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Review

Stable water isotope simulation in different reservoirs of Manaus, Brazil, by Community Land Model incorporating stable isotopic effect

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ABSTRACT: The daily and monthly variations of stable water isotopes in different reservoirs at Manaus, Brazil, are simulated and inter-compared in an equilibrium year, using the Community Land Model (CLM) involving the stable isotopic effects as a diagnostic tool for an in-depth understanding of the hydrometeorological processes. On the daily scale, the $\delta^{18}O$ in precipitation, vapour and surface runoff have clear seasonality, with marked negative correlations with the corresponding water amount. However, the $\delta^{18}O$ in surface dew displays marked positive correlation with dew amount. On the diurnal time scale, the $\delta^{18}O$ in precipitation displays an unclear diurnal variation and an unmarked correlation with the precipitation amount. However, the $\delta^{18}O$ in vapour keeps consistency with specific humidity. On the monthly time scale, the $\delta^{18}O$ in precipitation and surface runoff displays distinct bimodal seasonality, with two maxima in January and in July, and two minima in April and in October; Vapor displays a similar bimodal pattern, two maxima appear in January and August, and two minima in April and November. The amount effect simulated on the monthly time scale has consistency with the actual survey result at the Manaus station, from 1965 to 1990, set up by International Atomic Energy Agency (IAEA)/World Meteorological Organization (WMO). In addition, the slope (7.49) and the intercept (6.25) of the simulated meteoric water line (MWL) are all smaller than those of the actual mean MWL. However, compared with the annual MWL, the simulated MWL lies within the variation range of actual MWLs. Copyright © 2008 Royal Meteorological Society

KEY WORDS stable water isotope; CLM; simulation; fractionation; amount effect; meteoric water line

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1. Introduction

In the water cycle, the most important stable water isotopes are HDO and $H_2{}^{18}O$. The HDO and $H_2{}^{18}O$ are regarded as useful tracers for the study on the water cycle, because their abundance in water body reflects features of physical phase change. Precipitation is an important link to the water cycle. The heterogeneity in precipitation isotopes in time and space mirrors complex atmospheric behaviour. Many observation programmes on stable isotopes in precipitation have been carried out since 1964 (Danssaard, 1964), and a variety of models are used to explain the observed precipitation isotopes (Gat, 1991; Yoshimura *et al.*, 2006). However, most land-surface models excessively simplify or even neglect isotopes of vapour evaporated from the land surface. For example, the atmospheric general circulation models by Hoffmann *et al.* incorporated isotopes but assumed no isotopic fractionation from the land surface (Hoffmann *et al.*, 1998). Yoshimura *et al.* assumed constant isotopic compositions from evaporative sources though noted that this method introduced discrepancies between observation and simulation results (Yoshimura *et al.*, 2003). It cannot be denied that the land-surface evaporation will influence precipitation and mass exchange at the land surface. Henderson-Sellers *et al.* pointed out that the climatic shift of precipitation isotopes over Amazonian forest could not be reasonably explained without the land processes (Henderson-Sellers *et al.*, 2002).

Recently, a few land-surface models that incorporate stable water isotopes have been developed, in which the inter-comparison project of water isotopes in land-surface processes starting in 2004, iPILPS (isotopes in the project for inter-comparison of land-surface parameterization schemes), is the most famous (Henderson-Sellers *et al.*, 2006). The iPILPS experiment aims to identify and test different land-surface schemes that incorporate stable water isotopes, appraise the applicability of stable isotopic data in hydro-climatic study and water resources survey, identify observational data gaps required for

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evaluating the land-surface schemes with isotopes and apply stable water isotopic data to specific prediction of hydrometeorological processes.

This study incorporates stable water isotopes in Community Land Model (CLM) as a diagnostic tool, simulates and analyses variations of the stable water isotopes in different reservoirs on daily and monthly time scales at Manaus, Brazil. The simulated behaviours of stable isotopes in precipitation on monthly time scale have good consistency with the actual survey results at the Manaus station set up by International Atomic Energy Agency (IAEA)/World Meteorological Organization (WMO), showing that the simulation by the CLM incorporating stable water isotopes is reasonable.

2. Model description

2.1. CLM

As an important tool for simulating climatic change, the land-surface process model is intensively required. The CLM is developed from the Biosphere-Atmosphere Transfer Scheme (BATS), the Institute of Atmospheric Physics, Chinese Academy of Sciences land model (IAP94) and the National Center for Atmospheric Research (NCAR) Land Surface Model (NCAR LSM) (Dai *et al*., 2003). It is designed for coupling to climate and numerical weather prediction models, and can be described as the merging of a community-developed land model focusing on biogeophysics and a concurrent effort at NCAR to expand the NCAR LSM. The model is a simplified treatment of surface processes that reproduces at a minimal computational cost the essential characteristics of land–atmosphere interactions important for climate simulations and weather prediction. Therefore, the CLM is selected for this study.

The CLM takes into account ecological differences among vegetation types, hydraulic and thermal differences among soil types, and allows for multiple land cover types within a grid cell. Strictly speaking, CLM is a single point model. According to different physical processes, the model structure can be separated into two parts, the biogeophysical processes relating to vegetation cover at the surface and the physical processes relating to hydraulic and thermal transfer which include mainly radiation transfer, turbulence diffusion and thermal conduction in soil and so on. Detailed descriptions about CLM may refer to relational references and technical notes (Dai *et al*., 2003; Oleson *et al*., 2004).

2.2. Stable water isotope parameterization

The stable isotopic ratio incorporated into CLM is noted as

$$R_w = \frac{H_2{}^{18}O}{H_2{}^{16}O}\left(\text{or } R_w = \frac{HD^{16}O}{H_2{}^{16}O}\right) \tag{1}$$

The subscript w stands for the reservoir water, for example precipitation, runoff, or vapour.

There are two possible ways of mixing the reservoir water with the input. In a 'total mixing' scheme,

$$\begin{aligned} R_w(t) &= [N_1 R_w(t-1) + N_2 R_{\text{inputs}}(t)]/N \\ R_{\text{overflow}}(t) &= R_w(t) \\ N &= N_1 + N_2 \end{aligned} \tag{2}$$

where R_{inputs} is the stable isotopic ratio of any inputs, R_{overflow} is the ratio of overflow, that is, the water exceeding the maximum storage capacity of the reservoir, N_1 is the mass of water in reservoir, N_2 is the mass of input water, N is the total mass after mixing and t is the time.

In a 'partial mixing' scheme, namely, as $\max(N_1) \geq N$

$$\begin{aligned} R_w(t) &= [N_1 R_w(t-1) + N_2 R_{\text{inputs}}(t)]/N \\ R_{\text{overflow}}(t) &= R_{\text{inputs}}(t) \\ N &= N_1 + N_2 \end{aligned} \tag{3}$$

As phase change is generated, there will appear a fractionation effect of the stable isotope. As water evaporates, the stable isotopic ratio in residual water is

$$R_w(t) = R_w(0) f^{\frac{1}{\alpha}-1} \tag{4}$$

where $f = \frac{N_1(t)}{N_1(0)}$ is the fraction of residual water in the reservoir after evaporation event, $\alpha = \frac{R_w}{R_v}$ is the fractionation factor of stable isotopes between liquid and vapour, and R_v is the ratio in evaporated vapour.

As vapour condenses,

$$R_d = \alpha R_a \tag{5}$$

where R_d and R_a are stable isotopic ratios in dew and in atmospheric vapour, respectively.

The stable isotopic fractionation factor α is calculated for the below two vapour condenses, the process is usually regarded to be in equilibrium state. The equilibrium fractionation factors of ^{18}O and deuterium are respectively Equations (6) and (7) (Majoube, 1971a,b)

$$\alpha(^{18}O) = \exp\left(\frac{1.137}{T^2} \times 10^3 - \frac{0.4136}{T} - 2.0667 \times 10^{-3}\right) \tag{6}$$

$$\alpha(D) = \exp\left(\frac{28.844}{T^2} \times 10^3 - \frac{76.248}{T} + 52.612 \times 10^{-3}\right) \tag{7}$$

Secondly, as reservoir water evaporates in an unsaturated environment, a kinetic fractionation effect will be produced. The kinetic fractionation factors of ^{18}O and deuterium are respectively Equations (8) and (9) (Craig and Gordon, 1965)

$$\alpha_k(^{18}\mathrm{O}) = \exp\left[\frac{0.35041}{T^3} \times 10^6 - \frac{1.6664}{T^2} \times 10^3 + \frac{6.7123}{T} - 0.007685 + 28.4n(1-h) \times 10^{-6}\right] \quad (8)$$

$$\alpha_k(\mathrm{D}) = \exp\left[1.1588(T^3 \times 10^{-9}) - 1.6201(T^2 \times 10^{-6}) + 0.79484(T \times 10^{-3}) - 0.16104 + \frac{2999200}{T^3} + 25.0n(1-h) \times 10^{-6}\right] \quad (9)$$

where T(Kelvin) in Equations (6)–(9) is the temperature at condensation or evaporation surface, h is the relative humidity and n is the turbulence parameter that is related to the reservoir property (Gat, 1996).

As known, vegetation transpiration does not produce stable isotopic fractionation (Fischer, 2006); thus, the stable isotopic ratio in transpiration equals to that in root region, namely,

$$R_{\mathrm{tr}} = R_{\mathrm{root}} \quad (10)$$

Usually, stable isotopic ratio is expressed as parts per thousand of their deviation relative to the Vienna Standard Mean Ocean Water (V-SMOW),

$$\delta_{\mathrm{w}} = \left(\frac{R_{\mathrm{w}} - R_{\text{V-SMOW}}}{R_{\text{V-SMOW}}}\right) \times 1000 \quad (11)$$

where δ_{w} is the stable isotopic ratio in reservoir (δ^{18}O or δD) and $R_{\text{V-SMOW}}$ is the isotopic ratio in V-SMOW.

2.3. Experiment scheme

Three sites with different geophysical and climatic conditions are selected for iPILPS Phase 1 experiment. They are Munich, Germany (48.08 °N, 11.34 °E), Tumbarumba, Australia (35.49 °S, 148.01 °E) located in middle latitudes and Manaus, Brazil (3.08 °S, 60.01 °W) in tropical rainforest of South America.

Manaus, with an equatorial climate characterized by agreeable temperatures but plenty of rain and humidity, is situated in the heart of Amazonas, north of Brazil, more than 1450 km inland from the Atlantic. According to statistical data, the annual mean precipitation amount is about 2190 mm at Manaus, with the maximal monthly mean precipitation of 308 mm in April and the minimal mean precipitation of 52 mm in August; the annual mean temperature is 26.8 °C, with the highest monthly mean temperature of 27.9 °C in October and the lowest monthly mean temperature of 26.0 °C in March.

The survey of stable isotopes in precipitation shows that there is the marked negative correlation between monthly stable isotopic ratios in precipitation and precipitation amount at Manaus (Henderson-Sellers *et al.*, 2002). In view of that, some variational features of stable isotopes in precipitation at Manaus have comparability with those under monsoon climate in East Asia, the simulation experiment for stable water isotopes was carried out at Manaus. For some basic parameters of Manaus, see Table I.

Table I. Summary of basic parameters at Manaus, Brazil.

Soil type	Vegetation type	Leaf area index	Vegetation cover (%)
Clay loam	Broadleaf evergreen forest	5.014	90

Table II. Variables provided from REMOiso as forcing.

Variable	Units
Large-scale precipitation rate ($H_2{}^{16}O$)	kg m^{-2} s^{-1}
Convective precipitation rate ($H_2{}^{16}O$)	kg m^{-2} s^{-1}
Atmospheric bottom level temperature	K
Downward shortwave radiation onto surface	W m^{-2}
Atmospheric bottom level specific humidity ($H_2{}^{16}O$)	kg kg^{-1}
Atmospheric bottom level zonal wind	m s^{-1}
Atmospheric bottom level meridional wind	m s^{-1}
Atmospheric surface pressure	Pa
$H_2{}^{18}O$ in large-scale precipitation	kg m^{-2} s^{-1}
HDO in large-scale precipitation	kg m^{-2} s^{-1}
$H_2{}^{18}O$ in convective precipitation	kg m^{-2} s^{-1}
HDO in convective precipitation	kg m^{-2} s^{-1}
Specific humidity $H_2{}^{18}O$	kg kg^{-1}
Specific humidity HDO	kg kg^{-1}

The CLM simulation requires forcing that contains isotopes in precipitation, atmospheric vapour, etc. at high resolution (Table II). These forcing data are derived from output of REMOiso (Regional Model with isotopes) at 15-min time step for one ideal year (360 days) (Sturm *et al.*, 2005; Fischer and Sturm, 2006).

In CLM experiment, the model iterates a 1-year calculation until differences between the initial and final values decrease below 0.01 mm/year for water storage and 0.01 mm/year × $R_{\text{V-SMOW}}$ for isotopic species.

3. Simulation results

3.1. Daily variations of the δ^{18}O and water budgets in land-surface reservoirs

By aggregating the 15-min simulation at a daily time scale in an equilibrium year, the daily variations of δ^{18}O in land-surface reservoirs and the corresponding water storages for inter-comparison are obtained at Manaus (Figure 1).

In Figure 1, precipitation and vapour, as well as their stable isotopes, show obvious seasonality and the typical isotopic signature in evergreen tropic forest: the heavy

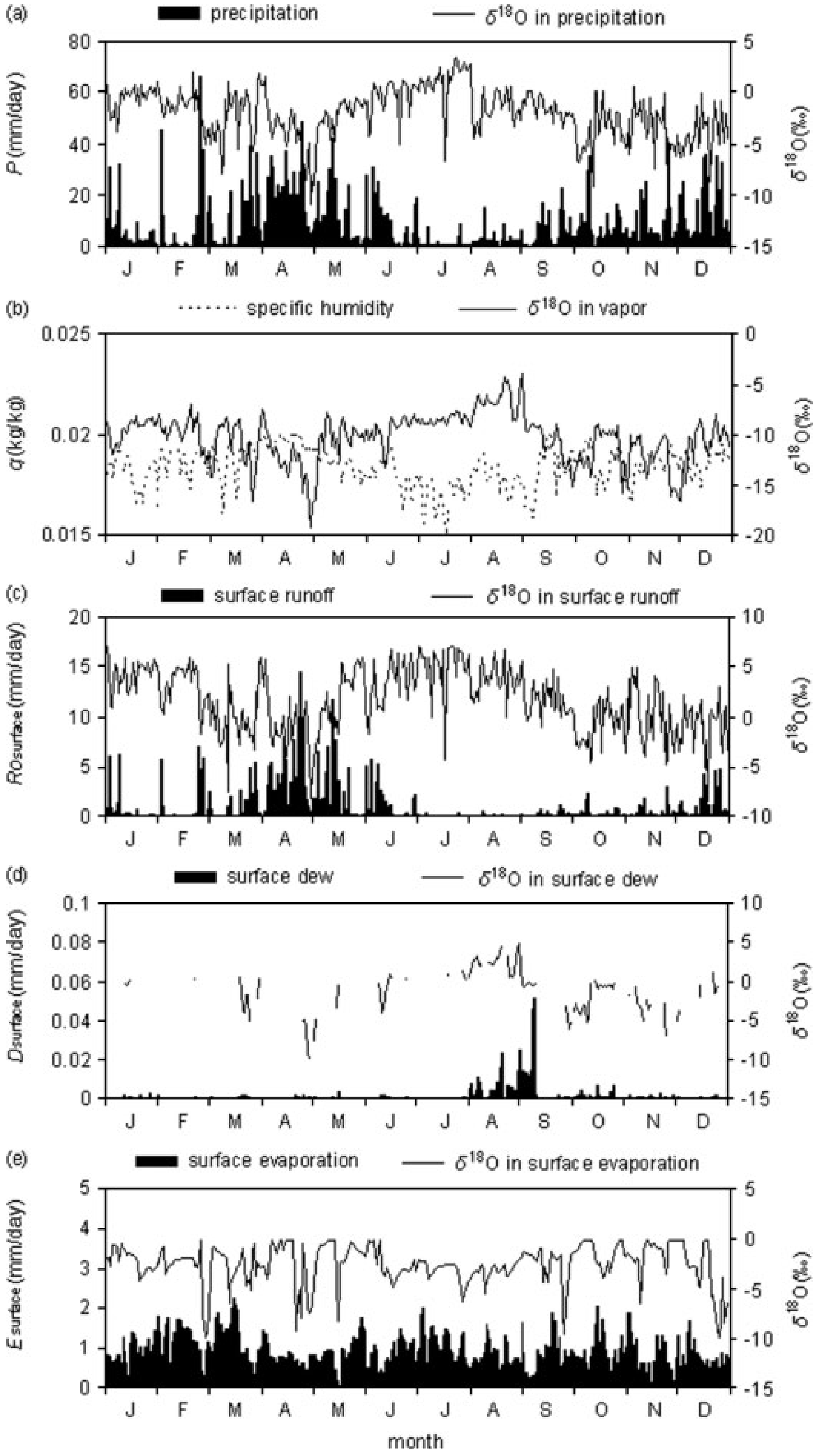

Figure 1. The daily variations of the δ^{18}O in precipitation (a), vapour (b), surface runoff (c), surface dew (d) and surface evaporation (e), with the corresponding water budgets at Manaus, Brazil.

rain is usually depleted in stable isotopes, whereas the light rain is usually enriched. The correlation coefficient between the δ^{18}O and precipitation achieves −0.28, exceeding 0.01 correlated confidence limits on t-test. Similarly, the δ^{18}O in vapour is lower in a high moist atmosphere (great q) than in a dry atmosphere (small q), with the negative correlation coefficient between them getting −0.46, exceeding 0.001 confidence limits. Compared with precipitation, vapour is isotopically depleted obviously.

Surface runoff depends on factors such as rainfall intensity, soil humidity, soil properties, and its variation keeps consistency with that of precipitation. Since surface runoff comes from rainfall primarily, the variation of δ^{18}O in surface runoff is similar to that in precipitation. In accordance with the amount effect, the isotopic composition in surface runoff is negatively correlated to the runoff amount (Figure 1(c)). The correlation coefficient between them gets −0.36, exceeding 0.001 confidence limits.

Unlike mentioned above, there is marked positive correlation between the δ^{18}O in surface dew and dew amount (Figure 1(d)), showing that the more the condensate, the higher the δ^{18}O in surface dew, and vice versa. Usually, the dry season has great temperature difference from daytime to night-time. The surface dew is easily generated owing to low temperature below the dew point at night or in early morning, and thus has high stable isotopic ratio because of high isotopic ratio in vapour in this season. In the atmosphere, condensation is generally treated as the Rayleigh equilibrium fractionation process that only is the function of temperature. In this process, stable isotopes in condensate are enriched. With the generation of condensation, the variation of the δ^{18}O in surface dew keeps very good consistency with that in vapour. Compared with vapour, surface dew is isotopically enriched.

Although evaporation is inverse process of condensation, no marked correlation exists between the δ^{18}O in evaporation and evaporation amount (Figure 1(e)). On one hand, such a result arises from that the evaporation draws water not only from surface storage directly but also from super-surface soil by capillarity, in which a part of the water is from the antecedent precipitation and another from groundwater. On the other hand, unlike condensation, surface evaporation is regarded as a disequilibrium process. Its fractionation is impacted by atmospheric humidity, wind speed, turbulence intensity and temperature and other factors.

3.2. Diurnal variations of the δ^{18}O and water budgets in land-surface reservoirs

Despite no marked distinction between summertime and wintertime in the tropics, intertropical convergence zone passes twice annually above the equatorial zone and reaches its southernmost and northernmost positions in January and in July, respectively. The seasonality in vapour origins and influencing factors will impact variation of stable water isotopes in reservoirs. Figures 2 and 3 show the δ^{18}O in surface storages and water budgets with hourly resolution created by averaging all the days in January and July.

On the diurnal timescale, the precipitation amount has obvious diurnal variation with the greater amount during the day than at night, no matter in January or in July. In contrast, the δ^{18}O in precipitation has an unclear diurnal variation and an indistinctive correlation with the precipitation amount.

The diurnal variation of atmospheric specific humidity is related to that of temperature. The specific humidity in daytime is always smaller than that at night-time, either in January or in July. On contrary to that in Figure 1, the δ^{18}O in vapour keeps consistency with specific humidity, with their correlated coefficient getting 0.76 in January, slightly lower than 0.81 in July. Evaporation is probably responsible for such a result. In daytime, more evaporated vapour reaches the atmosphere owing to stronger evaporating action than at night-time. Compared with condensation, the evaporation is isotopically depleted.

The diurnal variation of surface runoff relies on precipitation. The mean runoff in January is greater than that in July because of heavier precipitation in January than that in July. However, the δ^{18}O in runoff displays very different styles for the two months. In January, the δ^{18}O in runoff is positively correlated to the runoff amount with the correlated coefficient of 0.43, and shows bimodal distribution with two maxima around noon and midnight, and two minima in early morning and in nightfall, respectively. However, in July, no clear consistency exists between the δ^{18}O in runoff and runoff amount.

Condensation usually happens at night-time. There appear longer condensation time and greater condensation amount in January than in July due to higher specific humidity in January than in July. It can be seen from Figure 2(d) that the δ^{18}O in dew decreases continuously from midnight to early morning in January, which is attributed to the successive depletion of stable isotopes in atmospheric vapour for condensation fractionation.

Influenced by humidity and temperature, evaporation also shows marked diurnal variation. No matter in January or in July, the time of getting the maximal evaporation corresponds to the daily highest temperature and lowest humidity. Moreover, a weak negative correlation appears between the δ^{18}O in evaporated vapour and evaporation amount. At night-time, the δ^{18}O in evaporation keeps basically unchangeable on low evaporation. With increasing evaporation during the day, the δ^{18}O in evaporation decreases. However, in January, the δ^{18}O in evaporation increases after noon, which is presumably explained by successive enrichment in evaporating reservoir under the evaporation action.

3.3. Seasonal variations of the δ^{18}O and water budgets in land-surface reservoirs

By averaging the δ^{18}O in reservoirs and the daily specific humidity as well as aggregating the daily water budgets, the monthly variations of the δ^{18}O in reservoirs and corresponding water budgets are obtained (Figure 4).

On the monthly time scale, precipitation, specific humidity and surface runoff show all the obvious bimodal seasonality, with two maxima in April and in December and two minima in January and in July, respectively, which characterizes the climatic regime of equatorial zones. Correspondingly, the δ^{18}O in reservoirs also shows the bimodal seasonality, in which the variations of δ^{18}O in precipitation and surface runoff are comparable: their two maxima appear in January and in July and two minima in April and in October, with the negative correlation between stable isotopic ratio and water budget; as for the δ^{18}O in vapour,

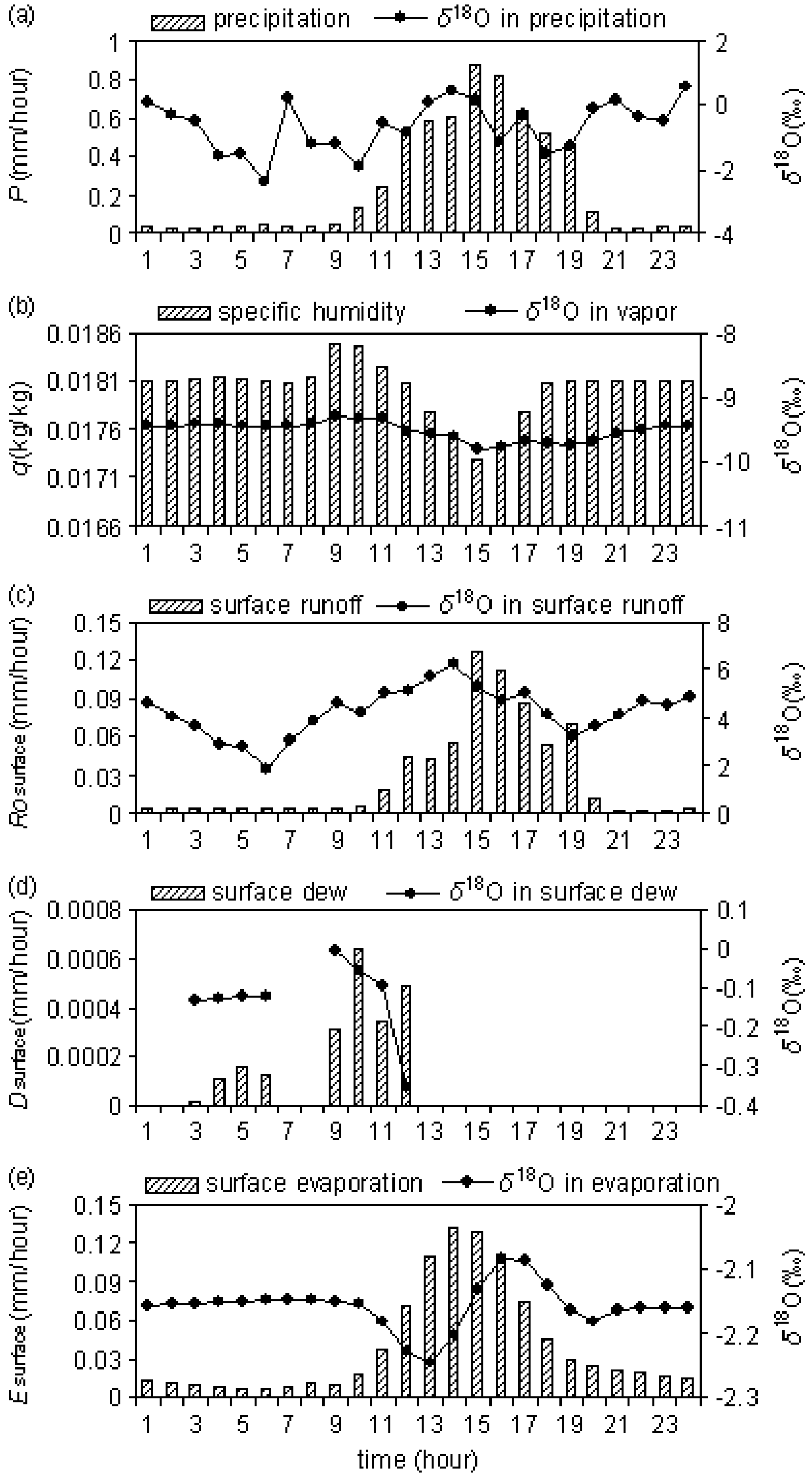

Figure 2. The mean diurnal variations of the δ^{18}O in precipitation (a), vapour (b), surface runoff (c), surface dew (d) and surface evaporation (e), with the corresponding water budgets in January at Manaus, Brazil.

two maxima appear in January and in August and two minima in April and in November, the second extremes being later than those in precipitation and in runoff.

Condensation happens mainly in August to October. The seasonality of the δ^{18}O in dew has very good consistency with that in vapour because dew is condensed from vapour directly. The magnitude can be determined by the Rayleigh equilibrium fraction process (Equation (5)). Dew is isotopically enriched markedly compared with vapour.

The magnitude of surface evaporation amount is related to atmospheric humidity. Comparing Figure 4(e) with (b), the evaporation is relatively small corresponding to two maximal specific humidity values in April and in December, but relatively great corresponding to the

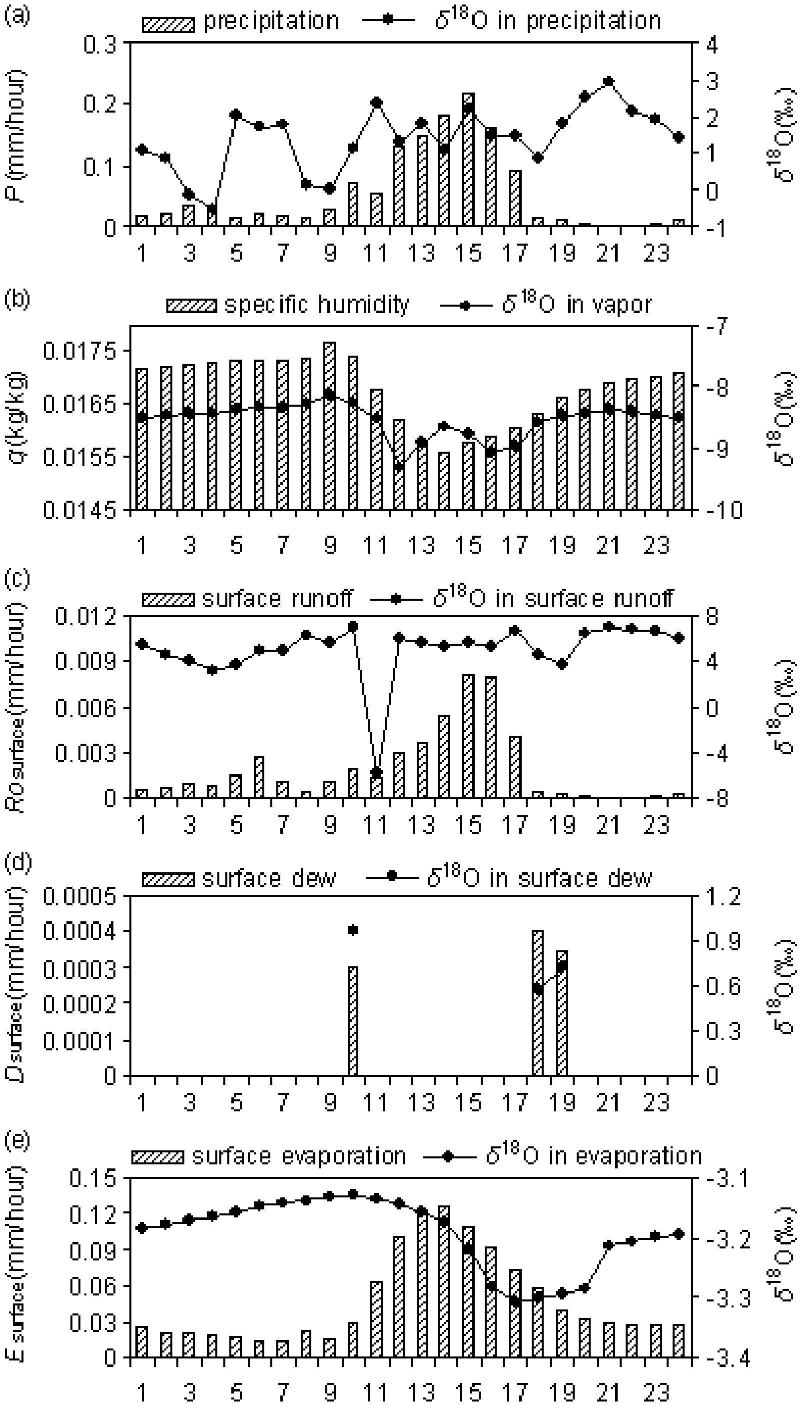

Figure 3. The mean diurnal variations of the δ^{18}O in precipitation (a), vapour (b), surface runoff (c), surface dew (d) and surface evaporation (e), with the corresponding water budgets in July at Manaus, Brazil.

minimal specific humidity in July. Unlike precipitation, specific humidity and condensation, evaporation shows the weak seasonality, and indistinctive correlation against the δ^{18}O in evaporated vapor.

3.4. Comparison between simulated and actual results

3.4.1. Comparison between actual and simulated amount effect

Manaus is one of sampling stations attached to the global survey network set up by the IAEA in co-operation with the WMO. There have been 26-year stable isotopic survey records from 1965 to 1990 (absent from 1993 to 1995) at Manaus. Based on these data, the correlation between monthly δ^{18}O in precipitation and precipitation is displayed in Figure 5.

From Figure 5, the maximal and minimal precipitations appear in April and August, respectively. On the contrary, the maximal δ^{18}O in precipitation appears in August, corresponding to the minimal precipitation amount; and the minimal δ^{18}O in precipitation in May,

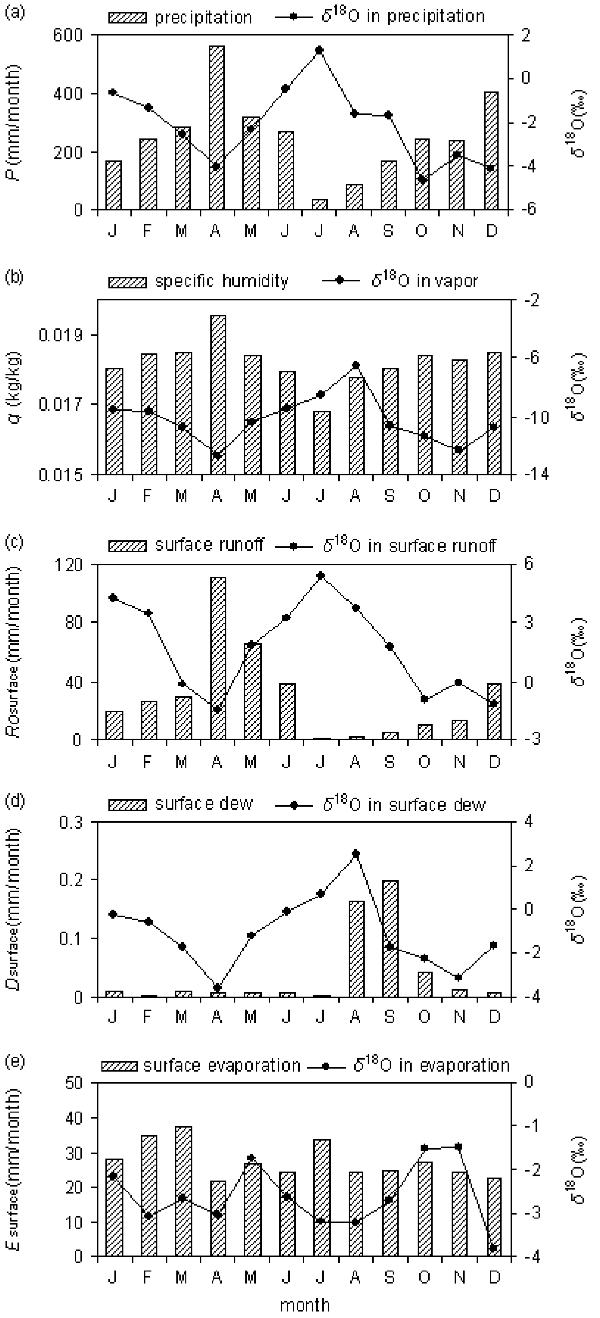

Figure 4. The mean seasonal variations of the δ^{18}O in reservoirs and corresponding water budgets at Manaus, Brazil; in precipitation (a), vapour (b), surface runoff (c), surface dew (d) and surface evaporation (e).

one month later than that of the minimal precipitation amount. Compared with Figure 4(a), the simulated seasonality of the δ^{18}O in precipitation and precipitation amount reproduces the actual temporal distribution.

On an average, the simulated precipitation amount is distinctly greater than actual ones, especially in April, June, October and December. Correspondingly, the simulated monthly δ^{18}O in precipitation during January through July is on the higher side and the time that the extreme appears is inconsistent with the actual one.

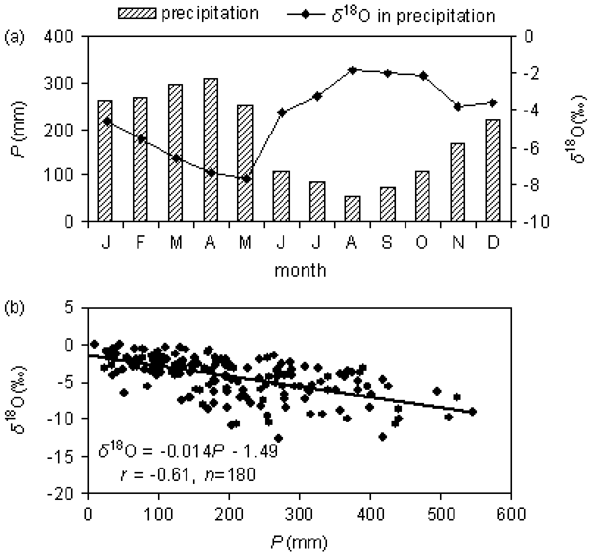

Figure 5. The actual seasonal variations of the δ^{18}O in precipitation and precipitation amount (a) and the scatter of the δ^{18}O in precipitation versus precipitation amount (b) at Manaus, Brazil (by the available data from 1965 to1990).

On the monthly timescale, the actual liner regression relation between the monthly δ^{18}O in precipitation and precipitation amount (Figure 5(b)) is

$$\delta^{18}\text{O}(\text{‰}) = -0.01P(\text{mm}) - 1.49 \quad n = 180 \quad r = -0.61 \qquad (12)$$

where n is sample size and r is the correlated coefficient. The correlated confidence limits exceed 0.001.

The simulated liner regression equation is

$$\delta^{18}\text{O}(\text{‰}) = -0.01P(\text{mm}) + 0.08 \quad n = 12 \quad r = -0.69 \qquad (13)$$

The correlated confidence limits exceed 0.02 (the figure omitted). There is a good consistency between simulated and actual results.

Because Figures 5(a) and 4(a) are the averages or aggregates for 26-year actual data and 15-min simulation in an equilibrium year, respectively, some discrepancies between them are inevitable.

3.4.2. *Comparison between actual and simulated MWLs*

The relationship between δD and δ^{18}O in atmospheric precipitation is called as the meteoric water line (MWL) (Craig, 1961; Danssaard, 1964). It has important indication for studying stable isotopic fractionation effect in the water cycle. The actual global MWL by Craig (1961) is

$$\delta\text{D}(\text{‰}) = 8.0\delta^{18}\text{O}(\text{‰}) + 10.0 \qquad (14)$$

The slope item 8.0 stands for the comparative relationship of fractionation rates between deuterium and oxygen-18, and the constant item 10.0 for the deviation degree of the deuterium from that in equilibrium state. They are controlled by all these phase-change processes from vapour evaporating from its origins to raindrops falling onto the land surface.

Based on the monthly IAEA/WMO δD and δ^{18}O in precipitation at Manaus, the annual MWLs, with the sample size above 9, are obtained (Table III). It can be seen that there are some differences among these WMLs, in which the smallest slope of these WMLs 5.66 appears in 1965, and the biggest one 9.58 in 1983; the smallest constant item 5.99 does in 1967, and the biggest one 16.79 in 1976.

By analysing the correlation between all monthly δD and δ^{18}O in precipitation at Manaus, the mean MWL is (Figure 6(a))

$$\delta\mathrm{D}(\text{‰}) = 8.14\delta^{18}\mathrm{O}(\text{‰}) + 12.96 \; n = 156 \; r = 0.98 \tag{15}$$

which is close to the global MWL, but with slightly greater slope and constant items. The simulated WML in an equilibrium year is:

$$\delta\mathrm{D}(\text{‰}) = 7.49\delta^{18}\mathrm{O}(\text{‰}) + 6.25 \; n = 12 \; r = 0.99 \tag{16}$$

It can be seen that the slope item 7.49 and the constant item 6.25 of the simulated MWL are all lower than those of the actual mean MWL. However, compared with annual MWLs in Table III, the simulated MWL lies within the variation range of actual MWLs.

4. Conclusions

This study incorporates stable water isotopes in CLM as a diagnostic tool, simulates and analyses the variations of stable water isotopes in different reservoirs on daily and monthly time scales at Manaus, Brazil. The simulated behaviours of stable isotopes in precipitation on monthly time scale are compared with actual survey result at Manaus station set up by IAEA/WMO. Some conclusions drawn are:

Table III. The actual MWLs at Manaus, Brazil.

Year	MWL	Correlated coefficient	Precipitation (mm)
1965	$\delta\mathrm{D} = 5.66\delta^{18}\mathrm{O} + 6.83$	0.97	1646
1966	$\delta\mathrm{D} = 7.19\delta^{18}\mathrm{O} + 9.59$	0.97	2332
1967	$\delta\mathrm{D} = 7.66\delta^{18}\mathrm{O} + 5.99$	0.99	2135
1968	$\delta\mathrm{D} = 6.55\delta^{18}\mathrm{O} + 11.76$	0.99	2420
1969	$\delta\mathrm{D} = 8.11\delta^{18}\mathrm{O} + 12.05$	0.98	2270
1973	$\delta\mathrm{D} = 8.07\delta^{18}\mathrm{O} + 14.97$	0.96	2568
1974	$\delta\mathrm{D} = 8.27\delta^{18}\mathrm{O} + 13.72$	0.99	2370
1975	$\delta\mathrm{D} = 7.74\delta^{18}\mathrm{O} + 11.19$	0.99	2326
1976	$\delta\mathrm{D} = 8.39\delta^{18}\mathrm{O} + 16.79$	0.99	2231
1982	$\delta\mathrm{D} = 7.93\delta^{18}\mathrm{O} + 10.31$	0.99	2237
1983	$\delta\mathrm{D} = 9.58\delta^{18}\mathrm{O} + 16.16$	0.98	2049
1984	$\delta\mathrm{D} = 8.52\delta^{18}\mathrm{O} + 14.54$	0.99	2093
1985	$\delta\mathrm{D} = 8.48\delta^{18}\mathrm{O} + 15.13$	0.93	2435
1986	$\delta\mathrm{D} = 8.49\delta^{18}\mathrm{O} + 11.94$	0.98	2418
1990	$\delta\mathrm{D} = 8.00\delta^{18}\mathrm{O} + 13.46$	0.98	1920

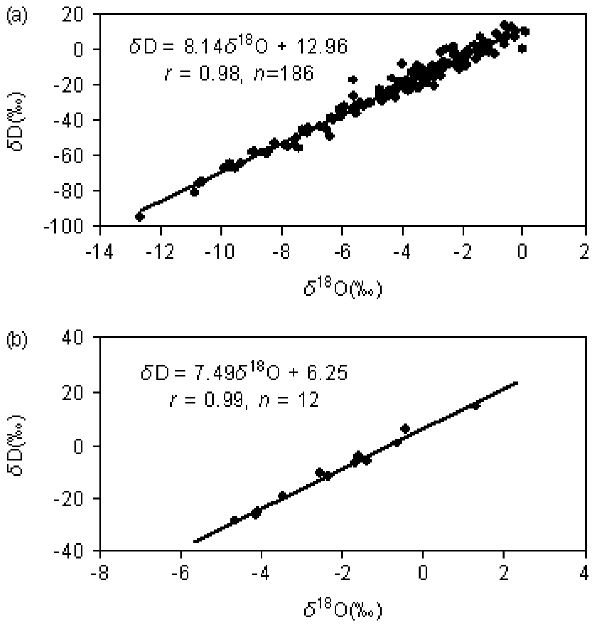

Figure 6. The comparison of actual MWL (a) with simulated MWL (b) at Manaus, Brazil.

1. The δ^{18}O in precipitation, vapour and surface runoff shows obvious seasonality and marked negative correlations with the corresponding water budget. Compared with precipitation, vapour is isotopically depleted.
2. There is marked positive correlation between the δ^{18}O in surface dew and dew amount. During condensation, the variations of the δ^{18}O in surface dew keep very good consistency with those in vapour. Although evaporation is an inverse process of condensation, no marked correlation exists between the δ^{18}O in evaporation and evaporation amount.
3. On the diurnal time scale, the δ^{18}O in precipitation has no obvious diurnal variation. However, the diurnal variation of the δ^{18}O in vapour keeps consistency with that of specific humidity.
4. On the seasonal time scale, precipitation, specific humidity and surface runoff show all the obvious bimodal seasonality with two maxima in April and December, and two minima in January and July. Correspondingly, the δ^{18}O in reservoirs also shows the bimodal seasonality, in which two maximal δ^{18}O values in precipitation and surface runoff appear in January and July, and two minimal δ^{18}O values in April and October. As for the δ^{18}O in vapour, two maxima appear in January and August, and two minima do in April and November.
5. Based on analyses of the actual data from the global survey network of IAEA/WMO, the simulated results are compared with actual results. On seasonal time scale, the simulated seasonality of the δ^{18}O in precipitation and precipitation amount, and the amount effect agree well with actual survey at Manaus. Moreover, the simulated MWL lies within the variation range of actual MWLs.

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