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Partitioning oak woodland evapotranspiration in the rocky mountainous area of North China was disturbed by foreign vapor, as estimated based on non-steady-state ^{18}O isotopic composition

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ABSTRACT

In terrestrial ecosystems, partitioning ecosystem-scale evapotranspiration (ET) between plant transpiration and soil evaporation remains a technical challenge. In this paper, we used a newly-developed laser-based isotope analyzer (OA-ICOS) and the Keeling plot approach to partition ET components of a *Quercus variabilis* plantation in a lithoid hilly area of north China. The results showed that, on day-of-year (DOY) 254, 257, and 263, the modeled leaf water ^{18}O composition ($\delta^{s}_{L,b}$) and observed leaf water ^{18}O composition ($\delta_{L,b}$) were in substantial agreement and had a significant linear correlation with coefficient of 0.96, indicating that Keeling plot approach and Graig–Gorden model can be used in portioning ET between plant transpiration and soil evaporation in a terrestrial forest ecosystem. Isotopic partition revealed that the percent contribution of transpiration to total ET increased from the morning, reached maximum values at noon, with maximum values of 91.19%, 86.30%, and 85.37% for DOY 254, 257, and 263, respectively, indicating the transpiration from *Q. variabilis* Blum contributed the most to the total ET in this forest ecosystem. On DOY 260, the stability stratification was unstable, which resulted from the foreign vapor concentration. The increased vapor concentration led to a 80.83% difference between $\delta^{s}_{L,b}$ and $\delta_{L,b}$. The correlation coefficients between $\delta^{s}_{L,b}$ and $\delta_{L,b}$ decreased from 0.96 to 0.43 when dataset on DOY 260 was included, indicating foreign vapor increased the uncertainty in the estimation of δ^{18}O of ET (δ_{ET}) and δ^{18}O of transpiration (δ_{T}) in the forest ecosystem. Path analysis results suggested that water vapor concentration was the major factor influencing the partitioning of ET with isotopic approach in the forest ecosystem. Since the increased water vapor concentration and decreased atmosphere δ^{18}O disturbed the estimation of δ_{E}, δ_{T}, δ_{ET}, the isotopic approach cannot be used in partitioning ET under this condition on DOY 260. Therefore, under most circumstances the isotopic approach can be used to partition ET for forest ecosystem in a non-steady state (NSS), while water vapor concentration may cause uncertainties.

1. Introduction

Evapotranspiration represents the total amount of water transferred to the atmosphere by transpiration from plants and evaporation from soil and other wet surfaces in a terrestrial ecosystem. Partition of these two components may contribute to understanding of water vapor exchange in the soil-plant-atmosphere continuum. Both evaporation and transpiration involve water phase transition. As a result, the water vapor contains less heavy isotopes compared with the soil water. Similarly, transpiration leads to enrichment of heavier isotopes in leaf water (Farquhar and Gan, 2003; Welp et al., 2008). Under strong transpiration or transpiration in an isotopic steady state (ISS) conditions, plant water vapor can become similar to xylem water in isotopic composition (Moreira et al., 1997; Harwood et al., 1998). Xylem water is a mixture of soil water absorbed by the root system from different depths (Flanagan et al., 1991; Wang and Yakir, 2000). Therefore, water vapor fluxes produced by soil evaporation (highly fractionated) and plant transpiration (less fractionated) have different H and O isotope compositions. This difference provides the basis for partitioning evaporation and transpiration components within an ecosystem (Yepez et al., 2003; Williams et al., 2004). This stable isotope-based technique has been used to quantitatively partition evaporation and transpiration components for farmland (Brunel et al., 1992; Yuan et al., 2010), shrub land (Xu et al., 2008), grassland (Yepez et al., 2005), and forest (Moreira et al., 1997). But these studies assumed that plant transpiration takes place in an isotopic steady state ($\delta_T = \delta_x$). Under field conditions, however,

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Table 1
Symbols used in the text.

ISS	Isotopic steady state
NSS	Non steady state
ET	Evapotranspiration (mol m^{-2} s^{-1})
δ_{ET}	$\delta^{18}O$ of evapotranspiration (‰)
δ_E	$\delta^{18}O$ of soil evaporation (‰)
δ_T	$\delta^{18}O$ of transpiration (‰)
$\delta_{L,e}$	$\delta^{18}O$ of leaf water at the site of evaporation calculated assuming isotopic non state conditions (‰)
$\delta_{L,b}$	$\delta^{18}O$ of bulk leaf water (‰)
$\delta^{s}_{L,b}$	$\delta^{18}O$ of leaf water at the site of evaporation calculated assuming isotopic steady state conditions (‰)
δ_s	$\delta^{18}O$ of soil water (‰)
δ_x	$\delta^{18}O$ of xylem water (‰)
δ_v	$\delta^{18}O$ of atmospheric water vapor (‰)
$\delta_{v,s}$	$\delta^{18}O$ of atmospheric water vapor at 0.1 m above ground (‰)
$\delta_{v,c}$	$\delta^{18}O$ of atmospheric water vapor at 11 m above ground (‰)
ε_{eq}	Equilibrium fractionation (‰)
ε_k	Kinetic fractionation (‰)
α^+	A temperature-dependent equilibrium fractionation coefficient
g	Leaf conductance (mol m^{-2} s^{-1})
W	The water concentration of the leaf (mol m^{-2})
w_i	The mole fraction of water vapor in leaf intercellular air spaces (mol mol^{-1})
ζ	Stability stratification

plant transpiration is affected by environmental factors often in a NSS state ($\delta_T \neq \delta_x$). Several studies have investigated the enrichment of heavy isotopes in leaf water under NSS conditions (Kahmen et al., 2008; Griffis et al., 2011b), these researches, however, focused on crops with flat underlying surface, and little research on forest with irregular surface has been reported. Forests play a major role within global water cycling. However, quantitative partitioning of water fluxes for forest ecosystems under NSS conditions has scarcely been reported because of more uncertainties compared with crop ecosystems, and various experimental difficulties, such as field measurement and the height of trees (Yepez et al., 2007).

Traditionally, atmospheric water vapor samples have been collected with cold traps (Helliker et al., 2002; Wang and Yakir, 2000). This technique can only collect samples transiently or discretely, and thus has poor time resolution (Lee et al., 2005). Additionally, cold trapping might involve incomplete cooling and thus create artificial isotopic enrichment. A new method called Off-Axis Integrated Cavity Output Spectroscopy (OA-ICOS) has been developed to continuously measure the heavy isotopic composition of oxygen ($\delta^{18}O$) and hydrogen ($\delta^{2}H$) which requires neither reference gas nor cryogenic cooling (Wang et al., 2009). In the present study, we analyzed water vapor isotope compositions in a *Quercus variabilis* Blume ecosystem in a stony hill region of north China with OA-ICOS. Our objectives were to: (1) calculate and validate plant transpiration δ_T under NSS conditions; (2) quantify the proportion of T/ET with the isotope model (Yakir and Sternberg, 2000, also see a description in the following Section 2); and (3) investigate the influence of foreign vapor on partitioning ET. Our goal was to provide baseline information for better understanding of water vapor exchange in such stony hill ecosystems in north China (Table 1).

2. Theory

Under natural conditions, water consists of isotopes of hydrogen (H, D) and oxygen (^{16}O, ^{18}O). The isotopic composition of water is characterized by its $H_2{}^{18}O/H_2{}^{16}O$ and $HD^{16}O/H_2{}^{16}O$ molar ratios. The ratios (δ, in ‰) is V-SMOW standard for water, expressed as

$$\delta^{18}O = \left(\frac{R_{sample}}{R_{standard}} - 1\right) \times 1000, \quad (1)$$

where R_{sample} and $R_{standard}$ are $^{18}O/^{16}O$ ratios of the sample and standard.

During soil water evaporation, soil water vapor has a reduced $\delta^{18}O$. The δ_E, referred to the soil water vapor $\delta^{18}O$, can be determined using the Graig–Gordon equation (Yakir and Sternberg, 2000):

$$\delta_E = \frac{\delta_s/\alpha^+ - h\delta_{v,s} - \varepsilon_{eq} - (1-h)\varepsilon_k}{(1-h) + (1-h)\varepsilon_k/1000}, \quad (2)$$

where δ_s is the isotopic composition of water in the evaporating surface, $\delta_{v,s}$ is the isotopic composition of atmospheric water vapor at 0.1 m above ground, h is atmospheric relative humidity, α^+ (>1) is a temperature-dependent equilibrium fractionation coefficient, $\varepsilon_{eq} = (1-1/\alpha^+) \times 1000$ represents the equilibrium fractionation effect, and ε_k is the kinetic fractionation coefficient. The ε_k for oxygen (H_2O: $H_2{}^{18}O$) is 1.0164 (Cappa et al., 2003). The α^+ can be calculated from Eq. (3) (Majoube, 1971):

$$^{18}O\alpha^+ = \frac{1.137(10^6/T^2) - 0.4516(10^3/T) - 2.0667}{1000} + 1, \quad (3)$$

where T is the soil temperature in Kelvin at 5 cm below the surface.

In the previous studies of quantitative partition of water vapor, leaf water vapor is usually assumed to have the same isotopic composition as stem water (ISS) (Yepez et al., 2003; Xu et al., 2008; Yuan et al., 2010). In the field, ISS consumption is frequently not satisfied, and plant transpiration actually occurs usually under NSS conditions. When water vapor diffuses from leaves to atmosphere, $H_2{}^{18}O$ enriched at the evaporation sites of stomata, showing an increased $\delta^{18}O$ at the sites of evaporative enrichment in leaves ($\delta_{L,e}$). Although $\delta_{L,e}$ itself is not directly measurable, the $\delta^{18}O$ of bulk leaf water ($\delta_{L,b}$) can be determined. The $\delta_{L,e}$ distribution in a leaf is heterogeneous, and $H_2{}^{18}O$ enriched at the evaporation sites is influenced by Péclet effect (Welp et al., 2008; Lai et al., 2006; Ripullone et al., 2008) to result in $\delta_{L,e} > \delta_{L,b}$. Therefore, for accurate determination of $\delta_{L,e}$, the Péclet effect and NSS must be considered. Under ISS conditions, isotope fractionation that occurs during leaf evaporation can be described by $\Delta_{L,s}$ (Welp et al., 2008) as

$$\Delta_{L,s} = \varepsilon^+ + \varepsilon_k + (\Delta_v - \varepsilon_k)h, \quad (4)$$

where h is relative humidity referenced to leaf temperature, ε_k is the kinetic fractionation coefficient, and ε^+ is the temperature-dependent equilibrium fractionation coefficient, given by

$$\varepsilon^+(‰) = 2.644 - 3.206\left(\frac{10^3}{T}\right) + 1.534\left(\frac{10^6}{T^2}\right), \quad (5)$$

where T is the leaf temperature (in K). Considering NSS conditions and the Péclet effect, $\Delta_{L,e}$ can be calculated using the NSS model developed by Farquhar and Cernusak (2005) as

$$\Delta_{L,e} = \Delta_{L,s} - \frac{\alpha_k\alpha^+}{gw_i}\frac{d(W(1-e^p)/(p)\Delta_{L,e})}{dt}, \quad (6)$$

$\delta_{L,e}$ can be calculated based on the measured value of $\delta_{L,b}$, using Eq. (7):

$$\Delta_{L,e} = \Delta_{L,s} - \frac{\alpha_k\alpha^+}{gw_i}\frac{d(W\Delta_{L,b})}{dt}, \quad (7)$$

where g is leaf conductance (mol m^{-2} s^{-1}), W is the water concentration of the leaf (mol m^{-2}), and w_i is the mole fraction of water vapor in leaf intercellular air spaces (mol mol^{-1}), t is time (s), and $\alpha_k = 1 + \varepsilon_k$, $\alpha^+ = 1 + \varepsilon^+$ (Barbour, 2007). The calculated $\delta_{L,e}$ is further