GLOBAL TRANSPIRATION FRACTION DERIVED FROM WATER ISOTOPOLOGUE DATASETS

Samuel Jonson Sutanto

[1)] Experimental Station for Sabo, Research Center for Water Resources, Ministry of Public Works and Housing Indonesia, Jl. Sabo 1, Sopalan Maguwoharjo, Depok Sleman, Phone: +62-274-886350, Fax: +62-274-885431, Yogyakarta 55282, Indonesia.

Corresponding email: sjsutanto@gmail.com

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ABSTRAK

Transpirasi sebagai salah satu komponen dari evapo-transpirasi memberikan kontribusi aliran masa yang besar dari permukaan tanah. Banyak metode telah di aplikasikan untuk menghitung besarnya komponen-komponen evapo-transpirasi ini. Sebagian besar kajian dilakukan pada skala lokal maupun regional dan hanya ada sedikit kajian yang menghitung besarnya transpirasi secara global. Transpirasi global yang dihitung baik dengan model maupun dengan teknik isotop menghasilkan nilai transpirasi yang rendah untuk model dan tinggi untuk teknik isotop dibandingkan dengan pengukuran. Hal ini mengindikasikan bahwa berapa nilai yang akurat dari transpirasi global masih merupakan tantangan. Oleh karena itu, kajian ini bertujuan untuk menghitung besarnya transpirasi global dan penguapan dari tanah dengan menggunakan teknik isotop yang datanya diambil dari hasil model GCM yang dilengkapi dengan modul isotop. Hasil perhitungan dengan teknik isotop nampak menjanjikan dengan nilai fraksi transpirasi sebesar 80% pada daerah yang ditumbuhi tanaman, and lebih rendah dari 50% pada daerah gurun. Fraksi transpirasi global dari hasil perhitungan adalah sebesar 69% dengan 43% air hujan yang jatuh menguap kembali ke atmosfer melalui transpirasi tanaman. Meskipun metode keseimbangan masa isotop dapat diaplikasikan ke TES satelit data, fraksi transpirasi yang dihasilkan masih menunjukkan adanya masalah pada resolusi data yang dihasilkan. Dengan adanya peningkatan resolusi satelit pada masa yang akan datang, perhitungan komponen-komponen evapo-transpirasi dengan menggunakan data isotop akan lebih akurat.

***Katakunci**: Transpirasi, isotop, model GCM, keterbatasan metoda, prospek*

ABSTRACT

Transpiration as one of the evapo-transpiration components contributes the large quantity of flux from the continental surface. Many approaches have been carried out to quantify different components of evapo-transpiration. Most of these studies were carried out on local or regional scale and only few study estimates transpiration fraction in global perspective. The global transpiration fraction calculated using modeling and isotope-based approach results relatively low and high transpiration fraction, respectively. This indicates that the accurate value of global transpiration fraction still remains a challenge. The goal of the present study is, therefore, to provide a fully global distributed transpiration and soil evaporation fraction from the isotope-based method using the outputs from Isotope-enabled GCM. The result of global distributed transpiration fraction seems promising, with high transpiration fraction of 80% over vegetated regions and lower than 50% over deserted area. The transpiration fraction globally is approximately 69%; with 43% water from the precipitation is transported back to the atmosphere via transpiration. Although the isotope mass balance method can be applied to the satellite datasets, the transpiration fraction derived from satellite water isotopologue measurement still faces difficulties in instrument resolution. With improvements of satellite resolution in the future, the robust calculation of evapo-transpiration components using water isotopologue datasets will become visible.

***Keywords:** Transpiration, isotope, GCM model, limitations, future outlook*

INTRODUCTION

An important process controlling energy and mass exchange between the terrestrial ecosystem and the atmosphere is evapo-transpiration (Seneviratne et al., 2010). Plant transpiration as one of the main components in evapo-transpiration contributes the large quantity of flux from the continental surface. Many approaches have been carried out to quantify the transpiration flux from the total evapo-transpiration or from direct measurements. However, the accuracy of each method still remains a challenge since continental evapo-transpiration is one of the most uncertain fluxes in the global water balance (Dolman and de Jeu, 2010; Miralles, et al., 2011).

The earliest attempt to quantify direct transpiration was conducted by Huber (1932) using heat to sense the sap movement in the stem xylem as a tracer. The idea was developed further by Huber and Schmidt (1937) and Dixon (1937). The used of heat to determine continuous tree water use was brought to more widespread use by Granier (1985). Since then, there are three well-known sapflow methods using heat transfer in the stem as the main component, which are Heat Field Deformation (HFD), Heat Pulse Velocity (HPV), and Thermal Dissipation (TD) (Steppe et al., 2010). A combination of some techniques, such as Bowen ratio, Eddy covariance and lysimeter, with the sapflow method is commonly used to partition the evapo-transpiration flux into soil evaporation and transpiration (Herbst et al., 1996; Roupsard et al., 2006; Mitchell et al., 2009; Cavanaugh et al., 2011).

Stable isotopes have been used for the first time by Calder et al. (1986, 1992) to measure the transpiration rates. They measured the movement of deuterated water in stem as a tracer. Just in the beginning of 90s, the use of stable isotopes to separate the evapo-transpiration flux into soil evaporation and transpiration have become more extensive. The distinct fractionation process between light and heavy stable isotopologues makes the separation analysis possible. The lighter isotopologues will evaporate first from bare ground or open water and leave the heavy isotopologues. On the other hand, the transport of water from roots and stems during plant water uptake does not modify the isotopic composition of source of water (Ehleringer and Dawson , 1992; Clark and Fritz, 1997; Kendall and McDonnell , 1998; Mook, 2000). This robust isotopic difference between evaporation and transpiration processes makes the water isotope technique an interesting tool to quantify the soil evaporation and transpiration fluxes from the total evapo-transpiration.

Most of the partitioning studies using both the hydrometric measurements and the stable isotopes techniques were carried out on local or regional scale. Many of them were conducted on local scale (Sutanto et al., 2012; Williams et al., 2004; Wenninger et al., 2010; Xu et al., 2008; Yepez et al., 2003; Herbst et al., 2006; Roupsard et al., 2006) and only few of them were conducted on the catchment scale (Lee and Veizer, 2003; Lee et al., 2010; Jasechko et al., 2013). A global partitioning study using the Isotope-based method and the hydrometric method is still a challenging task due to the lack of global observation of evapo-transpiration components (Sutanto et al., 2014). The flux network (FLUXNET) installed across the world provides continuously terrestrial ecosystem dataset over long periods and at sub-daily time scale (Blyth et al., 2010). However, the FLUXNET networks do not have fully-distributed spatial coverage. It does have high-density network over America and Europe but less dense network in Latin America, Africa, Australia and Asia. Together with satellite measurements, FLUXNET has been used to improve the land-surface model parameterization and model inputs, resulting in more reliable global evapo-transpiration partitioning analysis (Lawrence et al., 2007; Dirmeyer et al., 2006).

For global scale partitioning analysis, the land-surface model in general simulates the transpiration fraction values around 40-80% of the total evaporation (Lawrence et al., 2007; Choudhury and DiGirolamo, 1998; Dirmeyer et al., 2006; Miralles et al., 2011) and the isotope-based method simulates the transpiration fraction values around 35-90% (Jasechko et al., 2013; Coenders-Gerrits et al., 2014; Schlesinger and Jasechko, 2014). However, the global results from the isotope-based method may not necessarily be representative for all continental areas since the calculations from Jasechko et al. (2013) and Coenders-Gerrits et al. (2014) were only derived from the isotopic composition of big lake water bodies. The isotope-based method suffers a difficulty in global partitioning analysis due to the limitation of spatial data resolution. To overcome this problem, Isotope-enabled Global Climate Model (Iso-GCM) has been used to simulate the isotopic composition of precipitation and recently water vapor. The validation of the model results was carried out using the GNIP (Global Network of Isotopes in Precipitation) network for precipitation (Brown et al., 2006; Hoffmann et al., 1998; Noone and Simmonds, 2002; Yoshimura et al., 2008) and satellite observation for water vapor (Yoshimura et al., 2011; Frankenberg et al., 2009; Risi et al., 2012; Sutanto et al., 2015). The studies conclude that the Iso-GCM can realistically produce the isotopic pattern in the precipitation and in the water vapor compared to observations. Therefore, the main objective of this study is to provide a fully global distributed partitioning analysis results from the isotope-based method using the outputs from a well-known Isotope enabled GCM and to compare the results with others for validation. The limitations of the method and future outlook are also discussed

At a first step, we use the isotope mass balance method applied to the ECHAM4 Iso-GCM outputs to separate the total evapo-transpiration flux into soil evaporation and transpiration component. This method has been widely used on local scale or in situ ground based observation and proved to be a useful method in evapo-transpiration partitioning application. In this paper, we focus on the application of the isotope mass balance method and validation of the results as well as the limitation of the method. In the next paper, the impact of El Niño Southern Oscillation (ENSO) on plant transpiration in some regions (Indonesia, Amazon and Europe) is discussed. Moreover, we will also present the correlation of drought and transpiration fraction at the chosen regions.

The paper is organized as follows. Section 2 describes the ECHAM4 model and partitioning method. This section also gives brief explanation about water isotopologues and the isotope mass balance method. In section 3 we present the evapo-transpiration separation results, model validation, the possible errors and limitations of the method that we use, and future outlook. Conclusion is presented in section 4.

METHODOLOGY

2.1. The ECHAM4 model

All data used in this study were simulated from the ECHAM4 model, developed at the Max-Planck Institute for Meteorology in Hamburg. It is an atmospheric general circulation model (Röckner et al., 1996), which was used within the CMIP4 (Coupled Model Intercomparison Project; Covey et al., 2003) model inter-comparison study. The ECHAM4 model uses the spectral transform method for the "dry dynamics" (i.e., wind, temperature) but uses a semi-Lagrangian transport scheme for humidity and tracers such as the water isotopes (Williamson and Rasch 1994). The land surface scheme is based on a simple "bucket scheme" balancing the fluxes of heat and moisture over continental surfaces. Vegetation coverage and its influence on evapo-transpiration and runoff in dependence of the soils water holding capacity are parameterized in a highly idealized way. Water isotope physics module was implemented in the model with fractionation effects at the surface only during evaporation from interception water and snow and not from bare soil (Hoffmann et al., 1998). Each phase change is associated with temperature- and humidity-dependent isotopic fractionation processes, which are well known from laboratory experiments. The water isotope module allows therefore computing the full 4-dimensional distribution of the water isotopologues in all simulated water reservoirs of the ECHAM model.

ECHAM4 used in this study has spatial resolution of 2.8° by 2.8° (spectral resolution T42) and vertical sigma-pressure hybrid resolution of 19 layers extending from the surface to 10 hPa. The ECHAM4 model was "nudged" by a spectral nudging technique (von Storch et al., 2000). This procedure guarantees a good representation of past atmospheric conditions since the simulated wind fields are forced to be close to the ERA40 reanalysis data. It has been shown that this approach considerably improves both, precipitation and its isotopic composition for specific months, compared to climatological means. The ECHAM data used in this paper was nudged over the period 2001 with 6 hours temporal resolution. Detailed information about ECHAM4 Iso-GCM can be found in Hoffmann et al. (1998).

2.2 Partitioning method

2.2.1 Water isotopologues

Isotopes are atoms of the same element that have the same numbers of protons and electrons but different numbers of neutrons, meaning that the various isotopes have different masses. When different elements (atoms) combine to form certain molecules (e.g., two hydrogen atoms and one oxygen atom combine to form a water molecule) several isotopic varieties of the same chemical molecule can be made out of the different isotopes; these isotopically different molecules are called isotopologues. The most common isotopologue of water (HHO) consists of two hydrogen atoms with mass number one (1H) and an oxygen atom with mass number 16 (^{16}O). When a light isotope of hydrogen or oxygen in water is replaced by a heavy isotope (e.g., ^{2}H, ^{17}O, ^{18}O), various heavy water isotopologues with one or more heavy isotopes can be formed. For atmospheric and hydrological cycle studies, the water isotopologues, $^{1}H_2{}^{18}O$ (^{18}O = oxygen 18) and $^{1}H^{2}H^{16}O$ or HDO (^{2}H = D, Deuterium), are commonly used. These water isotopologues are categorized as stable water isotopologues.

One attractive method to partition the evapo-transpiration flux into its individual components is the use of isotope measurements. This is because the isotopic composition undergoes characteristic changes (isotope fractionation) during a number of physical processes, importantly phase changes and diffusion. As an important example, evaporation changes the isotopic composition of soil water, whereas transpiration via plants does not strongly modify the isotopic composition of root water (Ehleringer and Dawson 1992; Williams et al. 2004). Using these distinct isotope signatures allows the water isotopologue technique to qualitatively distinguish between soil evaporation and transpiration, for example studies by Yepez et al. (2003) in Arizona USA, Robertson and Gazis (2006) in Washington USA, and Sutanto et al. (2012) in the Netherlands.

The isotopic composition of a sample is commonly expressed as a fractional difference of an isotope ratio R relative to a standard ratio and reported in delta (δ) notation:

$$\delta = \left(\frac{R_{sample}}{R_{standard}}\right) - 1 \quad (1)$$

where R is the heavy-to-light isotope ratio of a certain element in the sample or standard. In the case of water, $^2R = {^2H}/{^1H}$ is the deuterium-to-hydrogen ratio in the sample, and the international standard ratio is Vienna Standard Mean Ocean Water (VSMOW) with ^{2}RVSMOW = 155.76 ± 0.05 x 10^{-6} and ^{18}RVSMOW = 2005.2 ± 0.45 x 10^{-6} (Craig, 1961). As isotope variations are generally small, they are normally expressed as per mil (‰) difference from the standard being used. A higher or more positive δ values means that the heavy isotope content is higher than in the standard (also called "heavier"), and vice versa for lower or more negative δ values.

2.2.2 Isotope mass balance

Partitioning of evaporation using isotope-based method is commonly conducted using an isotope mass balance method. This method is based on a simple two-source mixing model, where evapo-transpiration flux (F_{ET}) is consisted of soil evaporation (F_E) and transpiration (F_T) components.

$$F_{ET} = F_E + F_T \quad (2)$$

where F stands for flux. Subscript ET, E, and T stand for evapo-transpiration, soil evaporation, and transpiration, respectively. Each flux in Eq. 2 has its isotopic composition characteristic (δ), thus the soil evaporation and transpiration can be calculated based on the derivation of Eq. 2:

$$F_{ET}\delta_{ET} = F_E\delta_E + F_T\delta_T \quad (3)$$

$$F_E = \frac{\delta_T - \delta_{ET}}{\delta_T - \delta_E} F_{ET} \quad (4)$$

$$F_T = \frac{\delta_{ET} - \delta_E}{\delta_T - \delta_E} F_{ET} \quad (5)$$

Isotopic composition of each component can be obtained from direct measurements or calculated using empirical methods. The fractions of soil evaporation and transpiration can be simulated using Eq. 4 and 5 without F_{ET} component. To calculate the flux, F_{ET} can be directly calculated using hydrometric methods (e.g., Bowen ratio, lysimeter, eddy covariance) or from the model (for this study).

2.2.3 Estimation of the isotopic composition of soil evaporation

To calculate the isotopic composition of soil evaporation (δ_E), the empirical Craig-Gordon formulation is often used (Craig and Gordon, 1965):

$$\delta_E = \frac{\delta_e\alpha - h\delta_V - \varepsilon_{eq} - (1-h)\varepsilon_k}{(1-h) + (1-h)\varepsilon_k/1000} \quad (6)$$

where δ_e represents the $\delta^{18}O$ or δD values of soil water (δ_{SW}) at evaporation front (here we use δD soil water at top layer), δ_V is the δ value of the background atmospheric water vapor (here we use δD_V at near the surface, ~1000 hPa), α is the temperature-dependent equilibrium fractionation factor between liquid and vapor, ε_{eq} = 1000(1-1/α), ε_k = 1000(α_k-1), α_k is the kinetic fractionation associated with diffusion of water through the soil, and h is relative humidity (taken from the model at ~1000 hPa).

α for equilibrium fractionation factor (α) and kinetic fractionation factor (α_k) from liquid to vapor can be calculated according to Clark and Fritz (1997):

$$10^3 ln\alpha = \frac{10^6 a}{T^2} + \frac{10^3 b}{T} + c \quad (7)$$

$$10^3 ln\alpha = \theta 25(1-h)‰ \quad (8)$$

where T is temperature in kelvin (we use T at 2 m high); constants a, b and c for ^{2}H are a = 24.844, b = −76.248 and c = 52.612, θ is 0.5 for open water evaporation and 1 for soil and leaf cover (Kendall and McDonnell, 1998).

2.2.4 Estimation of the isotopic composition of transpiration

Researchers extent the use of Craig-Gordon empirical method to estimate the isotopic composition of transpired moisture from plants (δ_T). The method is similar with the original Craig-Gordon model. However, the δ_e to calculate δ_E is replaced by the isotopic ratio of liquid water at the evaporating front in leaf (δ_L). The formula is described as follow:

$$\delta_T = \frac{\delta_L\alpha - h\delta_V - \varepsilon_{eq} - (1-h)\varepsilon_k}{(1-h) + (1-h)\varepsilon_k/1000} \quad (9)$$

In Eq. 9, the isotopic composition of transpiration can be calculated if the values of δ_L are known. In this case, the δ_L is not measurable and direct measurement of δ_T is difficult. Using steady state assumption, where the isotopic composition of leaf water is equal to the isotopic composition of xylem or twig, the Craig-Gordon model has been modified to estimate δ_L. Then δ_L in steady state can be expressed as:

$$\delta_L \approx \delta_{SW} + \varepsilon_{eq} + \varepsilon_k + h(\delta_V - \varepsilon_k - \delta_{SW}) \quad (10)$$

2.2.5 Estimation of the isotopic composition of evapo-transpiration

The last unknown parameter to partition evapo-transpiration flux is the isotopic composition of evapo-transpiration (δ_{ET}). The mass balance mixing relationship such as Keeling plot analysis can be used to determine the δ_{ET} (Keeling, 1961). The concept is that the combination of some background amount of substance from sources is generally considered as the atmosphere concentration of this substance. Thus, the relationship between the isotopic values of air samples at different heights above the ground and the inverse of the concentration of the substance is linear:

$$\delta_D = (\delta_b - \delta_{ET}) w_b \left(\frac{1}{w}\right) + \delta_{ET} \qquad (11)$$

where δD is the isotopic composition of vapor collected from the boundary layer, δ_b is the isotopic composition of the atmospheric background, w_b is the atmospheric background of water vapor concentration, w is the water vapor concentration in the boundary layer. When we use the water vapor for Keeling plot analysis, the y intercept reflects the isotopic values of evapo-transpiration (Williams et al., 2004; Yepez et al., 2003). For Keeling plot analysis, we use the water vapor mixing ratio (H_2O) and δD at 1000-850 hPa (near surface to boundary layer) at 6 hours interval for 1 day.

RESULTS AND DISCUSSION

3.1 Global evaporation fraction

Global soil evaporation and transpiration fraction are calculated using the Craig-Gordon model (Eq. 6 and 9, respectively) in monthly basis. We use monthly δD values from the ECHAM4 model outputs, except for δ_{ET}. We average the daily δ_{ET} results from Keeling plot analysis to monthly δ_{ET}. Since the isotope mass balance method disentangles evapo-transpiration fraction into two evaporation components: transpiration and soil evaporation, this study, therefore, will only discuss the results based on these two evaporation contributors. The sum of transpiration fraction and soil evaporation fraction should be equal to 100%.

3.1.1 Annual T and E fraction

Partitioning results using isotope mass balance method are presented in Figure 2. Figure 2 shows that in general the transpiration fraction is higher for vegetated area and lower for deserted area. Southeast Asia, West Europe, North US, and North Amazon produce high transpiration fraction. On the other hand, Mediterranean continent, Sahara dessert, and Australia have higher soil evaporation fraction. This is expected since these regions are mostly covered by sand and less vegetation.

In Australia, although both transpiration and soil evaporation are high, the most dominant evaporation flux in this continent is still soil evaporation (~70%). The Australian dessert contributes higher soil evaporation yearly. In Amazon region, well known with one of the biggest rain forest, the transpiration fraction (~75%) is two times higher than soil evaporation fraction (~30%). A very clear distinct result between transpiration and soil evaporation fraction is seen over Indonesia region. In this region, the transpiration fraction (~80%) is four times higher than the soil evaporation fraction (~20%). The tropical climate, abundant precipitation, and dense vegetation are the causative factors for high transpiration fraction in Indonesia.

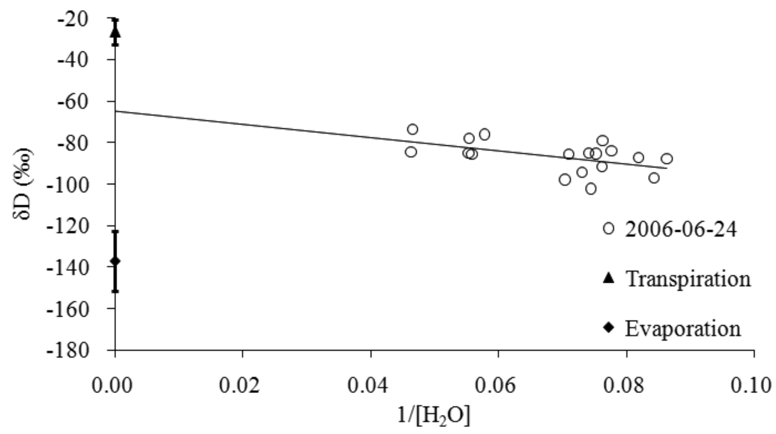


Figure 1 Keeling plots for water vapor isotopologue at different heights. The isotopic composition of evapo-transpiration is -58 ‰ (y intercept; Xu et al., 2008). *δD_T is around -30 to -20 ‰ and δD_E is around -165 to -150 ‰ (Y-axis).*

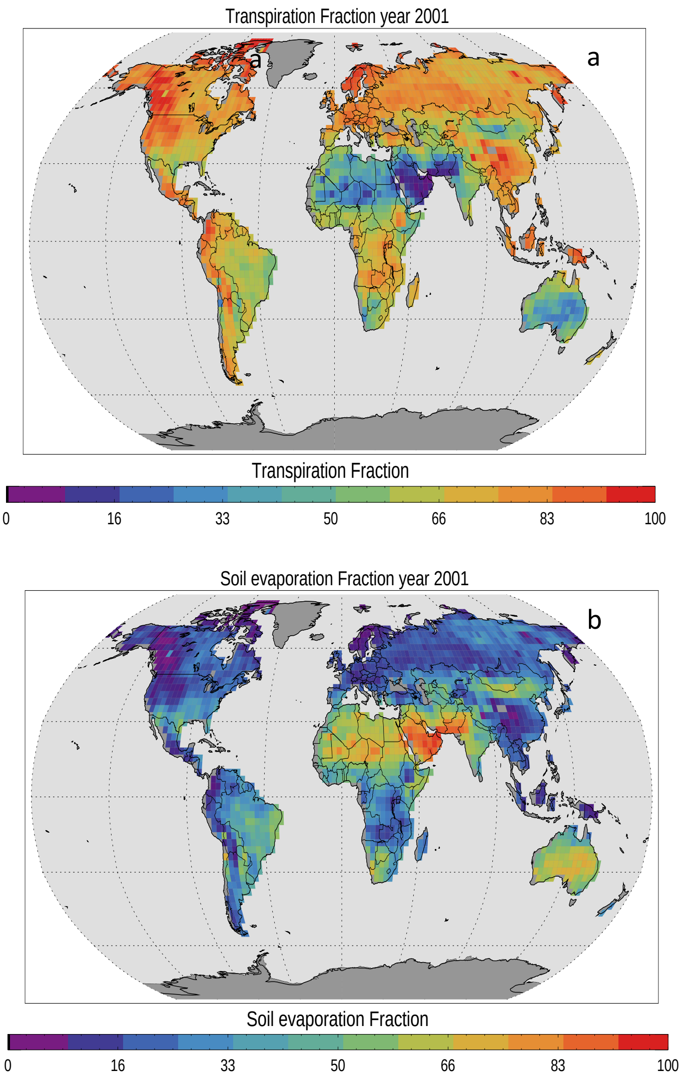
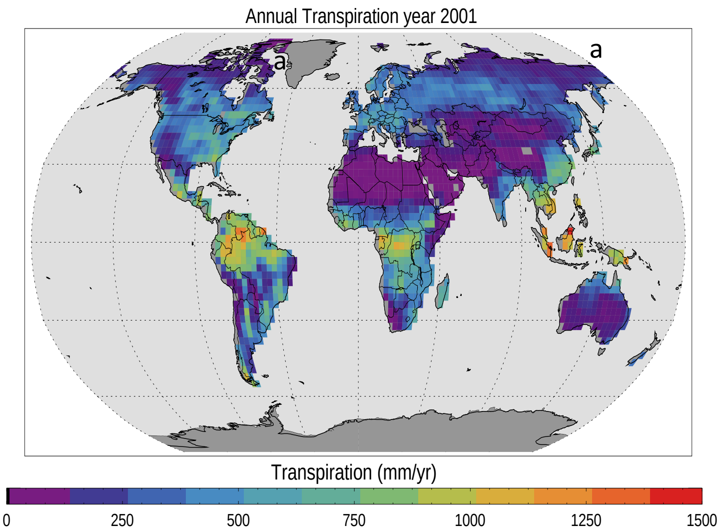


Figure 2 Global transpiration fraction (a) and global soil evaporation fraction (b) calculated using isotope-based method. Purple color stands for small fraction and red color stands for high fraction.

Using the global transpiration and soil evaporation fraction, the transpiration and soil evaporation amount can be calculated by multiplying these fractions with total evapo-transpiration flux simulated by the model (Fig. 3). The highest transpiration rate is occurred mainly in the tropical regions such as Amazon, Africa and Indonesia. The heart of Amazon rain forest and North Borneo rain forest produce the highest transpiration amount of more than 1200 mm/year, which is nearly identical with value reported by Kumagai et al. (2004) in Bornean rainforest (1193.1 mm/year). In average, the transpiration rate in both Amazon and Borneo is around 900-1200 mm/year, which is similar with a study from Bruijnzeel (1990) in humid tropical forests (885-1285 mm/year).

Compared with transpiration, high soil evaporation amount is visible in South America, Brazil, and Central Africa although maximum soil evaporation rate is only around 300 mm/year. There are some regions where both transpiration and soil evaporation amount are low. These regions are deserted regions such as Great Basin Desert in US, Sahara and Mediterranean desert, Australia, and Gobi desert in China, as well as cold-icy regions such as Alaskan and Himalayas. We exclude the results from North and South Pole (Greenland and Antarctica).



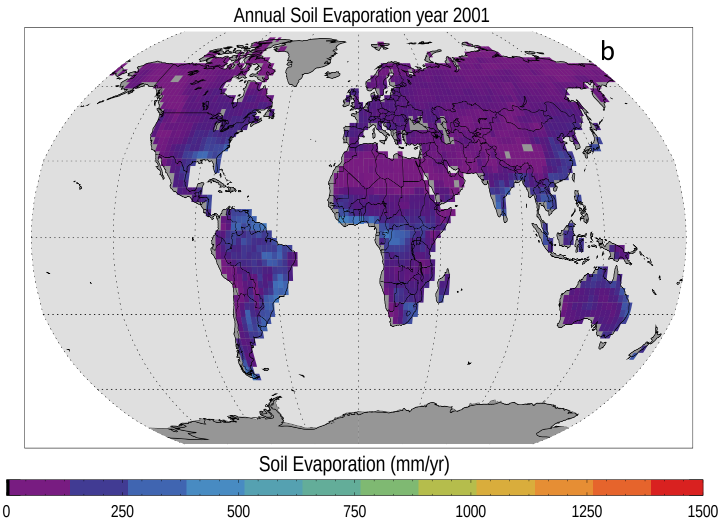


Figure 3 Global transpiration rate (a) and global soil evaporation rate (b) calculated from each component fraction multiplied by evapo-transpiration flux in mm/year. Purple color stands for small transpiration/soil evaporation rates and red color stands for high small transpiration/soil evaporation rates.

3.1.2 Seasonal T and E fraction

In previous section we discussed the annual transpiration and soil evaporation, separated using the isotope mass balance method. In addition, the monthly datasets that we use in the analysis allow us to look at in detail the seasonality fractions of both soil evaporation and transpiration during summer (June, July, August; JJA) and winter (December, January, February; DJF). The impacts of seasonality on the partitioning results due to summer and winter are presented in Figure 4.

Transpiration fraction during summer is extremely high (>80%) in the Northern hemisphere as well as in China as expected (Fig. 4a). The result agrees well with previous studies using in situ isotope based method and hydrometric method. Studies in North US show that the transpiration fraction during summer is around 76-88% for Savanna woodland and grass (Yepez et al., 2003, Ferreti et al., 2003). In European countries, the transpiration fraction can be up to 97% (Herbst et al., 1996). In china during summer, the transpiration fraction of wheat and oaks are 80 and 96%, respectively (Xu et al., 2008, Zhang et al., 2011). This indicates that our result is comparably good with in situ measurement based studies.

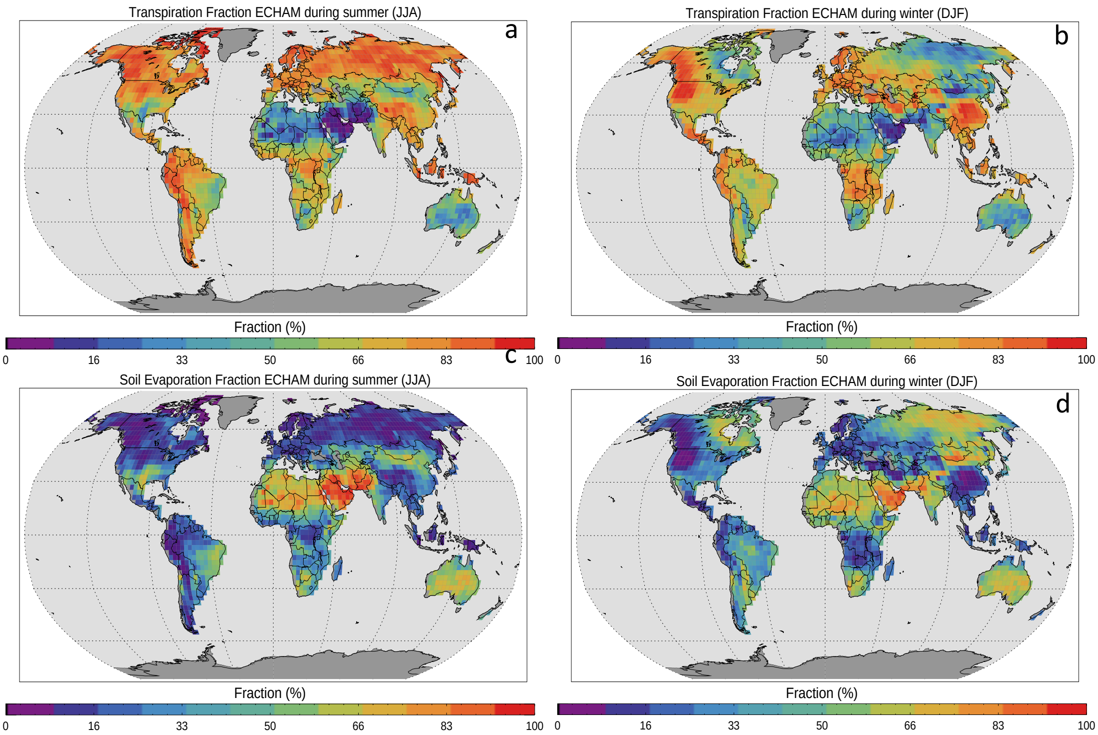


Figure 4 Seasonal comparison of transpiration fraction during summer (a) and during winter (b); and of soil evaporation fraction during summer (c) and during winter (d). Purple color stands for small fraction and red color stands for high fraction.

During winter, our analysis shows the reduction of transpiration fraction to below 50% in Siberian and North-east US (Fig. 4b). This is plausible since during this season the plants are dormant or under less favorable condition, and only few winter plant can transpire water (i.e., pine trees). In Europe, the transpiration fraction is somewhat still higher, with fraction of 50-70%. The reason why the transpiration fraction is high during winter in this region remains unclear. It is hardly to find any literatures discuss about plant transpiration fraction during winter when the temperature may reach below zero degree. Many studies discuss plant transpiration during summer, spring and few during autumn.

Another interesting point to look at is the transpiration fraction in the tropical rainforest during winter, Amazon and Indonesia. In these regions, the transpiration fraction is lower during winter/rainy season than during summer. One of possible reasons is due to the less availability of latent heat flux for transpiration. There are many low level clouds during rainy season that may reflect back the radiation from the sun to the atmosphere, meant less incoming solar radiation absorbed by the surface (in average globally around 161 Wm^{-2}) and less of that energy transferred back to the atmosphere as latent heat through evapo-transpiration (in average globally around 80 Wm^{-2}; Trenberth et al., 2009). Higher transpiration fraction during summer/dry season indicates that transpiration is not limited by the availability of soil moisture in these regions. Abundant precipitation each year and deep root trees make water accessible for trees transpiration for the whole year.

Over the deserted areas, the soil evaporation is highest during summer and higher during winter (Fig. 4c and d). However, over Australian desert, it seems that soil evaporation is slightly higher during winter than summer. This is obvious since Northern hemisphere winter means Southern hemisphere summer. A clear distinct result is seen in the North regions such as Siberian and Northeast US. Soil evaporation fraction in these regions is quite high during winter and relatively unusual. Snow most likely covering the surface and we expect no soil evaporation during this season. One process that may occur during winter season is sublimation, where snow evaporates and transforms into water vapor due to increase of temperature. Though this process is similar to evaporation, it is exactly not evaporation where liquid phase becomes air. Sublimation is the conversion between solid and gaseous phases of matter. Nevertheless, one should note that the isotope mass balance method calculate the evapo-transpiration fraction into transpiration and soil evaporation. There is no other component in this analysis such as canopy evaporation or even sublimation. Thus the evapo-transpiration fraction needs to be divided between these two components only and sublimation process may be added to soil evaporation fraction.

3.2 Validation of the results

3.2.1 Point sampling in Amazon

To validate our results, we average and plot all δD components over the Amazon basin for comparison with other studies (15N-15S, 75W-45W; Fig. 5). This study and two references that we use are based on the Steady State Assumption (SSA). Using SSA assumption, therefore, the isotopic composition of stem water is equal to the isotopic composition of soil water. δ_{SW} simulated by the ECHAM4 model (-37.7 ‰) is consistent with literatures, which is in between δ_{SW} values from other studies conducted by Yakir and Sternberg (2000), and Zhang et al. (2010) (-32 and -48 ‰, respectively). δ_E calculated using the Craig-Gordon model in our study is more depleted compared with the references. The differences in humidity and temperature data for calculation may responsible for the difference in δ_E analysis.

The modification of Craig-Gordon model to estimate the isotopic composition of transpiration produces higher δ_T value than the soil water isotopic value. In two references, however, they assume δ_T is equal to δ_{SW} due to SSA condition (Fig. 5). Under SSA condition, there is no isotopic fractionation during water uptake by roots and transport in stem (Ehleringer and Dawson, 1992). This assumption is not perfectly met in nature. An isotopic enrichment may occur in the leaf due to the effect of equilibrium and kinetic fractionations that take place during evaporation from leaf surface (Flanagan and Ehleringer, 1991). Thus the non-fractionated water from roots may mix with the fractionated water from leaf and produces isotopically enriched water for transpiration. Using the modification Craig-Gordon model, the isotopic composition of transpire water is ~7 ‰ higher than soil water. In general all δD values in our study is consistent with previous studies.

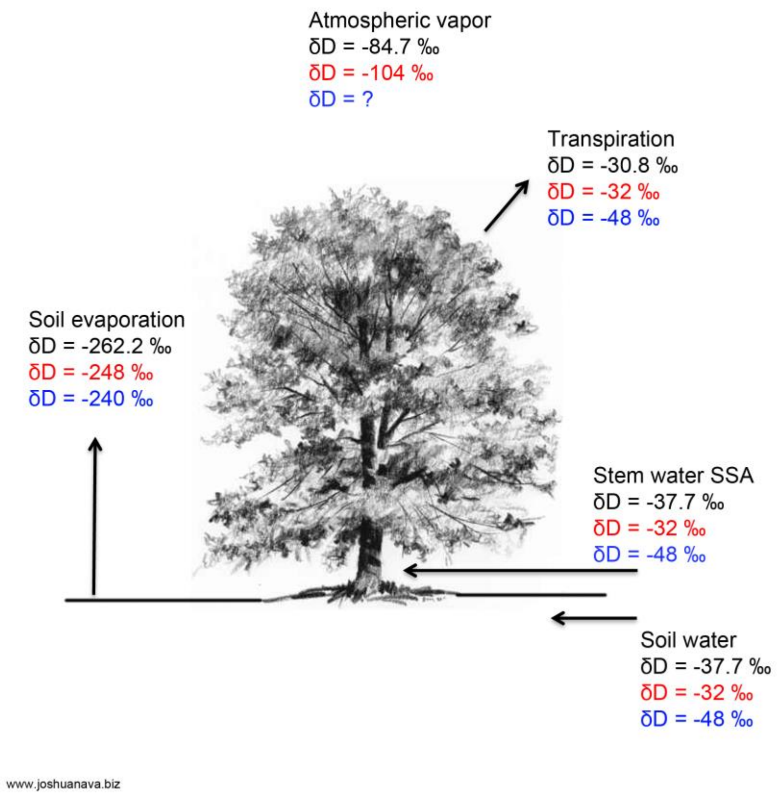


Figure 5 Isotopic values for all components. Black color is δD values calculated using ECHAM4 outputs, red color is δD values taken from Yakir and Sternberg (2000), and blue color is δD values taken from Zhang et al. (2010). There is no δD value for atmospheric vapor in Zhang et al. (2010).

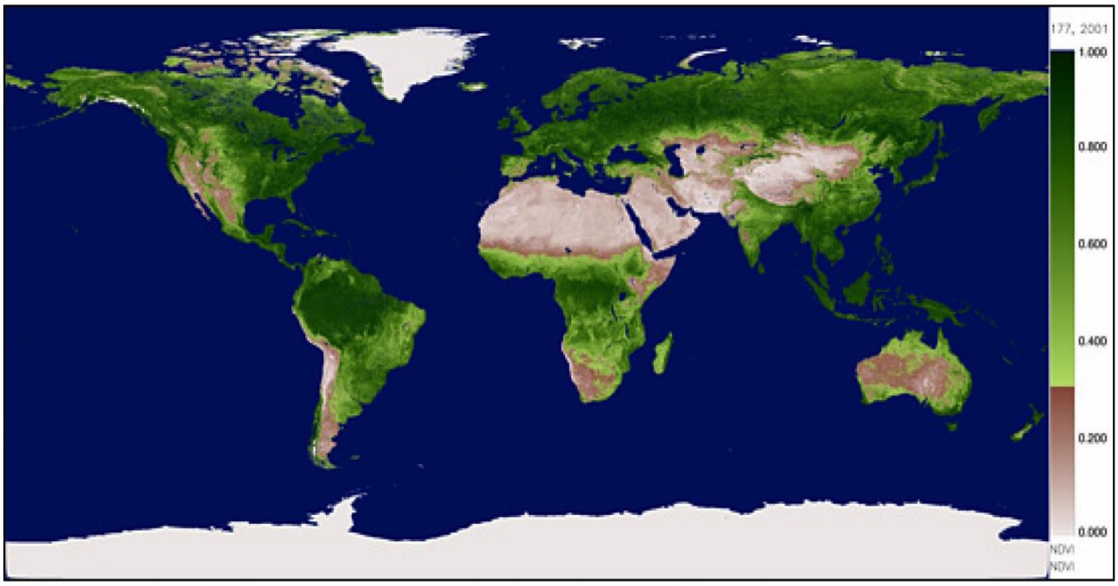
3.2.2 Comparison with vegetation index

Instead of comparing the isotopic calculation from different components, this section will compare the Normalized Different Vegetation Index (NVDI) measured by Moderate Resolution Imaging Spectroradiometer (MODIS) instrument onboard Terra/Aqua satellite with the simulated transpiration fraction. The NVDI is a well-known and simple index to asses whether the targeting area contains live, green vegetation or not. This method uses the highest absorption and reflectance regions of chlorophyll to detect the vegetation and it is sensitive to the photosynthetically active vegetation (Bulcock and Jewitt, 2010). The regions where contain dense vegetation yield high NVDI values.

Figure 6 shows the regions with high vegetation density as measured by MODIS. Regions with high vegetation density are Amazon, Europe and Russia, Indonesia, Central Africa, and East US. These regions are also presented as the regions with high transpiration fraction (Fig. 2a). The dense vegetation over these regions causes the trees canopy covering the whole surface thus bare soil is hardly exposed to the sun. This condition is favorable for high transpiration fraction compared to soil evaporation.

3.2.3 Comparison with assemble measurements

A recent study by Schlesinger and Jasechko (2014) summarizes the transpiration fraction for some vegetation types, from the tropics to Mediterranean shrubland (Fig. 7). The transpiration fraction can vary from 40% in the Mediterranean area to 80% in the tropical rainforest. Our results correspond well with their results. The transpiration fraction in the rainforest, for example, is in the range of 60-80%, and the result from isotope mass balance also shows similar amount (65-85%, see Fig. 2a for Amazon and Indonesia). For temperate grassland, the transpiration fraction is around 40 to 80%, and again the result of this study is situated in their ranges, 55-80%. Though most of our results are in agreement with Schlesinger and Jasechko (2014) study, the transpiration fraction in the deserted region from our study, however, is on the lowest boundary of their error bar. It can reach 20% in the Mediterranean desert. For global transpiration fraction, Schlesinger and Jasechko (2014) conclude that the contribution of transpiration to total evapo-transpiration is 61% and 39% of incident precipitation returns to atmosphere via evapo-transpiration.



Filled Normalized Difference Vegetative Index (NDVI) Maps 16-Day Averages

Figure 6 Normalized Different Vegetation Index (NDVI) taken from MODIS instrument in June. Source: http://modis-atmos.gsfc.nasa.gov/NDVI. Dark green color stands for high NVDI and white-pink color stands for low NVDI.

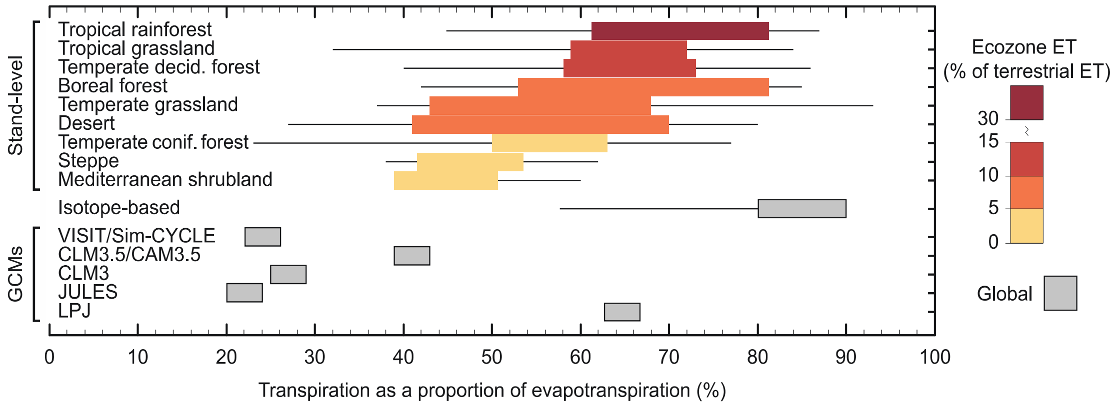


Figure 7 Transpiration fraction measurements in some land use types and global transpiration fraction from model and isotope-based method (Schlesinger and Jasechko, 2014).

Table 1 Transpiration fractions and water return to the atmosphere via transpiration. The amount of transpiration and precipitation is annual average.

No	Location	Coordinate	Transpiration fraction (%)	Transpiration amount (mm)	Precipitation amount overland (mm)	water return through transpiration (%)
1	Global	global	69	541,094	1,250,530	43
2	Amazon	15N-15S, 75W-45W	71	69,437	143,912	48
3	Indonesia	7.5N-10S, 90E-150E	87	19,519	68,643	28
4	Europe	40N-75N, 10W-40E	84	47,441	81,618	58

On the contrary, GCM models produce lower transpiration fraction than measurements and isotope-based method. In general, GCM models estimate transpiration fraction less than 50% (Sim-CYCLE, CLM3.5, CLM3, and JULES, Fig. 7). Only LPJ model and GLEAM model (Miralles et al., 2011) estimate transpiration fraction around 60-80%. The relatively lower transpiration fraction simulated in current models needs a reassessment of the model parameterizations.

In the following we average the transpiration fraction for global, in Amazon, in Indonesia and in Europe (Table 1). Table 1 shows that the global transpiration fraction calculated using the isotope mass balance method is 69% and 43% water at the surface returns to the atmosphere through transpiration. Our results are slightly higher than the previous results (Schlesinger and Jasechko, 2014). However, these results are lower than the global transpiration fraction calculated using the published isotope-based method (80-90%) and in agreement with LPJ model. A study by Blyth and Harding (2011) showed that the fraction of rain evaporated back to the atmosphere varies from 15-85% and using a water balance method, Oki and Kanae (2006) concluded that the land precipitation transported back to the atmosphere through evapo-transpiration may extent as much as 60%. Our result is in the range of their studies. Indonesia, which has the highest transpiration fraction around 87%, returns only 28% of the precipitation water back to the atmosphere. This indicates that there is much water available for storage or moving back to the ocean.

3.3 Limitations and possible errors

In the previous sections we have shown that the global transpiration fraction can be calculated using the isotope mass balance method, where all δD values are taken from the ECHAM4 model outputs. The results seem promising although each method has its own limitations and possible errors. The main source of inaccuracy in the isotope mass balance method is the SSA assumption that is normally used. Under the SSA assumption, plants transpire water, which is isotopically equal with soil water isotopologue. In fact, the isotopic composition of leaf water changes overtime. The SSA may produce a reasonable δ_T during the day when stomata are fully open (Lai et al., 2006; Farquhar and Cernusak, 2005). This may be the reason why our results are marginally higher than the measurements. More detailed information about limitation of the SSA assumption can be found in Sutanto et al. (2014).

The use of Keeling plot method with six hours temporal resolution datasets may lead to the significant errors. This method is based on the linear relationship between inverse concentration and background water vapor. For daily scale analysis, four point datasets are not adequate to produce a robust straight-line correlation since it is subject to change if there is an outlier and the data. Moreover, this method is very sensitive to the changes in the boundary layer because of some processes such as mixing of water vapor due to advection, precipitation events, and entrainment (Zhang et al., 2010). In addition, the assumption of a constant δET over the sampling periods is not always valid in the nature.

Despite some limitations in the methods that we use, the use of model outputs as bases for analysis may contribute the highest errors. Model is a model and it cannot fully represent the complexity of the nature although the model was validated with measurements (Hoffmann et al., 1998; Sutanto et al., 2015). In addition, the model does not compute strong fractionation effects over land, and the entire separation approaches are based fundamentally on significant differences of the isotopic fluxes from bare soil and transpiration. This argument makes the separation works using the GCM model outputs relatively insubstantial.

3.4 Future outlook

The biggest challenging for global surface fluxes analysis is due to lacking in temporal and spatial data coverage. The Intergovernmental Panel on Climate Change (IPCC) technical report on water and climate points out that evaporation is the weakest link in the hydrological cycle. There are lacks of publicly available global evaporation data compared with rainfall and river discharge. The flux network (FLUXNET; Baldocchi et al., 2001) installed across the world in different climatic zones hopefully can help to increase the spatial observation coverage. There are more than 400 flux towers operated independently or as part of regional networks.

Although FLUXNET is undoubting as the most comprehensive terrestrial ecosystem dataset currently available (Blyth et al., 2010), the coverage of the networks is dense at the area of interest such as Amazon, or in the well developed countries such as US, Europe, and Australia. For the rest of the world, the coverage of FLUXNET network is still poor. Remote sensing instruments installed on satellite platforms can measure surface fluxes with global coverage. However, measurement using satellite instruments is also inadequate to produce high temporal resolution datasets.

Figure 8 shows the global transpiration fraction analysis calculated from TES satellite datasets. We use daily TES δD and climatological datasets and aggregate them into monthly datasets. TES δET is obtained by using the Keeling plot analysis every two weeks. δ_{SW} for analysis is obtained from the ECHAM4 model since there is no global measurement of soil water isotopic composition. For this analysis, therefore, we use all measurement data except for δ_{SW}. The purpose of this analysis is to discover the possibility of using water isotopologue measured by satellite instrument.

CONCLUSION

We demonstrated that the isotope mass balance method could be applied using the isotope-enabled ECHAM4 model outputs to estimate global transpiration and soil evaporation fraction. Regions with dense vegetation may have high transpiration fraction up to 80%. On the contrary, transpiration fraction over deserted regions may extent to lower than 20%. During summer, the transpiration fraction, especially in Europe, US, and Indonesia, increases up to 90% and deceases to less than 50% when the snow is covering the surface. In average the global transpiration fraction estimated using isotope mass balance using the ECHAM model outputs is 69%. This result is in agreement with past studies using assemble of ground measurements and models, indicating the transpiration fraction of 61-80%. Globally, around 43% water from the precipitation is transported back to the atmosphere through transpiration.

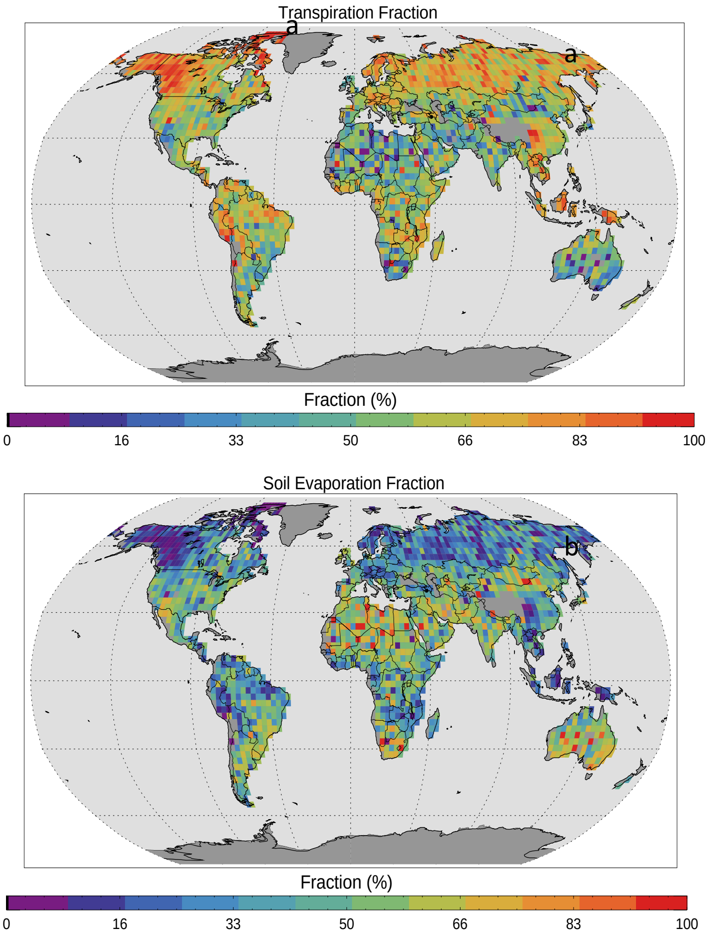


Figure 8 Global transpiration fraction (a) and soil evaporation (b) calculated using TES datasets. Purple color stands for small fraction and red color stands for high fraction.

Although the results are promising, the isotope mass balance method to separate the evapo-transpiration fraction into its individual components has some limitations. The SSA assumption and the Keeling plot method are conjectured to contribute the highest inaccuracy. The use of model outputs itself may also contribute the highest errors since the model cannot fully represent the complexity of the nature.

The use of satellite measurements to estimate the transpiration fraction is very promising. However, a new satellite instrument, which can measure water isotopologue at greatly improved precision and temporal as well as spatial resolution, is absolutely needed. In addition, new FLUXNET towers installed in the remote regions are desirable for model validation and to complement the previous datasets.

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