ENERGY
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Evaporative Processes on Vegetation: An Inside Look

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Abstract

While evaporation is the largest water consumer of terrestrial water, its importance is often (limitedly) linked to increasing crop productivities. As a consequence, our knowledge of the evaporation process is highly biased by agricultural settings, and results in erroneous estimates of evaporation for other land surfaces and especially for forest systems. The reason why crop and forest systems differ has to do with the vegetation height and what is happening in the space between the plant top and surface. Forests are multi-layered systems, where under the tallest tree species, lower vegetation layers are present. These lower vegetation layers transpire, but at a different rate then the main vegetation, since the atmospheric conditions are different under the canopy. Additionally, the sub-vegetation layers, and also the forest floor, intercept water. Next to different atmospheric conditions per layer, the interception process is highly complex due to differences in interception capacity and a time delay caused by the cascade of water when water flows from the top canopy down to the forest floor. Lastly, forests also have the capacity to store heat and vapor in the air column, biomass, and soil. While this energy storage can be up to 110 W/m^2 it is often neglected in evaporation models. To get a better understanding of what is happening inside a forest, for the purpose of evaporation modeling, we should make use of new sensing techniques that allow identifying the rainfall, energy, and evaporation partitioning. This will help to improve evaporation estimates for tall vegetation, like forest, and allow spatial up scaling.

Keywords

Evaporation • Heat and vapor storage • Remote sensing • Interception • Forest

3.1 Evaporation: Farmers' Wisdom or not?

Evaporation is, after precipitation, one of the largest fluxes of the water balance: globally 55–80% of the annual rainfall evaporates from the land surface (Gleick 1993). Nonetheless, hydrologists historically tend to focus on the relationship between rainfall and streamflow and consider evaporation as a residual flux (Harrigan and Berghuijs 2016). The result of this strong focus on rainfall–runoff relations, combined with the difficulties of measuring evaporation at the right temporal and spatial scale, is that knowledge on evaporation is underdeveloped (Brutsaert 1986; Oki 2006; Zhao et al. 2013). For agricultural areas this knowledge gap is smaller. Since farmers want to optimize crop production, information on crop behavior in relation to atmospheric conditions, soil moisture conditions, and supplied irrigation is required. Therefore, many extant evaporation studies focus on (well-watered) crops and they form the basis of many evaporation equations that are still used to date (e.g., Doorenbos and Pruitt 1977; Hargreaves and Samani 1982; Monteith 1965; Priestley and Taylor 1972). These crop-derived relations are, after some minor adjustments, used for other land surfaces as well and directly or indirectly incorporated into models that provide evaporation estimates (e.g., Allen et al. 1998; de Bruin and Lablans 1998; Konukcu 2007; Thom and Oliver 1977; Wright 1982). This approach seems to work reasonably well for most short vegetation covers

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J. T. Van Stan, II et al. (eds.), *Precipitation Partitioning by Vegetation*,
https://doi.org/10.1007/978-3-030-29702-2_3

but not for forested ecosystems, as will be shown on the basis of some studies that validate their evaporation estimates with other (independent) evaporation estimates. For these purposes, we distinguish two types of evaporation models: hydrological models and meteorological (RS) models (where RS refers to remote sensing, since these models often use RS data as input).

To assess the performance of these models, preferably independent "ground truth" data is used as a benchmark. Eddy covariance (EC) systems are currently seen as the best method to continuously measure evaporation (Wang and Dickinson 2012) and are used worldwide, e.g., FLUXNET (Baldocchi et al. 2001). EC-systems should be installed far above the vegetated surface (e.g., on a tower) and links high frequency measurements (>20 Hz) of water vapor and CO_2 concentrations to deviations in vertical wind velocity to estimate an ecosystem-scale flux. Depending on the slope, wind speed, and direction, the upwind area (i.e., footprint) where vapor originates from varies, which is problematic when the land cover is not uniform. Although it is commonly acknowledged that EC-systems have problems with varying footprints (Mu et al. 2011) and the non-closure of the energy balance (Stoy et al. 2013; Twine et al. 2000; Wilson et al. 2002), they are frequently used for calibrating and validating evaporation models as it is the best method available. Another frequently used method to assess model performance is to cross-compare evaporation estimates. Hence, in this case we compare the outcome of hydrological and meteorological models with EC observations and each other.

- **Hydrological and meteorological models versus EC**
  Morales et al. (2005) compared four process-based models (RHESSys, GOTILWA+, LPJ-GUESS, and ORCHIDEE) to EC observations in 15 European forests. They looked both at the water and the carbon fluxes and concluded that model performance varied greatly per location (RMSE[1]: 10–100 gC m^{-2} $month^{-1}$ and 50–300 mm/month, respectively) and that there was not a universal model that performed for all cases. Furthermore, they found that frequently the models overestimated the latent heat flux by a factor 1.3–2 (±20–80 W/m^2). This overestimation for forests was also found by Wang et al. (2015), who compared (among others) evaporation estimates of a VIC-model (Liang et al. 1994) to EC observations for different land types. For most forests an overestimation of a factor of 1.1–1.7 was found. In Ershadi et al. (2014), the remotely sensed SEBS model (Su 2002), Priestley and Taylor, Penman-Monteith (Brutsaert 2005), and an advection-aridity model (Parlange and Katul 1992) were compared with EC-data for different land cover types. They found that all models overestimated the evaporation flux (RMSE: 64–105 W/m^2) and that for different model performance metrics, all models performed worst for evergreen needle forest in comparison to, e.g., grassland or cropland. Also Hu et al. (2015) found for Europe that the operational MOD16 (Mu et al. 2011) and LSA-SAF MSG Eta (Ghilain et al. 2011) models performed best for crop- and grassland (RMSE: 0.47–0.72 mm/day) and overestimated the evaporation flux in complex canopies in summer (RMSE: 0.34–1.57 mm/day). Similar results are found by Ha et al. (2015), who found large uncertainties (R^2: 0.60–0.84) in four pine forests in the USA and showed that most models overestimate evaporation (RMSE: 15–23 mm/month). More recent Land Surface Models (LSMs), who implemented complex modeling schemes to model the water and energy fluxes, still show large uncertainties for forests. For example JULES (Best et al. 2011) compared their results to 10 FLUXNET sites and found that JULES overestimates evaporation in temperate forests with a RMSE varying between 15 and 30 W/m^2 (Blyth et al. 2011). Similar results are found by the CLM4 land surface model (Lawrence et al. 2011), where forest had on an hourly basis a RMSE of 34–49 W/m^2.
- **Intermodel comparison**
  Several hydrological models show discrepancies between simulated evaporation estimates and the SEBAL-algorithm (Bastiaanssen et al. 1998) for forests. For example, Immerzeel and Droogers (2008) compared (monthly) evaporation estimates of a SWAT model (Arnold et al. 1998) for a catchment in India and found the largest bias for (evergreen) forests (bias of –50 to 100 mm/month and average 40 mm/month). Similar results are found by Schuurmans et al. (2011), who ran a coupled groundwater and unsaturated zone model (MetaSWAP) for the Netherlands and found differences up to 4–5 mm/day for forests in comparison to 0–4-mm/day for other land classes. And Winsemius et al. (2008) tried to constrain the model parameters of a semi-distributed conceptual HBV-like model (Bergström 1995) with the SEBAL-algorithm and found that for forested areas this was difficult, indicating that forest systems are likely not yet modeled correctly. For testing the performance of Land Surface Models, special benchmarking platforms have been developed (e.g., ILAMB (Luo et al. 2012), PILPS (Pitman 2003), SUMMA (Clark et al. 2015a), CLASS (Verseghy et al. 1993)). As mentioned before, LSMs try to represent many biophysical and hydrologic processes. The downside of this is that parameterizing these LSMs becomes difficult, and therefore PILPS was initiated. But also in PILPS they found larger deviations in latent heat of ±50 W/m^2 for forests in comparison to ±20 W/m^2 for grass (Henderson-Sellers et al. 1995).

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[1] RMSE: root mean square error.

Hence, overall we see the common observation that short vegetation and cropland is often reasonable well-modeled, while the latent heat flux above forests is not well-modeled and often overestimated in comparison to EC observations. The hydrological models (VIC, SWAT, MetaSWAP) vary between –45 W/m^2 and 142 W/m^2, the meteorological models (SEBS, MOD16) between 10 and 105 W/m^2, and the LSMs (JULES, CLM4, PILPS) have a discrepancy of ±50 W/m^2. Thus the "farmers wisdom" on crop evaporation clearly cannot be transferred one-to-one onto forest systems.

3.2 By the Way, Which "Evaporation" Do We Mean?

The causes for the errors and mismatches between major evaporation models and the observations can originate from conceptual errors in the models as well as drawbacks in the measuring technique. However, to understand the causes it is important to first clearly define what is meant by evaporation, since in the literature many misconceptions exist whether only transpiration is meant, or that interception and soil evaporation are included as well (Savenije 2004).

Here we define total evaporation (E_{tot}) as the sum of transpiration (E_t), interception evaporation (E_i), soil evaporation (E_s), and open water evaporation (E_o), all with dimension [L T^{-1}] (Shuttleworth 1993)

$$E_{\text{tot}} = E_t + E_i + E_s + E_o \tag{3.1}$$

Although transpiration is the most dominant evaporative flux (Coenders-Gerrits et al. 2014; Schlesinger and Jasechko 2014; Sutanto et al. 2014; Wei et al. 2017) interception evaporation should not be underestimated, and at times it may be the dominant flux, especially in forested areas (see e.g., Fig. 5 in Wang-Erlandsson et al. 2014). A literature review by Miralles et al. (2010) has shown that the canopy can intercept 8–34% of rainfall.

Interception is present in both crops and forests; however, it can already partly explain the mismatch between the evaporation models and EC observations. First, neglecting interception will lead to an underestimation of evaporation in hydrological models. The results found by Mueller et al. (2013), Liu et al. (2016) showed that especially wet catchments perform worse. Neglecting the interception process in hydrological models causes an underestimation of total evaporation if not compensated by an increase in transpiration. Since more water enters the unsaturated zone than in reality or by any calibration, transpiration can be overestimated (Van den Hoof et al. 2013). But that would mean the model is conceptually wrong, which has large consequences for studies dealing with, e.g., carbon exchange and climate and/or land use change models. The second reason for the mismatch is that, next to hydrological models that often overlook interception, many eddy covariance systems also ignore interception. Many EC-systems are open-path systems, which measure the gas concentrations in situ by an optical sensor (as opposite of closed-path systems that draw air through an intake tube and analyze the sample not directly at the sample location). These optical open-path sensors do not work properly if the optical path is obscured like in the case if they are wet (Hirschi et al. 2017), resulting in the evaporation shortly after a rainfall event not being observed so long as the open-path analyzer is wet.

However, as said, interception is both present in croplands and forests, so it cannot be the only reason why evaporation models show worse model performance for forest systems. Hence the question remains: why can we not use the farmers' wisdom to model forest evaporation? Are not trees basically supersized crops?

3.3 Why Don't Forests Act like a Giant Crop Field?

Yes, in a way trees are just supersized crops; but, only when one looks at the transpiration of the main tree species, and even then differences occur due to different water use strategies (rooting depth, crop rotation). In forests, the space between the canopy and the ground allows other vegetation species to grow (Fig. 3.1). These species transpire and intercept water; however, since the atmospheric conditions under the main canopy are different, quantifying its magnitude is not straightforward for understory and ground vegetation. On top of that, heat and vapor can be stored in the space between the canopy and ground, which affects the entire water and energy balance. Both the effect of additional understory and ground vegetation, and heat and vapor storage are not (or less) present in crop systems and explain (at least partly) why crop-concepts cannot be used directly for forest systems.

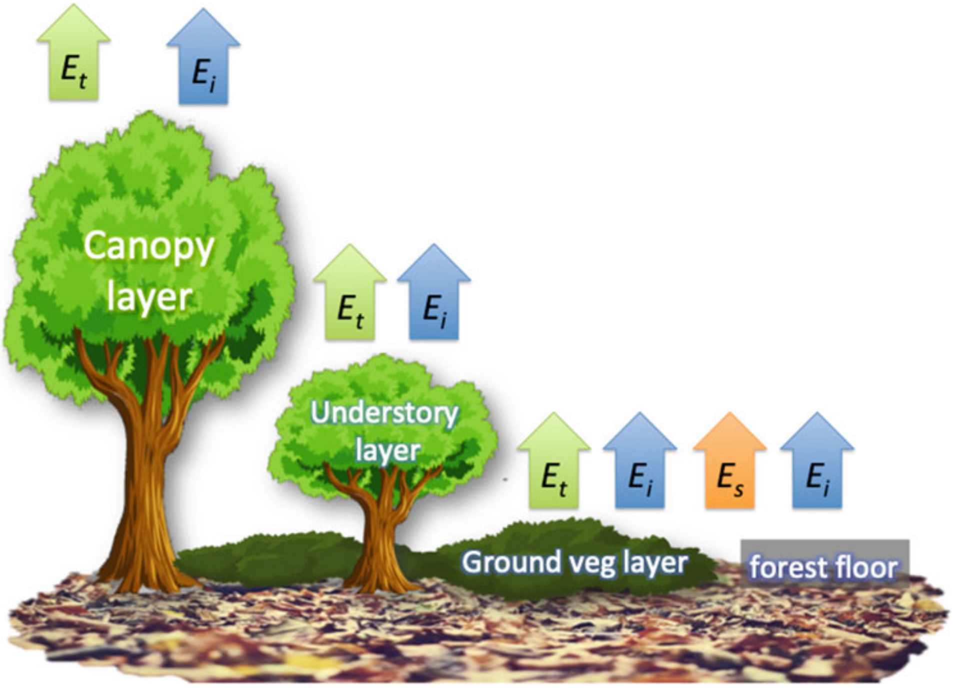

Fig. 3.1 Schematization of forest layering and its sources of transpiration (E_t), interception evaporation (E_i), and soil evaporation (E_s). *Note* forest floor interception evaporation is distinguished from soil evaporation by the fact that soil evaporation refers to water that is stored in the root zone (De Groen and Savenije 2006)

3.3.1 The Waterfall of Interception Storages

Canopy interception is often well modeled by Rutter-like models (Rutter et al. 1971; Valente et al. 1997), where the interception process (I) is modeled by a simple threshold model, whereby interception evaporation (E_i [L/T]) takes place as long as the interception storage (S [L]) is not emptied

$$I = E_i + \frac{\mathrm{d}S}{\mathrm{d}t} \tag{3.2}$$

Often the interception storage capacity (S) is derived from vegetation characteristics like LAI and the interception evaporation is a function of the potential evaporation.

If one wants to include the interception of the understory and forest floor as well, one cannot simply increase the interception storage capacity for two reasons. First, there is a sequence in the storages. Once a storm begins, the canopy “bucket” must be filled to initiate throughfall (including splash losses, see Chap. 12), then throughfall can fill the understory “bucket”, followed by the “bucket” of the ground vegetation and lastly the forest floor (Fig. 1.2). This filling and spilling, causes a cascade of interception storages, which causes a shift in time (Gerrits et al. 2010). For more details on water storage of vegetation see Chap. 2. Second, the potential evaporation below the canopy is lower than above. Radiation is less (Jarvis et al. 1976; Rauner 1976), wind is often reduced, and some energy is already consumed by evaporation of the intercepted canopy water, thus changing the air temperature and humidity the understory is exposed to. This lower potential atmospheric demand is often used to argue that forest floor interception is negligible; however, this lower atmospheric demand is compensated by the often-larger storage capacity of the forest floor (see values in Fig. 3.2) (Breuer et al. 2003; Bulcock and Jewitt 2012; Gerrits and Savenije 2011a, b; Kittredge 1948). This results in residence times of several hours to days for forest floor interception in comparison to less than an hour for intercepted canopy water (Baird and Wilby 1999; Gerrits et al. 2007, 2009; Li et al. 2017; Wang-Erlandsson et al. 2014). The interaction between the canopy and forest floor interception, also results in a reduced effect on the phenology. Often it is thought, that in winter time interception is zero for deciduous trees, because the trees do not have leaves; however, the leaves are on the forest floor where water can still evaporate (despite its low potential evaporation), because of the high water content (Gerrits 2010; Gerrits and Savenije 2011a, b; Van Stan et al. 2017).

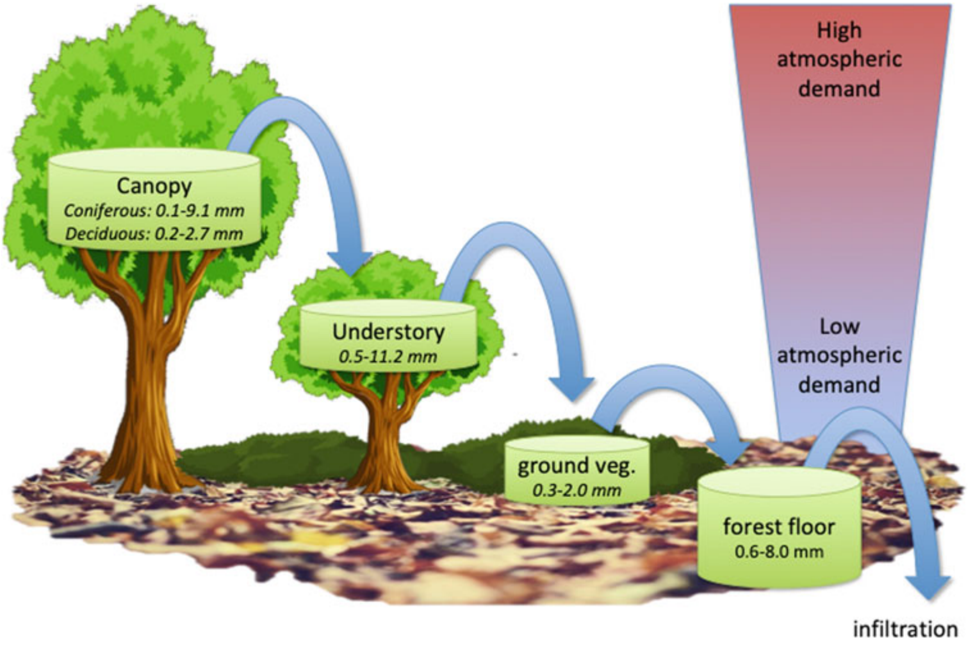

Fig. 3.2 Cascade of interception storages (S) in relation to atmospheric demand. Values in 'buckets' indicates minimum and maximum storage capacity as summarized by (Breuer et al. 2003) and (Gerrits and Savenije 2011a, b)

An elegant attempt to model both canopy as forest floor interception at the global scale, that also takes into account the reduced potential evaporation, has been done by Wang-Erlandsson et al. (2014). In their STEAM model they showed that globally for different forest types. 16–20% of the rainfall was intercepted by the canopy and 4–14% by the forest floor, resulting in a total interception of 22–34% of rainfall. For crop systems these values were much lower: 13%, 3%, and 16% for canopy, forest floor, and total interception, respectively. This indicates that in forest systems the below-canopy interception is more important in comparison to crop systems.

3.3.2 Energy Hotel

Remotely sensed evaporation products use different algorithms to estimate the latent heat flux, although they share some similarities. The basis of most products is the energy balance

$$R_n = \rho\lambda E + H + \frac{\mathrm{d}}{\mathrm{d}t}\left(\sum Q\right) \tag{3.3}$$

where R_n is the net radiation, $\rho\lambda E$ the latent heat (or evaporation expressed in W m^{-2}), H the sensible heat flux, and $\frac{\mathrm{d}}{\mathrm{d}_t}(\sum Q)$ the storage flux, all with unit (W m^{-2}). Generally, the $\sum Q$-term is set equal to the ground heat flux ($\mathrm{d}Q_g/\mathrm{d}t$). However, studies that investigated the non-closure of the energy balance of EC-systems, already indicated that only considering the ground heat flux is not sufficient. Following Oke (1987) two terms are missing: advective energy and a storage term. If we consider extensive forested areas, the advective energy is usually neglected, hence only the storage term remains. Foken (2008) and McCaughey and Saxton (1988) considered three types of storages:

- storage of heat and vapor in the air below the flux measurements (Q_h and Q_e, respectively),
- storage in the vegetation (Q_b), and
- energy required for photosynthesis (Q_p).

This was confirmed by several other studies (Mayocchi and Bristow 1995; Meyers and Hollinger 2004; Oliphant et al. 2004).

Hence, after neglecting advected energy and including the storage terms the energy balance above a forest at height z (m) can be defined as (e.g., Barr et al. 1994; McCaughey 1985):

$$R_n = \rho\lambda E + H + \frac{\mathrm{d}}{\mathrm{d}t}\left(Q_g + Q_h + Q_b + Q_p\right) \tag{3.4}$$

all with units W m^{-2}.

To estimate the first four storage terms, information on the thermal properties and state of the ground, air, and biomass is required. For the ground heat flux ($\mathrm{d}Q_g/\mathrm{d}t$) these are the ground temperature gradient ($\mathrm{d}T_g$, [K]) over depth ($\mathrm{d}z$, [m]) and λ the soil thermal conductivity [W K^{-1} m^{-1}] and z [m] the measuring depth and C the soil heat capacity [J m^{-3} K^{-1}] (Brotzge and Crawford 2003):

$$\frac{\mathrm{d}Q_g}{\mathrm{d}t} = -\lambda\frac{\mathrm{d}T_g}{\mathrm{d}z} + zC\frac{\mathrm{d}T_g}{\mathrm{d}t} \tag{3.5}$$

The heat and latent heat storage rates are defined as (Barr et al. 1994; McCaughey 1985)

$$\frac{\mathrm{d}Q_h}{\mathrm{d}t} = \int_o^z \rho_a c_p \frac{\mathrm{d}T_a}{\mathrm{d}t}\mathrm{d}z \tag{3.6}$$

$$\frac{\mathrm{d}Q_e}{\mathrm{d}t} = \int_o^z \rho_a \lambda \frac{\mathrm{d}q}{\mathrm{d}t}\mathrm{d}z \tag{3.7}$$

with ρ_a the density of air [kg m^{-3}], c_p the specific heat of air [J kg^{-1} K^{-1}], T_a air temperature [K], λ the latent heat of vaporization [J kg^{-1}], q the water vapor mixing ratio [kg kg^{-1}], and t the time [s].

And similar for the biomass heat storage rates (McCaughey and Saxton 1988):

$$\frac{\mathrm{d}Q_b}{\mathrm{d}t} = \int_o^{h_c} \rho_b c_b \frac{\mathrm{d}T_b}{\mathrm{d}t}\mathrm{d}z \tag{3.8}$$

where ρ_b the density of the biomass [kg m^{-3}], C_b the specific heat of biomass [J kg^{-1} K^{-1}], T_b biomass temperature [K].

The last storage term in Eq. 3.4 is related to the energy used for photosynthesis Q_p is estimated as ±422 kJ per mole fixed CO_2 (Masseroni et al. 2014; Meyers and Hollinger 2004; Nobel 1974). Although Blanken et al. (1997) illustrated that Q_p can be 23% of $\sum Q$ on clear sunny days, it is often neglected, since it is difficult to measure and it was found to be less than 3% of $\sum Q$ in the middle of the day (Jarvis et al. 1976; Tajchman 1981; Thom 1975).

In Table 3.1 an overview is given of the magnitude of the other storage terms. The difficulty with the individual storage terms is that–unlike the sensible and latent heat flux– the storage terms do not follow the net radiation pattern. Only the ground heat flux is a percentage of the net radiation once a time lag is included; however, the biomass storage peaks before noon. And the sensible and latent heat storage are peaking just before sunrise, where the latent heat storage becomes already negative two hours after, and the sensible heat storage just before sunset (Lindroth et al. 2010; Oliphant et al. 2004).

Hence the space between the top of the canopy and the forest floor is like a hotel, where energy can be stored during the day. For EC-systems these storage terms are not the primary cause of incorrect evaporation estimates, since only wind and vapor information is used (it is only partly responsible for the non-closure of the energy balance (Foken 2008)). However, it is important for RS-products. Ignoring the storage terms implies that more energy is attributed to the latent and/or sensible heat. Especially, for forests the storage terms can be significant, since there is a large air column where heat and vapor can be stored in comparison to, e.g., crop or grassland. As shown in Table 3.1 these storage terms have the same order of magnitude or even bigger than the ground heat flux. This might then also explain why RS-algorithms compare better to ground observations in non-forested areas.

Table 3.1 Literature study overview from magnitude of daily minimum and maximum storage rate terms

Vegetation type	Height [m]	Country	dQ_g/dt [W m^{-2}]	dQ_h/dt [W m^{-2}]	dQ_e/dt [W m^{-2}]	dQ_b/dt [W m^{-2}]	dQ_p/dt [W m^{-2}]	$d\sum Q/dt$ [W m^{-2}]	R_{net} [W m^{-2}]	References
Eucalyptus forest	40	Australia	–10 to 48	–75 to 50	–25 to 25	–50 to 61	NA	NA	NA	(Haverd et al. 2007)
Maize crop Soybean	3 0.9	Illinois, USA	–15 to 25 –5 to 17	NA NA	NA NA	–5 to 20 0 to 7	10 to 25 5 to 10	±80 ±35	±500 NA	(Meyers and Hollinger 2004)
Maple, beech, oak	27	Indiana, USA	–10 to 30	–10 to 20	–5 to 0	–10 to 20	NA	–30 to 60	NA	(Oliphant et al. 2004)
Mixed forest	20	Ontario, Canada	–25 to 90	–30 to 45	–70 to 30	–10 to 17	NA	–60 to 110	–60 to 720	(McCaughey and Saxton 1988)
Mixed forest	26–30	The Netherlands	–5 to 15	–50 to 70	–40 to 30	NA	NA	–50 to 70[a]	–80 to 650	(Schilperoort et al. 2018)
Mixed pine and spruce	28	Sweden	–5 to 15	–15 to 15	–8 to 6	–20 to 22	NA	–35 to 45	–50 to 400	(Lindroth et al. 2010)
Sorghum	1.14	Texas, USA	–25 to 60	–5 to 5	–5 to 10	–10 to 10	–10 to 13	NA	–25 to 600	(Kutikoff et al. 2019)
Tropical forest	14–25	Brazil	–25 to 0	–20 to 40		–20 to 40	NA	–50 to 70[a]	–20 to 700	(dos Santos Michiles and Gielow 2008)
Tropical forest	35	Brazil					NA	–80 to 80[a]	NA	(Moore and Fisch 1986)
Young larch	10.6	Eastern Siberia	0–50				NA	–25 to 100	–50 to 550	(Tanaka et al. 2008)

[a] Excluding Q_g

3.4 Outlook

To improve our knowledge on forest evaporation we should invest in studying what is happening in and underneath the forest canopy, and not neglect the space where water, vapor, and heat can be stored and released. However, this is not easy to achieve, as often observation techniques are limited or extremely expensive (Arya 2001; Tajchman 1981). Some attempts have been made to measure turbulent fluxes of momentum, heat, and vapor directly within forests (Baldocchi and Meyers 1988; Bergström and Högström 1989; Verma et al. 1986). However, the vertical spatial resolution was limited to a few points in height, while the space between the top of the canopy and forest floor is highly variable (Allen and Lemon 1976; Arya 2001; Patton et al. 2010; Rauner 1976). On top of that, this space is also interacting in a complex way with the air above the canopy. So can it be that at certain times of the day vapor originating for the understory is simply transported vertically through the canopy, while at other times it is stored in this space or is transported horizontally and finds another way to the atmosphere, e.g., near a forest edge or gap. Meaning that sometimes the below and above canopy air are entirely decoupled from each other, and at other times turbulent exchange takes place (Alekseychik et al. 2013; Belcher et al. 2008; Göckede et al. 2007; Thomas et al. 2017). This complex turbulent behavior is difficult to model, and even Large-Eddy Simulations (LES), which are currently the best numerical tool to simulate this, has shortcomings in dealing with these complex flows (Dellwik et al. 2019; Patton et al. 2010). The importance of detailed information on turbulent fluxes follows from the work of Clark et al. (2015b) where they showed how sensitive their model was for changes in below-canopy wind parameters.

Fortunately, new opportunities have arisen with new sensing techniques. Distributed Temperature Sensing (DTS) is one of these techniques, whereby continuously (up to 1 Hz) temperature is measured at a high spatial resolution (0.25 m) along a fiber optic cable (Selker et al. 2006). As shown by Euser et al. (2014) and Schilperoort et al. (2018), DTS can also be used to measure vertical temperature and moisture profiles, from which the latent and sensible heat flux can be derived as well as heat storage. And more recently, Van Ramshorst (2019) and Sayde et al. (2015) showed the application of wind profile measurements, by actively heating the fiber optic cable like a hot wire anemometer. Combining the temperature, vapor and wind profiles allows studying turbulence fluxes of momentum, heat and vapor within the forest layer at a high spatial and temporal resolution.

Additionally, LiDAR-information can help to better understand forest structure to estimate turbulent flows (Boudreault et al. 2015), vegetation characteristics like LAI (Zhao and Popescu 2009), DBH, height (e.g., Brede et al. 2017), interception storage capacity (e.g., Berezowski et al. 2015; Roth et al. 2007) and/or the heat stored in the biomass. For the latter objective thermal infrared imagery might also be a possible tool (Garai et al. 2010; Pfister et al. 2010; Voortman et al. 2016).

In addition to looking at the turbulence structure within and underneath the canopy, knowing how evaporation is partitioned between transpiration, interception, and soil evaporation is a key element to improve understanding of forest evaporation processes (Blyth and Harding 2011; Dubbert et al. 2013; Lawrence et al. 2007; Van den Hoof et al. 2013; Wang and Yakir 2000). One of the main methods to achieve this is by means of stable water isotopes either sampled from the surface (Kool et al. 2014; Soderberg et al. 2012) or derived from satellites (Steinwagner et al. 2007; Sutanto et al. 2015). Stable water isotopes are considered to be ideal tracers because of their natural occurrence and their ability to distinguish water evaporated from the soil and/or wet surface (i.e. canopy or forest floor) and water that has been transpired (Ehleringer and Dawson 1992; Fekete et al. 2006; Gat 2010; Kendall and McDonnell 1998). The first process causes physical fractionation (kinetic), while with root water uptake this isotopic fractionation does not occur (Williams et al. 2004). Hence after reaching steady state, the isotopic signature will be similar to the soil water. This methodology appears to work rather well for both canopy as the forest floor (Giuditta et al. 2018; Griffis 2013; Moreira et al. 1997; Rothfuss et al. 2010, 2015; Sutanto et al. 2012; Wenninger et al. 2010); however, it is costly and laborious, has a low temporal resolution, and some of the model assumption are questioned (Farquhar and Cernusak 2005; Lai et al. 2006; Rothfuss et al. 2010; Sutanto et al. 2014). Fortunately, with current developments in isotope measuring devices like improved accuracy and direct air samplers, where uncertain cold trap systems become redundant (Jiménez-Rodríguez et al. 2018; Rhee et al. 2004), new opportunities arise to disentangle the various evaporation components. A great example of the added value of isotopes is the study of Wei et al. (2018), where they included isotopes information in a combined LSM-LES-Cloud Modeling System model.

Combining knowledge on rainfall, energy, and evaporation partitioning will help to model the complex system that is present from the top of the vegetation to the forest floor. This model can explain how heat, energy, and water are transported from the top of the canopy to the unsaturated zone and vice versa. In the end, this will lead to improved evaporation estimates for tall vegetation, like forests, and allow upscaling by means of (thermal) remote sensing algorithms that can only observe the top of the canopy.

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