Isotopic signature of evapotranspiration flux and its use for partitioning evaporation/transpiration components

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Abstract

Isotopic signature of evapotranspiration flux gives us useful information for tracing global atmospheric water cycle. It can be also used for partitioning evaporation/transpiration components. In the present study, isotopic measurements of water below and above the land-atmosphere interface were conducted in temperate grassland, central Japan, throughout a growing season to assess change in relative contribution of transpiration and its relation to vegetation growth. The isotopic composition of evapotranspiration flux (δ_{ET}) was determined by "Keeling plot" approach based on dual level measurements of water vapor isotopes. The δ_{ET} values were generally close to, but somewhat lower than, those for shallow soil water and antecedent rainfall, suggesting a little contribution of evaporation component affected by isotopic fractionation. Atmospheric water vapor even just above grass canopy was further depleted in heavy isotope by mixing with isotopically lighter vapor from upper air. The transpiration fraction T/ET (where T is the transpiration flux, and ET is the evapotranspiration flux) estimated from the observed isotopic signatures exhibited drastic increase in the first month of growing season and was more than 0.8 over the latter months. The seasonal change in transpiration fraction could be represented very well by a function of leaf area index (determination coefficient, $r^2 = 0.864$) as well as the fraction of latent heat flux to available energy. These results directly demonstrate that vegetation growth affects water and energy balance through the transpiration activity, and reemphasis that isotope tracer approach is useful for evaluating quantitatively the relationship among them.

Key words: stable isotope; tracer; evapotranspiration; atmospheric water vapor; soil-vegetation-atmosphere transfer schemes (SVATS)

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1. Introduction

Stable isotopes of hydrogen (Deuterium; D) and oxygen (18O) are useful tracers to reveal terrestrial branch (e.g., Clark and Fritz 1997; Kendall and McDonnell 1998) and atmospheric branch (e.g., Merlivat and Jouzel 1979; Smith 1992; Gat 2000; Yoshimura et al. 2003) of global water cycle. Isotopic composition of water vapor supplied from the earth's surface to the atmosphere is particularly important for tracing atmospheric water cycle. For instance, it provides essential information in identifying the origin of precipitating water (e.g., Gat and Carmi 1970; Yamanaka et al. 2002) or in evaluating precipitation recycling within a region (e.g., Salati et al. 1979; Ingraham and Taylor 1991). Unfortunately, there have been a limited number of isotopic measurements of atmospheric water vapor (e.g., White and Gedzelman 1984; Bariac et al. 1989; Walker and Brunel 1990; Brunel et al. 1992). In particular, we could find no studies on isotopes in water vapor evapotranspiring from land surface until mid-1990s. The reason is because isotopic composition of water vapor in the atmosphere even just above the land surface does not reflect that of evapotranspiring water vapor owing to vertical mixing.

Yakir and Wang (1996) first adopted the "Keeling plot" for determining the isotopic signature of evapotranspiration flux. He and Smith (1999a) examined its theoretical validity and redefined it as the "mixing line method". This approach was then applied to isotopic measurements of atmospheric water vapor above a forest (He and Smith, 1999a; Li et al., 2006), a coastal salt marsh (He et al., 2001), an agricultural grassland (He et al., 2003), a tallgrass prairie (Riley et al., 2003), a savanna woodland (Yepez et al., 2003), an orchard (Williams et al., 2004) and within a forest canopy (Harwood et. al., 1999) and experimental chambers covering bare ground and grassland (Yepez et al., 2005), making clear the variability of isotopic signature of evapotranspiration flux among different land covers and its relation to source waters.

Isotopic signature of evapotranspiration flux can also be used for evaluating relative contribution of plant transpiration T and soil evaporation E to total evapotranspiration ET (Moreira et al. 1997; Yakir and Sternberg 2000; Yepez et al. 2003). The transpiration fraction T/ET is an important parameter for understanding

evapotranspiration mechanisms and dynamics of ecosystem, and for validating soil-vegetation-atmosphere transfer schemes (SVATS) in atmospheric general circulation models (e.g., Henderson-Sellers et al. 1996) and ecosystem process models (e.g., Running and Coughlan 1988; Ito and Oikawa 2002). The transpiration fraction cannot be usually obtained from conventional measurements of turbulent transport or energy budget (Wang and Yakir 2000), so that the isotopic approach is the most promising. Previous studies, however, have evaluated the isotopic signature and transpiration fraction only within a month at most, because of complicated procedures in water vapor sampling and its isotopic determination. Therefore, there remain uncertainties about seasonal change in transpiration fraction and its relation to vegetation condition.

This paper presents a simple measurement strategy for enabling us long-term monitoring at grassland. The principal objectives are twofold: (1) to examine validity of methodology for determining isotopic signatures of evapotranspiration flux, and (2) to evaluate the seasonal change in transpiration fraction and its relation to vegetation growth.

2. Field experiment

Atmospheric water vapor sampling and micrometeorological measurements were made during April-September 2003 in the experimental field of Terrestrial Environment Research Center (TERC; 36.1° N, 140.1° E, 27.0 m elevation) of the University of Tsukuba, Japan. The annual precipitation and annual mean air temperature are approximately 1300 mm and 14°C, respectively. The field is flat and circular with diameter of 160 m, and is surrounded by pine stands and lawn (Fig. 1). Dominant species inside the field are *Solidago altissima, Miscanthus sinensis* and *Imperata cylindrica*. By mowing once a year in winter, progress of secondary succession from grassland to shrub or forest is prevented. The leaf area index (LAI) usually increases from April to July, and then keeps almost constant (Fig. 2). The course of LAI in 2003 is almost similar to that in normal year, except for large value in August. Inside the grass canopy, soil surface is covered by sparse litter composed of plants that have been mown and blighted in previous years. The top soil of approximately 2 m thickness is loamy, volcanic ash soil (*Andosol*) and is underlined by a clay layer. Water table usually exists just above the clay layer during the growing season.

A mast of 2.6-m height was installed at a northern part of the field to take longer fetch against dominant wind direction (southerly or southeasterly during the sampling period). Atmospheric water vapor was sampled at two levels of the mast for isotopic analysis. The sampling
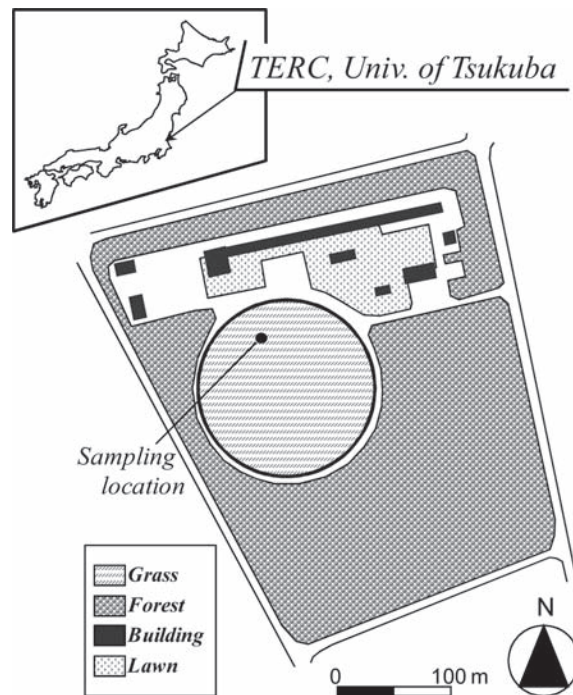
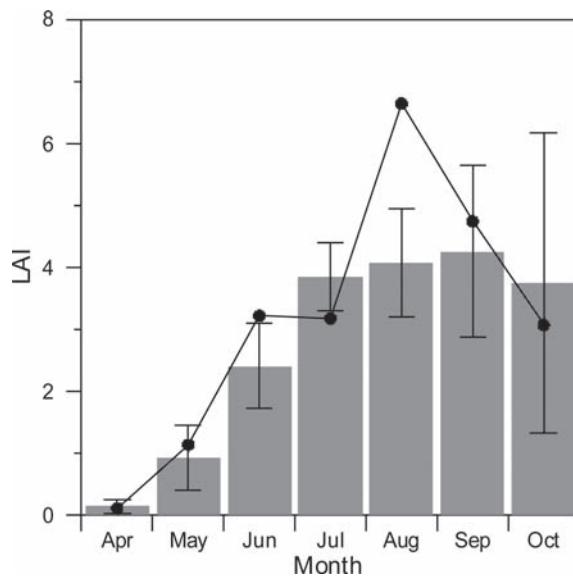

Fig. 1 Experimental site.

Fig. 2 Temporal evolution of the leaf area index (LAI) in 2003 (dots) and its long-term mean with standard deviation for a period from 1996 to 2002 (bars). Data sources: Tanaka and Oikawa (1998) for 1996 and 1997, Tanaka and Oikawa (1999) for 1998, Yokoyama and Oikawa (2000) for 1999, Yokoyama and Oikawa (2001) for 2000, Imasu et al. (2002) for 2001, Mo et al. (2003) for 2002, and Yokoyama et al. (2004) for 2003.

levels were changed with growth of vegetation (from 0.3 to 1.1 m at lower level and from 1.8 to 2.6 m at upper level; Table 1). The water vapor sampling system utilized in this study is similar to that of He et al. (2001, 2003) but

Table 1 Summary of all samplings.

SN[a]	DOY[b] (Date)	Time[c]	zl[d]	zu[e]	hg[f]	WD[g]	Rn[h]
1	114 (24 April)	15:15-16:30	30	180	10	SSE	187
2	129 (9 May)	14:45-16:35	30	180	20	ESE	238
3	142 (22 May)	9:35-11:35	40	200	30.9	ESE	472
4	142 (22 May)	14:00-16:00	40	200	30.9	E	319
5	155 (4 June)	9:10-11:15	50	210	38.1	S	184
6	155 (4 June)	14:00-16:00	50	210	38.1	S	35
7	170 (19 June)	9:00-11:00	80	240	61.2	SSE	416
8	170 (19 June)	14:00-16:10	80	240	61.2	S	297
9	184 (3 July)	9:45-11:45	90	250	64.3	SSE	284
10	184 (3 July)	13:20-15:30	90	250	64.3	ESE	129
11	201 (20 July)	9:15-11:15	110	260	69.2	E	209
12	201 (20 July)	14:10-15:20	110	260	69.2	ESE	174
13	254 (11 September)	7:40-8:40	110	260	72.5	SSE	255
14	254 (11 September)	9:40-10:40	110	260	72.5	S	300
15	254 (11 September)	11:45-12:45	110	260	72.5	SSE	455
16	254 (11 September)	13:45-14:45	110	260	72.5	SSW	325
17	254 (11 September)	15:45-16:50	110	260	72.5	S	87

[a] Sampling number   [b] Day of year   [c] Japanese Standard Time (JST)   [d] Lower sampling level, cm
[e] Upper sampling level, cm   [f] Grass height, cm   [g] Wind direction   [h] Net radiation, W/m2

modified for simplification (Fig. 3). Major modification is equipment for cryogenic trapping. We used an all-purpose Pyrex trap rather than sample canister which was filled with small aluminum balls. Therefore, surface area for condensation and thus vapor collecting efficiency may reduce more or less. However, we had confirmed by preliminary laboratory experiments (Tsunakawa and Yamanaka 2004) that almost all water vapor inflow could be trapped and isotopic composition of trapped water agreed with that of original water vapor with sufficient accuracy under flow rates less than 10 l/min. In the field sampling, we set the flow rate at approximately 3.75 l/min by a pump with flow controller (MP-2N, SHIBATA, Japan). While the amount of collected water vapor depends on absolute humidity of the atmosphere, water sample from 2 to 5 ml was obtained in each sampling. This sampling technique allows us to skip the procedure of sample transfer to another vessel through a vacuum line.

At the same two levels as in water vapor sampling, air temperature and relative humidity were measured by a ventilated thermometer (platinum resistance thermometer, PFT-01, PREDE, Japan) and hygrometer (capacitance type sensor, CHS-APS, TDK, Japan) for computing the water vapor mixing ratio Q. Net radiation R_n and soil heat flux G were also measured by net radiometer (Q7, REBS, USA) and heat flux plate (CN-81, Eiko, Japan), respectively, for determining evapotranspiration by energy balance/ Bowen ratio (EBBR) method (e.g., Brutsaert 1982, pp.210-212). Measurements of the micrometeorological components were done every 10 s (0.1 Hz) and mean values for 5 minutes were recorded using a data logger (CR10X,
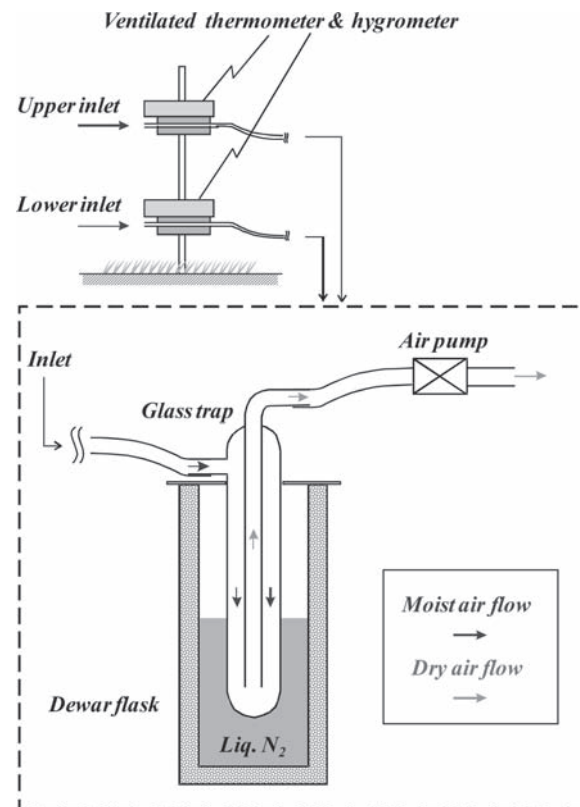

Fig. 3 Water vapor sampling system for isotopic analysis.

Campbell Sci. Inc., USA).

Soil water samples at five depths were collected on each sampling-day using suction lysimeters attached with ceramic cup on its tip (DIK-8390, DAIKI, Japan). For extracting of soil water, 100 ml syringes were utilized.

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Maximum suction level of this equipment is approximately 80 kPa. At the same depths, matric potentials of soil water were measured by pressure gauge type irrometers (DIK-3161, DAIKI, Japan). Precipitation sample was also collected every week using a precipitation collector, specially developed to avoid evaporation of stored water (Yamanaka et al. 2005).

For collected water samples, δD and $\delta^{18}O$ (‰; $\delta = (R_{sample}/R_{standard} - 1)\times10^3$; R is D/H or $^{18}O/^{16}O$ ratio and the standard is Vienna Standard Mean Ocean Water (VSMOW)) were determined by mass spectrometry with a stable isotope ratio mass spectrometer (MAT252, Thermo Finnigan) of the Hydrology Laboratory, University of Tsukuba. To prepare the sample in advance of isotopic measurement, we used the hydrogen gas equilibration using platinum catalyst (6 hours) for δD and the carbon dioxide gas equilibration (9 hours) for $\delta^{18}O$. This automatic equilibration system allows isotopic measurements of 20 samples (1 ml for each sample) for both hydrogen and oxygen within two days, and its measurement accuracy is ±1 ‰ for δD and ±0.1 ‰ for $\delta^{18}O$.

3. Analytical approach

3.1. Determination of isotopic signature of evapotranspiration flux

Isotope ratio of evapotranspiring water vapor or evapotranspiration flux (R_{ET}) is equivalent to the ratio of the flux of isotopic water (HDO or $H_2^{18}O$) to that of normal water (H_2O), thus (He and Smith, 1999a),

$$R_{ET} = j^{-1}\frac{F^i}{F} \tag{1}$$

where $j=19/18$ (20/18) is the ratio of the molecular weights of HDO/H2O ($H_2^{18}O/H_2O$), F and F^i are the evapotranspiration fluxes of normal water and isotopic water, respectively (hereafter superscript i denotes the quantity for the heavy isotope). The water vapor flux in the atmospheric boundary layer (ABL) can be expressed by flux-gradient relation (e.g., Garratt 1992, p.28)

$$F = -K\frac{d\overline{Q}}{dz} \tag{2}$$

where K is the eddy diffusivity, Q is the mixing ratio, z is the height, and overbar denotes the temporal mean value in certain period. In the same manner isotope flux is expressed as,

$$F^i = -K^i\frac{d\overline{Q^i}}{dz}. \tag{3}$$

Both normal and isotopic water vapors are passively transported by air turbulence within the ABL. Thus, we can assume $K = K^i$, and then isotope ratio of evapotranspiration flux is given (just like as Bowen ratio) by

$$R_{ET} = j^{-1}\frac{d\overline{Q^i}}{d\overline{Q}}. \tag{4}$$

Because the slope of a line in the Q^i - Q diagram is mathematically identical to the intercept of the corresponding line in the δ - $1/Q$ diagram (Appendix A&B of He and Smith 1999a), isotopic composition of evapotranspiring water vapor can be expressed in delta notation as,

$$\delta_{ET} = \frac{Q_l\delta_{vl} - Q_u\delta_{vu}}{Q_l - Q_u} \tag{5}$$

where δ_v is the isotopic composition of atmospheric water vapor, and subscripts u and l represent upper and lower sampling levels, respectively.

The δ - $1/Q$ diagram corresponds to so-called Keeling plot, which is derived from isotopic mass balance assuming an isotopically homogeneous background-atmosphere (Yakir and Wang 1996; Wang and Yakir 2000). The Keeling plot had been first suggested by Keeling (1958, 1961) to identify the sources that contributed to increases in atmospheric CO_2 in a forest canopy, and then have been utilized in numerous studies for carbon cycle (see for a review, Pataki et al. 2003). As compared to a derivation based on isotopic mass balance, the derivation from the flux-gradient relation given above can show more clearly limitations of this approach in application, for instance, on fetch requirement. Because the assumptions (i.e., flux-gradient relation and $K = K^i$) in deriving Eq. (5) are same as EBBR method, same order of fetch length (e.g., 100-300 times greater than measurement/sampling height; Brutsaert 1982, p.166) is required.

Here, we re-examines the fetch problem by another way. Considering an air column with an area of the base of 1 m², a height of 0.3 m, a density of 1.2 kg/m³ and a mixing ratio of 0.016 kg/kg, the initial content of water vapor within the column is computed to be 5.8 g. If the column moves laterally at a speed of 1 m/s over a distance of 160 m across the experimental field, in which latent

heat flux is 200 W/m², then water vapor supplied by evapotranspiration to the column is 12.8 g, being double initial vapor content. Thus, the air just above the canopy contains much enough water vapor evapotranspiring from the field. Even if isotopic composition of sampled water vapor at a height was affected by upper air isotopic composition through vertical mixing, its vertical gradient should reflect isotopic signature of evapotranspiration flux.

3.2. Evaluation of transpiration fraction

For grassland, evapotranspiration flux can be partitioned into two components: transpiration and soil evaporation,

$$ET = T + E. \tag{6}$$

As well as for normal water flux, isotope flux can be also distinguished between transpiration component and soil evaporation component,

$$ET \cdot \delta_{ET} = T \cdot \delta_T + E \cdot \delta_E \tag{7}$$

where δ_T and δ_E are the isotopic compositions of transpiration flux and soil evaporation flux, respectively.

Solving Eqs. (6) and (7) simultaneously, the following equation is given

$$\frac{T}{ET} = \frac{\delta_{ET} - \delta_E}{\delta_T - \delta_E}. \tag{8}$$

Consequently, transpiration fraction *T/ET* can be evaluated if we obtain the values of δ_T and δ_E in addition to δ_{ET}.

No isotopic fractionation occurs during water uptake and transport to the leaves inside the plant body (e.g., Ehleringer and Dawson 1992). Although leaf water becomes isotopically enriched due to evaporation, isotopic composition of transpiring water vapor from leaves is the same as that of soil water absorbed by root system and transported into the leaves under isotopic steady-state condition (White 1989, p.148; Flanagan et al. 1991). Therefore, δ_T can be given by isotopic composition of soil water in the active root zone. In resent years, several studies pose a question about the steady-state assumption (Farquhar and Cernusak, 2005; Yepez et al., 2005; Lai et al., 2006). Although the non-steady-state effect on δ_T is expected to depend on relative magnitudes of transpiration rate and leaf water pools that are affected by isotopic fractionation, the capacity of the pools has been still unknown. Therefore, we cannot assess non-steady state effect exactly at present, though it may affect δ_T and *T/ET*

estimations more or less.

For extremely dry soils, evaporation and thus water isotopic enrichment occur not at the soil surface, but at the bottom boundary of the dry surface layer (DSL) (Allison and Barnes 1983; Yamanaka and Yonetani 1999), and kinetic fractionation is strongly enhanced by molecular diffusion of water vapor through the DSL (Barnes and Allison 1983). However, soils covered by grass canopy are hardly ever put in such a situation. Therefore, we can utilize the Craig-Gordon model (Craig and Gordon 1965) or any alternatives (e.g., Flanagan et al. 1991; He and Smith 1999b) for estimating δ_E.

An arranged form of Craig-Gordon model is expressed as (Merlivat and Jouzel 1979: He and Smith 1999b),

$$\delta_E = \left\{ \kappa \frac{(1/\alpha_{eq})(10^{-3}\delta_s + 1) - h^*(10^{-3}\delta_v + 1)}{1 - h^*} - 1 \right\} \times 10^3 \tag{9}$$

where κ (= 0.9723 for oxygen or 0.9755 for hydrogen; Gat 1996) is the ratio of molecular diffusivities of isotopic water to normal water in the air, α_{eq} is the equilibrium fractionation factor, δ_s is the isotopic composition of surface soil water, δ_v is the isotopic composition of atmospheric water vapor, and h^* is the relative humidity defined as the ratio of the actual water vapor pressure *e* of the air to the saturation vapor pressure e_{sat} at the surface soil temperature. This model incorporates not only effects of equilibrium and kinetic fractionations but also an effect of isotope back flux (He and Smith 1999b). In the present analysis, kinetic fractionation effect is represented by κ. Although there are other alternatives including semi-empirical parameter for kinetic effect, we adopt theoretically simplest one. Besides, several assumptions in applying Eq. (9) will be discussed in 4.2.

4. Results and discussion
4.1. Isotopic signatures

In general, temporal change in isotopic composition of soil water largely depends on that of precipitation, and its amplitude decreases with depth. Such a tendency is observed also in this study (Figs. 4a and 4b). The δ value at 80 cm depth shows hardly any changes. In contrast, the values for the shallower depths temporally vary with isotopic input of precipitation. In September (day 254), considerably low value due to isotopically light rains was observed down to 20 cm depth, although soil water at 5 cm and 10 cm depths could not be extracted by suction lysimeters because of soil dryness. Although the δ of soil water at 20 cm depth was lower than that of precipitation, this would be due to overestimation of precipitation δ. A period from day 185 to 230, a portion of precipitation

was lost since rain collector became full, so that exact δ in precipitation for this period is expected to be lower than the measured value.

Temporal variations of δD of atmospheric water vapor at two levels are similar, but the value for the upper level is always smaller (i.e., isotopically lighter) than that for the lower level (Fig. 4c). The difference in δ_v values between morning and afternoon was not so large (less than 3‰) except for days 170 (approximately 30‰) and 201 (more than 7‰), showing that diurnal isotopic variability of atmospheric water vapor is usually small. Daily or seasonal variability of δ_v is greater than its diurnal variability and seems to be independent of that for soil water. The δ_v generally increases from spring to summer, while extremely low values were observed on day 170 (19 June). According to Yamanaka et al. (2001), isotopically lighter vapors, which are affected by heavy rainfall upwind, are sometimes brought into central Japan in a rainy period from June to early July. The exceptional low values may reflect such a vapor from upper air. As stated above, a drastic change in δ_v more than 30 ‰ was

observed only on this date. The flux-gradient relation is expected not to be valid under such conditions during the sampling duration, so that δ_{ET} for day 170 is eliminated from the latter analysis.

Figure 5 shows ranges of isotopic variation for each of precipitation, atmospheric water vapors, soil waters and evapotranspiration flux. The figure also includes data for hypothetical water vapor in isotopic equilibrium with precipitating water (hereafter abbreviated as VEP), calculated using equilibrium fractionation factors at 10 °C. We may take this value as a proxy for the upper air or free atmosphere. First we note that the average isotopic compositions for soil water are similar to the precipitation. In contrast, values for evapotranspiration flux are clearly

a) Precipitation

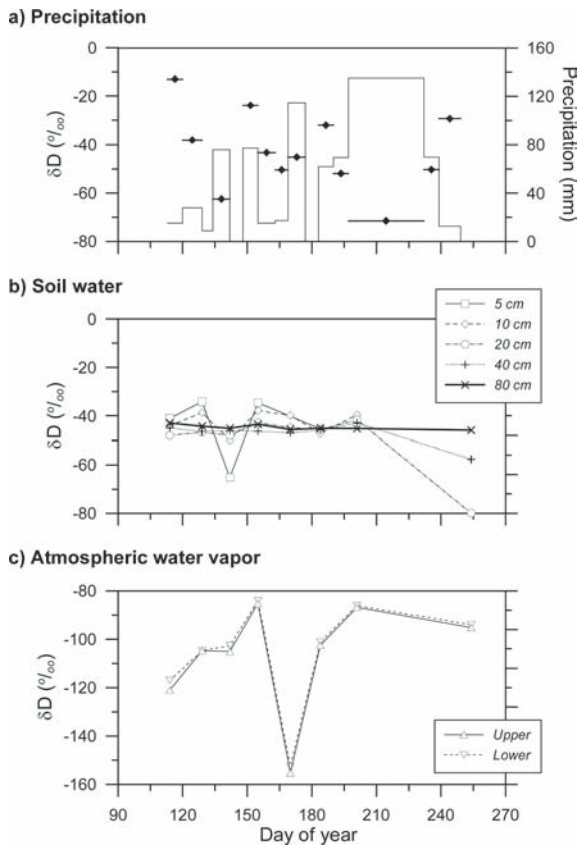b) Soil water

c) Atmospheric water vapor

Fig. 4 Temporal changes in (a) precipitation amount (step-like line) and isotopic composition of precipitation (diamond with bar), (b) isotopic composition of soil water at different depths, and (c) isotopic composition of atmospheric water vapor at two levels (daily averaged value).
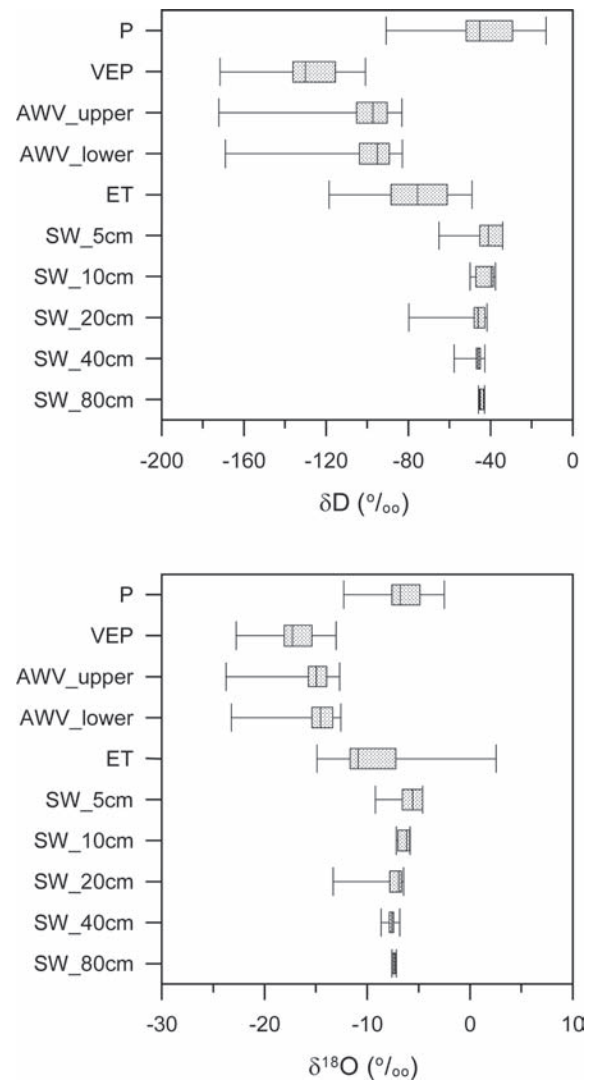

Fig. 5 Box-Whisker plots, which show the minimum, maximum, median, lower quartile, and upper quartile, for (top) δD and (bottom) $\delta^{18}O$ of precipitation (P), water vapor in isotopic equilibrium with precipitation (VEP), atmospheric water vapors (AWV), evapotranspiration flux (ET) and soil waters (SW).

lower than those for soil water indicating occurrence of isotopic fractionation. The data of atmospheric water vapor lie between those of evapotranspiration and VEP. This fact confirms an idea that the water vapor within the ABL is a mixture of relatively lighter vapor from upper air and heavier vapor generated by evapotranspiration from land surface. This idea is also supported by another fact that the δv for the upper level is always smaller than that for the lower level (Fig. 4c). Generally speaking, these results indicate that Eq. (5) successfully determines δ_{ET} with proper consideration of vertical mixing of the atmosphere.

Figure 6 shows estimated δET in each sampling. The variations of δ_{ET} for hydrogen and oxygen isotopes are not always similar. This dissimilarity may be partly due to kinetic fractionation during soil evaporation, which produces isotopically light vapor more effectively for oxygen than for hydrogen. However, estimated δ_{ET} also shows somewhat erratic variation including a positive δ value and considerably large diurnal changes. Many previous studies (Moreira et al., 1997; Harwood et al., 1999; Yepez et al., 2003; Li et al., 2006) have assumed constancy of δ_{ET} over a day. Williams et al. (2004) also showed that diurnal change in δET had same magnitude with that of estimation error. If we assume that diurnal change in δ_{ET} is fully due to an estimation error, its magnitude in the present study is estimated to reach 40‰ in δD and 7‰ in $\delta^{18}O$ at maximum; otherwise standard deviation of 7‰ in δD and 2.0‰ in $\delta^{18}O$ for five times sampling on day 254 would be a measure of the accuracy for δET. The errors in δ_{ET} may become serious in case of low humidity (i.e., large $1/Q$) or small gradient in δ_v and Q between the two sampling levels, because the errors are amplified by extrapolation of mixing line even if errors in δ_v and Q at each level are sufficiently small. The errors, however, tend to be random, so that they would be reduced if averaging several measurements. Otherwise, it is recommended to apply (simultaneous) multi-level measurement for constructing Keeling plot as in Yakir and

Wang (1996).

4.2. Transpiration fraction

The Craig-Gordon model (Eq. (9)) has some difficulties in practice. A matter of argument is how to set the values of δ_v and e. According to the definition, the values at the boundary between diffusive surface layer and turbulent mixing layer (He and Smith 1999b) should be appropriate, though it is not easy to measure them exactly at the boundary. In this study, δv and e used in Eq. (9) are taken from the lower sampling level, same as in He et al. (2001). This approximation would be minor problem because of small gradients of δ_v and e in the turbulent mixing layer. Another difficulty in this study is lack of isotopic measurement of surface soil in relatively dry conditions, due to limitation of suction lysimeters to extract soil water. Vertical gradient of isotopic composition of soil water is very small in this study except for day 254. On that day relatively large gradient was observed, while it was caused by precipitation input, not by soil evaporation. Considering the isotopic variation of precipitation (Fig. 4), isotopic composition of surface soil is expected not to be different very much from that at 20 cm depth. Therefore, in this study, the δ value at the shallowest depth where soil water could be extracted was adopted for δ_S for each day. The same value was also adopted for δ_T. This choice should be reasonable because herbaceous plants are expected to uptake soil water from shallow depth-zone in which water can be most easily extracted.

As noted in the preceding section, individual estimates of δ_{ET} can have considerable errors, so we compute ensemble mean of transpiration fraction (including those estimated both from hydrogen and oxygen isotopes) on each sampling day to reduce and quantify the error in T/ET evaluation. The mean T/ET ($\pm$s.d.) is shown in Fig. 7a as a function of date and in Fig. 7b as a function of LAI. The LAI value for each sampling day was estimated by interpolation of data of Yokoyama et al. (2004) using a spline function. The figures show that temporal evolution of T/ET is strongly controlled by vegetation growth. The relationship between T/ET and LAI was approximated very well by a power function ($T/ET = 0.891 LAI^{0.077}$, $r^2 = 0.864$). Unexpectedly, the T/ET is not so small (i.e., near 0.75) even in the earliest stage of growing season. This may be because the litter on the ground suppressed soil evaporation just same as straw mulch. Therefore, this feature is likely to be somewhat specific to the present experimental site.

Standard deviation of T/ET, which is shown as vertical bar in Fig. 7, ranges approximately from 0.05 to 0.1. These values indicate magnitude of the error in estimating T/ET and are relatively small in spite of large error in
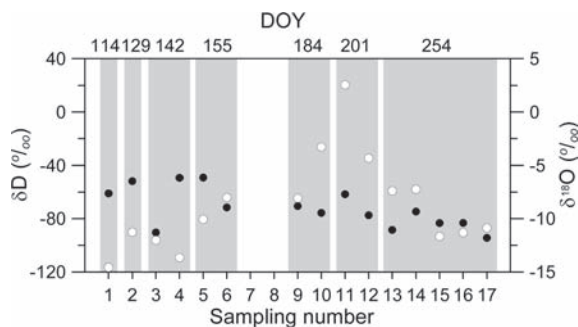

Fig. 6 Hydrogen (solid circle) and oxygen (open circle) isotopic composition of evapotranspiration flux for each sampling (except for values on DOY 170; see text).

δET estimation. As seen from Eq. (8), if the difference between δT and δE is large, then T/ET is less sensitive to δ_{ET}. In other words, when T/ET was far smaller than unity, the difference between δ_{ET} and δ_T is expected to be far greater than the estimation error in δ_{ET}. Thus, the isotopic approach adopted in this study is useful for evaluating large, seasonal change in T/ET, although improving of accuracy in δ_{ET} estimation is required for detecting and discussing small changes in T/ET.

In Fig. 7, the fraction of latent heat flux (lET; l is the latent heat for vaporization) to available energy (R_n - G) is also shown; it is conventionally called evaporative fraction, EF, (e.g., Sugita and Brutsaert 1991) but should be called "evapotranspirative fraction" in the present study to clearly distinguish evaporation and evapotranspiration. We can find a similarity between transpiration fraction and evapotranspirative fraction in response to vegetation growth. Indeed, good correlation ($r = 0.85$) between the two fractions was obtained. These facts suggest that partitioning of the available energy into the latent heat and sensible heat fluxes is controlled by vegetation conditions primarily through transpiration activity.

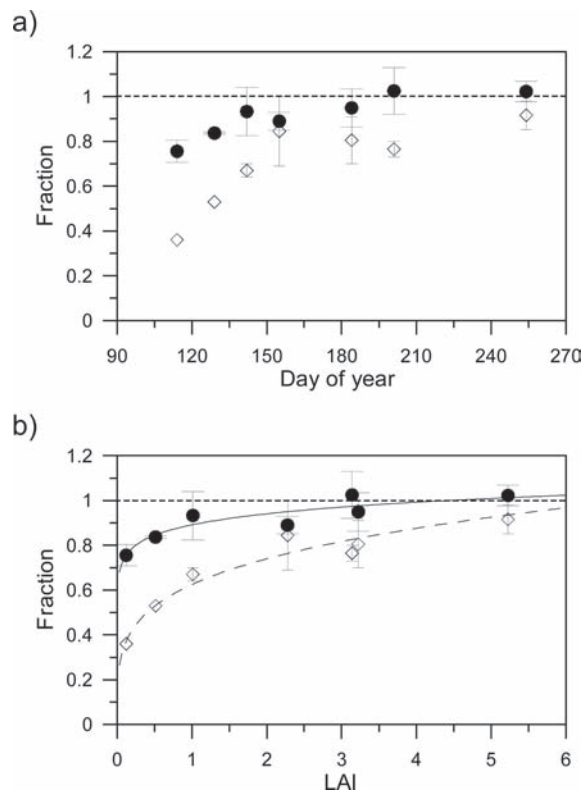It should be noted that in spite of high transpiration

Fig. 7 Transpiration fraction and evapotranspirative fraction as functions of (a) date and (b) LAI. Vertical bars denote standard deviation for each date. Regression lines are expressed as $T/ET = 0.891LAI^{0.077}$ ($r^2 = 0.864$) for transpiration fraction (solid line), and $EF = 0.625LAI^{0.244}$ ($r^2 = 0.964$) for evapotranspirative fraction (dashed line).

fraction throughout the growing season, the δ value of the evapotranspiration flux is lower than that of soil water (see Fig. 5) and thus transpiration flux. This can be explained by very low δ_E ($< -200‰$), that is, the effect of isotopic fractionation on δ_{ET} cannot be always negligible even if soil evaporation is a minor component. In other words, the computed transpiration fraction is sensitive to δ_E rather than determination of δ_{ET}. Thus, for reducing uncertainty in evaluating transpiration fraction, it is necessary to reexamine Craig-Gordon model and assumptions in its application under field situation.

5. Conclusion

Isotopic composition of evapotranspiration flux was close to, but somewhat depleted in heavy isotope than, that of source water (i.e., shallow soil water or antecedent rainfall), suggesting a little contribution of soil evaporation accompanied by isotopic fractionation. However, δ value of atmospheric water vapor even just above grass canopy was further lower than that of evapotranspiration flux by mixing with isotopically lighter vapor from upper air. All these results seem to be quite reasonable, while the determination of δ_{ET} by Keeling plot approach included non-negligible, potential errors up to 40 ‰ in δD and 7 ‰ in δ^{18}O at maximum.

In spite of non-negligible errors in δ_{ET} determination, transpiration fraction could be estimated with relatively small error of ±0.05 to 0.1, because this parameter is not sensitive to the accuracy of δ_{ET}. Seasonal change in the estimated transpiration fraction exhibited a close relationship with vegetation growth and could be represented by a function of LAI. The transpiration fraction was also linked very well to surface energy balance at the grassland surface. These facts explicitly demonstrate that change in vegetation condition can affect water and energy balance through the change in transpiration activity. Such a transpiration control in grassland ecosystem composed of various species and sizes of plants cannot be easily evaluated by any physical measurements (e.g., sap flow measurements) because of spatially and functionally heterogeneous vegetation (Schaeffer et al., 2000). Therefore, isotopic approach that can provide spatially-averaged transpiration fraction is very useful. Further studies for various types of vegetation under different climate must improve our understanding of the relationship between vegetation and climate.

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