, it ought to not cause much error while the remnant is ignored if the travel time is long enough. In this study, the mean monthly hyetograph was established by averaging the rainfall depths for each month during the period of 2000–2003. Secondly, the rainfalls during 2000–2003 are categorized into monsoon type and typhoon-related type which have very different isotopic compositions. Thirdly, the absent isotopic compositions during July–December need to be implemented. The isotopic compositions were copied from those of similar weather types for monsoon type precipitation. (1) October–December are specified by those with similar surface temperature, i.e., April, March and January, respectively. (2) July–September are assigned with June. In addition, the typhoon and typhoon-related precipitations are assigned with the weighted isotopic ratios during the period of 2003/6/8–2003/6/9. Fourthly, weighting the isotopic compositions with the corresponding daily rainfall depth derives the mean monthly isotopic compositions. The resultant monthly isotopic compositions are listed in Table 2. The isotopic ratios are highly depleted in November, which is traditionally not a typhoon season, but an unusually late typhoon, Xangsane, seriously pounded northern Taiwan during 2000/10/31–2000/11/2 and brought in 162 mm of rainfall to the HMGD. This typhoon event dominates the weighted isotopic compositions in November. Finally, the isotopic φ-index can be solved by trial and error. A trial φ value is first applied to the mean monthly hyetograph. The area below the trial φ-index for each month represents the amount of infiltration and is then weighted with isotopic composition. The summation of the weighted isotopic compositions is divided by the total infiltration, i.e., the area below the φ-index, to obtain isotopic compositions of hydrogen and oxygen which should be equal to those of groundwater through the iteration process.

As mentioned in the assumptions, the isotopic φ-index is not a constant due to the highly variable values during a hydrological year. To evaluate the pattern of the infiltration through a year, a uniform φ value is applied at first, and then the value of 264 mm/month is obtained for the mass balance of δD (Fig. 6a). The δD used here is derived from the mean isotopic ratios of groundwater sampled from the Hsinchu Science Park, the same location of precipitation samples, i.e., −41.5 and −6.5‰ for δD and δ^{18}O. The estimated φ value is higher than all of the monthly rainfall depths except September and is equivalent to an annual infiltration depth of 1,520 mm. It is not reasonable that this φ value causes 92% of the annual precipitation (1,646 mm) and 100% of the precipitations in 11 months to recharge the groundwater. Besides, the calculated isotopic ratios become depleted with increasing φ value (Fig. 6b) because higher rainfall depth has a lower isotopic ratio. The depleted isotopic compositions should be weighted more to get a reasonable isotopic φ-index. In other words, the amount of monthly infiltration during June–November with depleted isotopic compositions is more significant than that in other times.

Table 2 Monthly weather records during 2000–2003 and the weighted isotopic compositions for the study of infiltration in HMGD

Month	Surface temperature (°C)	Daily rainfall frequency (%)	Rainfall depth (mm)	Weighted δD (‰)	Weighted δ^{18}O (‰)
Jan	15.33	23	66.25	−9.31	−2.55
Feb	15.85	19	92.75	−3.47	−2.02
Mar	18.45	30	124.00	−11.35	−2.86
Apr	21.55	40	181.25	−17.04	−3.60
May	25.08	24	114.25	−40.99	−6.13
Jun	27.68	31	195.00	−52.93	−7.68
Jul	28.18	27	183.00	−56.16	−8.11
Aug	27.13	31	166.25	−61.66	−8.83
Sep	25.08	33	389.75	−62.67	−8.97
Oct	23.33	8	36.25	−60.37	−8.67
Nov	19.80	11	45.25	−64.27	−9.25
Dec	17.65	17	52.00	−9.31	−2.55

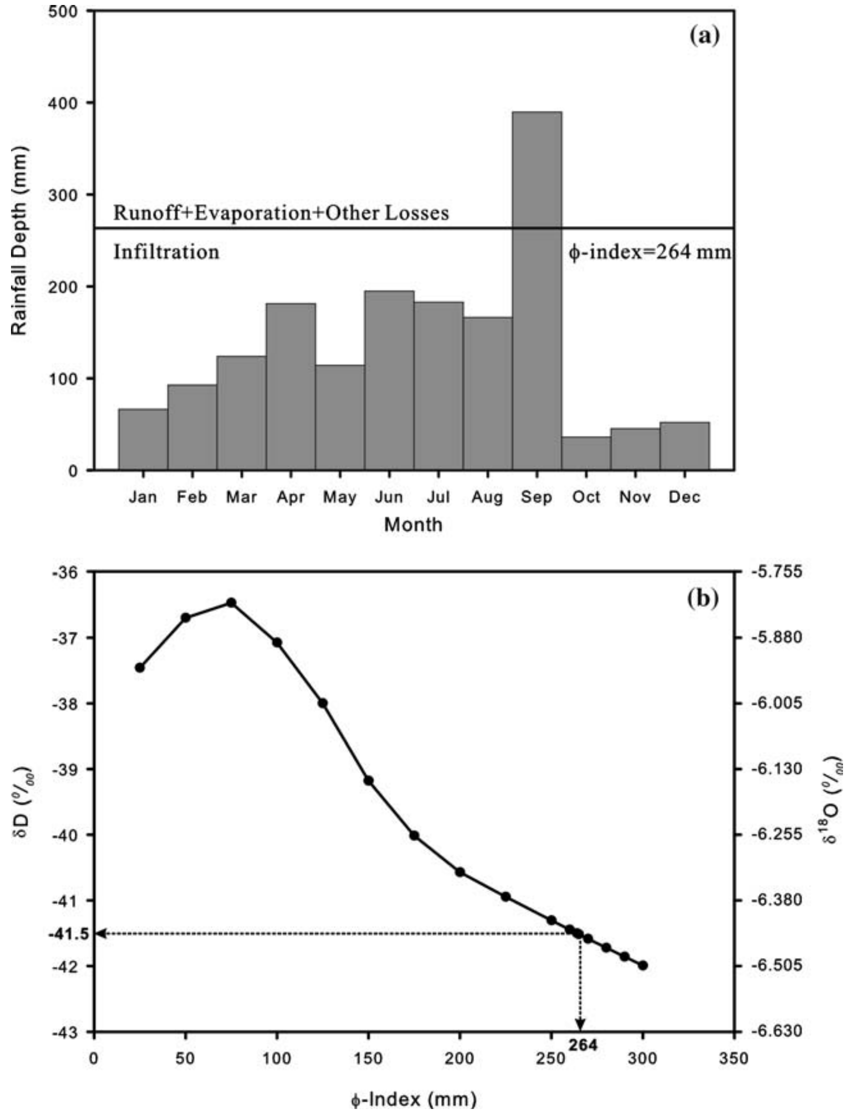

Fig. 6 **a** Monthly rainfall hyetograph. The uniform φ-index of 264 mm/month is derived by using $\delta D = -41.5‰$ as the target isotopic ratio. **b** Plot of calculated δD versus uniform φ-index. The target isotopic ratio becomes enriching with increasing uniform φ-index while φ-index is less than 60 mm/month because most of the months have enriched isotopic compositions. While the φ-index is greater than 60 mm/month and the depleted isotopic ratio is dominant, the target isotopic ratio decreases

In general, infiltration is generally proportional to the monthly rainfall depth or rainfall frequency (Table 2). Based on the assumptions, the calculated δD values are, respectively, -43.1 and $-36.7‰$. If the target isotopic composition of groundwater is, as mentioned, the result of depleted surface water after evaporation, then the infiltration derived from the ratio of rainfall depth is plausible and there is only a 3.7% difference with the target groundwater δD value. This result can be explained by the steep slope which minimizes the effect of the antecedent moisture condition of the soil. Subsequently, the proportion of recharge during May–September (5 months) can be summed up to comprise about 65% of the whole year's recharge, while the amount of monthly infiltration is assumed to be proportional to the monthly rainfall depth. A ratio of the amount of monthly infiltration during May–September, the same period mentioned above, to that during November–May is called the seasonal infiltration ratio. Under this situation, only the ratio can be derived. It is not possible to calculate the amount of recharge because the amount of monthly infiltration is assumedly dependent on each other. While using -41.5 and $-6.5‰$ for δD and $\delta^{18}O$ as the target groundwater isotopic ratios from the Hsinchu Science Park, the seasonal infiltration ratios are equal to 1.73 and 1.93, respectively. These two values agree with each other due to the highly correlated groundwater isotopic compositions. To sum up, the recharge during May–September makes up about 55–58% of the whole year infiltration. This value is 65% lower than that derived by the previous assumption because the target isotopic ratio of $-41.5‰$ is higher than that of $43.1‰$ in the previous case. To estimate conservatively, the recharge during the raining season (May–September) comprises at least 55% of the whole year infiltration.

Isotopic compositions of stream flow

Figure 7 presents a plot of the δD versus δ^{18}O values for 84 river waters collected in HMGD, showing a linear regression close to that of the precipitation samples (Appendix 3). The δD and δ^{18}O values define a line, $\delta D = 7.61\delta^{18}O + 9.62$, with a high correlation coefficient (r^2) of 0.97. As mentioned in the section Isotopic compositions of groundwater, the stream waters are supposed to be subject to evaporation loss. Under this situation, the isotopic regression line for the stream waters should evolve to another flatter line. By comparison with LMWL, the slope and interception of the regression line for stream water show only trivial differences. This also pertains to the evaporation processes under high humidity.

Stream water has three major components based on the speed of appearance after rainfall: surface runoff, interflow and groundwater (baseflow). The difference between arrival times for interflow and surface runoff is in the order of hours. As a result, they are both from recent storms with similar isotopic compositions. Therefore, from the point of view of isotopic composition, stream water can be conceived as being composed of groundwater and runoff, including surface runoff and interflow. It should be noted that the groundwater component in stream water comes principally from aquifers with variable elevations along the watershed.

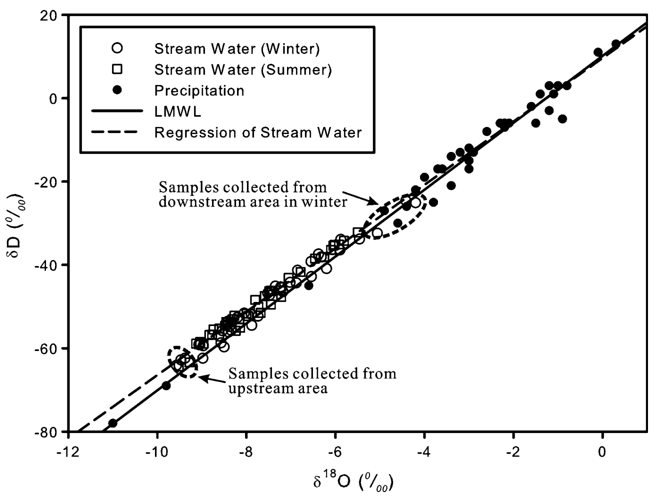

Fig. 7 Plot of δD versus δ^{18}O of the stream samples (*open symbols*) and precipitation samples (*solid circle*). The regression line of stream water samples is determined as $\delta D = 7.61\delta^{18}O + 9.62$ ($r^2 = 0.97$) from 84 samples collected within 11 watersheds

Toucian Creek, for example, in the northern HMGD, originates from the Hsuehshan Mountain Range which is over 2,000 m a.s.l. and its water samples have much lower isotopic ratios compared to other shorter creeks. However, the Keya Creek, originating under 100 m a.s.l. just south of Toucian Creek, has higher isotopic ratios which are similar to those of groundwater. These are illustrated by the enriching trend of isotopic compositions from upstream to downstream for most of the sampled creeks (Fig. 8) and can be explained by the heavier groundwater and low-elevation runoff becoming more dominant downstream. The dashed line in Fig. 8 also represents an interpolating boundary where the isotopic ratio of stream water exceeds that of groundwater. This line obviously shifts to the downstream area in the Toucian Creek watershed due to depleted precipitation at high elevation.

In this study, the stream waters were collected for both winter (February 2001) and summer (July 2001). A further discussion on the seasonal variation of isotopic compositions follows. To compare Fig. 8a and b, stream waters in summer are lighter than those in winter, and the δD value of −41.5‰ (dashed line in Fig. 8a) moves downstream in summer. This evidence seems to agree with depleted precipitations in summer. To demonstrate the seasonal variation in detail, δD differences between summer samples and winter samples were used as shown in Fig. 9. The summer samples are lighter than the winter samples only in the downstream area and the difference is more pronounced in the coastal area. By contrast, the summer samples are heavier than the winter samples in the upstream area. Of course, there is an exception in the southern HMGD. This area belongs to the Ta-an Creek watershed where hydrologic conditions are strongly influenced by a major upstream reservoir. Interflow is thought to be one of the factors causing spatial distribution of seasonal variation. Interflow runs laterally on top of layered down slopes, especially in places where the down hill topographic slope decreases such as the toe-slopes of hills or in topographically convergent areas, e.g., a creek valley in HMGD. It is more active in topsoil with higher moisture and higher hydraulic conductivity. Considering spatial and temporal conditions, the downstream area has a broad alluvial plain conducive to the development of interflow and interflow would be more active during the wet season (summer) due to the higher degree of saturation in the topsoil. Therefore, downstream samples in summer would have lower isotopic ratios due to lighter precipitations.

On the other hand, when the isotopic compositions in winter are more depleted than those in summer in the upstream area, values are in conflict with the case of precipitation. Stream waters were all sampled below 650 m a.s.l. in this study, but most of the creeks originated above 1,000 m a.s.l. Theoretically, there should be

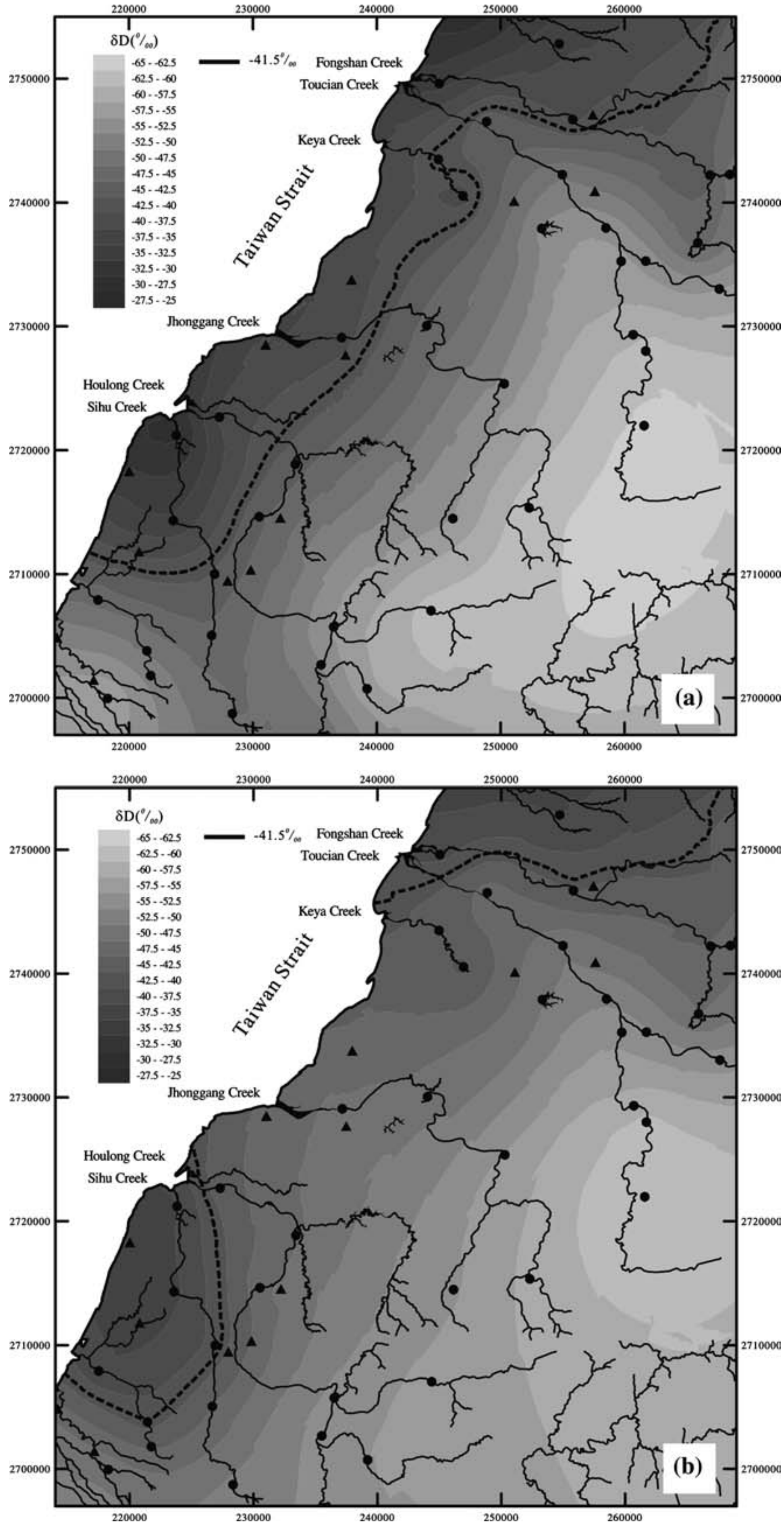

Fig. 8 The distribution of δD in stream waters collected in winter (**a**) and in summer (**b**). The data have been smoothed by the ordinary kriging interpolator. The *dashed line* represents the target groundwater isotopic ratio of −41.5‰. The distributions of δ^{18}O are not shown here due to the highly correlated linear relationship between δD and δ^{18}O

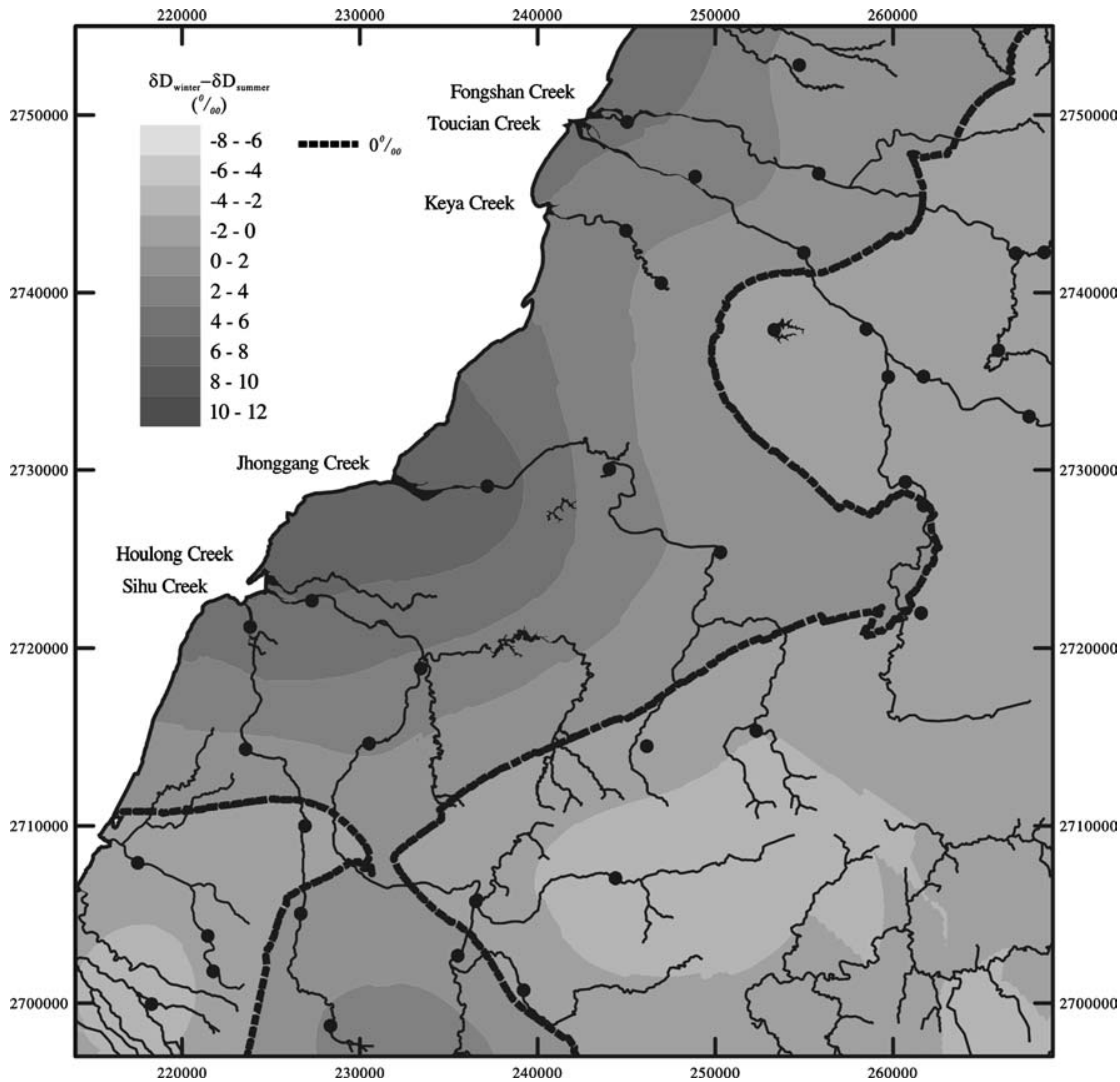

Fig. 9 Distribution of δD differences between winter samples and summer samples. The data have been smoothed by the ordinary kriging interpolator. The *dashed line* is the boundary of the isotopic compositions of stream water equal to those of groundwater. *Solid circles* show locations of the sampling sites

an intense altitude effect for the isotopic compositions of stream samples. According to the analyzed data, the most depleted stream water was collected at 650 m a.s.l. in the winter of 2001 and the isotopic ratios of oxygen and hydrogen are −9.47 and −62.8‰, respectively. This is in accordance with the theoretical trend of the altitude effect. However, the isotopic compositions of the typhoon type precipitation may be too negative to follow the altitude effect. These kind of data are not available for HMGD, but can still be demonstrated by the precipitation samples collected from Chiayi in southern Taiwan. Table 3 lists the isotopic compositions sampled from the plains (Chiayi City) and mountain ranges (Ali Mountain) during 2003/5/18–2003/6/18. The elevations are 75 m a.s.l. and 2,250 m a.s.l., respectively. During the sampling period, there was only one typhoon event on 2003/6/7. This typhoon event precipitated lighter water in the areas in the plains than in the mountain range. Whilst all other monsoon type precipitations follow the theoretical trend described by the altitude effect, which makes isotopic ratios depleted with increasing elevation, this observation of an opposite altitude effect for typhoon type precipitation provides an explanation as to why high elevation samples have depleted isotopic compositions in winter. Interflow developing in downstream areas and an opposite altitude effect for typhoon type precipitation are the major

Table 3 Comparison of precipitation isotopic compositions sampled from the plains (Chiayi City) and mountain ranges (Ali Mountain) during 2003/5/18–2003/6/18

Collecting date	Plain area (Chiayi city)		Mountain range (Ali mountain)	
	δD(‰)	δ^{18}O (‰)	δD (‰)	δ^{18}O (‰)
2003/5/18	−26	−4.3	−63	−9.1
2003/6/7	−96	−13.2	−66	−9.5
2003/6/10	−36	−5.8	−70	−10.0
2003/6/11	−21	−4.5	−37	−5.7
2003/6/12	−16	−3.8	−37	−5.7
2003/6/18	−16	−3.9	−60	−8.6
2003/7/9	−5	−1.7	−32	−6.0

reasons for the spatial distribution of seasonal variation in HMGD.

Conclusions

The study of isotopic compositions of hydrogen and oxygen in the water mass from HMGD reveal some important hydrological features.

1. The δD and δ^{18}O values in precipitations are highly related, following the equation $\delta D = 8.02\delta^{18}O + 10.16$. The line is perfectly fitted to the GMWL. In addition, the isotopic compositions of precipitations can be categorized into typhoon (depleted) type and monsoon (enriched) type. The groundwater isotopic compositions are nearly located on LMWL and the relationship on the isotopic compositions between precipitation and groundwater can be considered as a two-end-member mixing model.
2. By using the concept of the φ-index applied to isotopic compositions, it is concluded that the amount of monthly infiltration is roughly proportional to the monthly precipitation depth. The proportion of recharge during raining season (May–September) is conservatively estimated to comprise at least 55% of the whole year.
3. The calculated regression line of stream water is: $\delta D = 7.61\delta^{18}O + 9.62$. It is close to LMWL and apparently represents evaporation proceeding under high humidity. The isotopic compositions of stream water also show spatial distribution in seasonal variation, with depleted isotopic compositions in the downstream areas during summer due to more active interflow in the downstream areas in summer. However, there are enriched isotopic compositions in upstream areas during summer, which can be explained by an opposite altitude effect on isotopic compositions for typhoon-related precipitations.

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Appendix 1

Stable isotopic compositions of precipitations collected from the Hsinchu Science Park in HMGD

Sample number	Sampling date	δD (‰)	δ^{18}O (‰)
1	2003/1/9	−16.9	−3.7
2	2003/1/22	−7.1	−2.2
3	2003/2/4	−18.8	−4.0
4	2003/2/10	3.0	−1.2
5	2003/2/15	−6.1	−2.1
6	2003/3/3	−5.9	−2.2
7	2003/3/6	−13.1	−3.2
8	2003/3/9	−14.2	−3.4
9	2003/3/10	−5.9	−2.2
10	2003/3/19	−8.0	−2.6
11	2003/3/20	−16.8	−3.6
12	2003/3/23	−5.1	−0.9
13	2003/3/24	−6.2	−2.3
14	2003/3/29	−12.8	−2.9
15	2003/4/3	−26.7	−4.9
16	2003/4/4	−22.2	−4.2
17	2003/4/7	3.0	−1.0
18	2003/4/8	13.0	0.3
19	2003/4/9	−11.9	−3.0
20	2003/4/11	3.1	−0.8
21	2003/4/14	−2.0	−1.6
22	2003/4/15	0.9	−1.4
23	2003/4/25	1.1	−1.1
24	2003/5/1	11.0	−0.1
25	2003/5/4	−2.9	−1.2
26	2003/5/8	−25.1	−3.8
27	2003/5/15	−29.8	−4.6
28	2003/5/16	−45.3	−6.6
29	2003/5/17	−15.1	−3.0
30	2003/5/18	−25.9	−4.4
31	2003/6/8	−69.2	−9.8
32	2003/6/9	−78.2	−11.0
33	2003/6/12	−17.1	−3.0
34	2003/6/13	−6.0	−1.5
35	2003/6/18	−20.9	−3.4

Appendix 2

Stable isotopic compositions of groundwaters from HMGD

Sample number	Screen depth (m)	δD (‰)	$\delta^{18}O$ (‰)
1	36	−36.6	−5.8
2	100	−32.7	−5.7
3	176	−39.1	−6.2
4	234	−42.1	−6.4
5	76	−39.3	−6.5
6	114	−43.2	−6.6
7	150	−43.0	−6.5
8	63	−39.2	−6.8
9	108	−41.1	−7.0
10	64	−32.7	−6.0
11	36	−45.8	−6.9
12	85	−43.8	−6.9
13	134	−43.3	−6.8
14	163	−44.3	−6.8
15	21	−33.7	−5.9
16	99	−36.0	−6.1
17	132	−42.4	−6.4
18	18	−36.6	−6.2
19	107	−38.5	−6.2
20	53	−37.6	−6.6
21	99	−36.9	−6.5
22	26	−37.5	−5.8
23	20	−42.0	−6.6
24	66	−44.5	−6.8
25	24	−44.7	−6.8
26	51	−41.6	−6.4
27	85	−47.1	−7.1
28	48	−43.0	−7.1
29	106	−46.1	−7.6
30	22	−45.0	−6.7
31	146	−40.2	−6.2
32	58	−41.0	−6.2
33	107	−42.1	−6.8

Appendix 3

Stable isotopic compositions of stream waters from HMGD

Sample number	Sample altitude (m)	2001/2		2001/7	
		δD (‰)	$\delta^{18}O$ (‰)	δD (‰)	$\delta^{18}O$ (‰)
1	10	−24.6	−4.42	−34.2	−5.8
2	90	−38.1	−6.32	−35.4	−6.1
3	10	−32.4	−5.06	−36.4	−6.1
4	50	−36.4	−5.9	−38.5	−6.5
5	140	−39.2	−6.55	−43.1	−7.1
6	220	−52.3	−7.74	−46.3	−7.5
7	160	−41.3	−6.85	−41.7	−6.8
8	10	−47.3	−7.48	−53	−8.2
9	70	−51.6	−8.05	−54.6	−8.5
10	90	−55.7	−8.34	−53	−8.3
11	150	−56.1	−8.4	−56.8	−8.8
12	210	−59.1	−9.07	−58.5	−9.0
13	310	−62.4	−9.38	−63.4	−9.3
14	810	−64.1	−9.57	−64.4	−9.5
15	130	−54.8	−8.38	−52.2	−8.3
16	270	−55.8	−8.54	−53.7	−8.5
17	430	−59.4	−8.97	−55.5	−8.7
18	570	−59.2	−9.05	−58.9	−9.0
19	10	−45.3	−7.23	−46.3	−7.4
20	70	−33.9	−5.88	−35.1	−5.9
21	150	−54.3	−8.42	−48.4	−7.8
22	10	−37.4	−6.38	−49.5	−7.5
23	40	−46.4	−7.18	−47.5	−7.6
24	150	−52.6	−8.2	−55.8	−8.2
25	570	−53.5	−8.41	−53.6	−8.5
26	650	−62.8	−9.47	−58.9	−9.1
27	10	−40.9	−6.2	−47.7	−7.2
28	10	−44.2	−7.02	−46.2	−7.5
29	70	−45.5	−7.21	−50.1	−7.7
30	330	−57.1	−8.62	−53.8	−8.5
31	250	−51.9	−7.89	−55.1	−8.1
32	370	−45.1	−7.35	−52.3	−8.0
33	450	−58.7	−8.57	−55.3	−8.6
34	450	−62.4	−8.98	−56.7	−8.8
35	10	−25.1	−4.2	−32.2	−5.5
36	50	−33.8	−5.46	−35.2	−6.0
37	120	−42.8	−6.54	−38.1	−6.3
38	210	−44.3	−6.88	−46.1	−7.2
39	340	−46.8	−7.52	−51.5	−7.7
40	10	−54.5	−7.88	−53.6	−8.4
41	70	−59.7	−8.5	−51.7	−8.0
42	70	–	–	−34.1	−5.8

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