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Using stable isotopes to determine sources of fog drip in a tropical seasonal rain forest of Xishuangbanna, SW China

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Abstract

To identify the possible sources of fog drip, samples of rainfall, fog drip, throughfall, stemflow, stream water, river water, pond water and soil water were collected for 3 years (2002–2004) for stable isotopic analysis, at a tropical seasonal rain forest site in Xishuangbanna, Southwest China. We found that radiation fog is produced mainly through evaporation from pond, river, and soil, and through forest evapotranspiration. The analyses suggest that evaporation from the stream is limited. In addition, radiation fog produced during the dry season (low absolute humidity) contained more terrestrially recycled water than fog produced during the rainy season (high absolute humidity). Forest evapotranspiration appears to be the largest fraction, but a more intense sampling scheme will be needed to assign the relative contribution of the different sources of fog.

Keywords: Stable isotopes; Deuterium excess; Fog drip; Evapotranspiration; Moisture recycling; Tropical seasonal rain forest

1. Introduction

Parameterizing the exchange of water between the land surface and the atmosphere is a major challenge in Earth System modeling (Pitman and McAvaney, 2002; Henderson-Sellers et al., 2003; Pitman, 2003). Capturing the timing and extent of precipitation interception, storage and eventual loss to either runoff or evaporation will improve predictions of soil moisture, stream flow, ocean discharge and atmospheric water replenishment and, thus, model forecasts of droughts, floods, coastal ocean salinity and continental moisture recycling (Gat, 2000; Henderson-Sellers et al., 2002; Pitman and Narisma, 2005). There is an increasing body of work suggesting that the recycling of water to the atmosphere via evapotranspiration is very important for the maintenance of the water cycle in regional or global scales (Salati et al., 1979; Gat and Matsui, 1991; Gat et al., 1994; Martinelli et al., 1996; Gat and Airey, 2006), and that the condensational energy released by convective precipitation can be of sufficient magnitude to affect regional and global climate patterns (Henderson-Sellers et al., 2002; Zhao and Pitman, 2005; Gat and Airey, 2006). In addition, as recently demonstrated by some studies, forest cover has significant impact on local and regional-scale climate, and substituting forest for farmland or other land-use alters the energy balance and consequently the water balance (Wright et al., 1992; Henderson-Sellers et al., 2003; Pitman et al., 2004; Pitman and Narisma, 2005; Gat and Airey, 2006).

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Fog drip, the coalescing of fog droplets on foliage producing large drops that rain to the ground, has intrigued researchers for many years (Aravena et al., 1989; Ingraham and Matthews, 1995; Dawson, 1998; Bruijnzeel, 2001; Lawton et al., 2001). Past investigations have shown that water input or soil moisture is measurably higher around the tree canopies or in forest stands where the fog is stripped from the air mass and that when trees are lost or removed from the watershed both the water input from fog drip and the streamflow decline significantly (Ingwersen, 1985). Other studies (Weathers and Likens, 1997; Bruijnzeel, 2001) have demonstrated that ecosystem nutrient balance or other aspects of ecosystem biogeochemistry could be influenced by fog water inputs. Fog may also help ameliorate plant moisture stress by reducing canopy transpiration or evaporation from the habitat (Hutley et al., 1997), but little is known about the detailed origin of fog drip (Ingraham and Matthews, 1995; Martinelli et al., 1996).

In this study, we attempt to identify the possible sources of the fog drip in a tropical seasonal rain forest in Xishuangbanna, SW China, through analysis of the stable isotopic composition (hydrogen and oxygen) of the water in rainfall, throughfall, stemflow, stream, pond, and soil water inside the forest, as well as water in a river just near the forest. Previous study (Liu et al., 2004) showed that in this forest stand, the average amount of annual fog drip was 89.4 mm, which contributes an estimated 5% of the annual rainfall, with 86% of the fog drip occurring in the dry season (November–April) and representing 49% of total precipitation in the same period. The objective of this study is to explore the hypothesis that the sources of fog drip in this forest are mainly from terrestrially recycled water within the area, such as forest evapotranspiration. We also discuss the influences of the land cover change (LCC) on the fog formation since deforestation in this region has been extensive during the last 30 years (Cao et al., 2006; Li et al., 2006).

2. Materials and methods

2.1. Study site

This study was conducted at a tropical seasonal rain forest site (21°55′39″N, 101°15′55″E, 750 m a.s.l.) in Menglun town of Xishuangbanna Dai Autonomous Prefecture in southwestern China (Fig. 1). The site is approximately 800 km to the northeast of the Bay of Bengal and 600 km to the west of the Bay of Beibu. The study plot (dominated by *Pometia tomentosa* and *Terminalia myriocarpa*) is located on a flat area between two hills extending from east to west, and is a permanent plot dedicated to the long-term ecological research managed by the Xishuangbanna Tropical

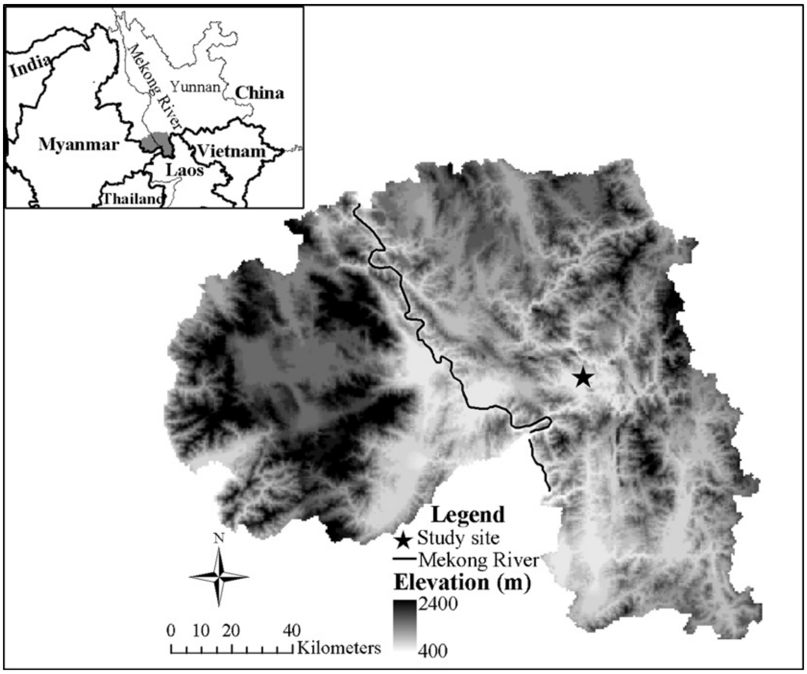

Fig. 1. Map showing location of the study site (21°55′39″N; 101°15′55″E; 750 m) in Yunnan Province, SW China, indicated by a solid star.

Rainforest Ecosystem Station, Chinese Academy of Sciences. The width of the flat is approximately 40 m. The slopes to the south and north of the site are about 20°. The soil is lateritic soil developed from siliceous rocks, such as granite and gneiss, with pH values of 4.5–5.5. A stream (about 1 m wide) winds through the site and the length of the valley is about 2 km. This valley is a typical site for the occurrence of tropical seasonal rain forest (Cao et al., 2006). This tropical rain forest differs from tropical Asian lowland rain forests in that some of its tree species are deciduous. Species richness is lower than those of Malesian rain forests, but higher than those of Australian and African rain forests, and similar to tropical forest on Barro Colorado Island, Panama (Cao et al., 2006; Zhu et al., 2006).

Xishuangbanna has mountain-valley topography with the Hengduan Mountains running north–south, and about 95% of the region is mountainous. The Mekong River flows through the center of the region, resulting in many river valleys and small basins. A large proportion of the forest in this region is tropical seasonal rain forest, which is primarily formed in wet valleys, lowlands and on low hills (less than 1000 m a.s.l.) where heavy radiation fogs frequently occur (Cao et al., 2006). The climate is strongly seasonal with two main air masses alternating during the year. Between May and October (rainy season), the tropical Southwest Monsoon from the Indian Ocean delivers about 85% of the annual rainfall, whereas the dry and cold air of the southern edges of the subtropical jet streams dominates the climate between November and April (dry season). The dry season includes a cool sub-season from November to February and a hot sub-season from March to April. The cool sub-season is characterized by highest frequency of radiation fogs during the night and morning. The hot-dry sub-season is characterized by dry and hot weather during the afternoon and with radiation fogs during the morning only. Radiation fogs occur nearly every day from November to April and are heaviest from midnight (23:00–02:00 h) until mid-morning (09:00–11:00 h) when the daily temperature difference is greatest. During the dry season, on average there is fog cover 37% of the time (Liu et al., 2004).

Long-term climate records collected for 40 years at a weather station (560 m a.s.l.) 5 km southeast from the study site show that the mean annual air temperature is 21.7 °C with a maximum monthly temperature of 25.7 °C for the hottest month (June) and a monthly minimum of 15.9 °C for the coldest month (January). Temperatures exceeding 38 °C often occur during March and April, and are always associated with low relative humidity (less than 40%). The mean annual rainfall is 1487 mm, of which 1294 mm (87%) occurs in the rainy season versus 193 mm (13%) in the dry season. Class A pan evaporation varies between 1000 and 1200 mm year^{-1}. The mean monthly relative humidity is 87%. The mean annual wind speed is 0.5 m s^{-1} (Liu et al., 2005).

2.2. Water sampling

Water samples for environmental isotope analysis were collected from fog drip, rainfall, throughfall, stemflow, stream water, river water, pond water and shallow soil water. Plastic funnel collectors (60 cm in diameter), each connected with a 1000-ml plastic bottle, were mounted 0.7 m above the forest floor to collect fog dripping from the canopy. Six of these collectors were placed in a fixed but random pattern on the forest floor to collect fog drip. To reduce the number of samples for isotope analysis, a volume of water proportional to the volume collected by each collector for each collection event were combined after separate measurement of the volumes collected. Fog drip sampling was performed at pre-dawn, at or near the peak of a fog drip event, but before isotopic fractionation had occurred from re-evaporation (Dawson, 1998). Fog drip samples were collected at 2-week intervals from January 2002 to December 2004 except June, July, August and September of 2002–2003 and July, August and September 2004 where fog drip was not available.

Rainfall samples were collected at monthly interval on an event basis, by using a collector consisting of a stainless steel funnel (collecting area of 314 cm^2) connected to a 2000-ml plastic bottle. The funnel was mounted on the top of a 72 m high meteorological tower. In addition, during the study period, rain sample was also collected at each rain event when rainfall exceeded 5 mm at the weather station 5 km southeast from the study site. Samples of rainfall were collected immediately after rainfall ceased, or in the early morning when rain fell overnight. However, no isotopically significant difference was found between rainfall samples collected concurrently at the two locations (Liu et al., 2005). Hence, rainfall samples at the weather station were used to represent that of the study site. Six of V-shape troughs (0.3 m × 2.0 m) placed in random pattern like fog drip collectors were used to collect throughfall. Stemflow was taken from 12 sample trees that were composed of three stems of small (<10 cm), moderate (10–30 cm), and larger (>30 cm) diameter classes for each of three dominant species. Stemflow for each tree was collected using spiral polyethylene collars around the tree trunk, connected to a plastic container (Wang et al., 2006). Samples of throughfall and stemflow were collected at monthly interval and concurrently on each sampling date on an event basis during the study period.

Six soil samples were obtained randomly within the site from cores collected to a depth of 20 cm beneath the canopy on each sampling day. The litters were taken out prior to the soil samples collection. The soil samples were stored in 200-ml glass bottle and transported to the laboratory and immediately frozen for soil water extraction. Streamflow, which origins from this forest stand, was monitored continuously at 90° V-notch gauging weirs at the outlet and stream water samples were collected at the outlet during non-storm runoff period (Liu et al., 2006). River water samples were taken from an important tributary (Luosuo River) of the Mekong River, which flows just beside the study plot. In addition, a small pond (0.4 ha) in the forest stand was also sampled to test the effects of its evaporation on the source of fog drip. Water samples for stream water, river water, pond water and soil water were collected at monthly interval and concurrently on each sampling date during the study period.

Soil water was extracted from the soil samples by cryogenic vacuum distillation (Ehleringer et al., 2000). Two glass tubes were attached to a vacuum pump in Y-shape configuration. Soil sample was placed in one tube and frozen by submerging the tube in liquid nitrogen. Both tubes were evacuated and then isolated from the vacuum line to create a closed U-shape configuration. The tube containing the sample was placed in boiling water, while the second tube was placed in liquid nitrogen to trap water. The extracts were sealed in vials and frozen at -20 °C until analysis.

Fog drip, rainfall, throughfall, stemflow, stream water, river water and pond water samples were collected in the 100-ml polysealed glass bottles, and transported to the laboratory and immediately frozen. During the period of observation, a total number of 59 samples for fog drip, 116 for rainfall at the weather station, 23 for throughfall, 17 for stemflow, 35 for soil water, and 36 for each of stream, river and pond water have been collected.

2.3. Isotopic analysis

The hydrogen- and oxygen-isotope analyses were performed at the Geochemistry Department, the Test Center of Lanzhou Branch, Chinese Academy of Sciences. The stable hydrogen and oxygen isotope composition was determined from a gas sample generated from pure liquid introduced into an isotope ratio mass spectrometer (Finnigan MAT252, Germany). Hydrogen- and oxygen-isotope data are reported in the standard delta (δ) notation (in parts per thousand, or ‰) where the ratios of the heavy to light stable isotopes in the sample (e.g. $^{2}H/^{1}H$ or $^{18}O/^{16}O$) is determined relative to Vienna-Standard Mean Ocean Water (V-SMOW; see Gat, 1996); these are denoted as δD and $\delta^{18}O$ for the H and O isotope ratios, respectively. The δD values of all water samples were determined using the zinc-reduction method outlined by Coleman et al. (1982). The $\delta^{18}O$ values of the water samples were determined using the carbon dioxide equilibration method outlined by Dugan et al. (1985). The precision ($\pm$S.D.) of oxygen and hydrogen isotope results are 0.2 and 2‰, respectively. Isotopic composition of each sample was analyzed at least twice to check for repeatability of analysis, and standards were analyzed to check for accuracy of analytical procedures.

2.4. Statistics

Volume-weighted mean isotopic composition was calculated for rainfall, throughfall, stemflow, fog drip and streamflow samples, and arithmetical mean for river, pond and soil water samples. Since very large variances are common, ranges and standard deviations are reported for isotopic composition.

As pointed out by Gat and Matsui (1991) and Gat (1996), the deviations from local meteoric water line (LMWL) produced during evaporation and condensation of vapor provide useful isotopic markers to trace the source of vapor. First, we can measure the slope described by the δD and $\delta^{18}O$ values of water bodies which are potential sources of vapor. If the slopes are significantly lower than that of the LMWL, we can conclude that a particular water body, such as stream, river, pond and soil, has undergone evaporation. Second, we can calculate the deuterium excess ($d = \delta D - 8\delta^{18}O$; Gat, 1996, 2000) of precipitation (rainfall and fog drip), by forcing a line with a slope of eight through the δD and $\delta^{18}O$ values of these waters, and determine whether they were produced by evaporation. If the d is significantly greater than that of the LMWL, we can conclude that the measured precipitation (rainfall and fog drip) was produced by increased admixtures of recycled evaporated moisture (Gat and Matsui, 1991; Gat, 1996; Martinelli et al., 1996).

The Student's t-test, with P-values of 0.05 and 0.01, was used to test the significance of d values larger or small than 8.9 (i.e. d of the LMWL; see Section 3.1), and slope of best-fit regression line smaller than 7.96 (i.e. slope of the LMWL). SPSS 10.0 for Windows was used for all statistical calculations.

Table 1
Summary of stable isotopic variability of waters by type in the tropical seasonal rain forest at Xishuangbanna, SW China during 2002–2004

Sample type	δD (‰)			δ^{18}O (‰)			d (‰)			n
	Mean	Minimum	Maximum	Mean	Minimum	Maximum	Mean	Minimum	Maximum	
Rainfall	−56.8 ± 15.1	−121.2	23.9	−7.9 ± 4.7	−16.9	3.1	8.9 ± 3.5	3.7	20.5	116
Throughfall	−51.3 ± 13.7	−118.3	22.8	−7.2 ± 3.3	−16.7	1.9	7.5 ± 2.7	4.8	17.4	23
Stemflow	−52.4 ± 12.2	−96.2	23.0	−7.3 ± 2.9	−16.7	2.2	7.3 ± 4.0	2.9	19.2	17
Fog drip	−11.5 ± 4.2	−30.5	29.4	−3.3 ± 1.8	−8.0	2.9	16.3 ± 3.5**	9.3	21.8	59
Stream	−49.7 ± 2.8	−51.2	−39.7	−6.9 ± 1.1	−8.4	−5.7	7.9 ± 2.0	4.1	13.3	36
River	−45.1 ± 2.3	−50.5	−42.1	−6.1 ± 0.9	−7.9	−5.8	6.9 ± 1.2*	4.7	15.9	36
Pond	−46.1 ± 1.7	−53.7	−37.5	−6.5 ± 1.3	−9.3	−5.1	6.8 ± 1.9*	3.5	17.5	36
Soil	−41.2 ± 5.0	−70.0	−9.2	−5.9 ± 2.0	−11.9	−2.2	7.1 ± 3.5	3.9	15.9	35

(*) and (**) denote the d (deuterium excess) is statistically lower or larger than 8.9 (i.e. d of local meteoric water) at the 0.05 and 0.01 level, respectively. Note that mean values of rain, throughfall, stemflow, fog drip and stream are annual volume-weighted means ± 1S.D., and those for river, pond and soil are times averaged means ± 1S.D.

3. Results

3.1. Rainfall, throughfall and stemflow

The stable isotope values (mean ± 1S.D.) in rainfall, throughfall, stemflow, fog drip, stream, river, pond and soil with minimum and maximum values are given in Table 1 along with their d values (deuterium excess; $d = \delta D - 8\delta^{18}O$; Gat, 1996, 2000). During the study period, the isotopic composition of rainfall is quite variable, with δ^{18}O ranging from −16.9 to +3.1‰ and δD from −121.2 to +23.9‰. A local meteoric water regression line (LMWL) drawn through these data is: $\delta D = (7.96 \pm 0.29)\delta^{18}O + (8.67 \pm 1.42)$ ($n = 116$; $R^2 = 0.97$; $P < 0.01$) (Table 2), which is virtually identical to the regional line ($\delta D = 7.92\delta^{18}O + 9.20$) obtained from 60 stations of the IAEA/WMO global network in the Southeast Asia (Araguás-Araguás et al., 1998). In addition, two regression lines were obtained to describe the isotopic data for different seasons: $\delta D = (7.85 \pm 0.30)\delta^{18}O + (7.21 \pm 1.57)$ ($n = 97$; $R^2 = 0.90$; $P < 0.01$) for the rainy season (May–October) and $\delta D = (8.15 \pm 0.22)\delta^{18}O + (13.79 \pm 1.31)$ ($n = 19$; $R^2 = 0.93$; $P < 0.01$) for the dry season (November–April) (Fig. 2). There is an inverse relationship between the mean monthly δ^{18}O and the monthly rainfall amount (P_{mon}, in millimeters) ($\delta^{18}O = (-0.029 \pm 0.014)P_{mon} + (-3.9 \pm 1.1)$; $n = 33$; $R^2 = 0.54$; $P < 0.05$), whereas the correlation with the mean monthly air temperature is virtually non-existent ($n = 33$; $R^2 = 0.18$; $P > 0.05$).

Concurrent sampling of rainfall, throughfall and stemflow was conducted for monthly intervals during the study period. Annual volume-weighted means differed insignificantly ($P > 0.05$) by 0.7 and 0.6‰ in δ^{18}O, and by 4.0 and 2.9‰ in δD in throughfall and stemflow versus rainfall (Table 1), respectively. The trend suggests a weak increase in the ratio of heavy isotopes in throughfall and

Table 2
Values of slope and interception with respective standard deviations, coefficient of best-fit regression and number of data points for the regression of $\delta^{18}O - \delta D$ in waters by type in the tropical forest site during 2002–2004

Sample type	Slope	Intercept	R^2	n
Rainfall	7.96 ± 0.29	8.67 ± 1.42	0.96	116
Throughfall	7.42 ± 0.34	5.64 ± 3.51	0.85	23
Stemflow	7.56 ± 0.22	6.92 ± 2.02	0.90	17
Fog drip	7.76 ± 0.15	14.63 ± 1.93	0.81	59
Stream	7.28 ± 0.12	7.15 ± 1.42	0.77	36
Pond	6.41 ± 0.16*	4.75 ± 1.08	0.84	36
River	6.33 ± 0.17*	4.95 ± 2.91	0.76	36
Soil	7.17 ± 0.45	7.02 ± 2.50	0.78	35

* Denotes the slope is statistically smaller than 7.96 (i.e. slope of LMWL) at the 0.05 level.

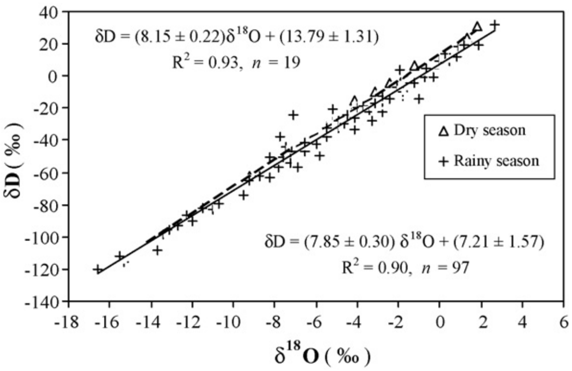

Fig. 2. δD vs. δ^{18}O relationship of rainfall samples for dry (November–April) and rainy (May–October) season collected at the tropical seasonal rain forest site during 2002–2004. The local meteoric water regression lines for dry (upper line) and rainy (below line) season drawn through these data are also shown.

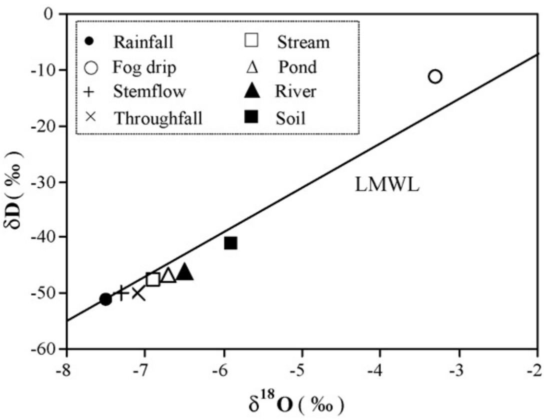

Fig. 3. Plot of δD vs. δ^{18}O of waters in rainfall, throughfall, stemflow, stream, pond, river and soil in comparison to fog drip collected at the tropical forest site shown along with the local meteoric water line (LMWL). Note that values mean the same in Table 1. Standard deviations for each of the waters are not shown for the sake of clarity, but given in Table 1. See text for discussion.

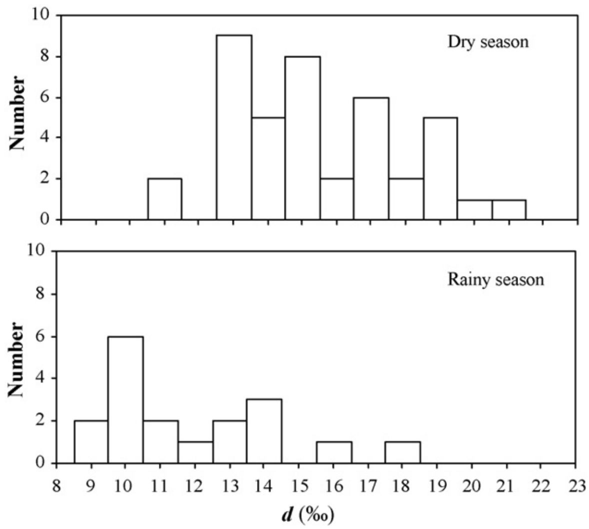

Fig. 4. Distribution of d (deuterium excess) in fog drip samples for dry and rainy season collected at the tropical forest site during 2002–2004. Note that samples collected during dry season have significantly higher d values compared with those in rainy season ($P < 0.05$).

stemflow due to an evaporation effect (Fig. 3). Consequently, d values of throughfall and stemflow are not significantly different from that of rainfall ($d = 8.9$) (Table 1). In addition, although insignificantly different ($P > 0.05$), the slopes of the regression lines for throughfall and stemflow are less than 7.96 (i.e. slope of local meteoric water line) (Table 2), suggesting that as rainfall travels through the canopy, some water is lost by evaporation. The minor differences in the isotopic composition between rainfall, throughfall and stemflow maybe attributed to volume-weighted calculation, which may reduce the difference (Kubota and Tsuboyama, 2003; Liu et al., 2006).

3.2. Fog drip

Compared to the rainfall, throughfall and stemflow, values of isotopic ratio for fog drip are more enriched, and ranged from −8.0 to +2.9‰ in δ^{18}O and −30.5 to +29.4‰ in δD, with a d value of 16.3, which is significantly greater than that of rainfall (Table 1). These data conform to the equation $\delta D = (7.76 \pm 0.15)\delta^{18}O + (14.63 \pm 1.93)$ ($n = 59$; $R^2 = 0.82$; $P < 0.05$), with a slope slightly less than 7.96 and an intercept greater that of LMWL (Table 2). The average value for fog drip is above the LMWL whereas the average throughfall and stemflow values trended below the LMWL (Table 1 and Fig. 3).

The range and distribution of d values for the fog drip are given in Fig. 4. Most dry season samples had d values greater than the average for rainfall ($d = 8.9 \pm 3.5$), with 75% of the samples between 13 and 17. During the rainy season, about 90% of the samples had values greater than 8.9, and 80% of the samples had values between 10 and 14. The d values of samples collected during the dry season were significantly higher than those collected

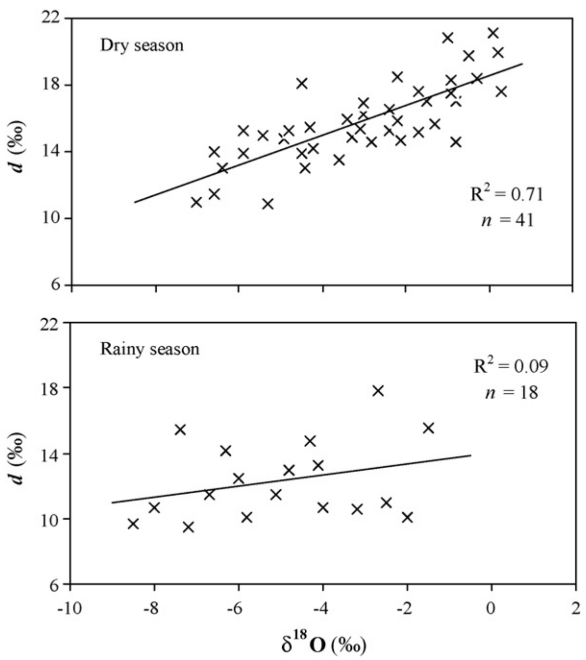

Fig. 5. Plot of δ^{18}O vs. d (deuterium excess) in fog drip samples for dry and rainy season collected at the tropical forest site during 2002–2004. The regression lines for dry and rainy season drawn through these data are also shown.

during the rainy season ($P < 0.05$). In addition, there was a significantly positive correlation ($n = 41$; $R^2 = 0.71$; $P < 0.01$) between $\delta^{18}O$ and d values during the dry season samples, but there was no significant relationship during the rainy season ($n = 18$; $R^2 = 0.09$; $P > 0.05$) (Fig. 5). This means that samples collected during the dry season are more enriched in $\delta^{18}O$ with more d excess than those of samples collected during the rainy season. This can arise because the relative contribution of recycled local moisture to fog during these two seasons is not likely to be the same pattern (Liu et al., 2005).

3.3. Stream, river, pond and soil water

The isotopic composition of surface waters sampled in stream, river, pond and soil is shown in Table 1 and Fig. 3. The soil samples tended to have a higher average value and a wider range in mean $\delta^{18}O$ and δD compared with the stream, river, and pond samples. For the river and pond waters, d values of both are significantly lower ($P < 0.05$) than that of rainfall ($d = 8.9$), while stream and soil waters were not significantly different from rainfall. The slopes of regression lines for the river and pond samples are significantly lower ($P < 0.05$) than 7.96 (i.e. slope of local meteoric water line), but there were no significantly differences in slopes for the stream and soil samples (Table 2).

However, there was a significant season effect (i.e. the dry season and the rainy season) on the slopes for the regression lines for $\delta^{18}O$ versus δD (Table 3). During the dry season, the slopes for the river, pond and soil are all significantly lower than 7.96 ($P < 0.05$), while the stream does not show significantly different from 7.96. During the rainy season, only water collected from the river and pond had slopes significantly lower than 7.96 ($P < 0.05$). Soil water has the same slope as the rainfall during the rainy season.

Table 3
Values of slope and interception with respective standard deviations, coefficient of best-fit regression and number of data points for the regression of $\delta^{18}O - \delta D$ in waters of dry (November–April) and rainy (May–October) season by type in the tropical forest site during 2002–2004

Sample type	Season	Slope	Intercept	R^2	n
Stream	Dry	7.16 ± 0.32	6.17 ± 0.45	0.79	18
	Rainy	7.82 ± 0.58	8.04 ± 1.12	0.85	18
Pond	Dry	$5.89 \pm 0.11^*$	3.63 ± 0.71	0.93	18
	Rainy	$6.82 \pm 0.45^*$	6.05 ± 0.71	0.81	18
River	Dry	$6.02 \pm 0.26^*$	-1.55 ± 1.81	0.67	18
	Rainy	$6.87 \pm 0.41^*$	6.04 ± 3.21	0.84	18
Soil	Dry	$6.14 \pm 0.16^*$	6.17 ± 1.08	0.65	17
	Rainy	7.96 ± 0.25	7.98 ± 1.25	0.82	18

* Denotes the slope is statistically smaller than 7.96 (i.e. slope of LMWL) at the 0.05 level.

4. Discussion

4.1. Isotopic ratios of rainfall, throughfall and stemflow

During the rainy season, the region receives precipitation predominantly from the Indian Ocean together with the Bay of Bengal, with relatively low $\delta^{18}O$ values (Fig. 6) and low d-excess values. The low $\delta^{18}O$ values should be due to the rain-out effect (Gat, 1996, 2000; Araguás-Araguás et al., 1998) of moist monsoon air masses since this region is over 800 km away from coast and the monsoon wind become weaker when they reach the Xishuangbana region. The air masses carrying large amounts of moisture move in a south-westerly direction and shed a significant fraction of their moisture during their transit in spells of heavy rainfall events. Such rainfalls continuously deplete heavy isotopes from the onward moving clouds. As a result, vapor clouds farther

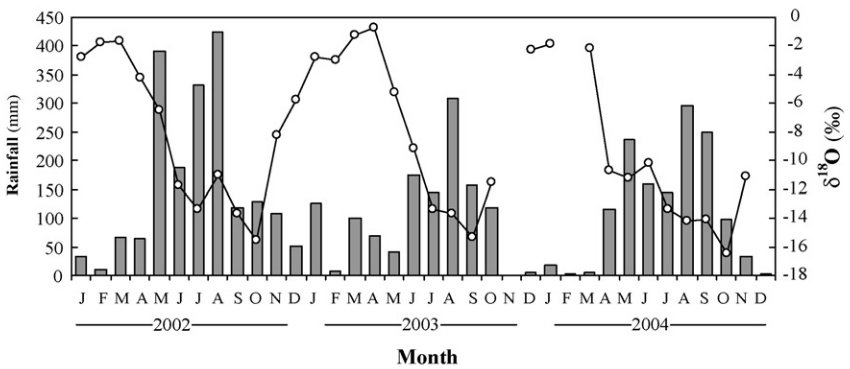

Fig. 6. Seasonal variability of rainfall (bars) and $\delta^{18}O$ (line) of rain samples (monthly means) at the tropical forest site during 2002–2004. Note that values for November 2003, February and December 2004 were not available since no rain occurred in these time.

away from the source are depleted isotopically and rainfall from such clouds are characterized by lower isotope values. Thus, isotopic depletion is expected to be more pronounced in August and September (with more rainfall), 2 months after the monsoon season has begun (Dalai et al., 2002). However, the $\delta^{18}O$ of rainfall in this region continues to drop, reaching the minimum in September or October when rainfall is already significantly lower than during the peak of monsoon (Fig. 6). Similar results were also observed in Yangon, Bangkok, and other stations in the monsoon-controlled region (Araguás-Araguás et al., 1998). In the Indian monsoon-controlled region, the influence of higher summer air temperatures on the overall isotopic fractionation during precipitation is overshadowed by the effects controlled by the amount of precipitation, i.e. the amount effect (Araguás-Araguás et al., 1998), as has been observed in our study ($\delta^{18}O = (-0.029 \pm 0.014)P_{mon} + (-3.9 \pm 1.1)$).

During the dry season, the cold air coming from the Central Asia seldom reaches the study area due to the topographic barrier of the Hengdwan Mountains to the north of the region, leaving this region with drier and clearer weather, less rainfall and low relative humidity. Values of isotopic ratio for this season are more enriched, with a narrower range compared to that of the rainy season (Fig. 2). These high isotope values should point to local, recycled moisture (nearby surface water bodies, evaporation of soil moisture and forest evapotranspiration) as major sources of rainfall during these months (Martinelli et al., 1996; Araguás-Araguás et al., 1998). The relative contribution of downwind evaporation to the local moisture should be another factor enhancing the enriched $\delta^{18}O$ values. However, partial evaporation of raindrops on the way to the rain gauge may also be an important factor leading to the relatively enriched $\delta^{18}O$ values (Gat et al., 1994; Araguás-Araguás et al., 1998; Gat and Airey, 2006). Previous study (Liu et al., 2006) at this site also showed that compared to the rainy season, the d values for the dry season were usually higher (>12), indicating that rainfall was mainly produced by the recycling of the moisture evaporated from surface water plus canopy transpiration (Salati et al., 1979; Gat and Matsui, 1991).

The slope of the regression line for the rainy season (Fig. 2) is virtually identical to the regional line obtained in the Southeast Asia (Araguás-Araguás et al., 1998) and reflects the fact that rainy months largely prevail in determining the yearly means. During the dry season, the slope of the LMWL tends to be slightly greater than 8. One reason for this may be the mixing of moisture from different sources (e.g. recycled moisture through local evaporation and moisture mass originating over Central Asia). Similar result was also found in another study in this region (Wei and Lin, 1994). IAEA (1992) also pointed out that slopes of LMWL larger than 8 are found in areas where the air masses responsible for the relatively enriched isotopic ratio values in rainfall are characterized by high deuterium excess values. Another reason may be partly attributed to the limited isotopic data of the dry season rainfall.

The reason to examine throughfall and stemflow resides in the fact that a significant proportion of rainfall can be trapped in the forest canopy, to be later returned to the atmosphere by evaporation (Bruijnzeel, 2001). If this intercepted rainfall is not completely evaporated, the fraction left behind falls on a local evaporation line (Gat and Matsui, 1991). However, the slopes for throughfall and stemflow obtained with the data from our site tend to be only slightly lower than 7.96 (Table 2). This difference maybe attributed to canopy effects such as storage, evaporation and rate of flushing of canopy water, and also to temporal isotopic variations of rain event (Gat and Matsui, 1991; Martinelli et al., 1996; Gat, 1996, 2000). Previous study at this forest site showed that compared to the rainfall, isotopic values for the throughfall and stemflow were more enriched on the basis of light rain event (Liu et al., 2006). When only light rain events are considered, the slope for the throughfall and stemflow is significantly lower ($P < 0.05$) than 7.96, which indicated that large amounts of precipitated water is returned to the atmosphere through evaporation (Liu et al., 2006).

4.2. Isotopic ratios of surface waters

During the dry season, the slopes for the river, pond and soil are all significantly less than 7.96 (Table 3), with the pond water being the lowest, suggesting that these three water bodies have had significant evaporation. The volume of the pond, which can be significantly reduced during the end of the dry season, also clearly shows that the evaporative flux is very important. The pond water level could also be reduced by lowering of the water table. However, it is not clear whether this effect exists since information of the potential connectivity between the pond and ground water is not available at present. In addition, the river can be one source of evaporated water because there are many subtributaries in this area and the water in these various channels almost stagnates during the dry season. Further, the intercepts for the river, pond and soil are all lower in the dry season than in the rainy season, indicating that these waters might have suffered more evaporation during the dry season than during the rainy season. As an alternative, the *d* values can be used to indicate water evaporation, i.e. values smaller than 8.9 (i.e. *d* of local meteoric water) should be produced in the residual water (Gat and Matsui, 1991; Gat, 1996; Martinelli et al., 1996). The *d* values for the river and pond (Table 1) are significantly lower than 8.9, also suggesting that the waters are subjected to evaporation. The forest soil water could be another possible source of evaporated water (Allison, 1982; Barnes and Allison, 1988; Martinelli et al., 1996). Owing to the large amount of litter covering the ground surface of the forest, the infiltration of rain water into soil is slow, enhancing the possibility of water loss by evaporation. However, the difference of isotopic composition between the soil water and the rainfall could not be solely attributed to evaporation from the soil because of the frequent occurrence of the dense radiation fog (Liu et al., 2005) and of the litter layer. Water contained in the litter layer may be more enriched than the soil water, but this effect should not be considered since the litters were taken out prior to the soil samples collection. An alternative explanation is that the soil water contains some fog drip water during the dry season. This is consistent with the empirical observation that nearly every morning during the dry season, especially during the cool-dry season (November–February) the wetness inside the forest is similar to a light rain owing to the dense fog. During fog event, fog drops cover tree leaves and the soil is wet by fog drip in the morning. Our empirical observation also shows that some fog drip can infiltrate into the forest soil through the litter layer. This enriched fog drip combined with the enriched soil water would explain why soil water tends to have a higher regression slope and intercept and a higher *d* value compared to the pond and river waters (Tables 1 and 3). However, the difference between the slope of the soil water and the pond water was not significant ($P < 0.05$). In addition, the degree of soil saturation (soil moisture content) could affect the slope of the soil water evaporation line (Barnes and Allison, 1988). But this effect remains to be confirmed in further research since information of the soil moisture content is not available at present. For the stream, the regression slope is only slightly lower than that of the LMWL and not significantly different, suggesting that the stream, which origins from and flows through this forest stand appears to have undergone the least evaporation since this study site is an elevated rainforest. The similarity in the slopes of the stream and LMWL also indicates that the isotopic composition of rainfall is well preserved in the stream since the stream water during the base-flow stage originates mainly from the groundwater, as suggested by Dalai et al. (2002).

During the rainy season, the slopes of waters collected from the river and pond are slightly higher than those of the dry season but significantly lower than 7.96 ($P < 0.05$), indicating that during the rainy season, the river and pond are still two potential sources of vapor to the atmosphere. It is also interesting to note that the soil water has the same slope as the rainfall during this time (Table 3), showing that the soil water does not appear to display evaporation effects. In this season, dense fogs and fog drip amounts were not observed as much as in the dry season (only 13% of the annual total), and its effect on the soil $\delta^{18}O$ and δD values should be small. A previous study at this forest site also shows that the soil water is not significantly different from the rainfall in mean $\delta^{18}O$ and δD values and hence can most reasonably be ascribed to that source (Liu et al., 2005). Study from Martinelli et al. (1996) in the Amazonian rain forest also pointed out that soil under the dense canopy in the rainy season has not undergone considerable evaporation.

4.3. Dependence of fog drip on recycled moisture

The stable isotopic compositions of fog and rain at the study site are consistently different as shown in Table 1. Similar result was also found in other previous studies (Aravena et al., 1989; Dawson, 1998). These stable isotopic differences can yield information concerning the origin of fog water (Ingraham and Matthews, 1995).

The magnitude of *d* is determined by conditions of the vapor source (e.g. relative humidity, temperature and wind speed over the evaporating surface) and moisture recycling in the area of precipitation (Gat and Matsui, 1991; Gat et al., 1994; Araguás-Araguás et al., 1998). Low humidity conditions at the source region enhance kinetic evaporation, resulting in a higher *d* in the resultant precipitation. On the other hand, precipitation resulting from a source at high humidity condition has *d* values lower. The radiation fog in our site, which is far away from coast (over 600 km inland), is mainly a result of strong nighttime radiative cooling at the lower atmosphere and generally associated with low wind speeds (<0.7 m s^{-1}) and region-wide air mass stagnation (Liu et al., 2004). The isotopically more enriched fog water with significantly higher *d* value indicates that the radiation fog in this site would contain water that has been terrestrially recycled by evaporation from surface waters (pond, river and soil) plus forest evapotranspiration since it is formed near the land surface. This recycled moisture mixes with water vapor already present in the lower atmosphere, which condenses producing fog that will then be characterized by values of $d > 8.9$. The range and distribution of d for the fog drip in the study site show that samples collected during the dry season have relatively higher d values compared with those in the rainy season (Fig. 4). This suggests that fogs in the dry season (low absolute humidity) contain more terrestrially recycled water than in the rainy season (high absolute humidity). Such an explanation is also consistent with the data in Fig. 5 because the samples collected during the dry season are more enriched in δ^{18}O with more d excess, while samples collected during the rainy season have not this trend. The same can be said of the rainfall during the dry season depicted in Fig. 2. It is seen from the data in Fig. 7 that, in general, the fog drip samples for pre-rain events, which denotes the fog occurs prior to the rain event, are more enriched in δ^{18}O with more d excess than those of the post-rain events, which denotes the fog occurs during 1–3 days following the rain event. Clearly, the pre-rain fogs produced under lower absolute humidity condition should contain more recycled moisture than the post-rain fogs under higher absolute humidity condition. One should note, however, as a point of caution that as the d parameter in precipitation also often changes on shorter time scale, shifts in the d parameters in an air mass cannot always be taken simply as a monitor of the evaporation part of the total evapotranspiration flux (Gat, 2000).

The δ^{18}O versus δD regression line of the fog drip in this site, which almost parallels to the LMWL but with a higher intercept, is also an indication that the condensed fog drip was produced by evaporation. Although water vapor provided through transpiration from plants would not change its isotope composition, it is still isotopically higher than the vapor generated by evaporation (Allison, 1982; Gat, 1996; Martinelli et al., 1996; Gat and Airey, 2006). A clear diurnal cycle in isotopic values of the vapor inside the forest was observed by Ribeiro et al. (1996) and Yepez et al. (2002), which showed that the isotopic values were lighter during the night than during the day for all forest strata and condensation of the vapor caused by a drop in the temperature during the night was the probable cause for this cycle. The increase of isotopic values observed during the day was attributed to the mixture of the remaining vapor from the night with vapor generated by transpiration, plus the re-evaporation of the previously condensed vapor during the day (Martinelli et al., 1996). With this understanding, we can propose that, with the atmosphere becoming gradually drier during the dry season, the condensed vapor (i.e. fog drip in this study) should become isotopically more enriched since the vapor generated by transpiration is just partially condensed during the night and the remaining will mix into the ambient air to join in the next day's cycle. Previous observation in this forest site, that the isotopic values of fog drip were more enriched at the end of the dry season compared to the early dry season (Liu et al., 2005), also confirms this hypothesis. However, the local isotopic signature could be transported out of this system, and the advected air moisture could also move in this system. Both these effects might be relatively lower since the radiation fog is mainly formed near the land surface and this area is generally calm with low wind speeds (mean annual wind speed 0.5 m s^{-1}). But these effects remain to be confirmed in further research.

Overall, the dense rain forest plays not only as an important source of its own moisture but also triggers the fog producing process dynamically since it continues to transpire throughout the year and transpiration is most likely the largest component of evapotranspirative flux, although we could not find the relative contribution of various evaporation sources to the radiation fog. However, climate change has occurred during the last several decades in this region, decreasing the annual number of fog events and duration, and warming temperatures (Li, 2001). Gong and Lin (1996) and Li et al. (2006) proposed that the observed climate change in this region could be partially attributed to the large-scale deforestation. There is an increasing body of work suggesting that substituting primary forest for farmland or other land-use might greatly alter the energy balance and the water balance and consequently the local and regional-scale climate (Wright et al., 1992; Henderson-Sellers et al., 2002; Pitman et al., 2004; Pitman and

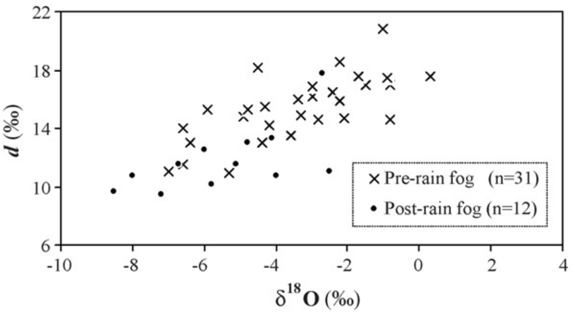

Fig. 7. Plot of δ^{18}O vs. d (deuterium excess) in fog drip samples for pre-rain events and post-rain events collected at the tropical forest site during 2002–2004. Note that pre-rain fog event denotes fog occurs prior to rain event, mostly occurred during dry season, and post-rain fog event denotes fog occurs during 1–3 days following rain event, mostly occurred during rainy season. See text for details.

Narisma, 2005; Gat and Airey, 2006), although the scale of this influence is difficult to determine but critical in our ability to predict climate feedbacks as a function of land cover change (Pitman and Narisma, 2005). There is also a report that large-scale deforestation could reduce cloud formation and increase cloud-base height (Lawton et al., 2001). Hence, it is reasonable to believe that converting densely multilayer rain forest with other land-use and land-cover patterns in this region will reduce fog formation and duration, and may have far-reaching negative impacts on this forest ecosystem itself.

5. Conclusions

The frequent occurrence of radiation fog appears to play an important role in the hydrological cycle of the tropical rain forest in Xishuangbanna. To evaluate the hydrological effect of fog on the forest, information about the sources of fog drip is needed. For this purpose, fog drip, rainfall, throughfall, stemflow, stream water, river water, pond water and soil water were collected at the rain forest site for stable isotopic analysis during 3 years period.

The fog drip, as compared with the other waters, was consistently more enriched, and yielded a higher *d* value. Radiation fog appears to be produced mainly through evaporation from a pond, river, and soil plus forest evapotranspiration, especially during the dry season. Our data suggest that radiation fog during the dry season (low absolute humidity) appeared to contain more terrestrially recycled water than during the rainy season (high absolute humidity). The analyses suggest that evaporation from the stream is limited. Although we could not identify the relative contribution of the different sources of fog, we believe that the rain forest contributes the largest fraction, but a more intense sampling scheme is necessary to answer to this question.

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