Simulations of Stable Isotopic Fractionation in Mixed Cloud in Middle Latitudes—Taking the Precipitation at Ürümqi as an Example

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ABSTRACT

The introduced mathematical model takes into account the role of the kinetic fractionation effect in a supersaturation environment at the ice surface as liquid and solid phases coexist in mixed cloud. Using the model, the temperature effect of stable isotopes in precipitation is simulated under different cooling conditions. The rate of change of δ^{18}O against temperature in the process of wet adiabatic cooling is smaller than in the process of isobaric cooling under the same humidity. The increasing supersaturation ratio at the ice surface, S_i, leads to the strengthening of the kinetic fractionation effect. The kinetic fractionation function makes the synthesis fractionation factor decreased and the change of δ^{18}O with temperature flatted, compared with that in the equilibrium state. The simulated results show that the slope parameter b and the intercept d of the meteoric water line (MWL), $\delta D = b\delta^{18}\mathrm{O}+d$, in wet adiabatic cooling are both greater than those in isobaric cooling. The global MWL lies between the two MWLs simulated under wet adiabatic and isobaric cooling processes, respectively. The magnitudes of b and d are directly proportional to S_i. The greater the S_i, the stronger the kinetic fractionation effect, and thus the greater the b and d, and vice versa. However, b and d have low sensitivity to the liquid-water contents in the cloud. Using the kinetic fractionation model, the variation of stable isotopes in precipitation at Ürümqi is simulated. The simulated stable isotopic ratio vs temperature and the δD vs δ^{18}O curves are very consistent with the actual regressions and MWL at Ürümqi, respectively.

Key words: stable isotopes, fractionation, kinetic effect, temperature effect, meteoric water line

1. Introduction

The behavior of stable isotopes in precipitation is intimately linked to the mechanism prevailing during cloud and precipitation formations. Temperature plays a very important role in phase change and is a main factor influencing isotopic change. The classic Rayleigh model is a useful tool in simulating the relationship between temperature and stable isotopic ratio in precipitation. The model assumes that the hydrometeor is immediately removed after its formation and that it leaves the air parcel in isotopic equilibrium with the vapor phase (Dansgaard, 1964). As stated above, a high fractionation is generated, which will lead to a high stable isotope/temperature slope.

The phase change that occurs in mixed cloud does not always remain in a complete equilibrium state in nature. The differences of vapor movement rate between liquid and sold phases, generated due to different humidity conditions, and of liquid-water content in cloud make the actual situation deviate more or less from the ideal results given by the Rayleigh model.

The stable isotopic fractionation model introduced in this study takes into account the kinetic fractiona-

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tion effect that occurs in the mixed cloud system and the influence of liquid-water content in cloud on stable isotopic fractionation. Using the improved model, the relationship between the stable isotopic ratio in precipitation and temperature at Ürümqi and the MWL are simulated. The results show that there is good consistency between the actual and simulated curves.

2. Basic model

It is assumed that a wet parcel does not mix with its outside environment, the liquid and air phases in the system remain in an equilibrium state, and the liquid water left in the system and that departing from it have the same stable isotopic ratio. Therefore,

$$dn_v + dn_l + dn_p = 0,$$
$$dn'_v + dn'_l + dn'_p = 0, \qquad (1)$$

where n_v, n_l, and n_p are the molecular amounts of vapor, liquid water, and precipitation, respectively; symbol $'$ is that of the corresponding stable isotopic species. In the equilibrium state,

$$\frac{n'_l}{n_l} = \alpha\frac{n'_v}{n_v}, \quad \frac{dn'_p}{dn_p} = \alpha\frac{n'_v}{n_v}. \qquad (2)$$

Here, α is the fractionation factor of the stable isotope. The stable isotopic ratio in vapor can be expressed as

$$R_v = \frac{n'_v}{n_v}. \qquad (3)$$

Therefore,

$$\frac{dn'_p}{dn_p} = \alpha R_v. \qquad (4)$$

According to Eqs. (1) through (4), we obtain

$$\frac{d\delta_p}{1+\delta_p} = \frac{M_v d\alpha + \alpha(\alpha-1)dM_v}{\alpha(M_v + \alpha M_l)}, \qquad (5)$$

in which, δ_p, deviated from the standard mean ocean water (SMOW) ($=R_p/R_{SMOW} - 1, R_p$ is the ratio in the water sample and R_{SMOW} is that in the SMOW), is the stable isotopic ratio in precipitation departing from the system, M_v the mixing ratio of the air parcel, and M_l the liquid-water content in the cloud.

If no liquid water remains in the system, namely M_l=0, then Eq. (5) will become the Rayleigh model,

$$\frac{d\delta_p}{1+\delta_p} = \frac{1}{\alpha}d\alpha + (\alpha-1)\frac{dM_v}{M_v}. \qquad (6)$$

It can be seen that δ_p is dependent on α and M_v. Because α and M_v respectively increase and decrease with decreasing temperature, and because the rate of change of α against temperature is smaller than that of M_v, the change of δ_p with temperature is roughly consistent with that of M_v. In the cooling process, $d\alpha > 0$ and $dM_v < 0$, thus, the sign of $d\delta_p$ is the same as that of dM_v. The greater the dM_v, the greater the $d\delta_p$, and vice versa.

As the cloud temperature descends below 0°C, the cloud is usually composed of supercooled water droplets, ice crystals, and vapor. The lower the temperature, the higher the proportion of the ice crystal in the cloud. The saturated or even the non-saturated environment for droplets has been probably supersaturated for ice crystals because of the lower saturated vapor pressure at the ice surface than at the water surface under the same temperature. The rapid growth of ice crystals will lead to the breakdown of the equilibrium state. Consequently, the stable isotopic fractionation is composed of two parts: the Rayleigh fractionation, namely the fractionation in the equilibrium state, and the kinetic fractionation. Define the supersaturation ratio at the ice surface as $S_i = e_v/e_i$ in which e_v is the vapor pressure in the system and e_i the saturation vapor pressure at the ice temperature. In fact, it is very difficult to measure ice temperature, but the atmospheric temperature is regarded as an approximate substitution under saturation, and M_v can be approximately expressed as

$$M_v = 0.622\frac{e_v}{P} = 0.622\frac{S_i e_i}{P}\ (\mathrm{g\ g^{-1}}), \qquad (7)$$

where P is pressure, then

$$\frac{dM_v}{M_v} = \frac{de_i}{e_i} + \frac{dS_i}{S_i} - \frac{dP}{P}. \qquad (8)$$

It is known that ice is of a low molecular diffusivity(Jouzel and Merlivat, 1984). It is assumed that the vapor flux pointing toward the ice surface F is directly proportional to $D(e_v - e_i)$, in which D is the atmospheric diffusion coefficient. As for stable isotopes,

$$e'_v = e_v R_{SMOW}(1+\delta_v)$$
$$e'_i = e_i R_{SMOW}(1+\delta_i)/\alpha_i, \qquad (9)$$

in which α_i is the equilibrium fractionation factor between the vapor and ice phases. The stable isotopic ratio in the ice phase δ_i is related to the ratio between vapor flux F and stable isotopic flux in vapor F'

$$1+\delta_i = (1/R_{SMOW})\left(\frac{F'}{F}\right), \qquad (10)$$

namely,

$$1+\delta_i = \frac{D'\left[e_v(1+\delta_v) - e_i(1+\delta_i)/\alpha_i\right]}{D(e_v - e_i)}. \qquad (11)$$

The above equation can be written as

$$1+\delta_i = \alpha_k\alpha_i(1+\delta_v), \qquad (12)$$

where

$$\alpha_k = \frac{S_i}{\alpha_i\left(\frac{D}{D'}\right)(S_i - 1) + 1}, \qquad (13)$$

α_k is the kinetic fractionation factor and its value is less than or equal to 1.0. As $S_i = 1.0, \alpha_k = 1.0$. The greater the S_i, the smaller the α_k, and thus the stronger the kinetic fractionation effect; the synthesis fractionation factor α_e ($=\alpha_k\alpha_i$) is smaller than α_i, which shows that stable isotopic fractionation is weakened under the kinetic effect; D/D' in air is based on the molecular weights,

$$\frac{D}{D'} = \left[\frac{m'(m+m_c)}{m(m'+m_c)}\right]^{\frac{1}{2}}, \tag{14}$$

where m_c is the molecular weight of gases in atmosphere; m and m' are those of pure water and stable isotopes, respectively. Actual measurements show that D/D'=1.0285 for oxygen-18, and 1.0251 for deuterium (Merlivat, 1978a, b).

Compared with $1+\delta_i=\alpha_i(1+\delta_v)$ in the equilibrium state, the α_i is replaced by the α_e in the kinetic model. Quantity α_e in the kinetic fractionation process is calculated as the function of temperature and supersaturation ratio at the ice surface. Setting $S_{i,1} = 1.2, S_{i,2} = 1.0 - 0.003t$, and $S_{i,3} = 0.05 + 0.969\exp(-0.008t)$, the relationships of the α_e with temperature and the supersaturation ratio at the ice surface are analyzed (Fig. 1). In addition, Fig. 1 also gives the variations, with temperature, of the fractionation factors of oxygen-18 α_i and α_l (between vapor and liquid phases).

From Fig. 1, it can be seen that

(i) α_i is always greater than α_l, showing that the δ_i in sublimated ice crystals is always greater than the δ_l in condensed droplets in the same air parcel and for the same temperature.

(ii) In the range of $t < 0$°C, α_e is always less than α_i, showing that fractionation of stable isotopes is weakened because of the generation of the kinetic effect. That is to say, the actual δ_i is smaller than that in the equilibrium state.

(iii) α_e decreases with increasing temperature for constant $S_{i,1}$. Corresponding to the variational S_i, α_e keeps either basically stable (such as $S_{i,2}$) or increasing (such as the $S_{i,3}$) with increasing temperature, showing that the synthesis fractionation factor α_e is controlled not only by temperature but also by the humidity condition to a great degree.

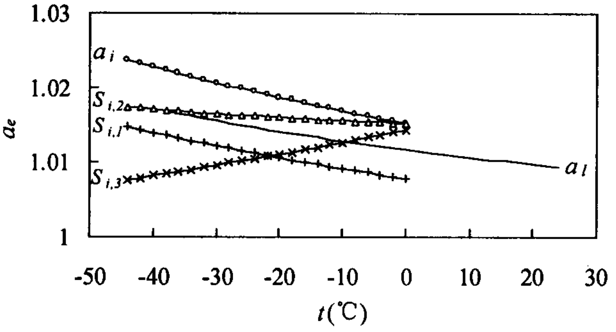

Fig. 1. The variations of fractionation factors of oxygen-18 with temperature.

3. The changes of the stable isotopic ratio in different phase changes

3.1 *Influences of different initial conditions on stable isotopes*

Usually, the stable isotopes in precipitation in middle latitudes are considered to originate mainly from tropic and subtropical oceans (Jouzel et al., 1987). Different temperature and humidity conditions in oceans will lead to different stable isotopic compositions in initially evaporated vapor and in initially condensed hydrometeors, and thus influence the thereafter successive changes of stable isotopic compositions in phase changes. The stable isotopic ratios in initially evaporated vapor at the ocean surface $\delta_{v,0}$ can be determined by (Merlivat and Jouzel, 1979)

$$\delta_{v,0} = \frac{(\delta_s+1)(1-k)}{\alpha_s(1-kf)} - 1, \tag{15}$$

where δ_s, assumed as 0, is the stable isotopic ratio in ocean surface water; α_s is the stable isotopic fractionation factor at the ocean surface temperature t_s; f is the relative humidity; and k is the roughness coefficient, related to diffusivity of vapor on the ocean surface. Parameter $k \approx 0.7‰$ on a relatively calm surface. After determining $\delta_{v,0}$, the stable isotopic ratio $\delta_{l,0}$ in initial hydrometeors can be calculated according to air humidity f, air temperature t_0, and surface pressure P_0. Figure 2 displays the changes of the $\delta^{18}O$ in the initial hydrometeors corresponding to different sea surface temperatures t_s (setting $t_0 = 29$°C, P_0=1010 hPa and f=80%) and different relative humidities f (setting t_0=29°C, P_0=1010 hPa and t_s=26°C).

Under the given conditions, the $\delta^{18}O_{l,0}$ is directly proportional to t_s and f, showing that more amounts of stable isotopic compositions in oceans can be fractionated into atmosphere in a warm and moist oceanic environment, and less in a cold and dry environment.

3.2 *The changes of $\delta^{18}O$ with temperature in different phase changes and under different humidities*

Knowing the $\delta^{18}O_{l,0}$ and the initial condensation state, the changes of $\delta^{18}O$ with temperature in the process of different phase changes and under different humidities can be calculated relying on the above model and the relationship between temperature and pressure. The integration starts from the initial state and is divided into two groups: the wet adiabatic cooling process I and the isobaric cooling process II. Figure 3 gives the δ^{18}O changes of the two groups in different phases and under different humidities. For each group, the considered supersaturation ratios are $S_{\rm i,1}$ and $S_{\rm i,2}$, and the corresponding equilibrium fractionation processes are Rayleigh-I and Rayleigh-II, respectively. From Fig. 3a, it can be found that (i) The δ^{18}O in the wet adiabatic cooling process is greater than in the isobaric process, showing that the fractionation rate of the latter is faster than that of the former.

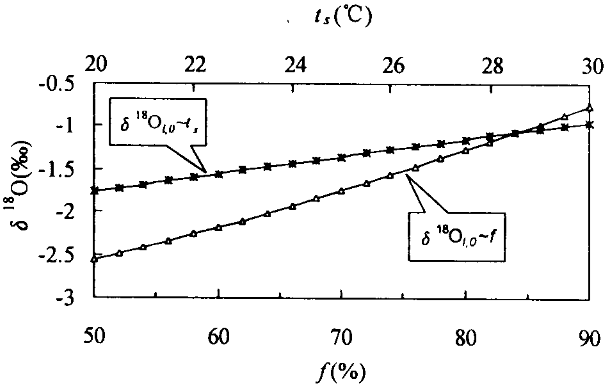

Fig. 2. Relationship of the δ^{18}O in the initial hydrometeors with sea surface temperature t_s and relative humidity f.

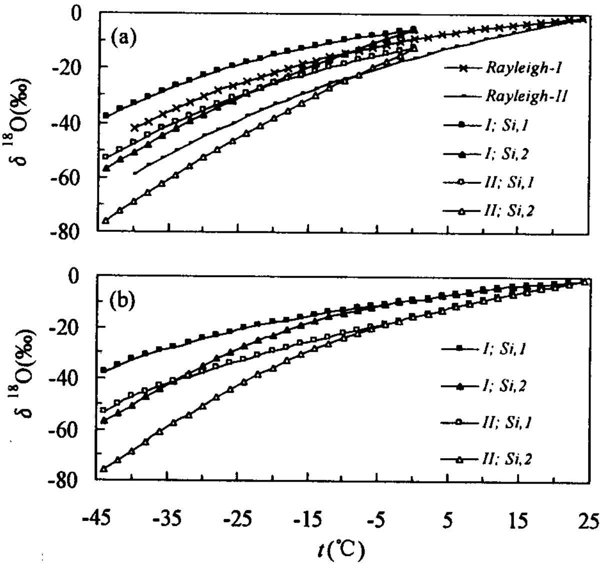

Fig. 3. Variations of the δ^{18}O with temperature (a) in liquid and solid phases, and (b) in precipitation under different supersaturation ratios at the ice surface.

(ii) Compared with the δ^{18}O in the liquid phase in the equilibrium process ($S_{\rm i}$=1), δ^{18}O in the solid phase is of a smaller change rate for greater $S_{\rm i}$ and a greater one for smaller $S_{\rm i}$.

(iv) With increasing $S_{\rm i}$, the kinetic fractionation effect is strengthened, and perhaps $\alpha_e \le 1$ because of the decrease of $\alpha_{\rm k}$, which means that the δ^{18}O in the solid phase will remain unchanged or increase with decreasing temperature. Such a situation had been observed in low-latitude oceans and coastal regions (Zhang and Yao, 1994).

As the temperature in the cloud layer descends below 0°C, supercooling droplets, ice crystals, and vapor coexist. The precipitation generated in this layer includes solid and liquid materials. Therefore, the δ^{18}O in precipitation is the weighted averages of the δ^{18}O in the liquid and solid materials. According to the features of mixed cloud, assuming that the constituent proportion R of ice crystals in the cloud display a linear increase from 0°C until −40°C, and the cloud is completely constituted by ice crystals below −40°C,

$$R=\begin{cases} 0 & t>0^\circ{\rm C},\\ -t/40 & -40^\circ{\rm C}<t\le 0^\circ{\rm C},\\ 1 & t\le -40^\circ{\rm C}, \end{cases} \qquad (16)$$

then, the stable isotopic ratio $\delta_{\rm p}$ in precipitation, generated in this layer, can be expressed as

$$\delta_{\rm p}=\delta_{\rm l}\,(1-R)+\delta_{\rm i}R. \qquad (17)$$

The calculated results are shown in Fig. 3b. The curves in the figure stand for the weighted averages of the δ^{18}O in the materials of the two phases. Their magnitudes are dependent on the temperature and humidity condition in the cloud.

In the figure, the changes of $d\delta_{\rm p}/dT$ with decreasing temperature under lower humidity (such as $S_{\rm i,2}$) are greater than under higher humidity (such as $S_{\rm i,1}$). The important function of the kinetic fractionation effect in mixed cloud is mirrored here.

4. The influence of liquid-water content in the cloud on δ^{18}O

Unlike the hypotheses made by the Rayleigh model, there is usually liquid water in cloud, especially in convective cloud with strong updraft. Assuming that t_0 is the temperature of the cloud base, the distribution of liquid-water content $M_{\rm l}$ of the cloud with temperature is expressed as

$$M_{\rm l}=\begin{cases} -c_1\int_{t_0}^{t} dM_{\rm v} & 0\le t,\\ -c_2\int_{t_0}^{-t} dM_{\rm v} & t<0,\\ 0 & t\le -t_0, \end{cases} \qquad (18)$$

in which c_1 and c_2 are the liquid-water content coefficients in cloud, and both greater than 0. They stand for the condensation amount still remaining in the cloud.

In the study, two cases are considered only in the wet adiabatic cooling process, case A: $c_1=1/3$ and $c_2=1/4$, and case B: $c_1=1/2$ and $c_2=1/3$.

Different liquid-water contents in the cloud are input into the Rayleigh model, noted as Rayleigh-A corresponding to case A and Rayleigh-B to case B. The supersaturation ratios are $S_{i,1}$ and $S_{i,2}$, respectively. Figure 4 gives the calculated results.

The liquid water contents in the cloud are of heavy $\delta^{18}O$ values because they come from the previous condensation process. Consequently, the curves of $\delta^{18}O$ against temperature have low slopes, compared with ones in which the liquid-water contents in the cloud are not considered. The more the liquid-water contents in the cloud, the lower the slope.

5. The relationship between δD and $\delta^{18}O$

The relationship between δD and $\delta^{18}O$ in atmospheric precipitation, $\delta D = b\delta^{18}O+d$, is called the meteoric water line (MWL). It has important significance for studying stable isotopic changes in the water cycle. Craig (1961) gave the global MWL

$$\delta D(‰) = 8.0\delta^{18}O(‰)+ 10.0. \tag{19}$$

As is known, the slope parameter of the MWL expression stands for the comparative relationship between two kinds of fractionation rates from oxygen-18 and deuterium, and the intercept term gives the deviation degree of actual deuterium from that in the equilibrium state. These are all controlled by phase-change processes from evaporation at the vapor origin site to the falling site. In the equilibrium state, b=8.0 and d=0.0.

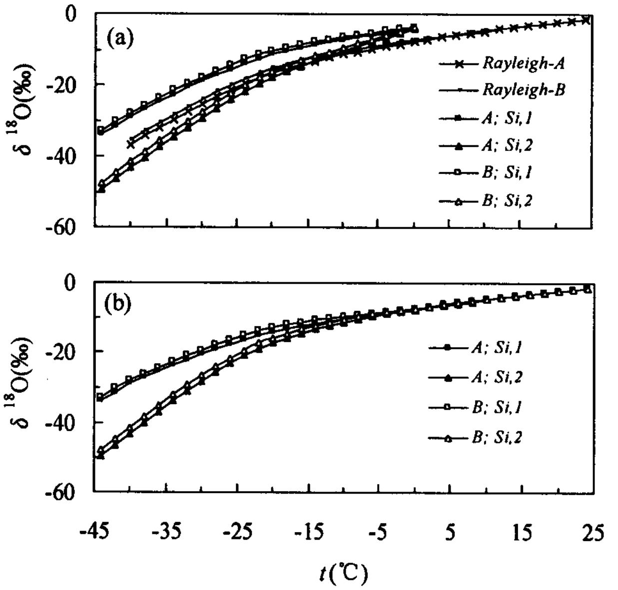

Fig. 4. Influence of liquid-water contents in cloud on $\delta^{18}O$ under different humidities in wet adiabatic cooling (a) $\delta^{18}O$ in different phases against temperature; (b) the weighted average $\delta^{18}O$ against temperature.

Using the kinetic model, the relationship between δD and $\delta^{18}O$ are theoretically analyzed under different cooling conditions, different supersaturation ratios at the ice surface, and different liquid-water contents in the cloud (Fig. 5). The cooling range is from the initial condensation temperature down to −30°C.

(i) the influence of the cooling conditions

It can be seen from Fig. 5a that the slope and the intercept in the wet adiabatic process I are greater than those in the isobaric process II. Moreover, the global MWL lies between the I and II lines, showing that the actual precipitation is generated neither in the wet adiabatic nor the isobaric cooling processes, but in the synthesis of both.

(ii) the influence of the supersaturation ratio in cloud

Because processes I and II have basically similar features, the influences of different supersaturation ratios ($S_{i,1}$ and $S_{i,2}$) and different liquid-water contents (case A and case B) on MWLs are analyzed here, only considering the situation in the wet adiabatic cooling process.

When inputting the influence of the supersaturation ratio at the ice surface into the kinetic model, the simulated results show that the b and the d have a marked positive correlation against S_i (Fig. 5b). Compared with MWL in the wet adiabatic cooling process under Rayleigh conditions, the supersaturation ratio S_i exerts a marked impact on MWL. All slopes are greater than 8.0, and intercepts greater than 10.0. Such a result shows that the δD and the $\delta^{18}O$ in the phase change have a strong sensitivity to the kinetic fractionation effect under supersaturation at the ice surface.

(iii) the influence of the liquid-water contents in cloud

Although the magnitude of the liquid-water contents in the cloud is directly proportional to those of b and d in MWL, its degree of influence is inferior to the supersaturation at the ice surface. Comparatively, the δD and the $\delta^{18}O$ in the phase change are of weak sensitivity to the variation of liquid-water contents in the cloud.

6. The stable isotopes at Ürümqi

The investigation of the stable isotopic changes in water cycle started in the early 1950s (Dansgaard, 1953). A large-scale and organized collection of precipitation samples was initiated in 1961 (Dansgaard, 1964). The International Atomic Energy Agency (IAEA) in co-operation with the World Meteorological Organization (WMO) set up a global network with more than 550 meteorological stations that continually collect monthly precipitation samples for the environmental isotopic compositions in precipitation and in additional meteorological elements. The main objective of the global survey program was to determine the temporal and spatial variations of environmental isotopes in precipitation and, consequently, to provide basic isotope data for the use of environmental isotopes in hydrological investigations for water resources inventory, planning, and development. China entered late into the global survey network set by IAEA/WMO. Ürümqi is one of the Chinese sampling stations in the global network. The station (43.78°N, 87.62°E, 918 m MSL) is in an arid and semi-arid inner continent of Northwest China, and is the most distant from the oceans of the world. The precipitation is mainly influenced by the westerlies fluctuation, and the vapor originating from the North Atlantic and the evaporation recharge of different water bodies on the way. Its mean annual precipitation amount is 283 mm with the annual amplitude only 28 mm and the annual mean temperature is 7.4°C with the annual amplitude of 37°C. Since it is located on the windward slope of the eastern Tianshan Mountains, the precipitation at Ürümqi is abundant and relatively well-distributed (Fig. 6a). It has the longer survey series of the two IAEA/WMO stations in Xinjiang (the another is Hetian station). Since 1985, an 11-year record has been obtained. For details on the sampling and the measurement of stable isotopes in precipitation at Ürümqi, refer to the paper of Zhang et al. (2002).

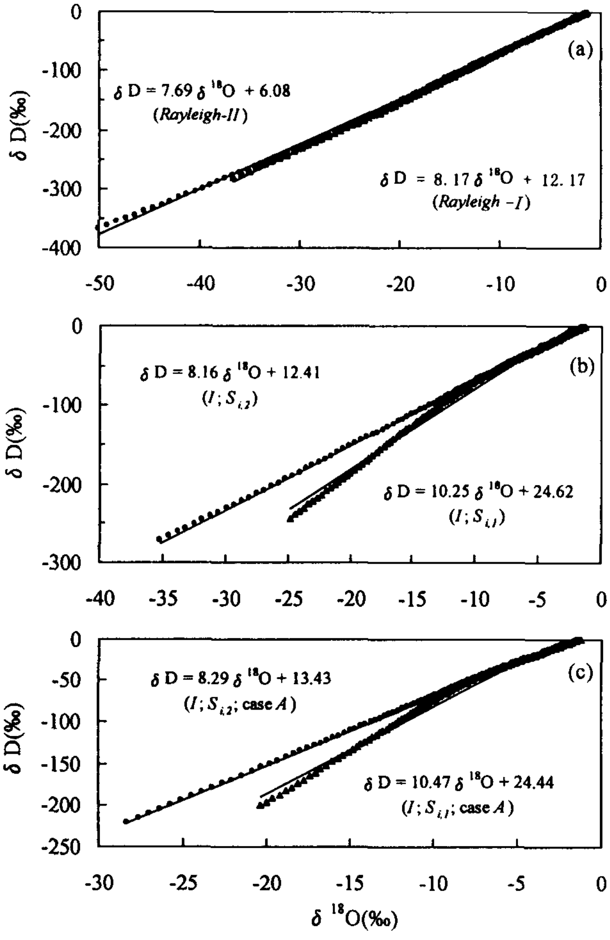

Fig. 5. The simulations of MWLs under (a) different cooling conditions, (b) different supersaturation ratios at the ice surface, and (c) different liquid-water contents in the cloud.

6.1 *The simulation of the temperature effect*

The temperature effect, namely the marked positive correlation between the stable isotopic ratio and temperature, is recognized as the most important feature on stable isotopic changes in precipitation. It mainly appears in mid-high latitude continents. The climatic information recorded in the snow and ice of mid-high latitudes can be reasonably interpreted using a simple linear relationship between the stable isotopic ratio and the temperature (Rozanski et al., 1992; Yao et al., 1996). According to the above analyses, one can see that the temperature effect arises from the fact that the fractionation of stable isotopes in the atmosphere is related to phase-change temperature in the process of phase change.

According to the variations of the monthly δ^{18}O and δD (Fig. 6a), and the scatter plots of the monthly δ^{18}O and δD against the mean monthly temperature (Figs. 6b and 6c), there is a marked temperature effect at Ürümqi. For oxygen-18,

$$\delta^{18}\mathrm{O}(‰) = 0.4131t(°\mathrm{C}) - 15.732 \qquad r = 0.8602, \quad (20)$$

and for deuterium,

$$\delta D(‰) = 3.0415t(°\mathrm{C}) - 112.14 \qquad r = 0.8567, \quad (21)$$

Both confidence limits exceed the 0.001 level.

Setting the stable isotopic ratios in the surface water in vapor origins $\delta^{18}\mathrm{O}_s$=0.0 and δD_s=0.0, t_0=29°C, P_0=1010 hPa, f=80%, t_s=26°C; the super saturation ratio at ice surface under the wet adiabatic cooling process S_i=1.0–0.003t, the liquid-water content coefficients in cloud c_1=1/5 and c_2=1/6, and the simulated curves L1 are obtained. They reveal the relationship between the stable isotopic ratios in the initial hydrometeors from ocean vapor and the condensation temperature. It can be seen that there are great diff-

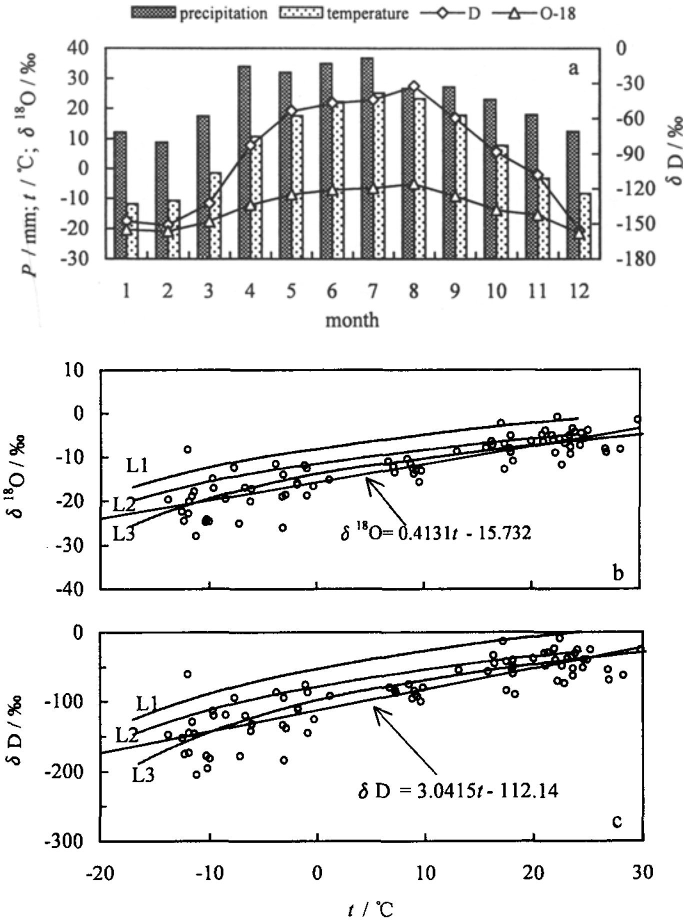

Fig. 6. Seasonal variations of the temperature, precipitation, δ^{18}O and δD at Ürümqi as well as the simulations of temperature effect.

erences between the simulated curves and the actual regression lines.

Keeping the other conditions unchanged, setting $\delta^{18}\mathrm{O_s}$=–3.5‰, and $\delta D_\mathrm{s} = 8.0 \times \delta^{18}\mathrm{O_s} = -28.0$‰, the simulated curves L2 are similar to the actual regressions to a certain degree.

In fact, it is impossible that the δ^{18}O and δD in the surface water of oceans is as light as the assumption. So, it can be concluded that the rainfall at Ürümqi is not the outcome of the initial condensation of ocean vapor. The stable isotopic compositions in the precipitation are greatly depleted after vapor undergoes long distance transportation. The mean depleted magnitude is about –3.5‰ for oxygen-18 and about –28.0‰ for deuterium.

It needs to be pointed out that the condensation temperature in the kinetic model is different from the sampling temperature. Assuming the mean condensation level is about 1000 m, the difference between the condensation temperature at the mean condensation level and the sampling temperature is about 6.5°C. The condensation temperature in the cloud is converted into the sampling temperature as the assumption, and consequently, the simulated curves L3 are in very good agreement with the actual regressions.

6.2 *The simulation of* MWL

According to the analyses of the relationship between the monthly δD and δ^{18}O in the precipitation at Ürümqi, the MWL is obtained (Fig. 7)

$$\delta D(‰) = 7.2892\delta^{18}\mathrm{O}(‰) + 2.7798 \quad r = 0.8567. \qquad (22)$$

Compared with the global WML, both the slope and the intercept of the Ürümqi's MWL are smaller.

The simulated curve L is obtained by plotting the relevant δD and δ^{18}O in simulated curves L2 (or L3) in Fig. 6b and 6c onto Fig. 7. It is quite consistent with the actual MWL. However, with the small difference between the two, the slope parameter and the intercept term of the simulated curve are both greater than the actual values.

Strictly speaking, curve L should be called a condensation line owing to mirroring the relationship between the δD and the δ^{18}O in hydrometeors in mixed cloud. The stable isotopic ratio in precipitation reaching the ground is different from that in raindrops departing from cloud base, influenced by evaporation and the isotopic exchange of raindrops with the surrounding atmosphere during falling (Zhang et al., 1998). Because of the relatively fast fractionation rate for deuterium, the slope parameter and the intercept term of the actual MWL are both smaller than those of the condensation line.

7. Conclusions

(1) The kinetic fractionation effect is controlled by the supersaturation ratio at the ice surface in mixed cloud. The greater the supersaturation ratio at the ice

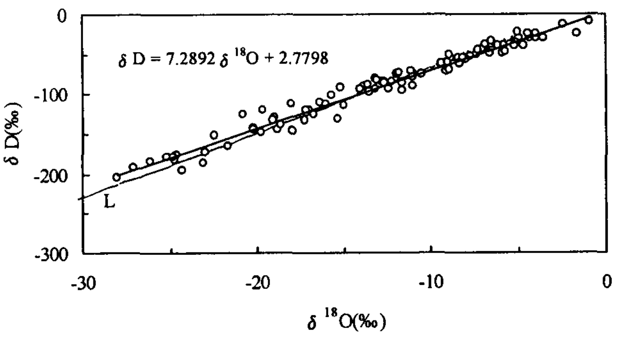

Fig. 7. The actual and simulated MWL of Ürümqi.

surface, the stronger the kinetic fractionation effect. The kinetic fractionation function makes the synthesis fractionation factor decreased and the change of stable isotopes with temperature flatted. The contribution of liquid-water contents in the cloud is only to change the magnitude of the stable isotopic ratio in water, but not the variation trend with temperature.

(2) The slope parameter b and the intercept term d in MWL, $\delta D = b\delta^{18}O + d$, in wet adiabatic cooling are both greater than those in isobaric cooling. Also, b and d are directly proportional to S_i. The greater the S_i, the stronger the kinetic fractionation effect, and thus the greater the b and d, and vice versa. However, the changes of b and d have weak sensitivity to the liquid-water contents in cloud.

(3) The precipitation at Ürümqi is not the outcome of the initial condensation of ocean vapor. The stable isotopic compositions in precipitation are greatly depleted after vapor undergoes long distance transportation. The mean depleted magnitude is about –3.5‰ for oxygen-18 and about –28.0‰ for deuterium.

(4) The simulated $\delta D/\delta^{18}O$ curve is quite consistent with the actual MWL at Ürümqi. The difference between the two is related to evaporation and the isotopic exchange of raindrops with the surrounding atmosphere during falling.

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中纬度地区混合云中稳定同位素分馏的数学模拟
—以乌鲁木齐降水为例

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摘　要

介绍的数学模型考虑了混合云中液、固态共存时以及冰面过饱和环境下稳定同位素的动力分馏效应。利用该数学模型，模拟了不同冷却条件下稳定同位素的温度效应。在相同的湿度条件下，湿绝热冷却过程中$\delta^{18}O$随温度的变化率小于等压冷却过程。冰面过饱和比S_i的增大意味着动力分馏效应的增大。与平衡态相比，它的作用使得稳定同位素的综合分馏系数减小，从而使得降水中$\delta^{18}O$随温度的变化趋缓。模拟显示，湿绝热冷却过程中大气水线(MWL：$= b\delta^{18}O + d$) 的斜率b和常数d均大于等压冷却过程。全球大气水线位于湿绝热冷却过程和等压冷却过程条件下分别模拟的两条大气水线之间。b 和d 的大小与S_i呈正比。S_i愈大，动力分馏效应愈强，b 和d也愈大。反之亦然。然而，b 和d 的大小对云中含水量的变化具有低的敏感性。利用动力分馏模式模拟了乌鲁木齐降水中稳定同位素的变化。模拟的稳定同位素比率–温度以及δD–$\delta^{18}O$曲线分别与乌鲁木齐实测的稳定同位素比率–温度回归线以及大气水线有非常好的一致性。

关键词: 稳定同位素, 分馏, 动力效应, 温度效应, 大气水线