

Using hydrogen and oxygen stable isotopes to estimate soil

Yang Li¹ and Siyuan Huo¹

¹Affiliation not available

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water evaporation loss under continuous evaporation conditions

Yang Li^{a,b}, Siyuan Huo^{a,b*}, Jing Guo^{c,d}, Jie Sun^e, Jie Pan^e, Deng Wang^{b,e}, Qiankun Tan^{a,b}, Bingbing Pei^{a,b}

a Hubei Key Laboratory of Petroleum Geochemistry and Environment, Yangtze University, Wuhan 430100, China

b College of Resources and Environment, Yangtze University, Wuhan 430100, China

c Hubei Key Laboratory of Resources and Eco-environment Geology, Wuhan 430030, China

d Geological Environmental Center of Hubei Province, Wuhan 430030, China

e Six Geological team of Hubei Geological Bureau, Xiaogan 432000, China

E-mail addresses: liyangyoung097@163.com (Yang Li); syhuo@yangtzeu.edu.cn (Siyuan Huo); guojing612@foxmail.com (Jing Guo); 946915781@qq.com (Jie Sun); whpanjie@126.com (Jie Pan); 1556793646@qq.com (Deng Wang); 2504218696@qq.com (Qiankun Tan); pb17740668729@163.com (Bingbing Pei);

ABSTRACT

The use of hydrogen and oxygen stable isotopes in estimating soil water evaporation loss under continuous evaporation conditions is crucial for gaining insight into soil water movement processes under different conditions. In this study, via high-frequency meteorological monitoring and continuous soil water measurements, we investigated the variation of hydrogen and oxygen stable isotopes and soil water fluxes with soil depth and time for soil water at different depths under continuous evaporation conditions. The precipitation isotope δ_{rain} and soil water flux changes were determined using the Craig–Gordon model. It was shown that a gaseous-dominated transport process dominated the isotopic fractionation of soil water in the surface layer 0–30 cm, and that both the $\delta^{18}\text{O}$ and $\delta^2\text{H}$ values, and the evaporative intensity decreased with soil depth. In terms of time dynamics, the evaporation loss of soil water varies continuously with seasons and is the highest during summer. The use of $\delta^{18}\text{O}$ to quantify the soil water evaporation loss provides a greater accuracy than that provided by $\delta^2\text{H}$. The relative errors in the evaporation loss calculated based on $\delta^{18}\text{O}$ and $\delta^2\text{H}$ were 13% and 34%, respectively. A sensitivity analysis of each parameter indicated that the relative error calculated by the model is primarily determined by temperature and relative humidity uncertainty. The sensitivity analysis reveals the critical evaporation intensity of soil water at various depths from unsteady to steady state evaporation. When the relative humidity changes by 1%, the evaporation loss fraction changes from 0.001 to 0.034. The results of this study are important for quantifying the soil water resources in arid and semi-arid areas without precipitation using stable isotopes of hydrogen and oxygen.

Keywords:

Soil water evaporation; Hydrogen and oxygen stable isotopes; Craig–Gordon (C–G) model; Evaporation loss; Uncertainty and sensitivity analysis

1. Introduction

Soil water evaporation is the process by which water in the soil layer below the surface to above the ground-water surface is transported to the atmosphere through the soil and plants by soil water suction. It is an integral part of the water transfer mechanism between atmospheric water, plant water, and groundwater. Soil water evaporation is affected by several factors such as temperature gradient (Liu *et al.* , 2020), soil-water transport mechanisms (Liu *et al.* , 2019), soil lithology (Hou *et al.* , 2018), vegetation (Lichner *et al.* , 2020), and salinity (Li *et al.* , 2021) , and thus its water transfer mechanisms are complex. The accurate quantification of soil water evaporation is crucial for the assessment of shallow groundwater resources (Grimaldi *et al.* , 2015), prevention and control of saline-alkali soil (He *et al.* , 2021), reconstruction of irrigated areas (Figuerola *et al.* , 2013), and evaluation of the ecological water demand (Jiang *et al.* , 2021), particularly in arid and semi-arid regions.

Currently, common methods for quantifying soil water evaporation include direct measurements using a lysimeter (Annelie *et al.* , 2021; Laura *et al.* , 2021), the formula method (Lehmann *et al.* , 2019), location flux method (Xing *et al.* , 2019; Tingting *et al.* , 2021), and numerical simulations (Ma *et al.* , 2019; Li and Shi, 2021). These methods are simple in application and differ greatly from the complex and variable natural conditions, and do not fully consider the vertical transport of water in the soil, which is not conducive to an in-depth understanding of the evaporation process and mechanism of soil water.

Naturally occurring stable isotopes (^{18}O and ^2H) have been widely used in soil water research, for instance, in the estimations of regional recharge (Koeniger *et al.* , 2016), infiltration and mixing (Stumpp and Maloszewski, 2010; Zhao *et al.* , 2013), plant water uptake (Koeniger *et al.* , 2010; Gaines *et al.* , 2016), evaporation (Gonfiantini *et al.* , 2018) and soil water transfer (Yang *et al.* , 2018), and the mutual transformation of surface water and atmospheric water (Li *et al.* , 2021), which is difficult to realize using other techniques. The Craig–Gordon (C–G) model is commonly used for quantifying evaporation from open water bodies. It is used to calculate the isotopic composition of evaporated water using information about the water that undergoes evaporation (i.e., temperature, relative humidity, and isotopic composition) (Skrzypek *et al.* , 2015; Gonfiantini *et al.* , 2018). In recent years, the model has also been used to quantify soil water evaporation at different elevations (Wei *et al.* , 2015; Yong *et al.* , 2020). However, in the case of arid and semi-arid regions that receive no rainfall for a long time, exploring the applicability of this model and determining the parameter changes that can affect the model is worth considering.

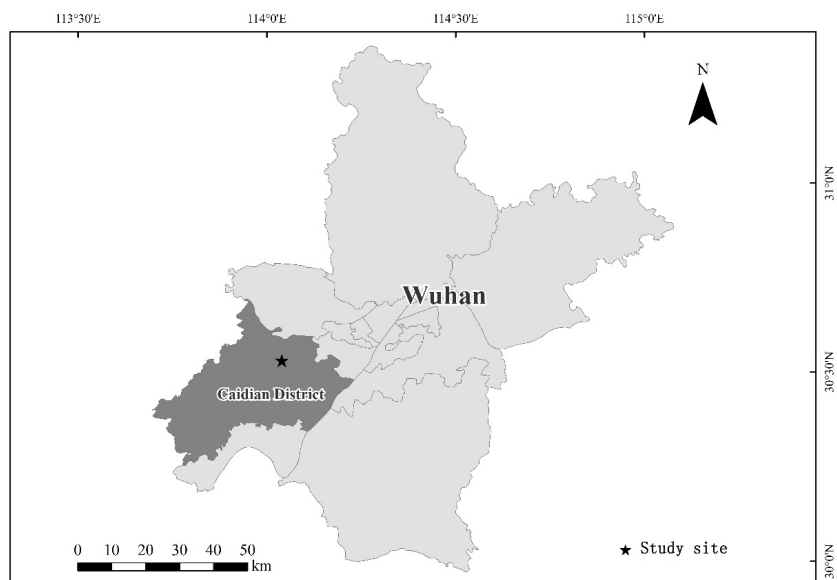
In this study, we created a continuous-evaporation condition and monitored the meteorological parameters (temperature and relative humidity), soil water fluxes (0–30 cm), and soil-water isotope data collected from Wuhan. The study period extended from May 2018 to June 2019. The monitoring data were compiled and analyzed to determine the variability of stable isotopes at different soil depths and the influencing factors and their interrelationships. The primary objectives of this study are as follows: (1) to determine the variation of hydrogen and oxygen stable isotopes in soil water and soil water fluxes with soil depth and time in a soil profile under continuous-evaporation conditions; (2) to test the applicability of the C–G model in quantifying soil water evaporation under continuous-evaporation conditions; and (3) to explore the potential uncertainties associated with the C–G model and the implications for quantifying evaporation losses. We anticipate that this will enhance our understanding of water cycle processes, as well as provide an estimate of soil water evaporation loss on different underlying surfaces. Furthermore, the study will contribute to a rational and scientific approach to the development and utilization of soil water resources in arid and semi-arid environments.

2. Materials and methods

2.1. Evaporation experiment

The test site is located at the Environment and Ecological Restoration Production and Research Base at

Yangtze University in China (114.04°N, 30.53°E) (Fig. 1). It is situated in a subtropical humid monsoon climate zone, with abundant rainfall and sunshine and four distinct seasons. The mean annual air temperature, mean annual precipitation, and mean annual evaporation are 16.9 °C, 1217 mm, and 934 mm, respectively. The precipitation exhibits evident seasonality and is primarily concentrated in June, July, and August. The soils in the surface layer of the test area are not non-homogeneous, and the lithologies are all silty loam soils with a water table depth of 1.9-2.8 m.



(2)

2.2 Μετεωρολογικά data ανδ $\delta^{18}\text{O}$ ανδ $\delta^2\text{H}$ ισοτοπες νη προσιπιτιον

During the test period, the daily average temperature, relative humidity, and evaporation data of Wuhan were obtained from the National Oceanic and Atmospheric Administration Climate Prediction Center (<https://www.cpc.ncep.noaa.gov/>) and NASA Earth Observations (<https://neo.sci.gsfc.nasa.gov/>).

Rainfall is the only source of soil water recharge in the study area. A comprehensive understanding of the composition of precipitation isotopes in the study area helps us to determine the initial conditions of soil water isotopes. In this study, we collected 115 monthly average precipitation isotope data in the Wuhan area, spanning from 1986 to 2013, with a couple months of missing data, and the available data were divided into three parts: 23 data from January 1986 to October 1992, 27 data from January 1996 to May 1998, and 21 data from September 2011 to May 2013. Data from January 1986 to May 1998 were obtained from the International Atomic Energy Agency (IAEA) (<https://nucleus.iaea.org/wiser/index.aspx>). The monitoring site is the Wuhan meteorological station (114.13°E, 30.62°N) and the monitoring frequency is measured month by month. The standard used was the Vienna standard average seawater isotope value, and the test precision for $\delta^{18}\text{O}$ and $\delta^2\text{H}$ was ± 0.1 respectively. Data from September 2011 to May 2013 were obtained from (Deng *et al.*, 2016). The monitoring site is located at the Wuhan University Engineering Department and Wuhan Institute of Biological Products dormitory (114.36degE, 30.54degN), and the monitoring frequency is measured month by month. The standard used was the Vienna standard average seawater isotope value, and the test precision for $\delta^{18}\text{O}$ and $\delta^2\text{H}$ was ± 0.2 respectively.

2.3 Quantification of soil water evaporation using the C-G model

When the stable isotope composition of water changes only because of isotopic fractionation during the evaporation process, and information about the water undergoing evaporation (ambient temperature, relative humidity, and isotopic composition) is known, the non-steady-state model in the C-G model (Bennett *et al.*, 2008; Horita *et al.*, 2008) is often used to calculate the evaporation losses of open water bodies. Theoretically, this method is also suitable for evaluating soil water evaporation under continuous-evaporation conditions. In our study, soil-water evaporation losses under continuous-evaporation conditions were quantified via three field experiments conducted from May 13, 2018 to July 12, 2018; July 12, 2018 to October 19, 2018; and January 5, 2019 to April 20, 2019.

In this scenario, the evaporation loss fraction of the soil water volume (f) can be calculated using the following equation (Stephen K. Hamilton, 2005). On multiplying f by the thickness of the soil layer and subsequently dividing it by the number of test days, we obtain the corresponding evaporation loss (mm/d) at different depths.

(3)

where $\delta_{P(\text{soil})}$ is the initial value of the soil-water sample isotope composition, $\delta_{L(\text{soil})}$ is the final value of the soil-water sample isotope composition, δ^* is the limiting factor for the isotope enrichment—this term is also known as the limit isotope composition enrichment (m denotes the correlation between the isotopic composition of evaporated water vapor and the isotopic composition of liquid water—this term is also known as the enrichment slope (G.B. Allison, 1982).

The change in the ratio of the difference between the soil-water isotope composition and the environmental limit isotope composition at the end and the initial time reflects the change in the remaining soil-water isotope composition value. A large value (for example, $f = 0.40$) indicates that 40% of the initial water content has evaporated.

(4)

where h is the average relative humidity (fraction), ϵ is the total isotope fractionation (composition of water in ambient air (

The most accurate method for determining δ_A is based on on-site measurements. However, owing to the difficulty associated with direct measurements in the field, the δ_A value is generally estimated based on the stable isotopic composition of the local rainfall (Gat.J.R 1995; Gibson and Reid, 2014).

(5)

where $\bar{\delta}$ is the average of the precipitation isotopes at the time of the two sampling periods before and after; $\epsilon+$ and $\alpha+$ are the equilibrium enrichment factor and equilibrium fractionation factor that change with temperature, respectively, as proposed by (Juske Horita, 1994):

(6)

^2H : (7)

^{18}O : (8)

where temperature (T) is expressed in Kelvin.

(9)

where α is the total isotopic fractionation factor (isotopic fractionation factor (

(10)

where n is a constant, which is related to the correlation between the molecular diffusion resistance and molecular diffusion coefficient and is usually considered to be 1 for non-mobile air layers (e.g., soil water evaporation and plant transpiration), ϕ is the ratio of the molecular diffusion coefficient to the total diffusion fractionation coefficient and is generally considered to be 1 for soil water evaporation. C_D is a parameter that describes the diffusion efficiency of the molecule and has a value of 25.1 for hydrogen and oxygen, respectively (GONFIANTINI, 1986; John Crusius, 2000).

(11)

In this study, precipitation isotopes are used to reflect the environmental changes during evaporation, i.e., in Eq. (5). In this way, the during the test period can be determined by knowing the precipitation isotope values for both sampling periods. In the absence of precipitation isotope data for Wuhan during the test period, precipitation isotope data for Wuhan collected from network monitoring data from 1986 to 1998 were used to replace precipitation isotope data during the test period in order to calculate rain. As shown in Fig. 2, we conclude that precipitation isotope values for the same period from 1986 to 1998 and from 2018 to 2019 can be considered similar if the past and current monthly averages of precipitation and temperature parameters in the same month were similar in the same study setting. Therefore, we used the following equations to calculate $\delta^{18}\text{O}$ and $\delta^2\text{H}$ for the months that satisfied this condition.

(12)

(13)

(14)

(15)

In these formulas, n is the total number of data collected in a particular month; j is the month j ; i is the data collected in a particular month i ; rain is the rainfall in mm; T is the temperature in $^\circ\text{C}$; ^{18}O , ^2H , and T are the oxygen 18 isotope value, deuterium isotope value, and temperature, respectively; and $\bar{\delta}$, $\epsilon+$, $\alpha+$, and C_D are the corresponding monthly mean values.

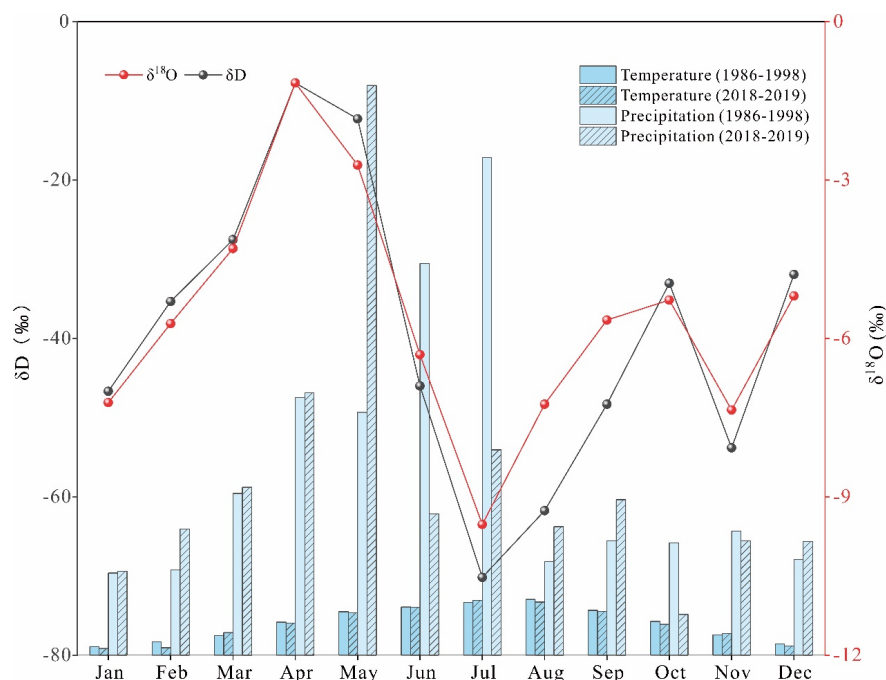


Fig. 2 Comparison chart of average temperature and precipitation between 1986–1998 and 2018–2019.

3. Results

3.1 Temporal variation of environmental variables

The daily variations in precipitation, air temperature, land surface evaporation, and humidity in the study area from May 2018 to June 2019 are presented in Fig. 3. The daily precipitation ranged from 0.01 mm to 146.4 mm. During the rainy season, the cumulative rainfall was 1079.49 mm, which accounted for 64.7% of the total rainfall during the test period, and the average daily rainfall was 8.7 mm. During the test, the daily temperature ranged from -2.38 °C to 32.67 °C, and the average daily temperature was 18.48 °C. The daily relative humidity ranged from 39.78% to 98.79%, and the daily average relative humidity was 74.53%. The daily evaporation ranged from 0.36 mm to 7.07 mm, and the average daily evaporation was 2.8 mm. The daily evaporation trend was consistent with that of the temperature. The evaporation in the rainy season was 755.69 mm, which accounted for 63.43% of the total evaporation, thus reflecting the rapid evaporation of atmospheric precipitation driven by the temperature in the rainy season.

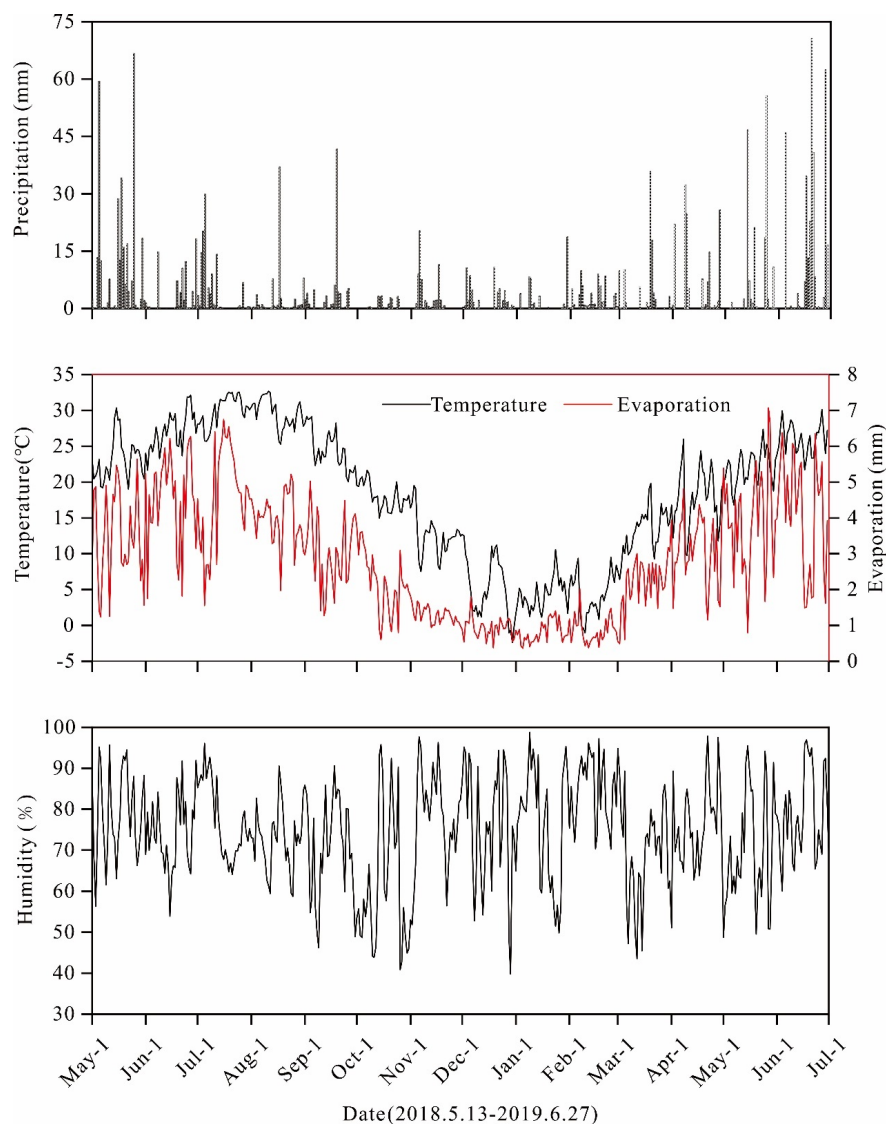


Fig. 3 Daily variations in precipitation, air temperature, evaporation, and humidity in the plot from May 2018 to Jun 2019. The shaded area indicates the rainy season.

3.2 $\delta^{18}\text{O}$ and $\delta^2\text{H}$ isotopes composition of precipitation in Ωνηαν

As a direct source of soil water in the test area, precipitation dramatically affects the soil water content. Moreover, the change in the precipitation reflects the shift in air environmental factors (temperature and humidity), which is vital for calculating the isotope composition enrichment limit. Herein, it was used as the basis for analyzing the $\delta^{18}\text{O}$ and $\delta^2\text{H}$ isotopes in soil water at the test site.

As shown in Table 1, the maximum values of $\delta^{18}\text{O}$ and $\delta^2\text{H}$ in the local precipitation isotopes continue to increase, whereas the minimum values initially increase and subsequently decrease. The overall average value exhibits an increasing trend. The concentration of the majority of the precipitation isotope samples was negative, and there were only five positive values. The reason for the occurrence of these five positive values may have been that the water vapor of these samples had different sources or experienced intense secondary evaporation during their decrease, thus resulting in the enrichment of heavy isotopes and a lack

of light isotopes.

Table 1

The general characteristics of isotopic composition of precipitation.

Types	Time	Sample amount	$\delta^{18}\text{O}$ (	$\delta^{18}\text{O}$ (	$\delta^{18}\text{O}$ (	$\delta^{18}\text{O}$ (	δD (	δD (	δD (
			Max.	Min.	Luffing	Mean	Max.	Min.	Luffing
Precipitation	1986-1992	23	-2.15	-12.18	10.03	-5.94	-6.10	-86.80	80.70
	1996-1998	27	0.12	-9.68	9.80	-5.37	4.20	-79.30	83.50
	2011-2013	21	3.06	-14.36	17.42	-4.82	14.30	-114.70	129.10

As shown in Fig. 4, the formula for the local meteoric water line (LMWL) established based on the data of precipitation isotopes obtained from 1986 to 2013 is $\delta^2\text{H} = 8.29\delta^{18}\text{O} + 7.44$, and the correlation coefficient is 0.93. Compared with the global meteoric water line (GMWL) formula proposed by the IAEA-World Meteorological Organization and Craig ($\delta^2\text{H} = 8\delta^{18}\text{O} + 10$), the collected 115 samples of precipitation isotopes are all located near the GMWL, and the majority of the samples are found to be on the upper left of the GMWL. In addition, the slope of the LMWL is 8.29, which is higher than the slope of the GMWL of 8. The shift in slope reflects a different degree of non-equilibrium fractionation in Wuhan than the global average during the experiment (Chen *et al.* , 2016).

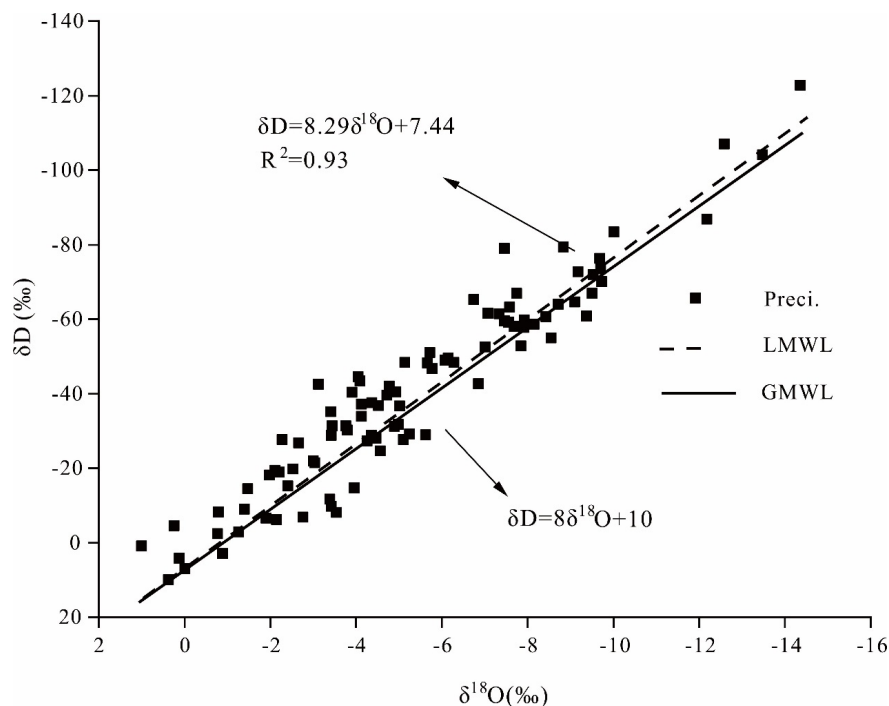


Fig. 4 Regional atmospheric waterline in Wuhan (data from IAEA (1986–1998) and (Deng *et al.* , 2016) (2011–2013)).

3.3 $\delta^{18}\text{O}$ and $\delta^2\text{H}$ isotope composition of soil water at the test site

Table 2 lists the general characteristics of the isotope composition of soil water. As can be observed, $\delta^{18}\text{O}$ in soil water varies from -9.53-59.34 during the test. The average values of $\delta^{18}\text{O}$ and $\delta^2\text{H}$ decreased gradually with depth, which indicates that the isotope fractionation caused by evaporation was more significant in the surface layer of the soil than that in the deeper layers.

Table 2

The general characteristics of isotopic composition of soil water.

Types	Date	Depth(cm)	Sample amount	$\delta^{18}\text{O}$ (	$\delta^{18}\text{O}$ (	$\delta^{18}\text{O}$ (	δD (	δD (	δD (
				Max.	Min.	Mean.	Max.	Min.	Me
Soil water	2018.5-2019.6	0	5	0.05	-5.18	-2.91	-31.83	-40.43	-37
		5	6	-3.47	-8.18	-5.47	-40.80	-55.72	-48
		10	6	-4.50	-8.65	-6.60	-42.44	-59.34	-50
		15	6	-6.68	-7.70	-7.21	-44.53	-53.90	-51
		20	5	-6.47	-8.38	-7.56	-45.12	-55.21	-50
		25	5	-6.93	-8.68	-7.88	-45.27	-55.73	-51
		30	5	-7.12	-9.53	-8.14	-46.23	-57.79	-52

Fig. 5 presents the relationship between $\delta^{18}\text{O}$ and $\delta^2\text{H}$ in soil water at different depths in the study area under continuous evaporation conditions. The results show that the soil water isotope points all fell near the LMWL. At 0-15 cm, the compositions of oxygen-18 were mostly higher than -8‰, the compositions of oxygen-18 were lighter than -7.5‰ and $\delta^2\text{H}$ in soil water decreased with increasing depth above 30 cm. A similar pattern has been observed in other semi-arid and arid regions (Barnes and Allison, 1983; Gat, J.R 1995), since water vapor-dominated transport controls the composition of isotopes near the surface. Moreover, the slope of the soil water evaporation line was lower than that of the atmospheric precipitation line for each profile, and the slope increased overall with depth. The results suggest that soil water evaporated continuously during the test period and that water vapor-dominated near-surface transport controlled isotopic composition. In general, the smaller the slope of the evaporation line, the stronger the non-equilibrium distribution of soil water, the greater the evaporation rate, and the smaller the amount of water remaining in the soil (Zimmermann et al., 1966).

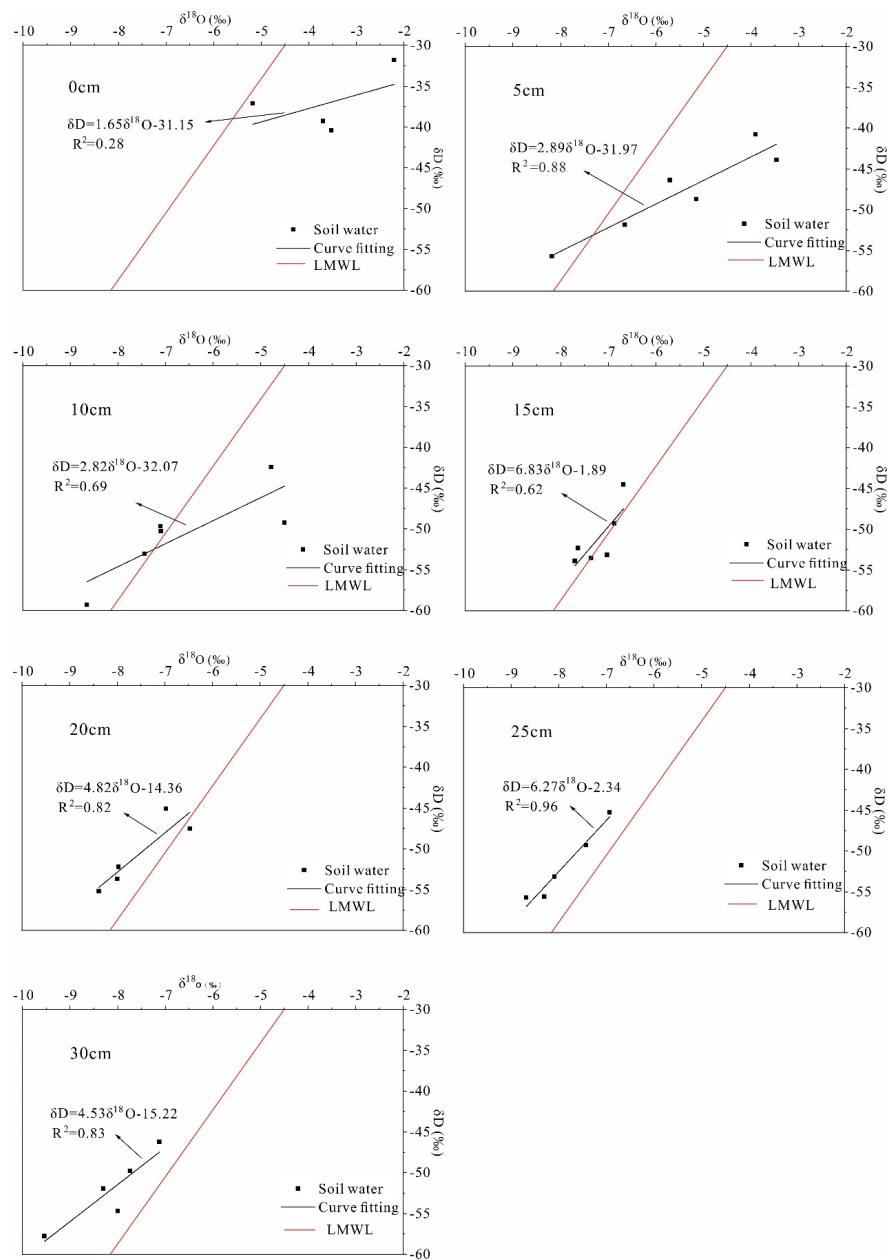
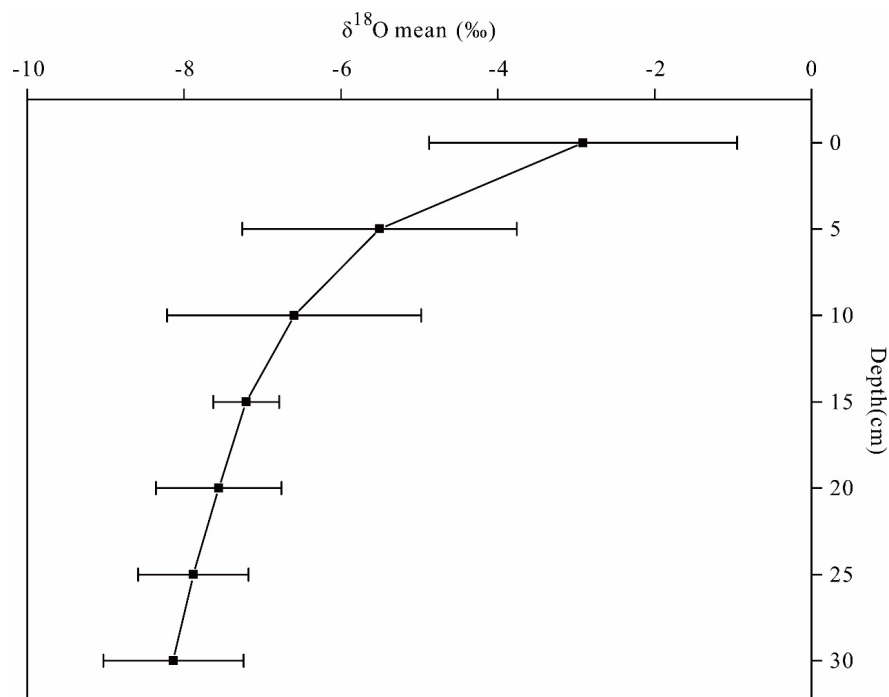


Fig. 5 Relationship diagram of soil water $\delta^{18}\text{O}$ and δD at different depths.



As shown in Fig. 6, along the soil profile, the mean value of $\delta^{18}\text{O}$ becomes increasingly negative with increasing soil depth, and the standard deviation of $\delta^{18}\text{O}$ decreases from 1.75 to 0.38 at the surface and stabilizes at 20 cm below the surface. $\delta^2\text{H}$ exhibits the same trend as $\delta^{18}\text{O}$. It is more of an effect of evaporation on topsoil. As soil depth increases, the influence of evaporation on soil water isotope fractionation diminishes. In the same manner as at the low rainfall point in Hawaii, soil water evaporation intensity is determined by atmospheric evaporative power and soil water conductivity (Jean C.C Hsieh, 1998).

Fig. 6 Variation map of $\delta^{18}\text{O}$ along the soil profile, one standard deviation is represented by the bars.

3.4 Quantifying soil water evaporation

The evaporation losses at different depths ranging from 0–30 cm at the sampling sites during the test period were obtained according to Equation 3 in Section 2.3 and are presented in Table 3. Owing to the collapse of the canopy caused by heavy snow in Wuhan on December 7, 2018, the experiment was divided into three periods to ensure continuous evaporation conditions during each period. In addition, the snowfall only changed the initial water content in the third stage without affecting the test results. As shown in Table 3, the evaporation loss calculated from ^2H for the surface layer of 0–5 cm during the period from May 13, 2018 to July 12, 2018 is negative, which may have resulted from the larger overall value of $\delta^2\text{H}$ than that of $\delta^{18}\text{O}$, which is likely to cause large errors in the calculation. There were only two pieces of data collected from July 12, 2018 to October 19, 2018 because the numbers of samples at other depths were small, and the instrument could not measure them. It should be noted that if f is 0, the calculated evaporation loss score is exceedingly small; however, this does not indicate that no water evaporated during the test. Fig. 7 shows that the evaporation intensity is stronger closer to the surface layer. As the soil depth increased, the evaporation loss of the soil water gradually decreased, and it tended to become stable at 15–20 cm, which is consistent with the drying front observed in Section 3.3. The evaporation losses at different soil depths were accumulated to obtain the evaporation loss at 5–30 cm from the surface of the sampling point based on $\delta^{18}\text{O}$ and $\delta^2\text{H}$. During the periods from May 13, 2018 to July 12, 2018 and January 5, 2019 to April 20, 2019, the calculated evaporation losses based on $\delta^{18}\text{O}$ were 1 mm and 14 mm, respectively, and they were

0 mm and 13.5 mm, respectively, based on $\delta^2\text{H}$.

Table 3

The calculation of soil water evaporation.

Date	Depth (cm)	Evaporation loss fraction ($\delta^{18}\text{O}$)	Evaporation loss $\delta^{18}\text{O}$ (mm)	Evaporation loss fra
2018.5.13-2018.7.12	5-10	0.02	1.00	-0.03
	10-15	0.00	0.00	0.00
	15-20	0.00	0.00	0.01
	20-25	0.00	0.00	0.01
	25-30	0.00	0.00	0.02
2018.7.12-2018.10.19	5-10	0.16	8.00	0.10
	10-15	0.08	4.00	0.02
2019.1.5-2019.4.20	5-10	0.09	4.50	0.08
	10-15	0.07	3.50	0.07
	15-20	0.04	2.00	0.04
	20-25	0.04	2.00	0.04
	25-30	0.04	2.00	0.04

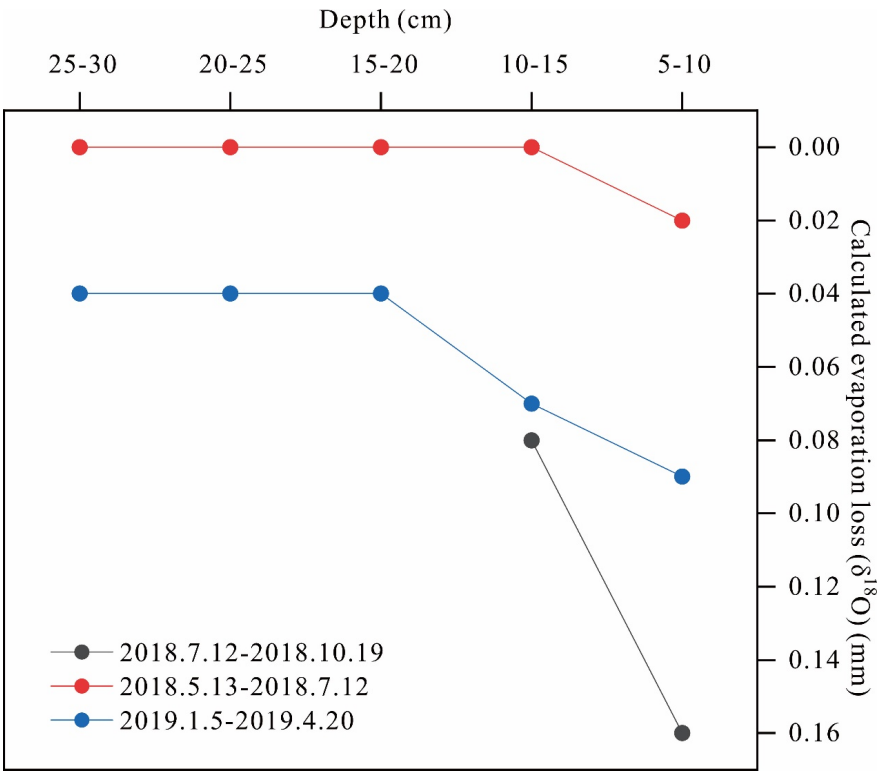
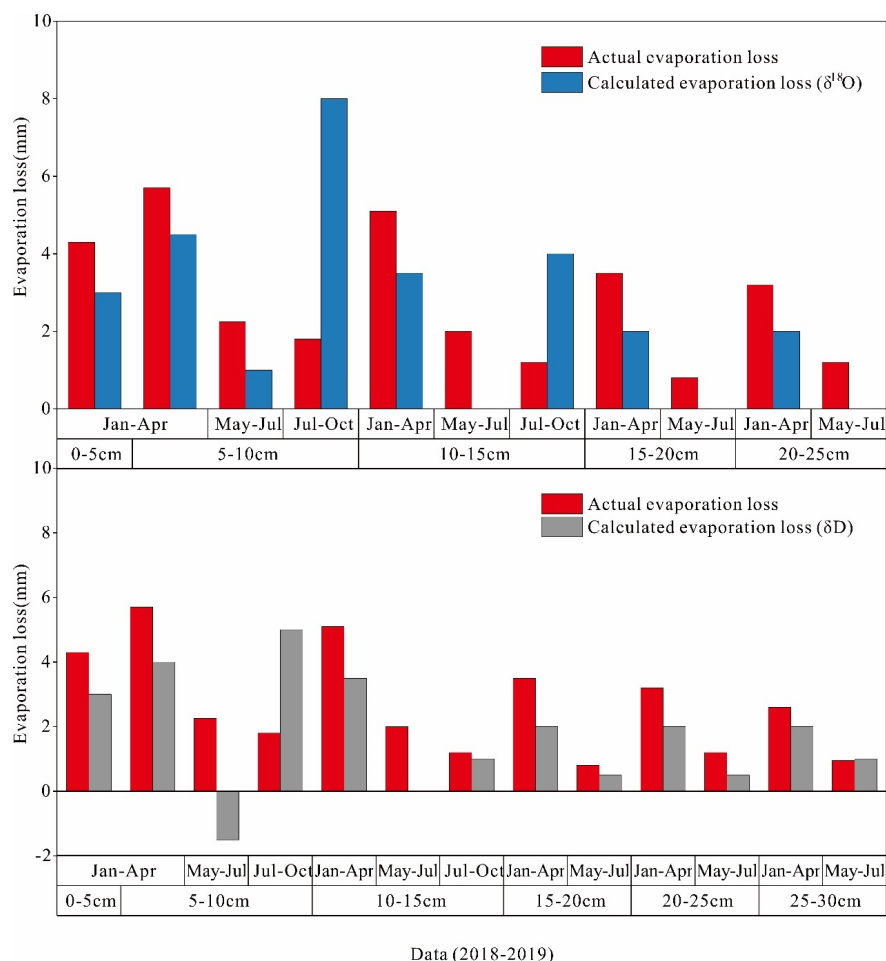


Fig. 7 Calculated evaporation loss varies with soil depth.

3.5 Comparison between calculated and measured values



To verify the accuracy of quantifying the soil water evaporation using the C-G model, we compared the actual evaporation losses with the calculated evaporation losses (Fig. 8). We observed that the calculated evaporation was close to the actual evaporation, and the variation trend followed by the soil water evaporation intensity with depth was approximately the same. During the entire test period, the relative error of the evaporation loss calculated based on $\delta^{18}\text{O}$ was 13%, and the relative error of the evaporation loss calculated based on $\delta^2\text{H}$ was 34%. The use of $\delta^{18}\text{O}$ to quantify the soil water evaporation loss resulted in a 21% greater accuracy than that resulting from $\delta^2\text{H}$. However, the approach involving the use of isotopes had a lower evaporation loss, which may have been caused by the approximate treatment of the values of various parameters in the model calculation. Considering that there were some unavoidable sources of errors, such as the absolute error caused by the instrument itself, in field experiments, it is feasible to use stable hydrogen and oxygen isotopes combined with a modified C-G model to quantify the evaporation of bare soil under continuous-evaporation conditions.

Fig. 8 Comparison of the actual evaporation loss and calculated evaporation loss at different soil depths and different periods from May 13, 2018 to June 27, 2019 (based on $\delta^{18}\text{O}$ and δD isotopes).

4. Discussion

4.1 The C-G model reveals the hydrological processes

The non-stationary model of the C–G model is understood as a non-linear model that can be applied to soils with closed isotope systems. The variation of the isotopic composition of input soil water (δ_P) and the isotopic composition of output soil water (δ_L) in the C–G model reflects the course of soil water composition in the environment at a given time and temperature and humidity (Fig. 5). In continuous evaporation conditions and at high frequencies of isotope measurements, evaporation exhibits a sinusoidal response for different soil depths at the same time (Fig. 8). Variation of evaporation f indicates that evaporation is not constant throughout the year, and f does not evolve as a simple evaporation process but as a competitive process of evaporation and isotopic mixing of multiple reservoirs (isotopic recharge of upper or lower soil water and evaporation of intermediate soil layers). In addition, the variation in evaporation f in different seasons reflects different isotopic processes - evaporation and mixing - that can be explained by different functions (Fig. 8).

4.2 The accuracy of the C–G model in quantifying soil evaporation

The lysimeter method is the only physical evaluation method that can be used to measure changes in soil moisture fluxes directly. Additionally, it can analyze the soil evaporation pattern of water at various soil depths, for different soil textures, and different crops growing under different climatic conditions (Annelie *et al.*, 2021; Laura *et al.*, 2021). However, the set water level in the lysimeter test is fixed, and a deviation results from the natural condition of water level change or the natural state of the water level rising and falling after continuous evaporation. The formula method is based on the observation results produced by the lysimeter; herein, the empirical formula is analyzed, summarized, and then applied to the estimation of soil evaporation without the consideration of associated mechanism problems such as the movement of water during the soil evaporation process; additionally, the result estimated by the experimental formula of evaporation is still different from the actual value (Lehmann *et al.*, 2019). The location flux method is based on the theory of soil hydrodynamics and involves an analysis of the evaporation process and calculation of the soil evaporation based on the characteristics of the water potential distribution of soil (Xing *et al.*, 2019; Tingting *et al.*, 2021). Although the physical interpretation of this method is clear, the applicable conditions are relatively simple and differ significantly from the complex and variable natural conditions; however, the observation requirements for soil water potential are high, and the application of this method is thus limited. The numerical simulation method is based on a large amount of experimental data and combines the principles of ground energy balance and soil hydrodynamics to establish a model for simulating the soil water movement (Ma *et al.*, 2019; Li and Shi, 2021). The limitation of this method is that more parameters are required for the calculation, and the uncertainty associated with the model is also high. Furthermore, the accuracy of the model in estimating the boundary conditions is limited. In contrast to previous research methods, quantifying soil water evaporation using the C–G model is theoretically applicable to any soil evaporation conditions. Quantification of soil moisture evaporation using the C–G model was similarly validated in the studies of Wei and Yong *et al.* (Wei *et al.*, 2015; Yong *et al.*, 2020). Unlike Wei and Yong *et al.* We studied a more specific case of soil evaporation estimation under continuous evaporation conditions, which is critical for the study of soil water resources in arid and semi-arid regions. Moreover, in this study, we used soil water content data obtained simultaneously in an experiment as a real measure of soil evaporation. We compared it with the evaluation results of the C–G model, thus validating the accuracy of the model (Fig. 8).

However, soil evaporation is influenced by several factors, such as salinity (C.J. Barnes, 1988), temperature gradients (McDonnell, 1998), soil-water transport mechanisms (Vincent Marc, 2001; Gazis and Feng, 2004), and soil layering (Brent D Newman, 1997). Sofer found that salinity generally reduces equilibrium fractionation, and the degree of reduction is dependent on the type of salt used. Indirectly, salinity affects relative humidity, which reduces kinetic fractionation (Stewart and Friedman, 1975; Z. Sofer, 1975). As Barnes found, temperature affects evaporation of unsaturated soil water primarily by affecting saturated water vapor density, which increases as temperature increases. As a rule of thumb, equilibrium fractionation coefficients, effective diffusion coefficients, and hydraulic conductivities cannot vary more than a few percent per degree Celsius with temperature (C.J. Barnes, 1989). Padilla demonstrated that soil water transport mechanisms produce different isotope profiles that alter soil evaporation calculations. In unsaturated soils,

piston flow is dominant when the water content is high, while preferential flow is dominant when the water content is low (Padilla *et al.*, 1999). Shurbaji found that clay layers possess lower hydraulic conductivity than sand layers, allowing a greater concentration of water to remain in clay layers, decreasing the rate of isotope transport. This resulted in significantly different isotope samples based on different soil layers (A.-R. Shurbaji, 1997). It is still undetermined if these factors affect the calculation of the model and if they can be used to modify the model. Thus, the manner in which the interaction between the parameters affects the model is yet to be determined. These issues should be addressed in a follow-up study.

4.3 The uncertainty analysis of the C-G model

For calculating evaporation losses, the C-G model can identify relative and absolute error sources. The absolute error of the experiment originates from the error of the measuring instrument itself. The stable isotope composition of water was frequently analyzed in the laboratory, and the errors associated with $\delta^{18}\text{O}$ and $\delta^2\text{H}$ were found to be $\pm 0.2\%$. When the difference between the maximum analytic uncertainty in the corresponding values and the valid values of two adjacent samples was the greatest ($+0.2\%$ of $0.4\delta_L$ for $\delta^{18}\text{O}$ and an increase of 1.2% the difference between those for $\delta^2\text{H}$). Considering the maximum expected uncertainty in the measured δ value, the maximum difference between the final calculation results of this experiment was 1.67% ($f_{\delta^{18}\text{O}}$) and 1.35% ($f_{\delta^2\text{H}}$). However, this difference depends on the relative change in the δ value owing to evaporation and impacts the samples that exhibit greater evaporation. It would be worth future research to develop a more accurate method of measuring isotopes.

The relative error of the experiment depended on the uncertainty associated with the parameters in the model calculation. We analyzed the sensitivity associated with each parameter of the model using the control variable method combined with the test data. As shown in Fig. 9, in terms of the sensitivity of the model calculation to a change in a single parameter, parameters such as h , δ_P , and δ_L have a greater influence on the calculation. The changes in T , δ_{rain} , and other parameters had almost no effect on the calculation of the model. (1) As h varies within the interval (0.5, 1), each 1% change in h results in a change in f of 0.006 to 0.892. As relative humidity increases, f decreases gradually until it reaches a fixed value. The surface soil is more strongly influenced by the atmosphere and its variation is more pronounced. Essentially, relative humidity (h) is the ratio of the actual water vapor pressure in the air to the saturated water vapor pressure at the same temperature, and it reflects how close the atmospheric water vapor pressure is to the saturated water vapor pressure at the current temperature. As the relative humidity increases, the saturation pressure of water vapor in the atmosphere increases, which makes water evaporation more difficult; therefore, evaporation intensity is negatively correlated with relative humidity. (2) As $\delta_{P(\text{O}^{18})}$ varies within the interval (-10, 0), each 0.5% change in h results in a change in f of 0.006 to 0.892. For each soil layer, f decreases followed by an increase and is then stabilized. The trend of $\delta_{L(\text{O}^{18})}$ is counter-intuitive to that of $\delta_{P(\text{O}^{18})}$. As $\delta_{L(\text{O}^{18})}$ varies within the interval (-10, 0), each 0.5% change in h results in a change in f of 0.019 to 0.462. f increases followed by a decrease and is then stabilized. In essence, the ratio of the difference between the final and limiting isotope values of soil water isotopes to the difference between the initial and limiting isotope values reflects the variation in soil water content. In addition, soil non-stationary evaporation isotopes' limiting evaporation capacity depends on soil water content and soil medium. In soils with a constant limiting isotopic composition, i.e., when the hydraulic medium of the soil itself is constant, the change in evaporation intensity is determined by changes in soil water content (G.R. Walker, 1988; C.J. Barnes, 1989). As soil water isotopic values become similar to their initial values, it indicates that the soil has changed from a non-steady state evaporation state to a steady state evaporation state. At this point, f describes the evaporation intensity of each soil layer at steady-state. (3) It should be noted that although the controlled variable method can reflect the sensitivity of a certain parameter to the model to a certain extent, it does not consider the synergistic relationship between the parameters. Unsaturated soil water movement is affected by temperature gradients primarily due to saturated water vapor density increasing rapidly with temperature. As the temperature increases with depth, the temperature gradient enhances evaporation. In addition, as the temperature decreases with depth, the temperature gradient reduces evaporation. Thus, a single temperature control does not show evaporation's response to temperature changes.

In summary, the temperature and relative humidity are the root causes of errors resulting in model calculations. The relative error in the experiment primarily depends on the uncertainty in the temperature and relative humidity.

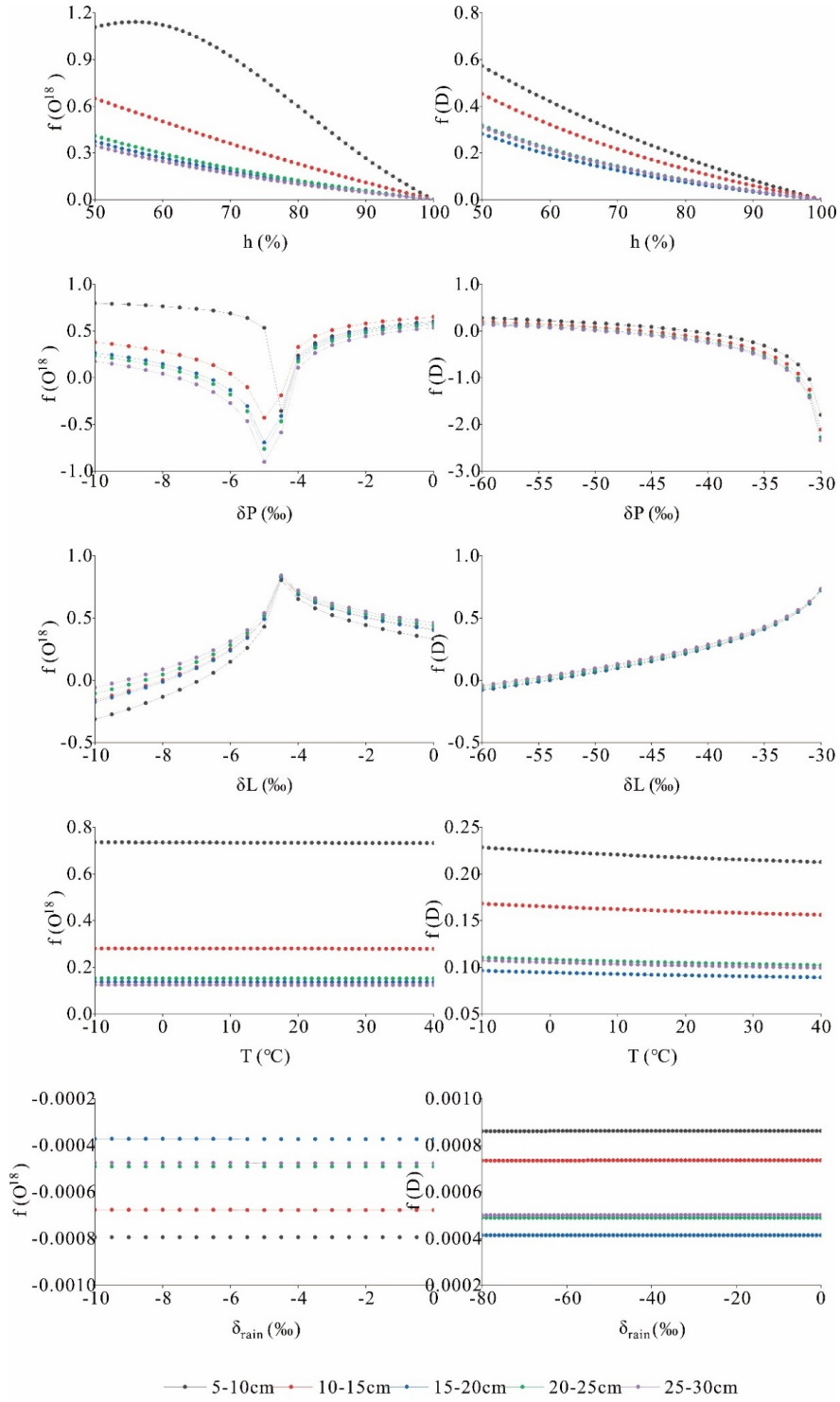


Fig. 9 Parameter sensitivity analysis.

5. Conclusions

Our research provided a qualitative and quantitative mutual demonstration of soil water stable isotope dynamics at different depths under continuous evaporation conditions. The results showed that $\delta^{18}\text{O}$ and $\delta^2\text{H}$ decreased with increasing soil depth from 0-30 cm. Most soil water was transported in a gaseous state. Within 10 cm of the soil surface layer, evaporation had the greatest effect.

Furthermore, $\delta^{18}\text{O}$ provided better results than $\delta^2\text{H}$ for estimating evaporation losses overall. During the calculation periods extending from May 13, 2018 to July 12, 2018 and January 5, 2019 to April 20, 2019, the evaporation loss within a range of 5–30 cm from the surface layer was 1 mm and 14 mm, respectively. The relative error of the evaporation loss calculated based on $\delta^{18}\text{O}$ and $\delta^2\text{H}$ was 13% and 34%, respectively. The use of $\delta^{18}\text{O}$ to quantify the soil evaporation loss resulted in an additional accuracy of 21% compared with that resulting from $\delta^2\text{H}$.

The control variables approach suggests that temperature and relative humidity constitute sensitive parameters of the model, and the critical intensity of evaporation for switching from unsteady to steady-state evaporation of soil water at different depths can be identified utilizing sensitivity analysis. Furthermore, changes in a single factor cannot be fully reflected in the model, and the parameters work synergistically.

The study focused primarily on the spatial and temporal heterogeneity of soil water isotope evaporation signals at different depths under conditions of continuous evaporation, and quantified the significant influence of evaporation losses. These results are valuable for understanding regional hydrological processes, and soil water resources planning in arid and semi-arid regions.

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