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HESS Opinions: A perspective on different approaches to determine the contribution of transpiration to the surface moisture fluxes

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Abstract

Current techniques to disentangle the total evaporative flux from the continental surface into a contribution evaporated from soils and canopy, or transpired by plants are under debate. Many isotope-based studies show that transpiration contributes generally more than 70 % to the total moisture fluxes, while other isotope-independent techniques lead to considerably smaller transpiration fractions. This paper provides a perspective on isotope-based vs. non isotope-based partitioning studies. Some partitioning results from isotope-based methods, hydrometric measurements, and modeling are presented for comparison. Moreover, the methodological aspects of the analysis of partitioning are discussed including their limitations, and explanations of possible discrepancies between the methods are briefly discussed. We conclude that every method has its own uncertainties and these may lead to a high bias in the results, e.g. instruments inaccuracy and error, some assumptions used in analyses, parameters calibration. A number of comparison studies using isotope-based methods and hydrometric measurements in the same plants and climatic conditions are consistent within the errors, however, models tend to produce lower transpiration fractions. The relatively low transpiration fractions in current state of the art land surface models calls for a reassessment of the skill of the underlying model parameterizations. The scarcity of global evaporation data makes calibration and validation of global isotope-independent and isotope-based results difficult. However, isotope enabled land-surface and global climate modeling studies allow the evaluation of the parameterization of land surface models by comparing the computed water isotopologue signals in the atmosphere with the available remote sensing and flux-based data sets. Future studies that allow this evaluation could provide a better understanding of the hydrological cycle in vegetated regions.

1 Introduction

Continental evaporation (including transpiration, soil evaporation and canopy evaporation) is an important process controlling energy and mass exchange between the terrestrial ecosystems and the atmosphere (Seneviratne et al., 2010). Transpiration is the largest contributor to the water flux from continental areas (Lawrence et al., 2007; Blyth and Harding, 2011; Jasechko et al., 2013). Therefore, many studies have been carried out to quantify the transpiration through direct measurements or techniques disentangling the various evaporation components. Direct measurement of transpiration has been performed for decades but the accuracy and separation of transpiration from total evaporation still remain a challenge. The fraction of transpiration to the total surface moisture flux obviously depends upon the nature and state of the surface, such as the presence of lakes or the seasonality of vegetation (growing in the spring or dormant in the fall). These conditions generate variability in each evaporation component and complicate the separation of the various components.

There are many methods and techniques to quantify this transpiration fraction. A conventional method for partitioning evaporation is the combination of hydrometric measurements such as sapflow to measure transpiration rate with other methods to calculate the total evaporation (Kelliher et al., 1992; Herbst et al., 1996; Roupsard et al., 2006; Mitchell et al., 2009; Cavanaugh et al., 2011). Soil evaporation is then calculated from the difference of total evaporation and transpiration. More recent methods analyze the isotopic composition of liquid water and water vapor (e.g. Yepez et al., 2003; Ferretti et al., 2003; Williams et al., 2004; Xu et al., 2008; Lai et al., 2006; Robertson and Gazis, 2006; Wang et al., 2010; Sutanto et al., 2012; Jasechko et al., 2013). This method has become common since measurements of stable isotopologues in precipitation and water vapor are relatively easy and robust. Other methods use global land-surface models (Choudhury and DiGirolamo, 1998; Dirmeyer et al., 2006; Oleson et al., 2004). In the land-surface models, the contribution of each evaporation component is principally simulated from the land surface scheme due to e.g. turbulent transfer, moisture limitations, and plant physiology. Although all methods above can estimate the transpiration fraction, results from those methods vary. In general, isotope-based methods tend to result in a higher transpiration fraction than other methods. Jasechko et al. (2013) estimated the global transpiration fraction to be 80–90 % of the total evaporation from land, which implies a more dominant role of transpiration than most other previous studies. The transpiration flux in that study was calculated using the isotope mass balance method in lake catchments where the isotopic composition of the evaporative flux was estimated using an evaporation model applied to the isotopic composition of the lake water.

Here we provide a perspective on different approaches for disentangling the different fluxes contributing to the total evaporation. Some partitioning results from isotope-based methods, hydrometric measurements, and modeling are presented for comparison. We also discuss briefly how the partitioning studies have been carried out, their limitations, and the possible causes of the discrepancies between those methods.

2 Methods to derive the transpiration fraction of total evaporation

Transpiration from plants can be measured directly using a combination of hydrometric techniques and empirical formula (Herbst et al., 1996; Roupsard et al., 2006; Kelliher et al., 1992; Cavanaugh et al., 2011). The advance of stable isotope technique to measure transpiration since the 1980s (Calder et al., 1986; Calder, 1992) has ben supported by the development of robust and accurate instruments. In the late 90's, analysis of evaporation fluxes using global models became a popular technique. The development of super computer and remotely sensed observations make these global model simulations possible.

2.1 Hydrometric method

Measurements of direct transpiration are available using hydrometric devices. A widely used quantity to determine continuous transpiration through tree stems and branches is sapflow (Granier, 1985). However, sapflow measurements need to be combined with methods measuring total evaporation to quantify the relative contribution of transpiration to the total evaporating flux (including soil evaporation and evaporation from intercepted water). Eddy covariance, Bowen ratio techniques, and lysimeters are commonly used in a combination with sapflow to calculate the total evaporation. Soil evaporation rate can be obtained from the difference between total evaporation and transpiration, from bare soil weighing-lysimeters, or from eddy covariance systems placed bellow the canopy.

Measurement of transpiration using the sapflow method is based on asymmetric heat transfer by upward or downward moving tissue moisture. The Thermal Dissipation Probe (TDP) heated needles are implanted in the active xylem up- and downstream from a pulse heat source. The velocity of water through the plant is minimal when the temperature difference between two sensors is maximal. On the other hand, the velocity of water increases when the temperature difference decreases.

Evaporation measured by the Eddy covariance system is defined as the flux of H_2O through a horizontal plane above the canopy. Simultaneous high frequency vertical wind speed and atmospheric water vapor measurements are needed since this method correlates fast fluctuations of vertical wind speed with fast fluctuations in atmospheric water vapour density (Nouri et al., 2013).

Another method to measure evaporation is Bowen ratio energy balance. This method is a micrometeorological method used to estimate latent heat flux, which is calculated by measurements of the temperature and humidity gradients above the canopy (Bowen, 1926). Unlike the Eddy Covariance and lysimeter methods it is an indirect method; it uses the ratio of the temperature and moisture gradient to partition the available energy (net radiation minus soil heat flux) over latent and sensible heat fluxes.

A lysimeter measures the evaporation loss by weighing an isolated soil sample. Ideally, lysimeters should contain undisturbed soil samples to represent the direct surroundings. Evaporation is calculated from the weight change over time, corrected for precipitation gains and percolation losses.

More methods to determine the evaporation exist (Brutsaert et al., 1982; Shuttleworth, 1993; Kool et al., 2014). A summary of the advantages and disadvantages of these hydrometric methods were described by Nouri et al. (2013) (Table 1).

2.2 Isotope-based method

The use of stable water isotopologues (principally $H_2^{18}O$ and $^1H^2HO$) as tracers present a new and important technique to enable fluxes tracing within soil–plant–atmosphere continuum (SPAC) system (Kendall and McDonnell, 1998; Mook, 2000; Zhang et al., 2010; Wenninger et al., 2010). Fractionation processes associated with water diffusion and phase changes affect the partial pressures of the different water isotopologue. The lighter isotopologues preferentially evaporate from bare ground and open water surfaces such as oceans and lakes, and leave the condensate phase enriched in the heavier isotopologue. In contrast, transpiration does not modify the isotopic composition of the remaining groundwater (Ehleringer and Dawson, 1992; Kendall and McDonnell, 1998; Tang and Feng, 2001; Williams et al., 2004). This distinct isotopic difference makes water isotope analysis an interesting diagnostic for a quantitative separation between the evaporation flux from bare soils, open water on the one hand, and plant transpiration on the other hand. The isotopic composition of water is commonly expressed as a fractional difference relative to standard, noted as "delta (δ)" notation in unit of permil (‰).

$$\delta = ((R_{\text{sample}}/R_{\text{standard}}) - 1) \tag{1}$$

For stable hydrogen isotopes, $R = {}^2H/{}^1H$ is the deuterium to hydrogen ratio in sample, or in a standard usually Vienna Standard Mean Ocean Water (VSMOW).

Early studies to measure transpiration using stable isotope measurements had been carried out on Eucalyptus trees by Calder et al. (1986, 1992); Calder (1992). They injected deuterated water (D) into tree roots and measured the transport of deuterium in stem water from roots to leaves as a tracer. Transpiration was calculated based on the total mass of tracer administered and the concentration in the stem water.

The isotope mass balance method is used to quantify the transpiration fraction. This method is based on a simple two-source mixing model, where evaporation (F_{ET}) is a sum of soil evaporation (F_{E}) and transpiration (F_{T}).

$$F_{\mathrm{ET}} = F_{\mathrm{T}} + F_{\mathrm{E}} \tag{2}$$

where F denotes a flux, and subscript ET stands for total evaporation, T for transpiration and E for soil evaporation. When each flux has its characteristic isotopic composition δ the isotopic mass balance is:

$$F_{\mathrm{ET}}\delta_{\mathrm{ET}} = F_{\mathrm{T}}\delta_{\mathrm{T}} + F_{\mathrm{E}}\delta_{\mathrm{E}} \tag{3}$$

Transpiration and evaporation fluxes can be derived from Eq. (3) according to (Zhang et al., 2010):

$$F_{\mathrm{E}} = \frac{\delta_{\mathrm{T}} - \delta_{\mathrm{ET}}}{\delta_{\mathrm{T}} - \delta_{\mathrm{E}}} F_{\mathrm{ET}} \tag{4}$$

$$F_{\mathrm{T}} = \frac{\delta_{\mathrm{ET}} - \delta_{\mathrm{E}}}{\delta_{\mathrm{T}} - \delta_{\mathrm{E}}} F_{\mathrm{ET}} \tag{5}$$

The total evaporation flux (F_{ET}) can be directly calculated using hydrometric methods as described above (Eddy covariance, Bowen ratio, lysimeter, etc).

Isotopic composition in each evaporation component can be obtained from direct measurements (e.g. using a bare soil lysimeter), and calculated using empirical methods. The Craig–Gordon formulation has often been used to calculate the isotopic composition of transpiration (δ_{T}) and soil evaporation or open water evaporation (δ_{E}; Craig and Gordon, 1965). The principle of this conceptual method is that the isotopic composition of the net soil evaporation or transpiration (extended Craig–Gordon model) flux can be derived as a function of environmental parameters (e.g. temperature, humidity). In this model, measurements of isotopic composition in the evaporating front of soil water (δ_e), isotopic composition in water vapor (δ_v), temperature, and relative humidity are required to calculate δ_E. For this method, the determination of the depth of the evaporating front and isotopic composition of water vapor are crucial. The soil evaporating front, which is defined as the transition zone between the liquid and vapor diffusion, can be determined from the profiles of soil water isotopic composition. This location (usually between 0.1 to 0.5 m, Barnes and Allison, 1983; Kendall and McDonnell, 1998; Clark and Fritz, 1997; Wang and Yakir, 2000; Sutanto et al., 2012) is clearly marked by the enrichment of heavy isotopes in the soil water above the front due to soil evaporation and depletion of heavy isotopes in the soil water bellow the front due to diffusion and capillary rise (see Fig. 2). This implies that detailed measurement of isotopic composition of soil water in vertical soil profile is needed. Isotopic composition of water vapor can be measured directly; sometimes it is assumed that δ_v is in equilibrium with precipitation.

Isotopic composition of transpiration can be estimated directly from measurement of water in leaves, or calculated based on an empirical method (e.g. Craig–Gordon method). Under steady state conditions, the isotopic composition of transpired water is assumed equal to the isotopic composition of plant source water, being stem water or soil water in the rooting zone. However, this assumption is generally not satisfied for short (hourly to daily) timescale and not valid under rapidly changing environmental conditions (Yepez et al., 2005; Lee et al., 2007; Xu et al., 2008; Zhang et al., 2010; Wang et al., 2012a). Some studies observed that a modeled transpiration under Steady State Assumption (SSA) is similar to measurements when stomata were relatively fully open. This condition is only met during the afternoon (Yepez et al., 2005; Farquhar and Cernusak, 2005; Lai et al., 2006; Zhang et al., 2010). Some methods exist to estimate isotopic composition of leaves water (δ_L) in non-steady state conditions. The modified steady state Craig–Gordon model, a non-steady state model proposed by Dongmann et al. (1974), and the Farquhar–Cernusak model (Farquhar and Cernusak, 2005) can be used to calculate δ_L (see also Zhang et al., 2010). Wang et al. (2010, 2012a) describe a method to measure δ_T using a customized leaf chamber in a 100 % N_2 atmosphere. This method requires a measurement of water vapor isotopic composition and water vapor concentrations to estimate the δ_L signal.

A method commonly used to estimate isotopic composition of total evaporation (δ_{ET}) is the Keeling plot approach (Keeling, 1958). It expresses a mass balance relationship by plotting the isotopic values of air samples at different heights above the ground (δ_V) against the inverse of concentration of the substance of interest e.g. water vapor mixing ratios (Yepez et al., 2003; Xu et al., 2008; Zhang et al., 2010). This results in a linear relationship where the vertical intercept reflects the isotopic composition of total evaporation (δ_{ET}). This method has three assumptions: (1) there is no loss of water vapor from the system apart from turbulent mixing with the atmosphere, (2) the atmospheric concentration of vapor in the system combines the input from background vapor and an additional component from local evaporation, and (3) the contribution of the two sources does not change during measurements. Again, this method assumes steady state for δ_{ET} over the sampling period, which is not always valid in nature.

2.3 Evaporation components in global land-surface models

In land-surface models, a simple water balance method is commonly applied to calculate surface fluxes. Evaporation as one of the water balance components is calculated as a sum of soil evaporation, transpiration and canopy interception. Each of these fluxes can be estimated separately using a collection of methods. The most common methods to calculate soil evaporation and transpiration are the ones developed by Priestley–Taylor (Priestly and Taylor, 1972) and Penman–Monteith (Monteith, 1981), respectively. Horton's model adopted for partial canopy cover can be used to calculate interception (Horton, 1919). All these methods require many parameters such as climatological and vegetation parameters, which can be obtained from the satellite data and measurements. An example for this partitioning work can be seen in Choudhury and DiGirolamo (1998).

The development of remote sensing technology has improved the representation of vegetation in these models. In recent land-surface models, LAI (Leaf Area Index) is an important component in formulating soil evaporation, transpiration and canopy evaporation. Lawrence and Slingo (2004) described in detail the use of LAI in the land-surface model MOSES2 (Met Office Surface Exchange Scheme) to estimate individual evaporation components. In this model, the distribution of soil evaporation and transpiration is controlled by two MOSES2 extinction parameters, which are the extinction coefficient for photosynthetically active radiation (k_{par}) and a shading factor controlling the fraction of the surface that is exposed to the atmosphere above the canopy (k_{sh}). Adjusting these two parameters leads to adjusting the transpiration fraction. Increasing k_{par} means decreasing the fraction of light for photosynthesis and therefore will decrease the transpiration fraction. On the other hand, decreasing k_{sh} increases the fraction of surface exposed to the atmosphere and therefore will increase the contribution of soil evaporation. The precise values for these two parameters to use in land-surface models remains a source of uncertainty (Lawrence and Slingo, 2004).

Another attempt to separate the different evaporation terms is applied using the CLM3 model (Community Land Model Version 3 Collins et al., 2006). Lawrence et al. (2007) noted that this model partitions the global evaporation fraction in an un-realistic manner, giving 13 % ET (evapo-transpiration), 44 % ES (soil evaporation) and 43 % EC (canopy evaporation). In their study, they tried to improve the fraction of transpiration in the CLM3 model by modifying several parameters and performing a sensitivity analysis. These efforts increase the final transpiration fraction in the CLM3 model from 13 % to 44 %. However, this value is still somewhat lower than results from the Global Soil Wetness Project 2 (GSWP2; Dirmeyer et al., 2006), which calculates a transpiration fraction of 48 %.

3 Some results from past partitioning works

Several studies partition the evaporation fraction into basically soil evaporation and transpiration. Only few studies take into account the evaporation from intercepted water. Here, we summarize some partitioning studies from the three different methods for comparison (Fig. 1). It is seen that in general, hydrometric and isotope-based methods give higher transpiration fraction values than land surface models. On average, hydrometric method calculates transpiration fractions exceeding 50 % while isotope-based method produces transpiration fraction higher than 70 %. Global land models estimate the transpiration fraction to be less than 50 %. However, not all results represent global annual averages. These studies have been carried out for different surface types, different climatic conditions, and different seasons. The global modeling methods give global annual averaged results that are lower than the contribution of transpiration fraction from total evaporation derived from the isotope-based method from Jasechko et al. (2013) (80–90 %).

The portion of transpiration varies as function of many factors such as radiation, air temperature, air humidity, wind, soil water content, crop characteristic, etc. For cultivated land, also the crop development stage, environment, irrigation practice, and crop management all strongly influence the transpiration rate. Under optimal conditions the transpiration flux may reach more than 90 % of the total evaporation, but it can drop to values lower than soil evaporation when plants are in a dormant or under less favorable conditions (Robertson and Gazis, 2006; Cavanaugh et al., 2011).

This temporal variation has been identified in many studies (including isotope-based partitioning studies), which show that transpiration during the growing season or summer is a dominant water flux compared to other fluxes, in general more than 70 % of the total evaporation (Lee et al., 2010; Wang et al., 2010, 2012a, b; Wenninger et al., 2010; Zhang et al., 2011; Sutanto et al., 2012), with some studies (Robertson and Gazis, 2006; Xu et al., 2008) reporting transpiration fractions exceeding 90 % of the total evaporation flux. For crops (Allen et al., 1998), the transpiration flux may be more than 90 % of the total evaporation at full crop cover. However, during autumn (September–November), the transpiration fraction drops to below 30 %, the remaining moisture flux being soil evaporation (Robertson and Gazis, 2006).

A seasonal cycle of hydrological recycling rates modulated by surface evaporation is evident over many of the mid-latitudes regions (Dirmeyer and Brubaker, 2007). At higher temperatures, the fractionation processes associated with evaporation and condensation lead to higher isotopic enrichment of precipitation during summer than winter, known as "the temperature effect". In Europe, for example, intensified recycling in spring/summer with a comparably enriched isotope value, and a reduced contribution of local evaporation to the atmospheric water vapor during autumn/winter is prominently visible in the water isotopologues. In summer, a large evapo-transpired fraction leads to a small spatial isotope enrichment gradient from the Atlantic coast to the interior of the continent. A steeper spatial gradient of depleted heavy isotopologue in winter appears from a strongly reduced contribution of continental water sources (evaporation fluxes) to the incoming water vapor masses (Rozanski et al., 1982; Hoffmann et al., 1998).

Different plant types exhibit a different transpiration fraction under similar climatic condition. Examples have been reported in some isotope studies in US, Europe and China. Studies from Yepez et al. (2003), and Ferretti et al. (2003) in Arizona and Colorado show that in summer, the transpiration fractions from Savanna woodland and grass are 85 % and 76–87 %, respectively. In Washington, partitioning results in early summer from Robertson and Gazis (2006) give a transpiration fraction of 57–88 % for a steppe-forest region. In China during summer, the maximum transpiration fractions of oaks and wheat are 96 and 80 %, respectively (Xu et al., 2008; Zhang et al., 2011). Hydrometric methods result in much lower transpiration fractions in Arizona US. A study from Cavanaugh et al. (2011) during summer in Shrubland area partitions transpiration fraction of 42–47 %. This is very low compared to an isotope-based study (85 %) in the same region although different plant types are examined. In Europe during summer, results from an isotope-based method to partition transpiration fraction from grass and a hydrometric method to partition transpiration fraction from maize are very comparable, yielding 78 % transpiration fraction for grass (Sutanto et al., 2012) and 77–97 % for maize (Herbst et al., 1996). Sutanto et al. (2012) also shows that the total evaporation fluxes analyzed from isotope-based and hydrometric methods are similar (3.4 and 3.5 mm d^{-1}, respectively). A comparison of isotope-based and hydrometric methods in an olive tree site is presented by Williams et al. (2004). Their study shows an agreement between two approaches for the midday period (10.30 a.m. to 12.30 p.m.). However, slightly different values are found for late-afternoon period (14.30 p.m. to 16.30 p.m.), with the differences of 4 % for transpiration and 15 % for soil transpiration.

For global scale partitioning analysis, only few studies separate the transpiration fraction from the evaporation. The study from Jasechko et al. (2013) is the first one that reports a global transpiration fraction based on an isotope approach. This study concluded that the transpiration fraction is 80 % up to 90 %. However, this may not necessarily be representative for all continental areas since the calculation is derived from the isotopic composition of lake water. A recent compilation study from Schlesinger and Jasechko (2014) results only 61 % (±15 % s.d.) transpiration fraction. Coenders-Gerrits et al. (2014) estimates the transpiration fraction of 35–80 % globally using another choice of input data from Jasechko et al. (2013) method. These two recent studies indicate that the estimate of transpiration fraction 80–90 % is far too unrealistic. Existing model and field estimates of the transpiration fraction do not suggest values higher than 80 % (GSWP-2, CLM3). Based on observations, Haverd et al. (2013) estimates that 50 % of the Australian rainfall is lost as bare soil evaporation. A relatively high transpiration fraction of 80 % is simulated by GLEAM (Global Land-surface Evaporation: the Amsterdam Methodology; Miralles et al., 2011).

This systematic difference between isotope-based estimates and models is supported by a local study from Sutanto et al. (2012) who compared results from an isotope mass balance method to a soil–water model (the HYDRUS-1D model) in a grass-covered lysimeter site. The soil water model estimates the transpiration fraction to be systematically lower than calculated with the isotope mass balance method (soil model:

64 %; isotope method: 78 %). Other field measurement programs estimate the fraction of transpiration to be around 52 % globally, 65–76 % for forest and 60 % for grasses (Choudhury and DiGirolamo, 1998; Blyth and Harding, 2011).

4 Possible discrepancies

What can explain these systematic discrepancies between the isotope and non-isotope methods? Each technique has its own limitations and these may lead to a high bias in the results. The main source of inaccuracy in lysimeter methods is the edge-flow of water that can occur in the contact area between bucket and soil. This produces significant errors in hydraulic conductivity calculation, infiltration rates, and saturated soil at the bottom of the lysimeter. Sapflow measurements suffer from leaking through the contact between tree and probes, probe misalignment, geometry of stems, the spatial variation of thermal conductivity, and calibration difficulties. Some studies (Allen et al., 2011; Burba and Anderson, 2007; Billesbach, 2011; Nagler et al., 2005) review the errors in Eddy covariance techniques, which include the assumptions used in calculation, instrument problems, frequency response, sensor time delays, noise, instrument calibration, etc. For the Bowen ratio method, a main source of uncertainty is the limited instrument accuracy to detect (sometimes very small) gradients (Todd et al., 2000; Nouri et al., 2013; Herbst et al., 1996).

For the isotope approach, the assumptions underlying the Keeling plot approach used to calculate isotopic composition of evaporation as described in Sect. 2.2 are rarely met perfectly in nature. The isotopic composition of water vapor is not constant during the period of interest, which can be caused by e.g. loss of water due to condensation, influence of air mass advection, entrainment at the boundary layer, etc. The steady state assumption (SSA) is shown to be fairly valid during midday hours only and is not satisfied in many field conditions when canopy climatic conditions are highly variable (Dongmann et al., 1974; Flanagan et al., 1991; Farquhar and Cernusak, 2005; Yepez et al., 2005). According to this assumption, the isotope signature of the source water will not be altered by transpiration or by canopy and litter interception from low vegetation such as grass. However, Gehrels et al. (1998) showed that the isotopic composition of soil water in forested areas is heavier than in non-forested areas, where the isotope signature of soil water is similar to the respective precipitation. Frequency and amount of rain will effectively modify the fractionation of interception and soil water. Over vegetated areas precipitation is partly intercepted by the canopy where it is exposed to open evaporation and accompanying fractionation. If followed by a new shower before the interception water is evaporated entirely, the enriched interception water is washed off and may cause an enrichment of soil water similar to a situation without the existence of a canopy with large leaves. In addition, the SSA in the Craig–Gordon formula leads to an overestimation of the isotopic composition of leaf-water compared to the measurements, which becomes higher as the transpiration rate increases (Flanagan et al., 1991; Lai et al., 2006). A failure to calculate the isotopic enrichment of leaf-water precisely will produce a bias in the transpiration fraction analysis. Therefore the assumption of steady state non-fractionating transpiration fluxes and a fractionation of all remaining surface fluxes should be critically reassessed.

The difference in the effective fractionation of the water isotopologues between transpiration on one hand and evaporation from bare soils and lakes on the other hand is key for the isotopic method to deliver quantitative constraints. Several studies (Yepez et al., 2005; Lai et al., 2006; Lee et al., 2007; Cuntz et al., 2007; Ogée et al., 2007) have shown that both fluxes (from non-fractionating vegetation and fractionating soils and open waters) show a strong seasonal and diurnal variability. For instance, in a dry season vegetation with deep roots has access to water from deeper soil layers or groundwater (Moreira et al., 1997; Wang and Yakir, 2000; Lee et al., 2007). The isotopic composition of groundwater depends on the isotopic input during the main recharge season, i.e. it is more enriched during summer than winter. Due to the fact that groundwater recharge is usually stronger in winter than in summer, the deep water will generally be more depleted than the upper layer/near surface water. Thus, the assumption that is often used in the isotope-based method, stating deeper water reflects the isotope composition of precipitation, is not necessary true. In addition, near surface water (accessible to short rooted plants) has often undergone isotopic enrichment by evaporation from bare soils. Hence extensive transpiration from vegetation with deep roots will generally result in an atmospheric water vapor signature that is more depleted of heavy isotopes than transpiration from superficial sources (see Fig. 2), and may lead to an overestimation of the transpiration fraction when this is not taken into account. In isotope-based methods the isotopic values of transpiration usually only uses isotopic composition of shallow soil water as source water.

Such effects might also contribute to the spatial patterns of isotopes in atmospheric water vapor that are becoming available by satellite-based remote sensing techniques. Water vapor over the Amazon basin, for example, shows less deuterium enrichment than over tropical Africa in two independent data sets, TES (Worden et al., 2012) and SCIAMACHY (Frankenberg et al., 2009, Fig. 3). This evidence may relate to the contribution of shallow and deep soil water to the transpiration flux: rain forest in the Amazon has access to deeper ground water reservoirs (Miguez-Macho and Fan, 2012) than most vegetation types in Africa.

Another factor that may contribute to a bias in the calculation of the transpiration fraction using the isotopic composition of open water (e.g. lakes) is the assumed similarity of isotopic enrichment processes for soil evaporation and open water. Haverd et al. (2011) found that after the first phase of soil drying, the value of the isotopic enrichment of evaporated soil water rapidly reaches a steady state value, which is approximately equal to that of the transpired water vapor. So the top soil layer may act as an isotopically enriched conduct for water from deeper layers without strong overall fractionation between the soil water and evaporated water, similar to the enrichments in leaves during transpiration. This does not apply to the isotopic enrichment of lakes.

Obviously, land surface model based estimates can be strongly biased due to systematic errors and poor representation of relevant processes. The representation of tiling bare ground vs. vegetated fractions in models deviates strongly from the true organization of vegetated or bare fractions of natural surfaces. Litter or moss layers are often missing, and the coarse vertical discretization of the soil does not allow steep moisture gradients near the surface. The relatively low transpiration fraction in current state of the art land surface models (Dirmeyer et al., 2006) calls for a reassessment of the skill of the underlying model parameterizations. A proper physical interpretation of (tuned) model parameters is still a difficult task (Lawrence and Slingo, 2004).

5 Conclusions

Different approaches exist to infer the transpiration fraction of the total continental evaporation flux. In general, estimates based on water isotope balance calculations tend to allocate a fairly large contribution of transpiration to the total moisture flux (generally exceeding 70 % of the total moisture fluxes). These results are at or above the upper estimates of the transpiration fraction using other, isotope-independent techniques (models, in situ physical flux data). The comparability is limited due to incomplete temporal and spatial coverage of most isotope-based analyses. However, a few studies that compare estimates of evaporation at the same location and conditions using the isotope-based and hydrometric methods show that the results are in fairly good agreement. The relatively low transpiration fraction in land-surface models shows that a reassessment of model parameterizations must be carried out.

The lack of global evaporation partitioning estimates using the isotope-based method represents a challenge and validation of global isotope-independent techniques using isotope-based results remain a challenge. Continued measurements of global isotopic composition of soil water and water vapor around leaves are needed to put the isotope-based results into context. Isotope enabled land-surface and global climate modeling studies such as iPILPS (isotopes in the Project for Intercomparison of Land-surface Parameterization Schemes; Henderson-Sellers et al., 2005), and SWING (Stable Water Isotope Intercomparison Group; Noone, 2007) allow the evaluation of the parameterization of land surface models by comparing the computed water isotopologue signals in the atmosphere with the available remote sensing and flux-based data sets.

A validation of modeled water isotoplogue signals in the atmosphere is needed. We suggest that future studies should perform this evaluation in order to provide a better understanding of the hydrological cycle in vegetated regions.

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Table 1. Summary of hydrometric methods (Adopted from Nouri et al., 2013).

Methods	Advantages	Disadvantages
Sapflow	– direct measurement of transpiration – cheap – manual or automatic measurement – individual plant coefficient can be determined – accurate technique to study plant eco-physiology and water scheduling	– point-based measurement – representative only for one vegetation type – some errors may occur from the space between probes, the variable geometry of stems, the variation of heat ratio, and uncertainty in measuring soil and understory evaporation
Eddy covariance	– direct method to study mixed vegetation – suitable for large areas – feasible to predict evaporation and no need for specific evaporation equation for different species	– expensive – need well-trained operators in electronics – need a monitoring tower above canopy and a uniform fetch
Bowen ratio energy balance	– simple and cheap – can measure evaporation even from non-watered plantation surface – proved to be a robust method in many forestry areas	– need uniform fetch for accurate measurement of net radiation and soil heat flux – need an adequate elevation above the canopy to collect the required meteorological data – sensitive to the bias of instrument gradient and energy balance
Lysimeter	– easy for inspection – manual or automatic measurement – high accuracy for undisturbed soil	– point-based measurement – very sensitive to different vegetation conditions inside and outside lysimeter – not practical for mixed vegetation types at large spacing – must consider plant root development – may have edge-flow effect – difficult to install and maintenance

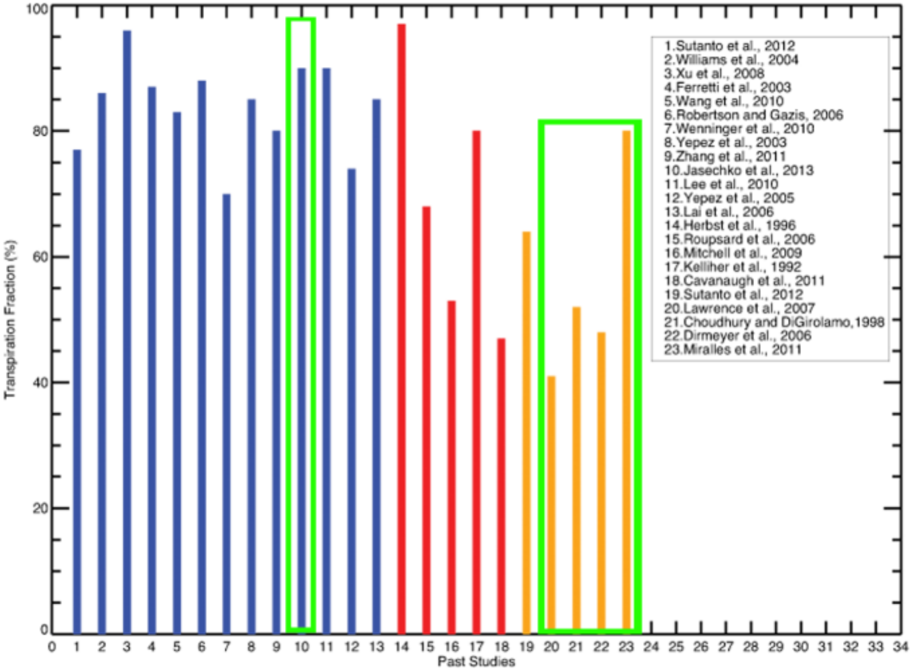

Fig. 1. Transpiration fraction from several past partitioning studies. The plot shows maximum values of transpiration fraction from those studies. Blue color is for isotope-based methods, red color is for hydrometric methods, and orange is for land-surface model results. The studies inside the green boxes present global average values.

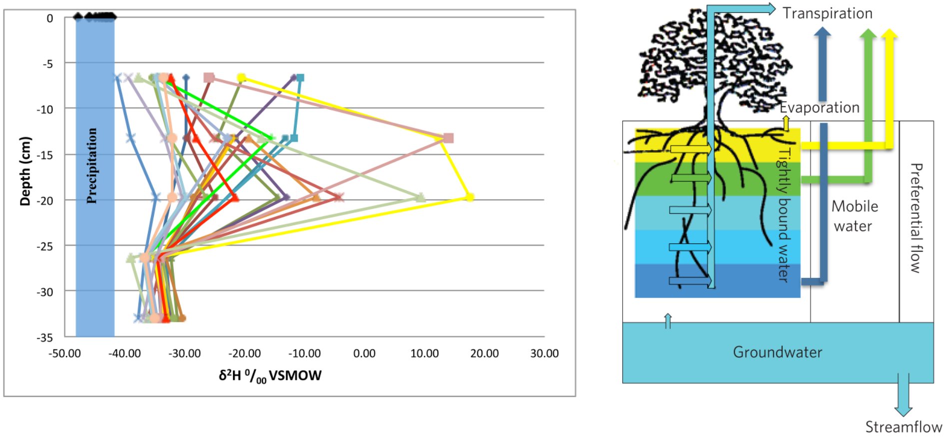

Fig. 2. (Left): Soil water isotope profiles and evaporating front in a lysimeter experiment from a collection of soil samples from Sutanto et al. (2012); Different colors represent different sampling periods. (Right): A transpiration illustration from Brooks et al. (2010); Different colors indicate different soil water isotopic values, with enriched isotopic values close to the surface and more depleted isotopic values in the bottom layers.

Title Page
Abstract
Introduction
Conclusions
References
Tables
Figures
◀▎
▶▎
◀
▶
Back
Close
Full Screen / Esc
Printer-friendly Version
Interactive Discussion

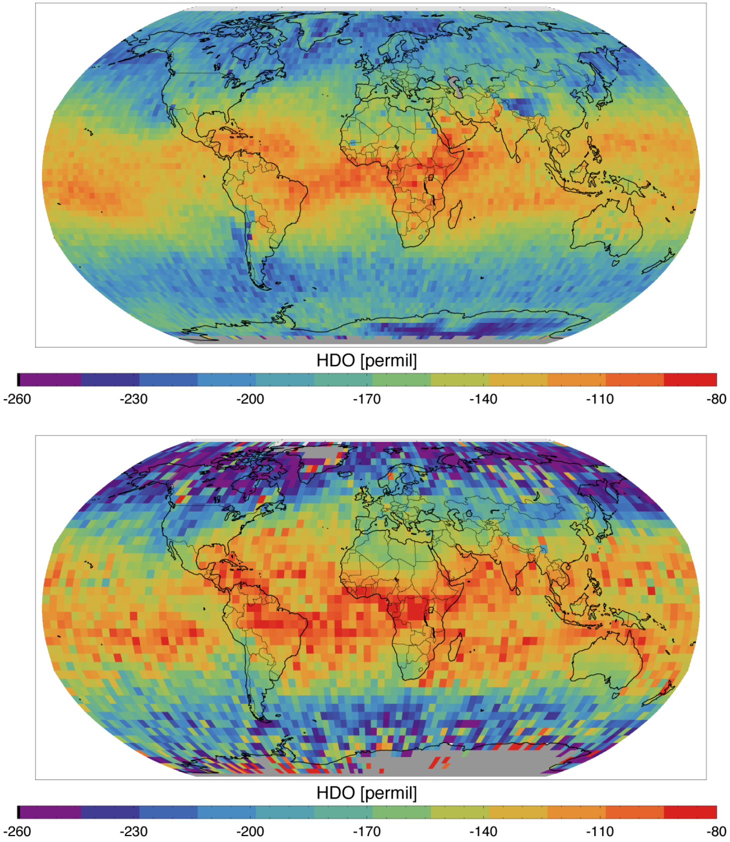

Fig. 3. Water vapor isotopic composition (δD) measured from TES (version 5 year 2006, weighted averaged from 925 to 450 hPa, left) and SCIAMACHY (average over 2003 to 2005, total column, right) satellites. Red colors indicate an enrichment in heavy isotopes and blue and purple colors indicate a depletion in heavy isotopes.