A method for estimating the contribution of evaporative vapor from Nam Co to local atmospheric vapor based on stable isotopes of water bodies

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During the summer monsoon season, the moisture of precipitation events in southern and central regions of the Tibetan Plateau is mainly moisture from the Indian Ocean transported by the Indian monsoon and terrestrial vapor derived from the surface of the Tibetan Plateau. However, the respective contributions of these two types of moisture are not clear. From June to September, the excess deuterium values of precipitation and river water in the Nam Co basin are higher than those for the southern Tibetan Plateau. This reflects the mixing of evaporation from Nam Co and local atmospheric vapor. On the basis of theory for estimating the contribution of evaporative vapor from surface water bodies to atmospheric vapor and relative stable isotopes in water bodies (precipitation, river water, atmospheric moisture and lake water), this study preliminarily estimates that the average contribution of evaporation from the Lake Nam Co to local atmospheric vapor has varied from 28.4% to 31.1% during the summer monsoon season in recent years.

atmospheric vapor, stable isotopes, excess deuterium, evaporation, Nam Co

Studies on terrestrial vapor recycling are important in fully determining the various sources of moisture and temporal and spatial characteristics of precipitation in different districts of the Tibetan Plateau (TP). During spring and summer, as a strong source of heat, much latent heat and sensible heat is derived from the surface of the TP [1–3], and it not only plays an important role in the formation of the Indian monsoon [4–6] but also affects the global atmospheric cycle [7,8]. Meanwhile, with the output of much heat from the surface and the strengthening of atmospheric convection during the day [9,10], there is distinct terrestrial vapor recycling, and much atmospheric vapor derived from the surface of the TP enters the atmosphere. Notably, the numerous lakes on the TP generate considerable evaporative vapor.

The total area of these lakes is nearly 44993.3 km^2 [11]. Therefore, the contribution of surface evaporative vapor and especially the vapor from lakes, to atmospheric vapor on the TP cannot be ignored. In this regard, qualitative studies have been carried out. For example, Yang et al. [12] found that there was nearly a water balance among precipitation, evaporation and osmotic water flowing through soil on the western TP owing to local water recycling, which was the major form of water cycling. Yao et al. [13] studied the $\delta^{18}O$ characteristics of snow in the region of Tanggula Mountain and found that moisture in some snow events came from evaporative vapor of the surface of the inner TP. From comparisons among the stable isotopic compositions in precipitation, moisture near ground, and vapor released by transpiration, Kurita and Yamada [14] considered that local evaporative vapor contributed to most precipitation

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events during the summer season of 2004. However, quantitative studies on the contribution of terrestrial moisture to total atmospheric moisture on the TP are lacking.

Water stable isotopes, especially those in atmospheric moisture, have been useful in quantitatively and qualitatively investigating terrestrial vapor recycling in different regions around the world. In the Amazon basin, via the construction of a steady state evapotranspiration model, Gat and Matsui [15] calculated that evaporation from the surface water bodies provides approximately 20%–40% local atmospheric vapor. Analogously, in the Great Lakes region, Gat et al. [16] preliminarily estimated the contribution of evaporation from the Great Lakes to the continental atmosphere. Worden et al. [17] considered that moisture in the troposphere in the tropical zone is mainly derived from oceans and continents through comparison of stable isotope levels in the moisture of the troposphere and precipitation. In eastern Japan, Yamanaka and Shimizu [18] collected vapor samples at different sites in the Kasumigaura lake region, and calculated the proportions of various vapor sources and the contribution of evaporation by the lake to the local atmosphere. Additionally, similar studies have been carried out for the Mediterranean Sea [19], Madagascar off the coast of Africa [20], the Slave River Delta in Canada [21], and California and Nevada in the United States [22].

As the second largest saline lake in China, Nam Co (30°30′–30°55′N, 90°16′–91°03′E, 4719 m a.s.l.) has an area of nearly 2000 km^2 [23] and is located on the central TP and the north piedmont of the Nyainqentanglha Range, mostly in Dangxiong county and Bange county, Tibet. The Nam Co basin possesses all Tibetan subsystems (such as atmosphere, glaciers, snow cover, lakes, frozen earth, and vegetation), and provides advantages for investigating land–atmosphere reciprocity and its effect on climate and atmospheric cycling on the TP. Through numerical modeling, Lü et al. [24] found that the presence of Nam Co depresses the height of the atmospheric boundary layer above and leads to limited moisture and heat preservation in the low zone of the boundary layer. Thus, evaporation from Nam Co can fully mix with input vapor from other regions. In addition, it has been found that the evaporation flux from Nam Co is strong during the thawing season, especially after the monsoon season (autumn and beginning of winter) [25]. All the above suggests that the cycling of evaporative vapor from the Nam Co basin, especially the evaporation of Nam Co lake water, may affect the atmospheric system of the Nam Co basin.

With the above background, this study discusses a method that uses data of stable water isotopes, including $\delta^{18}O$, δ^2H and excess deuterium for water bodies in the Nam Co basin and adjacent regions, to preliminarily estimate the average contribution of evaporation from Nam Co to local atmospheric vapor.

1 Theoretical background

The stable isotopic compositions in marine vapor are controlled by reciprocal processes between air and sea [26,27]. These processes ensure excess deuterium (*d* excess, which is defined as $d=\delta^2H–8\delta^{18}O$ [28]) in the marine vapor. During the transportation of marine vapor by an air mass, the stable isotopes in vapor become more and more depleted owing to continuous precipitation, and the *d* value remains constant if no terrestrial vapor enters the air mass [29]. However, in reality, some precipitation returns to the atmosphere through the evaporation of surface water bodies and transpiration of vegetation. The vapor from transpiration does not change the stable isotopic compositions in former vapor because there is no isotopic fractionation during transpiration [16,30], while evaporative vapor from the surface returns air with a higher *d* value, and mixes with former vapor with a lower *d* value quickly. This leads to the actual *d* value of mixed vapor being higher than that of former vapor [16]. Thus, we can make use of the difference between the *d* values of former and mixed vapor to estimate the contribution of evaporation from surface water bodies to the continental atmosphere [15,20,22]. To evaluate the contribution of evaporation from a water body to the local atmosphere, we need to know the *d* values of the former vapor (usually the input vapor from upwind regions) and local vapor (mixed vapor) in the region of the water body.

Supposing that the contribution of evaporation from a water body is x ($0 \leqslant x < 1$), then the contribution of vapor from the upwind area is $1–x$, and d_A in local atmospheric vapor can be expressed as

$$d_A = d_A'(1-x) + d_E x, \tag{1}$$

where d_A, d_A', and d_E are the *d* values of local atmospheric vapor, input vapor from upwind regions and evaporative vapor from the water body, respectively. Rearranging eq. (1), the formula for calculating the contribution (x) of evaporation from the water body is

$$x = \frac{d_A - d_A'}{d_E - d_A}. \tag{2}$$

d_E can be calculated from the stable isotopic composition (δ_E), which is usually estimated employing a simplified Craig–Gordon [26] water evaporation model as [31]

$$\delta_E = \frac{(\delta_W - \varepsilon^*)/\alpha^* - h\delta_A - \varepsilon_K}{1 - h + \varepsilon_K}$$

$$\approx \frac{(\delta_W - \varepsilon^*)/\alpha^* - h\delta_A - \varepsilon_K}{1 - h}, \tag{3}$$

where δ_A and δ_W are stable isotopic compositions of local atmospheric vapor and water of the water body, respectively. h is the relative humidity upon the water surface. The liquid-vapor equilibrium fractionation factor α^* can be

calculated using empirical equations based on the air temperature at the interface of the water surface and air [32]. ε^* describes equilibrium separation between the liquid and vapor phases, and $\varepsilon^* = \alpha^* - 1$. ε_K can be evaluated with a function of the boundary layer conditions and the humidity deficit [33]: $\varepsilon_K = C_K(1-h)$ with C_K being commonly determined as 0.0125 for hydrogen and 0.0142 for oxygen under typical evaporation conditions for a lake [33].

Expressing eq. (3) for ^{18}O and ^{2}H, respectively, we obtain

$$\delta_E(^{18}O) = \frac{(\delta_W(^{18}O) - \varepsilon^*(^{18}O))/\alpha^*(^{18}O) - h\delta_A(^{18}O) - \varepsilon_K(^{18}O)}{1-h}, \quad (4)$$

$$\delta_E(^2H) = \frac{(\delta_W(^2H) - \varepsilon^*(^2H))/\alpha^*(^2H) - h\delta_A(^2H) - \varepsilon_K(^2H)}{1-h}. \quad (5)$$

According to the definition of d [28] and the empirical equation for ε_K [33], after substituting $\delta_E(^2H)=d_E+8\delta_E(^{18}O)$, $\delta_A(^2H)=d_A+8\delta_A(^{18}O)$, $\varepsilon_K(^2H)=0.0125(1-h)$ and $\varepsilon_K(^{18}O)=0.0142$ $(1-h)$ into eq. (5), the combination of eqs. (4) and (5) yields

$$d_E = \frac{A - hd_A - B}{1-h} + 0.1011, \quad (6)$$

where $A = \dfrac{\delta_W(^2H)}{\alpha^*(^2H)} - \dfrac{8\delta_W(^{18}O)}{\alpha^*(^{18}O)}$ and $B = \dfrac{\varepsilon^*(^2H)}{\alpha^*(^2H)} - \dfrac{8\varepsilon^*(^{18}O)}{\alpha^*(^{18}O)}$. Eq. (6) can be rearranged as

$$d_E - d_A = \frac{A - d_A - B}{1-h} + 0.1011. \quad (7)$$

Finally, substituting eq. (7) into eq. (2), we get

$$x = \frac{(d_A - d'_A)(1-h)}{A - d_A - B + 0.1011(1-h)}. \quad (8)$$

It is notable that all relative parameters used in the equations are presented in decimal notation; e.g. $d_A = 0.010$ and $h = 0.6$.

Summarizing the above descriptions, if we know the values of h, α^*, δ_W, d_A and $(d_A-d'_A)$, the contribution (x) of evaporation from the water body can be estimated using eq. (8).

In this study, to avoid the effect of an elevated d value for precipitation vapor resulting from the formation of ice and snow at super low air temperature [34], we only estimate the contribution during the summer monsoon season from isotopic data of rainfall from June to September.

2 Results and discussion

2.1 Stable isotopic compositions in the Nam Co basin and its southern districts in summer

Previous studies show that precipitation moisture on the

southern and central TP is mainly moisture from the Indian Ocean transported by the Indian monsoon along two typical trajectories (Figure 1) that pass through the Arabian Sea and the Bay of Bengal [35,36]. Table 1 presents mean stable isotopic data (δ^{18}O, δ^2H and d) for the summer precipitation recorded at stations along the two trajectories (Figure 1). Looking at Table 1 from south to north, if we set the Himalayas as the boundary, the mean δ^{18}O, δ^2H and d of the precipitation recorded at stations located on the Indian subcontinent (Kozikode, Maner basin, Bombay, Allahabad, New Delhi and Shillong) are −9.0‰ to −1.0‰, −64.0‰ to −2.0‰, and 8.7‰ to 11.0‰, respectively, while the mean δ^{18}O, δ^2H and d (−20.0‰ to −15.0‰, −153.0‰ to −120.0‰ and 4.0‰ to 5.0‰, respectively) for precipitation recorded at stations located in the Himalayas and on the southern TP (Nyalam, Tingri and Lhasa) are clearly lower than values for the Indian stations. The reason for these clear differences between δ^{18}O and δ^2H values of precipitation south and north of the Himalayas is the obvious depletion of ^{18}O and ^{2}H in monsoon vapor after plentiful precipitation when long-distance vapor from the Indian Ocean crosses the Himalayas [35,37]. There is no current explanation as to why the mean d value of precipitation recorded at stations in the Himalayas and on the southern TP (4.5‰) is lower than that recorded at stations in the southern Himalayas (9.0‰). Theoretically, because we are considering the same precipitation moisture from the Indian Ocean, the d values for the precipitation recorded at the stations mentioned above should be similar.

In Figure 1 and Table 1, the mean d values of summer precipitation recorded at each station on the southern TP (Nyalam, Tingri and Lhasa) from 1998 to 2001 are similar. This suggests that the d values of moisture at these stations are also similar. Because the Nam Co station located on the central TP has the same moisture from the Indian Ocean and is near the upwind three stations on the southern TP, especially the Lhasa station (only 120 km distant) (Figure 1), the d value of the input vapor in the Nam Co basin should be similar to that of the vapor at these three stations, especially that at the Lhasa station. Thus, if there is no evaporation from surface water bodies in the Nam Co basin to mix with the input vapor, the d value of the local vapor should be the same as that of the input vapor. The mean d value of the summer precipitation in the Nam Co basin should also be similar to that recorded at the three stations on the southern TP. However, the mean d value of summer precipitation at the Nam Co station from 2005 to 2008 (11.1‰) is 6.6‰ higher than that at the three stations on the southern TP (4.5‰). This suggests that there should be evaporative vapor with a higher d value from surface water bodies in the Nam Co basin that mixes with the local atmosphere. Meanwhile, a previous study considered that the evaporation flux from Nam Co holds the majority of the total evaporation flux from the surface water bodies in the Nam Co basin [38]. Therefore, the higher d values of summer
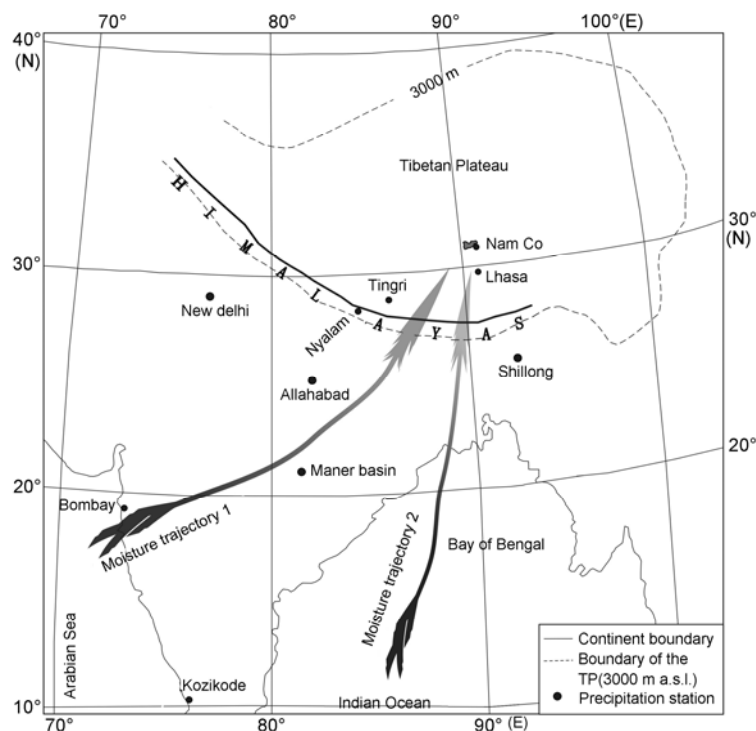
Figure 1 Distribution of stations for collecting precipitation samples on the southern TP and in Indian subcontinent. The two typical trajectories of the Indian monsoon to transport moisture from the Indian Ocean are shown: (1) the moisture trajectory through the Arabian Sea and (2) the moisture trajectory through the Bay of Bengal. These two trajectories are confirmed by backward air trajectories (500 m above the ground), which arrive at the Nam Co basin within 10 days and are modeled by the HYSPLIT air trajectory model provided by the NOAA on the basis of GDAS historical meteorological data for June to September from 2006 to 2009 (http://ready.arl.noaa.gov/hysplit-bin/trajtype.pl?runtype=archive).

Table 1 Stable isotopes in summer precipitation recorded at stations in this study

Station	Data period	Mean δ^{18}O (‰)	Mean δ^2H (‰)	Mean d (‰)	Data source
The south district of the Himalayas					
Bombay	1961–1977	−1.4±1.1	−2.4±8.7	8.7±4.0	
Kozikode	1998–1999	−2.9±1.9	−12.6±13.5	10.2±2.1	
New Delhi	1961–1996	−5.7±4.5	−36.1±33.1	7.8±6.7	[39]
Shillong	1969–1978	−7.1±4.0	−45.6±30.0	10.8±5.2	
Allahabad	1980	−9.0±2.7	−63.4±20.2	8.6±2.3	
Maner basin	1977	−4.7±2.5	−29.4±19.0	8.1±2.8	
The central and south regions of the TP					
Tingri	2000	−19.5±4.2	−152.0±32.3	4.1±4.0	[40]
Nyalam	1998–2001	−15.7	−120.4	5.0	[35]
Lhasa	1998–2001	−18.4	−142.7	4.2	
Nam Co	2005–2008	−19.6	−146.6	11.1	this study

precipitation recorded at the Nam Co station must result from the mixing of evaporation from Nam Co and input vapor from upwind regions.

2.2 Excess deuterium of river water in the Nam Co basin

The characteristic of higher d values of summer precipitation in the Nam Co basin are reflected by the d values of river water in the Nam Co basin. The mean d value of river water samples (12.4‰) is clearly higher than that of river water samples (6.9‰) in the upwind regions (Table 2). Furthermore, it is especially clear that the d values of water in rivers No. 2, No. 3, No. 4, No. 5 and No. 6 are higher than 13.0‰. This is because the main source of these rivers is the meltwater of glaciers and snowpack in the Nam Co basin [41]. The mean d value of meltwater in the Nam Co basin is about 14.0% [38]. However, the mean d value of river water for rivers No.1, No.7 and No. 8 is 8.7‰, which is only a little higher than 6.9‰. This is because the two

main sources of these rivers are precipitation and wetlands, and the supplies of plentiful wetland water with an extremely low mean d value (−1.4‰) weaken the high d value information in the precipitation [38].

2.3   Stable isotopic compositions of Nam Co lake water

The mean $\delta^{18}O$ and δ^2H values of Nam Co lake water (−6.7‰ and −67.7‰) are far higher than those of river water in the Nam Co basin (−16.4‰ and −119.1‰) (Tables 2 and 3). This is due to the gradual enrichment of ^{18}O and ^2H in the lake water resulting from long-term strong evaporation [43]. In addition, Table 3 shows that there is little monthly and yearly variation. Thus, we conclude that the $\delta^{18}O$ and δ^2H values of Nam Co lake water are steady and the mean stable isotopic compositions of Nam Co lake water from 2005 to 2008 are the mean values for lake water in recent years.

2.4   Stable isotopic compositions of atmospheric vapor in the Nam Co basin

It is difficult to measure δ_A directly because of the dynamic nature of the atmosphere [30,31] and strong seasonality [44]. In this study, we use the vapor–precipitation balance to estimate the stable isotopic composition (δ_A) in the vapor of the Nam Co basin during the summer monsoon season.

This method has been successfully employed in many studies for many districts [16,21,45,46]. The calculation is [46]

$$\delta_A = (\delta_P - \varepsilon^*)/\alpha^*, \tag{9}$$

where δ_P is the mean stable isotopic composition of summer precipitation. From the mean air temperature during summer and the mean $\delta^{18}O$ and δ^2H values of summer precipitation recorded at the Nam Co station from 2005 to 2008 (−19.6‰ and −146.6‰, respectively; see Table 1), the mean $\delta^{18}O$ and δ^2H values (δ_A) of the vapor in the Nam Co basin are determined as −30.2‰ and −224.1‰, respectively. Accordingly, the d value of the vapor is 17.6‰.

2.5   Contribution of evaporation from Nam Co to local atmospheric vapor

(1) Estimation of the average contribution during summer. During summer, the annual d values of the precipitation at the Nam Co station for 2005, 2006, 2007 and 2008 are 10.2‰, 12.5‰, 10.7‰ and 11.3‰, respectively. Thus, the mean d value of summer precipitation at the Nam Co station from 2005 to 2008 can be considered equal to the mean d value of summer precipitation at the Nam Co station for the past 10 years. The difference between the d value of the precipitation at the Nam Co station (11.1‰) and the mean d value of the

Table 2   Mean d values of river water samples in the Nam Co basin and southern upwind regions

River No.	River name	$\delta^{18}O$ (‰)	δ^2H (‰)	d (‰)	Data source
		Nam Co basin			
1	Niyaqu	−17.2	−128.5	9.0	
2	Langma	−17.3	−124.4	14.1	
3	Zhaoshaqu	−17.2	−122.8	15.0	
4	Qugaqie	−16.1	−115.8	13.2	
5	Lanong	−15.2	−106.5	14.7	this study
6	Bilang	−15.5	−107.7	16.2	
7	Aangqu	−17.1	−128.4	8.1	
8	Boqu	−16.0	−118.7	9.2	
Mean value		−16.4	−119.1	12.4	
	The upwind regions of the Nam Co basin (the north of the Himalayas)				
	Lhasa River	–	–	~8.5	
	Dangxiong River	–	–	~7.5	
	Yarlung Zangbo River	–	–	~7.5	
	Shigatse River	–	–	~6.5	[42]
	Tingri River	–	–	~6.0	
	Langlongcuo Pass River	–	–	~5.5	
Mean value		–		~6.9	

Table 3 $\delta_W{}^{18}O$ and $\delta_W{}^2H$ of Nam Co lake water in summer from 2005 to 2008

Year	$\delta_W{}^{18}O$, $\delta_W{}^2H$ in June (‰)	$\delta_W{}^{18}O$, $\delta_W{}^2H$ in July (‰)	$\delta_W{}^{18}O$, $\delta_W{}^2H$ in August (‰)	$\delta_W{}^{18}O$, $\delta_W{}^2H$ in September (‰)	$\delta_W{}^{18}O$, $\delta_W{}^2H$ in summer (‰)	Data source
2005	–	–	−6.5, −64.0	−7.1, −69.7	−6.8, −66.8	
2006	−6.7, −67.6	−6.8, −64.9	−6.5, −63.6	–	−6.6, −65.4	
2007	−6.4, −67.0	−6.4, −67.4	−6.6, −68.5	−7.0, −71.0	−6.6, −68.5	this study
2008	−6.5, −67.5	−6.7, −69.0	−7.0, −70.7	−7.1, −72.3	−6.8, −69.9	
The mean $\delta_W{}^{18}O$, $\delta_W{}^2H$ value from 2005 to 2008					−6.7, −67.7	

precipitation at the Nyalam, Tingri and Lhasa stations (4.5‰) might be the difference between the mean d values of precipitation in the Nam Co basin and its upwind region; i.e. $(d_A-d'_A) = 6.6‰$. As presented above, the mean $\delta_W{}^{18}O$ and $\delta_W{}^2H$ of summer Nam Co lake water and the mean d_A value of the atmospheric vapor in the Nam Co basin from 2005 to 2008 are –6.7‰, –67.7‰ and 17.6‰, respectively. Meanwhile, from the mean air temperature (7.8°C) during summer from 2005 to 2008, the parameters A and B in eq. (8) are calculated as –0.00853 and 0.004231, respectively. Substituting these data and the average evaporation-weighted relative humidity (h is 61%–62%) at the Nam Co station into eq. (8), the contribution of evaporation from Nam Co to local atmospheric vapor is estimated to be 28.4%–31.1%. According to the estimation, there is plentiful evaporation from Nam Co to mix with vapor from the upwind region and increase the d value of local vapor and precipitation in the Nam Co basin. Additionally, the estimation made in this study is similar to the results of previous similar studies (Table 4).

　　(2) Error analysis.　According to eq. (8), the estimation of the contribution strongly depends on the exactness of parameters h, d_A and $(d_A-d'_A)$. The $(d_A-d'_A)$ value is based on field measurements made over several years and it is believable and exact. The d_A value is estimated employing the vapor–precipitation balance, and relative humidity (h) of the lake water surface is substituted with the relative humidity of air recorded at the Nam Co station; therefore, these two parameters have certain inevitable errors. In Table 5, an increase or decrease in $(d_A-d'_A)$ by 5% results in a change in the estimated contribution (x) of 5% or –5%. Relative humidity (h) is the most critical parameter in the estimation. If h increases or decreases by 5%, the estimated contribution (x) changes by 39.7% or –19.5%. The parameter d_A is another critical factor of the estimation result, and an increase or decrease in d_A by 5% results in a decrease or increase in

the estimated contribution (x) by 5%. Summarizing the analysis above, h and d_A may be the main sources of error in the estimation of the contribution (x).

3　Conclusions and prospects

During the summer monsoon season, owing to long-distance transportation of moisture from the Indian Ocean, the $\delta^{18}O$, δ^2H and d values of precipitation on the Indian subcontinent are clearly higher than those in the northern Himalayas (the southern TP). The d values of summer precipitation recorded at Nyalam, Tingri and Lhasa stations are very similar, while the mean d value of the summer precipitation at the Nam Co station located on the central TP (11.1‰) is higher than that at those three stations. This suggests mixing of evaporation from Nam Co and the local atmospheric vapor. Meanwhile, the mean d value of river water samples in the Nam Co basin (12.4‰) is also higher than that in the upwind regions (6.9‰). Additionally, this study preliminarily estimates that the average contribution rate of evaporative vapor from Nam Co to local atmospheric vapor has varied from 28.4% to 31.1% during the summer monsoon season in recent years. The estimation of the contribution is a preliminary result based on limited stable isotopes in water bodies and meteorological data that can be obtained at present. Thus, its exactness and validity require more relative data (especially field measurements of atmospheric vapor) and further study for improvements and validation.

Table 4　Comparison of the results of similar studies for the contribution of evaporation from a surface large water body to continental vapor based on stable isotopes

Contribution (%)	Study region	Data source
20–40	Amazon basin	[15]
~20	Northern and central California	[22]
5–16	North American Great Lakes	[16]
≥16–50	Ihorty lake, Madagascar	[20]
10–20	Kasumigaura lake, Japan	[18]
0–45	Great Slave River Delta, Canada	[21]
28.4–31.1	Nam Co basin, Tibet, China	this study

Table 5　Percentage changes in the contribution of evaporation from Nam Co to local vapor for a change in input parameters by ±5%

Parameters (change by 5%/–5%)	$d_A-d'_A$	h	d_A
Corresponding change of contribution (%)	5/–5	39.7/–19.5	–8.8/10.7

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