Below-Cloud Evaporation of Precipitation Isotopes over Mountains, Oases, and Deserts in Arid Areas

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ABSTRACT: As raindrops fall from the cloud base to the ground, evaporation below those clouds affects the rain's isotope ratio, reduces precipitation in arid areas, and impacts the local climate. Therefore, in arid areas with scarce water resources and fragile ecological environments, the below-cloud evaporation is an issue of great concern. Based on 406 event-based precipitation samples collected from nine stations in the Shiyang River basin (SRB) in the northwest arid area, global meteorological water line (GMWL) and local meteorological water line (LMWL) are compared, and the Stewart model is used to study the effect of spatial and temporal variation of below-cloud evaporation on isotope values in different geomorphic units at the SRB. Furthermore, factors influencing below-cloud evaporation are analyzed. The results show that 1) the change of d-excess (Δd) in precipitation at the SRB and the residual ratio of raindrop evaporation (f) vary in time and space. With regard to temporal variation, the intensity of below-cloud evaporation is described by summer < autumn < winter < spring. Regarding spatial variation, the below-cloud evaporation in mountain areas is weaker than in oases and deserts. The intensity of below-cloud evaporation in mountain areas increases with decreasing altitude, and the below-cloud evaporation in oasis and desert areas is affected by local climatic conditions. 2) Below-cloud evaporation is also affected by local transpiration evaporation, especially around reservoirs. Reservoirs increase the relative humidity of the air nearby, weakening below-cloud evaporation. This study deepens our understanding of the water cycle process in arid areas.

KEYWORDS: Atmosphere; Complex terrain; Rivers; Evaporation; Hydrometeorology; Isotopic analysis

1. Introduction

Studying the evaporation of precipitation below clouds helps us to understand the water cycle of a region, as well as the microscopic model, mechanism, and evolution information of the water cycle (Dansgaard 1964; Aggarwal et al. 2016; Zwart et al. 2016; Craig 1961). Water vapor condenses in the atmosphere as clouds and then falls; below-cloud evaporation refers to the evaporation that occurs as precipitation travels to the ground (Froehlich et al. 2008). Generally, when water vapor saturation in the atmosphere is lower, below-cloud evaporation is greater (S. Wang et al. 2016). Since Rayleigh proposed the Rayleigh fractionation model (Dansgaard 1964; Dansgaard et al. 1993), it has also been applied to studying the regional water cycle (Druhan and Maher 2017). However, relevant studies have shown that different degrees of below-cloud evaporation occur when raindrops fall from the cloud base to the ground (S. Wang et al. 2016; Pan et al. 2018), and the below-cloud evaporation affects the isotope ratio in raindrops, causing changes in the slope and intercept of the local meteorological water line (LMWL) (Sun et al. 2020; Zhao et al. 2015).

The spatiotemporal changes of δD and $\delta^{18}O$ values in precipitation depend on variation in the water vapor sources, topographic features, surface transpiration and evaporation, local meteorological factors, and moisture rainout from

source evaporation to precipitating region en route (Sun et al. 2020; Wang et al. 2017; Salamalikis et al. 2016; Stumpp et al. 2014; Crawford et al. 2017; Zhu et al. 2016; Li and Garzone 2017; Kong and Pang 2016; Guan et al. 2009). After analyzing the impact of meteorological factors and geographic units on stable isotopes in precipitation, the significance of those stable isotopes to the geographical environment is discussed (Dansgaard et al. 1993; Johnsen et al. 1972; Lorius et al. 1985; Kreutz et al. 2003). In recent years, studies have found that below-cloud evaporation has a significant impact on rain isotopes in the local area (Peng et al. 2007), whereby geomorphologic features have a significant effect on the below-cloud evaporation, showing that the intensity of below-cloud evaporation is higher in plains than in mountainous areas (Froehlich et al. 2008). Stable isotopes in central and eastern Asia are affected by below-cloud evaporation, especially in the arid areas of central Asia and the northern areas of the Qinghai-Tibet Plateau (Kong et al. 2013; S. Wang et al. 2016; Yuan et al. 2019; Wang et al. 2017). According to previous studies (Peng et al. 2011; Froehlich et al. 2008), precipitation d-excess ($d\text{-excess} = \delta^2H - 8\delta^{18}O$) usually shows a strong positive correlation with ground temperature. Below-cloud evaporation in the Alps was studied and, under the assumption of a fixed raindrop diameter and cloud-base altitude, it was found that for every 1% increase in f (raindrop evaporation remaining ratio), d-excess decreases by about 1‰, linearly (Froehlich et al. 2008). When raindrops fall through dry air masses, they tend to evaporate more easily (Stewart 1975). Rain falling to the ground from the cloud base shows higher $\delta^{18}O$ and δD values (Peng et al. 2007). At the same time, due to the influence of kinetic isotope fractionation, the slope and intercept of the LMWL line will both decrease (Li et al. 2016a; Peng et al. 2007).

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FIG. 1. Sketch map of the study area and the distribution of sampling points.

The Shiyang River basin (SRB) topography is divided into the mountain area of the upper reaches, the oasis area in the middle reaches, and the desert area in the lower reaches. It is well suited to study below-cloud evaporation under different mountains, oases, and deserts under different geomorphic units. Previous studies focused on precipitation sources, moisture recycling, and below-cloud evaporation of precipitation in arid areas (Kong et al. 2013; S. Wang et al. 2016; Pan et al. 2018; Sun et al. 2020; Zhu et al. 2016; Li and Garzzone 2017; Kong and Pang 2016). However, because it is difficult to collect samples in high-altitude areas and there is little precipitation in deserts, there are few comparisons of stable isotope in precipitation and below-cloud evaporation for mountains, oases, and deserts. Studies on the precipitation isotope of below-cloud evaporation in arid areas often rely on a few stations in specific areas. Most previous studies focused on the arid basin or a certain mountainous area. Thus, they have ignored the vast geographical and environmental differences between mountainous areas, oases, and deserts in arid areas. In our study, based on intensive sampling data in the SRB area, the temporal and spatial differences in precipitation isotopes of below-cloud evaporation in mountains, oases, and deserts were analyzed.

2. Study area

The SRB (36°29'–39°27'N and 101°41'–104°16'E; Fig. 1) covers an area of approximately 41 600 km² and is located in the east of the Hexi Corridor, west of Wushaoling and south of the Qilian Mountains on the northern edge of the Qinghai–Tibet Plateau. The SRB has a high elevation in the south, 2000–5000 m above sea level, and a low elevation in the north, where the altitude is less than 2000 m (L. Wang et al. 2016). It inclines

from southwest to northeast, and the entire basin can be divided into three geomorphic units from south to north: the Qilian Mountains in the upper reaches, the Hexi Corridor Plain in the middle reaches (low hilly areas), and desert areas in the lower reaches (Zhu et al. 2019).

Lenglongling is in the eastern part of the Qilian Mountains, and the northern slope of Lenglongling is the source of the Shiyang River. The valley rivers, which are mainly composed of abundant snowmelt from the surrounding mountains and atmospheric precipitation, form the Shiyang River after the river confluence. The annual average runoff of the SRB is 15.91×10^8 m³, and the runoff area is about 1.1×10^4 km² (Zhu et al. 2018). Most of the watersheds are present in the arid areas. The temperatures in January are the yearly lowest (average temperature: −7.8°C), and in July are the highest yearly (average temperature: 21.5°C); the annual average temperature is 8.07°C, and annual hours of sunshine range between 2700 and 3030 h. The average annual precipitation is 154.8 mm, mainly from May to September, accounting for 83% of the annual precipitation, and annual evaporation is 2000–3000 mm (Ma et al. 2013). There is a large area of permafrost in the upper Qilian Mountains and a small number of glaciers in Lenglongling, the highest point of the Qilian Mountains. The soil in the middle and lower reaches of the Qilian Mountains is zonal desert soil. Wuwei and Minqin basins are oasis areas, greatly affected by artificial irrigation. The elevation of 1775–2700 m is the mountain area in the upper reaches of the Shiyang River, with hills, intermountain basins, and a concentration of small piedmont plains (Ma et al. 2013; Li et al. 2016b). The annual average temperatures of the mountains, oases, and deserts are 4.9°, 10.4°, and 12°C; relative humidity is 61.3%, 42.9%, and 37.8%, while precipitation is 536.5, 112.5,

TABLE 1. Basic information of sampling points in SRB. Note: temperature, precipitation, and relative humidity are average values during the sampling period, the sampling date is October 2016–April 2018, “M” stands for mountain areas, “O” stands for oasis areas, and “D” stands for deserts.

Sampling station and abbreviation		Geographical parameter			Meteorological parameters			No. of samples
		Lon (°E)	Lat (°N)	Altitude (m)	<i>T</i> (°C)	<i>P</i> (mm)	RH (%)	
Lenglongling	M1	101°50′	37°33′	3553	−0.19	595.1	69.2	89
Ningchan	M2	101°53′	37°41′	2721	3.34	431.9	66.6	54
Yangjiazhuang	M5	102°48′	37°25′	2357	6.81	1029.5	50.42	37
Huajian	M3	102°00′	37°50′	2323	6.6	363.5	60.4	53
Xiying	M4	102°11′	37°53′	2097	7.9	262.5	59.8	58
Yangxiaba	O1	102°41′	38°2′	1489	10.8	29.6	46.4	21
Wuwei	O2	102°4′	37°55′	1531	10.15	186.5	42.4	52
Minqin	O3	103°05′	38°38′	1367	10.3	121.4	39.8	24
Qingtuhu	D	103°38′	39°08′	1308	12	79.7	37.8	18

and 79.7 mm. From the mountains to the oases to the deserts, the temperature rises, relative humidity decreases, and precipitation decreases (Zhu et al. 2019).

3. Data sources and research methods

a. Sample collection and analysis

Yangjiazhuang, Lenglongling, Xiying, Huajian, and Ningchan in the mountain areas of the SRB; Yangxiaba, Wuwei, and Minqin in the oasis area; and Qingtuhu in the desert area were selected as precipitation collection points (Table 1). The sampling period was from October 2016 to April 2018. Rainwater samples were collected immediately after each period of precipitation using a plastic funnel bottle, putting it into a 50-mL polyethylene sampling bottle, closing the bottle cap tightly, and then sealing the bottle mouth with a sealing film. All precipitation samples were placed in cold storage until their analysis to minimize evaporation; 406 samples of atmospheric precipitation were collected. Air temperature, atmospheric pressure, precipitation, and relative humidity during precipitation events were measured every hour using automatic weather stations. The meteorological data we analyzed were obtained from automatic weather stations on the ground. All the content collected data were calculated by the improved Stewart model.

The δD and $\delta^{18}O$ of all samples were measured using DLT-100 isotope analyzer (Los Gatos Research, United States) at the isotope laboratory of Northwest Normal University. Each precipitation sample and isotope standard sample was injected six times. The first two injections were discarded and the average result of the last four injections used in the analysis. The isotope analysis results are expressed by the symbol “ δ ” and relative to Vienna Standard Mean Ocean Water (V-SMOW):

$$\delta_{\text{sample}} (\%) = (R_{\text{sample}}/R_{\text{V-SMOW}} - 1) \times 1000, \quad (1)$$

where R_{sample} is the ratio of $^{18}O/^{16}O$ or $^2H/^1H$ in the samples, and R_{standard} is the ratio of $^{18}O/^{16}O$ or $^2H/^1H$ in V-SMOW (Froehlich et al. 2008). The δ^2H of atmospheric precipitation in the study area is mostly in the range from -97.3‰ to -9.2‰ . Therefore, LGR3C, LGR4C, and LGR5C were selected as reference standard samples. The three standard samples tested

were: LGR3C ($\delta^{18}O = -13.39\text{‰}$, $\delta D = -97.3\text{‰}$), LGR4C ($\delta^{18}O = -7.94\text{‰}$, $\delta D = -51.6\text{‰}$), LGR5C ($\delta^{18}O = -2.69\text{‰}$, $\delta D = -9.2\text{‰}$) (Xu et al. 2020). The precision is $\pm 0.6\text{‰}$ for δ^2H and $\pm 0.2\text{‰}$ for $\delta^{18}O$ (Zhu et al. 2019).

b. Stewart model research

Under the influence of the below-cloud evaporation, the stable isotope values in raindrops at the cloud base are different from those on the ground. Below-cloud evaporation can be represented by the isotope difference between raindrops at the base of the cloud and raindrops falling to the ground. This assumes that precipitation at the cloud base and the surrounding water vapor reach a state of fractionation equilibrium (Froehlich et al. 2008). The difference between d-excess in the ground and d-excess in the cloud base can be expressed by Δd (Stewart 1975):

$$\Delta d = \left(1 - \frac{2\gamma}{\alpha}\right)(f^{2\beta} - 1) - 8\left(1 - \frac{18\gamma}{18\alpha}\right)(f^{18\beta} - 1), \quad (2)$$

where 2γ , 18γ , 2β , and 18β are defined by Stewart (1975); 2α and 18α are equilibrium fractionation coefficients (Friedman and O’Neil 1977; Criss 1999); “2” is used for δ^2H and “18” for $\delta^{18}O$; f is the raindrop evaporation residual ratio. The calculation method of 2γ , 18γ , 2β , and 18β (Stewart 1975) is

$$2\gamma = \frac{2\alpha RH}{1 - 2\alpha(^2D/^2D')^n(1 - RH)}, \quad (3a)$$

$$18\gamma = \frac{18\alpha RH}{1 - 18\alpha(^{18}D/^{18}D')^n(1 - RH)}, \quad (3b)$$

$$2\beta = 1 - 2\alpha \frac{(^2D/^2D')^n(1 - RH)}{2\alpha(^2D/^2D')^n(1 - RH)}, \quad \text{and} \quad (3c)$$

$$18\beta = 1 - 18\alpha \frac{(^{18}D/^{18}D')^n(1 - RH)}{18\alpha(^{18}D/^{18}D')^n(1 - RH)}. \quad (3d)$$

According to Stewart’s research, in these formulas, 2α and 18α are equilibrium fractionation coefficients (Friedman and O’Neil 1977; Criss 1999), RH is relative humidity (the unit of RH is the decimal equivalent), $^2D/^2D'$ and $^{18}D/^{18}D'$ are 1.024 and 1.0289, and n is 0.58 (Merlivat 1970; Stewart 1975).

Parameters $^2\alpha$ and $^{18}\alpha$ are

$$^2\alpha = \exp\left(\frac{24.844 \times 10^3}{T^2} - \frac{76.248}{T} + 52.612 \times 10^{-3}\right), \quad \text{and} \quad (4a)$$

$$^{18}\alpha = \exp\left(\frac{1.137 \times 10^3}{T^2} - \frac{0.4156}{T} + 2.0667 \times 10^{-3}\right). \quad (4b)$$

In the above formulas, T is the condensation temperature (K).

The raindrop evaporation remaining ratio f is the remaining mass of raindrops after evaporation as a percentage of the original mass of raindrops falling from the cloud base, and is expressed as follows:

$$f = \frac{m_{\text{end}}}{m_{\text{end}} + m_{\text{ev}}}, \quad (5)$$

where m_{ev} is the raindrop mass lost by evaporation under the cloud and m_{end} the raindrop mass when they fall (Kinzer and Gunn 1951; S. Wang et al. 2016). The mass m_{ev} is defined as

$$m_{\text{ev}} = r_{\text{ev}} t, \quad (6)$$

where r_{ev} is the evaporation rate of raindrops and t is the time it takes for the rain to fall from the cloud base to the ground:

$$t = \frac{H}{v_{\text{end}}}, \quad (7)$$

where H is the distance from the cloud base to the ground, that is, the distance over which the raindrop falls, and v_{end} is the final speed at which raindrops fall:

$$v_{\text{end}} = 9.58e^{0.0354H} \left\{ 1 - \exp\left[-\left(\frac{D}{1.77}\right)^{1.147}\right] \right\}. \quad (8)$$

In the above formula, v_{end} is the final velocity of the falling raindrops (m s^{-1}) (Yang 2011), H is the height of the cloud base (km) (Best 1950), and D is the diameter of raindrops (mm).

In previous studies, precipitation height was mostly set at a default of 1500 m (850 hPa) (Wang et al. 2009). However, in many alpine mountain areas, precipitation height is often less than 1500 m due to the influence of terrain and meteorological factors. Therefore, in this study, the cloud-base height (H_{cb}) was calculated by the Laplace pressure formula:

$$H = 18400 \left(1 + \frac{T_{\text{mean}}}{273} \right) \log \frac{P}{P_{\text{LCL}}}, \quad (9)$$

where T_{mean} is the average temperature between the condensation height and Earth's surface ($^{\circ}\text{C}$), where the arithmetic mean value of Earth's surface temperature and the lifting condensation level (LCL) temperature is used to express P and P_{LCL} is the atmospheric pressure at the surface and at the condensation level, respectively:

$$P_{\text{LCL}} = P \left(\frac{T_{\text{LCL}}}{T} \right)^{3.5}, \quad (10)$$

where P is the ground pressure of the observation point (hPa), and T_{LCL} and T are condensation height and surface temperature ($^{\circ}\text{K}$), respectively.

The cloud-base height is determined as the LCL. The temperature at the condensation level (T_{LCL}) is estimated using the following formula (Barnes 1968):

$$T_{\text{LCL}} = T_d - (0.001296T_d + 0.1963)(T - T_d), \quad (11)$$

where T_d and T are dewpoint temperature and air temperature at the point of observation ($^{\circ}\text{C}$), respectively.

The raindrop diameter is estimated using the following formula (S. Wang et al. 2016):

$$D = \sqrt[3]{0.69AI^p}, \quad (12)$$

where D is the raindrop diameter (mm), I is the rainfall intensity (mm h^{-1}), and the parameters n , A , and p are 2.25, 1.30, and 0.232, respectively. The term I represents the intensity of precipitation, that is, the amount of precipitation per hour: $I = P/24$ (P stands for daily precipitation).

The raindrop evaporation rate (r_{ev}) is the mass of raindrops evaporated per unit time, expressed as the product of Q_1 and Q_2 (Kinzer and Gunn 1951):

$$r_{\text{ev}} = Q_1(T, D)Q_2(T, h), \quad (13)$$

where Q_1 (cm) is a function of ambient temperature in degrees Celsius and raindrop diameter in cm; Q_2 ($\text{g cm}^{-1} \text{s}^{-1}$) represents a function of ambient temperature in degrees Celsius and relative humidity as a percentage. The values of Q_1 and Q_2 for specific conditions (i.e., $T = 0^{\circ}, 10^{\circ}, 20^{\circ}, 30^{\circ}$, and 40°C ; $D = 0.01, 0.02, 0.03, \dots, 0.04$ cm and $h = 10\%, 20\%, 30\%, \dots, 100\%$) were presented by Kinzer and Gunn (1951). To acquire Q_1 and Q_2 at specific meteorological conditions for each precipitation event in our study, a bilinear interpolation method was used.

While the shape of raindrops undergoes complex changes during the process of falling (Thurai et al. 2009; Szakáll et al. 2010), in the simulation of precipitation raindrops are generally assumed to be a sphere (Yang 2011) and in turn the mass of raindrops falling to the ground can be expressed as

$$m_{\text{end}} = \frac{4}{3} \pi r_{\text{end}}^3 \rho, \quad (14)$$

where m_{end} is the mass of the raindrops as they reach the ground, r_{end} is the radius of the raindrops when they reach the ground and ρ is the density of water.

According to the meteorological data of nine sampling points in the SRB, the Stewart model was used to calculate f and Δd . The typical parameter values in the formulas are shown in Table 2.

c. Cloud-base height (H_{cb}) test method

The LCL of convective clouds can be estimated using the Romps (2017) equation. The necessary input parameters required to estimate LCL obtained from various satellite data. We recalculated the height of the cloud base using the following equation and compared it to the earlier calculation:

TABLE 2. Typical values of various parameters at each sampling point in SRB from October 2016 to April 2018; these parameters are used in the Stewart model.

Sampling locations	D (mm)	H_{LCL} (m)	V_{end} (m s ⁻¹)	t (s)	r_{ev} (g s ⁻¹)	M_{ev} (g)	M_{end} (g)	f	Δd
Mountains									
M ₁ (Lenglongling)	0.75	611.47	3.04	213.83	0.39	91.24	247.90	71.37	-60.06
M ₂ (Ningchan)	0.79	591.17	3.19	196.25	0.48	104.02	296.79	73.01	-55.48
M ₃ (Haujian)	0.73	1063.82	3.01	370.83	0.78	287.78	237.00	47.76	-176.90
M ₄ (Xiying)	0.75	681.45	3.06	233.17	0.56	134.71	237.57	64.98	-82.95
M ₅ (Yangjiazhuang)	1.07	866.14	4.21	209.49	1.21	275.97	702.51	72.17	-52.22
Oasis									
O ₁ (Yangxiaba)	0.76	1080.38	3.12	367.82	1.05	414.46	247.54	44.76	-333.07
O ₂ (Wuwei)	0.80	691.88	3.24	221.07	0.90	242.56	301.47	61.98	-123.79
O ₃ (Minqin)	0.79	726.53	3.22	232.18	1.05	293.38	283.08	59.71	-164.58
Desert									
D (Qingtuhu)	0.66	741.09	2.68	316.54	0.60	227.88	195.95	53.76	-366.83

$$Z_{\text{LCL}} = Z + \frac{C_{pm}}{g} (T - T_{\text{LCL}}), \quad (15a)$$

$$C_{pm} = (1 - q_v)C_{pa} + q_v C_{pv}, \quad (15b)$$

$$C_{pa} = C_{va} + R_a, \text{ and} \quad (15c)$$

$$C_{pv} = C_{vv} + R_v. \quad (15d)$$

These equations give the parcel's temperature and height at its LCL (T_{LCL} and Z_{LCL}) in terms of its initial temperature and height (T and Z), the unit in T_{LCL} and T is centigrade (°C), the unit in Z_{LCL} and Z is height (m), C_{pm} is its specific heat capacity at constant pressure, q_v is the mass fraction of water vapor, C_{pa} is the specific heat capacity at constant pressure for dry air, C_{pv} is the specific heat capacity of water vapor at constant pressure, the parameters C_{va} , R_a , C_{vv} , R_v , and g are 719, 287.04, 1418, and 461 J kg⁻¹ K⁻¹ and 9.81 m s⁻², respectively,

$$T_{\text{LCL}} = c[W_{-1}(\text{RH}_l^{1/a} \text{ce}^c)]^{-1}T, \quad (16a)$$

$$a = \frac{C_{pm}}{R_m} + \frac{C_{vl} - C_{pv}}{R_v}, \quad (16b)$$

$$b = -\frac{E_{0v} - (c_{vv} - c_{vl})T_{\text{trip}}}{R_v T}, \quad (16c)$$

$$c = b/a, \quad \text{and} \quad (16d)$$

$$R_m = (1 - q_v)R_a + q_v R_v, \quad (16e)$$

where the unit is percent; R_m is the air parcel's specific gas constant; and the parameters E_{0v} , T_{trip} , and C_{vl} are 2.3740×10^6 J kg⁻¹, 273.16 K, and 4119 J kg⁻¹ K⁻¹, respectively.

4. Results and analysis

a. Stable isotope characteristics of below-cloud evaporation in precipitation

Rainwater subjected to below-cloud evaporation shows a $\delta^{18}\text{O}$ and δD enrichment, providing qualitative information about whether raindrops have undergone evaporation when they fall from the bottom of the cloud to the ground (Kong and

Pang 2016). Figure 2 shows the LMWL of mountains, oases, and deserts in SRB from October 2016 to April 2018. From the mountainous area to the oases and then to the desert area, the slope of the LMWL is 7.96 (± 0.075), 7.84 (± 0.185), and 6.37 (± 0.506), respectively. Compared with the global meteorological water line (GMWL) ($\delta\text{D} = 8\delta^{18}\text{O} + 10$) and the LMWL in China ($\delta\text{D} = 7.9\delta^{18}\text{O} + 8.2$) (Zhang and Wang 2016; Yuan et al. 2019; Zhu et al. 2016; Li et al. 2015), the slope of the LMWL in mountains, oases, and deserts is smaller than the slope of the GMWL and of the LMWL in China. This is mainly related to the changes in altitude of mountains, oases, and deserts. The lower the altitude, the stronger the corresponding below-cloud evaporation. Comparing the three topographies of the SRB, the slope and intercept of the LMWL vary greatly in the mountains, oases, and deserts. From the perspective of the slope of the LMWL, that of the mountain area is higher than that of the oasis and of the desert. It shows that moving from the mountain to the oasis and then to the desert, the below-cloud evaporation intensity progressively increases. Generally speaking, the SRB still follows the principle that the below-cloud evaporation in the upper reaches of the mountain is less than that of the oasis and the desert, but there are also specific changes affected by local meteorological factors.

From Table 3, we can see that the slope and intercept of the LMWL increase systematically with the increase of rainfall in the basin; the slope increases from 7.48 (± 0.21) to 7.64 (± 0.17), and the intercept increases from 11.8 (± 0.07) to 13.09 (± 0.14). The smaller the slope and intercept, the stronger the below-cloud evaporation. It can thus be inferred that the kinetic isotope fractionation caused by intense below-cloud evaporation mainly occurs during small rainfall events (Yuan et al. 2019; Zhao et al. 2015). From Table 3, it can be seen that the slope and intercept of the LMWL based on samples ≥ 10 mm are 7.74 (± 0.21) and 14.31 (± 1.8), respectively, on average, which are higher than the slope and intercept of samples of the LMWL based on 0–10 mm. Therefore, the below-cloud evaporation of raindrop diameters ≥ 10 mm is weak. With the increase of air temperature, $\delta^{18}\text{O}$ and δD increase, with the average values of $\delta^{18}\text{O}$ and δD being -5.07‰ and -26.89‰ , respectively, at temperatures $\geq 15^\circ\text{C}$. The slope of the LMWL dropped to 7.65 (± 0.2),

FIG. 2. The LMWL in the (a) mountain, (b) oasis, and (c) desert areas of the SRB from October 2016 to April 2018.

indicating that the higher the temperature, the stronger the below-cloud evaporation.

Regarding the impact of RH on the LMWL, when RH is less than 70% the slope and intercept are $7.27 (\pm 0.19)$ and $9.44 (\pm 1.64)$, respectively, so the below-cloud evaporation is the strongest in this range. This is because, with the increase of RH, the water vapor in the air gradually saturates and the water pressure increases, which causes the below-cloud evaporation to gradually weaken (Yuan et al. 2019). Finally, comparing the effects of

different raindrop diameters on below-cloud evaporation, the $\delta^{18}\text{O}$ and δD increase with increases in raindrop diameters, as is expected. When $D \geq 1$ mm, the slope and intercept of the LMWL are less than those calculated for $D < 1$ mm. Therefore, we speculate that the larger the raindrop diameter, the stronger the below-cloud evaporation. To sum up, the smaller the precipitation and relative humidity, the higher the temperature, the larger the raindrop diameter and the stronger the below-cloud evaporation.

TABLE 3. Precipitation isotopes and LMWL for different precipitation, temperature, relative humidity, and raindrop diameter in SRB.

Meteorological elements	Variation range	δD average	$\delta^{18}\text{O}$ average	Slope	Intercept	R^2
Precipitation	$P < 5$ mm	-52.79	-8.64	$7.48 (\pm 0.21)$	$11.8 (\pm 2.07)$	0.94
	$5 \text{ mm} \leq P < 10$ mm	-40.49	-7.01	$7.64 (\pm 0.17)$	$13.09 (\pm 1.47)$	0.96
	$P \geq 10$ mm	-45.70	-7.75	$7.74 (\pm 0.21)$	$14.31 (\pm 1.8)$	0.95
Temperature	$T < 0^\circ\text{C}$	-75.46	-12.11	$8.71 (\pm 0.5)$	$30.05 (\pm 6.3)$	0.96
	$0^\circ\text{C} \leq T < 10^\circ\text{C}$	-58.50	-9.68	$8.03 (\pm 0.18)$	$19.22 (\pm 1.9)$	0.95
	$10^\circ\text{C} \leq T < 15^\circ\text{C}$	-36.52	-6.04	$7.93 (\pm 0.3)$	$11.35 (\pm 2.18)$	0.93
	$T \geq 15^\circ\text{C}$	-26.89	-5.07	$7.65 (\pm 0.2)$	$11.89 (\pm 1.23)$	0.96
Relative humidity	$\text{RH} < 70\%$	-40.61	-6.88	$7.27 (\pm 0.19)$	$9.44 (\pm 1.64)$	0.94
	$70\% \leq \text{RH} < 85\%$	-46.00	-7.89	$7.79 (\pm 0.18)$	$15.41 (\pm 1.62)$	0.93
	$\text{RH} \geq 85\%$	-56.91	-9.19	$7.94 (\pm 0.19)$	$16.07 (\pm 2.03)$	0.97
Raindrop diameter	$D < 0.8$ mm	-50.41	-8.28	$7.44 (\pm 0.19)$	$11.19 (\pm 1.77)$	0.97
	$0.8 \text{ mm} \leq D < 1.2$ mm	-42.83	-7.40	$7.69 (\pm 0.15)$	$14 (\pm 1.24)$	0.98
	$D \geq 1.2$ mm	-49.69	-8.20	$8.76 (\pm 0.36)$	$22.15 (\pm 3.2)$	0.99

FIG. 3. Time variation of (a) f (raindrop evaporation residual ratio) and (b) Δd in precipitation at the SRB from October 2016 to April 2018. The box represents 25%–75% percentile, the line in the box represents median (50th percentile), the required line indicates 90th and 10th percentile, and the point indicates the 95th and 5th percentile. (The number of samples from January to December is 10, 12, 33, 24, 27, 42, 53, 85, 50, 42, 18, 11).

b. Quantification of below-cloud evaporation in precipitation stable isotopes

1) TIME VARIATION OF BELOW-CLOUD EVAPORATION INTENSITY AT THE SRB

As is shown in Figs. 3, f and Δd in the study area change with time, and f fluctuates between 37.48% and 101.34%. From June to October, f is higher than 60%, and f increases from June, reaching its maximum in September. f is lower from November to May, and less than 60%. From November to May of the following year, the f from December to February of the second year (winter) is higher than that from March to May (spring), which indicates that the below-cloud evaporation in spring is stronger than in winter. The below-cloud evaporation at the SRB is greatly affected by local precipitation and temperature. The effect of precipitation is dominant throughout the year. From June to October, the months with higher precipitation, below-cloud evaporation is weaker than during the other months. From June to October, the greater the precipitation, the weaker the below-cloud evaporation. There is less precipitation from November to May of the following year, so for the whole year, the below-cloud evaporation is weak and the variation of Δd is similar to the variation of f . Due to the complex climatic conditions at the SRB, the temporal characteristics of below-cloud evaporation show an interesting pattern. The intensity of below-cloud evaporation is summer < autumn < winter < spring.

2) COMPARISON OF BELOW-CLOUD EVAPORATION IN MOUNTAINS, OASES, AND DESERTS

Figure 4 shows the spatial variation of below-cloud evaporation in the mountains, oases, and deserts of the arid inland river basin. From Fig. 4, it can be seen that the f in the mountain ranges between 51% and 82.92%, in the oasis between 37.04% and 77.51% and in the desert between 35.78% and 85.37%. Consequently, the range of f and Δd in the mountain is small, and the change in range of f and Δd in the desert is the largest,

indicating that the annual change of below-cloud evaporation in the mountains is smaller than that in the oases and deserts. The greatest change in the year is due to the fixed source of water vapor in the mountainous area and the complex water vapor source in the desert areas (Zhu et al. 2019). It has been discussed previously that the below-cloud evaporation in the SRB follows the pattern of summer < autumn < winter < spring, while the precipitation in the desert area is mainly concentrated in summer and autumn. On an annual basis, the below-cloud evaporation at the SRB is as follows: the mountain areas have the weakest below-cloud evaporation and the difference between the intensity of below-cloud evaporation between the oasis and the desert areas is small. The spatial trends of f and Δd are largely controlled by local meteorological factors. Therefore, it is difficult to characterize any specific spatial trend.

3) RELATIONSHIP BETWEEN F AND ΔD

The variation of f (raindrop evaporation residual ratio) is similar to the change of Δd value, as shown in Fig. 5, and there is a correlation between f and Δd with a slope of about 2.5‰ per percent [standard deviation (SD) = 11.9, $P < 0.05$]. When f is higher than 90%, there is a linear correlation between f and Δd , with a slope of about 1.23‰ per percent (SD = 0.15, $P < 0.05$). That is, Δd decreases by about 1.23‰ (SD = 0.15, $P < 0.05$) for every 1% increase in f , which indicates that when f is higher (higher relative humidity, lower temperature, larger raindrop radius), the below-cloud evaporation of falling raindrops is weaker, and f and Δd show a relatively obvious linear relationship. However, as f decreases (relative humidity decreases, air temperature increases, and raindrop radius decreases), the linear relationship of 1.23‰ per percent (SD = 0.15, $P < 0.05$) between the two gradually weakens, and when f decreases below 45%, Δd gradually deviates from the linear fit.

Previous studies have found that when f is 95% or more in the Tianshan region, there is a linear correlation with a slope of about 1‰ per percent. However, f may be much lower than 95% in the

FIG. 4. Time variation of (a) f and (b) Δd in precipitation in SRB from October 2016 to April 2018. The box represents 25%–75% percentile, the line in the box represents median (50th percentile), the required line indicates 90th and 10th percentile, and the point indicates the 95th and 5th percentile.

arid environment, and some smaller raindrops may completely evaporate before reaching the ground (S. Wang et al. 2016). In previous studies, many scholars have generally assumed that the correlation between the f and Δd is 1‰ per percent but based on the current analysis, the value ascribed to Δd needs to recognize the specifics of the environments in the study area.

5. Discussion

a. Influence of meteorological factors

The relationship between f (raindrop evaporation residual ratio) and Δd is controlled by meteorological conditions, such

as the temperature, relative humidity, rainfall, and raindrop diameter (S. Wang et al. 2016; Pan et al. 2018; Sun et al. 2020). In general, when the temperature is low, relative humidity is high, precipitation is abundant and raindrop diameter is large. Thus f is relatively high, Δd is close to 0, the correlation between f and Δd is relatively strong, and the slope of the fitted logarithmic relationship is relatively low. In contrast, in a drought environment with high temperatures, low relative humidity, low precipitation and a small raindrop diameter, f is relatively small and Δd is relatively large, the correlation between the two is weak and the slope of the logarithmic relationship is relatively high (S. Wang et al. 2016).

FIG. 5. Relationship between f and Δd in precipitation in SRB from October 2016 to October 2017. Note: The box in the top right corner represents the relationship between f and Δd when $f > 90$.

FIG. 6. The relationship between f and Δd in precipitation under different meteorological conditions at the SRB from April to October 2017.

We analyzed the sensitivity of below-cloud evaporation to the variation of meteorological factors from April to October 2017, where precipitation is concentrated. When the air temperature is below 0°C , f and Δd are close to a logarithmic relationship distribution and the slope is 157.67% per percent ($P < 0.05$). When the temperature is between 0° and 10°C , the

slope is 173.28% per percent ($P < 0.05$), and when the temperature is higher than 10°C , the slope reaches 179.93% per percent ($P < 0.05$) (Figs. 6a–c). When the relative humidity is lower than 70% , the correlation between f and Δd is lower, with a slope as high as 200.98% per percent ($P < 0.05$). When the relative humidity is between 70% and 85% , the correlation

FIG. 7. Changes of altitude, relative humidity, temperature, and f at sampling points in the upper reaches of the SRB from October 2016 to April 2018.

between them increases and the slope decreases gradually, and when the relative humidity is higher than 85%, the correlation between them is at its highest, with a slope of 137.31‰ per percent ($P < 0.05$) (Figs. 6d–f). When precipitation is less than 5 mm, f has a high correlation with Δd , with a slope of 181.93‰ per percent ($P < 0.05$), and when the precipitation is between 5 and 10 mm, the correlation between f and Δd decreases, and the slope is slightly higher than 173.71‰ per percent ($P < 0.05$). When the precipitation is higher than 10 mm, f and Δd also increase, and the slope also increases to 171.63‰ per percent ($P < 0.05$) (Figs. 6g–i). When the raindrop diameter is less than 0.7 mm, the correlation between the f and Δd is low and the slope is high, reaching 182.67‰ per percent ($P < 0.05$), and when the raindrop diameter is between 0.7 and 1 mm, the logarithmic relationship between the two is enhanced, the slope is 177.04‰ per percent ($P < 0.05$). Finally, when the raindrop diameter is greater than 1 mm, the strength of correlation between the two indices is continuously increased, with the slope reaching 158.54‰ per percent ($P < 0.05$) (Figs. 6j–l).

b. Influence of reservoir on below-cloud evaporation

The below-cloud evaporation in the upper reaches of the Shiyang River follows the altitude effect of a decrease in f with decreasing altitude (Kong et al. 2013; Kong and Pang 2016). At the Xiying Reservoir, located at a lower elevation in the mountain areas at the upper reaches, f suddenly rises. By analyzing the topographic and geomorphic features around the sampling point, we found that the sampling point is close to the Xiying Reservoir; the distance between the two is 4.4 km. The Xiying Reservoir is the largest artificial water body in the basin, with a total capacity of 23.5 million m^3 . Differences in underlying surfaces may lead to significant differences in the contribution of recycled water vapor. A large open water surface will increase the water vapor recycling ratio in the surrounding area (S. Wang et al. 2016; Zhu et al. 2019; Bowen et al. 2012; Gibson and Edwards 2002). The contribution to evaporation from lakes and reservoirs in

North America's Great Lakes and the contribution from the Qinghai Lake to atmospheric precipitation is 15.7% and 23.4%, respectively (Zhu et al. 2019; Gat et al. 1994; Xu et al. 2011). The contribution of water vapor evaporation to meteoric precipitation in the Yinchuan and Pingliang lakes at the edge of the monsoon is 25%, which indicates that the lake reservoir evaporation on the edge of the monsoon has a stronger influence on local meteorological elements than that in other regions (Zhu et al. 2019; Qian et al. 2014). In SRB, the average contribution of plant transpiration, land surface evaporation and advected vapor to precipitation accounted for 11%, 6%, and 83% in the mountain regions, for 21%, 7%, and 72% in the oasis regions, and for 10%, 5%, and 85% in the desert regions, respectively. The average contribution to surface evaporation water vapor of the Xiying (M4) station is the largest in the study area, with a contribution fraction of 9% (Zhu et al. 2019). Therefore, the Xiying Reservoir's existence may be the reason for the sudden increase of f at the sampling point. We compared the temperature, relative humidity, and f of each sampling location at different altitudes. Except at the M4 sampling point, f in the mountain area decreased with decreasing altitude (Fig. 7). However, a sudden change occurs in Xiying, which further indicates that the difference in underlying surface properties causes the change in f near the Xiying Reservoir.

c. Uncertainty analysis

Due to the small amount of precipitation in oasis and desert areas, the number of samples collected is small in the study area. In addition, precipitation is concentrated in May–October. Thus, there are limited data for analyzing the monthly variations of below-cloud evaporation under different underlying surface conditions (Figs. 3 and 4). In previous studies, the premise of calculating the below-cloud evaporation with the Stewart model was that the system is in a state of fractionation equilibrium (S. Wang et al. 2016; Sun et al. 2020). Our study also assumes that precipitation

TABLE 4. Comparison of cloud-base heights, f , and Δd obtained by two calculation methods.

Sampling locations	Stewart model			Verification		
	H_{LCL} (m)	f	Δd	Z_{LCL} (m)	f	Δd
Mountains						
M1 (Lenglongling)	611.47	71.37	−60.06	613.96	72.03	−57.20
M2 (Ningchan)	591.17	73.01	−55.48	594.82	73.08	−55.65
M3 (Haujian)	1063.82	47.76	−176.90	1071.29	47.39	−187.96
M4 (Xiyang)	681.45	64.98	−82.95	685.96	64.52	−84.08
M5 (Yangjiazhuang)	866.14	72.17	−52.22	873.17	72.04	−52.67
Oasis						
O1 (Yangxiaba)	1080.38	44.76	−333.07	1093.98	44.53	−337.54
O2 (Wuwei)	691.88	61.98	−123.79	701.56	62.95	−122.25
O3 (Minqin)	726.53	59.71	−164.58	737.12	59.49	−165.54
Desert						
D (Qingtuhu)	741.09	53.76	−366.83	750.33	53.60	−371.39

from the cloud base and the surrounding water vapor reach a state of fractionation equilibrium. However, this assumption is somewhat different from reality, which may cause calculation errors. We use ground meteorological data, including temperature, precipitation, relative humidity, and pressure, to calculate the cloud-base height, raindrop diameter, condensation height and the temperature of different precipitation events using the improved Stewart model, and then calculate the raindrop evaporation remaining ratio f in the SRB.

The raindrops will evaporate at any height from the condensate level to the ground, and there will be a significant amount of stratiform (large scale) rain occurring in areas much higher than the condensate level. Therefore, we used the equation (Romps 2017) to estimate the lifting condensation level (LCL)/lowest levels of convective clouds, compared it with the previously calculated value. As shown in Table 4, H_{LCL} is the cloud-base height calculated by the previous estimation method, and Z_{LCL} is the cloud-base height estimated by the equation (Romps 2017). It can be concluded that Z_{LCL} is slightly larger than H_{LCL} . The values of f and Δd are not much different. The higher the cloud-base height, the larger error between the two calculation methods, which verifies the reliability of our calculations to a certain extent. In addition, the lack of comparison between the observation data of the radar rangefinder and the condensation height and raindrop diameter calculated by the Stewart model may lead to errors in the calculations, and thus substantial differences between the model results and reality. Therefore, the diameter observations of high-accuracy high-resolution rain droplets need to be improved in our future observations.

6. Conclusions

Based mainly on the precipitation stable isotope data and meteorological data of the SRB, we studied the spatiotemporal variability of below-cloud evaporation isotopes in the mountains, oases, and deserts of the arid areas. The below-cloud evaporation in the SRB follows the characteristics of summer, having less evaporation than autumn, which has less than winter, with spring having the least evaporation of all seasons.

Moreover, the below-cloud evaporation in summer and autumn follows the precipitation and temperature effects in winter and spring. In terms of spatial changes, the below-cloud evaporation in the mountains of the SRB is weaker than that in the oasis and in the desert areas. The below-cloud evaporation in the oases and deserts changes significantly during the year and is mainly controlled by the local microclimate.

Under low temperature, high relative humidity, high precipitation intensity, and a large raindrop diameter, the slope and intercept of the LMWL are small, f is large, Δd is close to 0, the linear relationship between f and Δd is strong, and below-cloud evaporation is weak. When f is higher than 90%, there is a linear correlation between f and Δd , with a slope of about 1.23‰ per percent, which indicates that the assumption of a relationship between f and Δd with a slope of 1‰ per percent in arid areas that also needs to take into account the environmental differences in the study area. Indeed, while our study found that local transpiration and evaporation will increase relative humidity, the below-cloud evaporation does not follow the altitude effect due to the influence of reservoirs in mountain areas. Moreover, the farmland and sheltering forests weakened the below-cloud evaporation in summer. However, the current understanding of this phenomenon is limited, and future research needs to strengthen understanding of the influence of ground features on below-cloud evaporation.

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Data availability statement. The data that support the findings of this study are available on request from the corresponding

author, stable isotope data are not publicly available due to privacy or ethical restrictions, and meteorological data can be obtained on China Meteorological Data Network (<http://data.cma.cn>).

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