Chapter 5
A Synthesis of Forest Evaporation Fluxes – from Days to Years – as Measured with Eddy Covariance

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5.1 Introduction and History

The annual water budget of a forested landscape is the sum of precipitation minus the sum of evaporation, runoff, storage, and leakage. The evaporation term, which is the subject of this chapter, comprises the sum of plant transpiration and evaporation from the soil/litter system and rainfall/dew intercepted by the foliage.

The literature on "forest evaporation" is vast; at the time of this writing, it contains over 1,100 references, according to a query of the Web of Science. Most of the long-term measurements (years to decades) on forest evaporation are based on forest catchment studies, which evaluate evaporation as a residual of the water balance (Swank and Douglass 1974; Bosch and Hewlett 1982; Komatsu et al. 2007) or by measuring changes in soil water balance and rain interception (Calder 1998). These budget approaches have merit in evaluating forest evaporation because they are relatively inexpensive and they can evaluate water budgets over long time periods, across large geographic areas, and in complex terrain. On the other hand, evaporation sums derived from hydrological water balances are limited in their ability to extract information on biophysical controls of forest evaporation on hourly and daily timescales. Water balance methods are also unable to provide information on the partitioning of evaporation according to transpiration and soil and re-evaporation of intercepted rainfall and dew.

Another segment of this literature uses micrometeorological techniques to produce direct measurements of forest evaporation. Rapid growth in the application of micrometeorological methods over forests occurred over the past 30 years because of its ability to measure fluxes of water vapor directly, in situ, at the stand scale and with minimal interference. But the majority of these studies and the many fine reviews and syntheses on the topic of "forest evaporation" using "micrometeorological methods" are confined to short campaigns during the heart of the growing season (Jarvis et al. 1976; Jarvis and McNaughton 1986; Black and Kelliher 1989; Kelliher et al. 1993; Komatsu et al. 2007; Tanaka et al. 2008). To our knowledge only the review by Tanaka et al. (2008) focuses on long-term evaporation measurements and it concentrates on evaporation from tropical forests.

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The earliest measurements of water vapor exchange between forests and the atmosphere relied on the flux-gradient method (an indirect technique that evaluates flux densities of H_2O as the product of a turbulent diffusivity (K) and the vertical gradient of H_2O concentration, dq/dz), rather than the eddy covariance technique, due to a lacking of fast responding anemometers and H_2O sensors (Denmead 1969; Droppo and Hamilton 1973; Stewart and Thom 1973; Black 1979). Application of flux-gradient theory over tall vegetation was found to be problematic at the onset (Raupach 1979). Over tall forests, vertical gradients of H_2O are small and difficult to resolve because turbulent mixing is vigorous at the canopy–atmosphere interface (Black and McNaughton 1971; Stewart and Thom 1973; Hicks et al. 1975). Secondly, use of Monin–Obukhov similarity theory to calculate eddy exchange coefficients (K) is invalid above forests (Raupach 1979). This occurs because turbulent transport is enhanced in the roughness sublayer over the forest – large shear at the canopy–atmosphere interface causes nonlocal transport to occur (Raupach et al. 1996). By the mid-1970s, additional studies on evaporation over forests would need to wait for technical developments that would permit use of the eddy covariance technique.

The earliest eddy covariance measurements of water vapor exchange over forests occurred between the mid-1970s and early 1980s (Hicks et al. 1975; Spittlehouse and Black 1979; Shuttleworth et al. 1984; Verma et al. 1986). This advance was made possible with a wave of technological improvements that included three-dimensional sonic anemometers, fast-responding ultraviolet hygrometers (krypton and Lyman-alpha) (Buck 1976), infrared spectrometers (Hyson and Hicks 1975; Raupach 1978), and personal computers. The execution of the ABRACOS project in Brazil (Shuttleworth et al. 1984) and the BOREAS project in Canada (Sellers et al. 1995) heralded a new era of routine and long-term measurements of evaporation from forests by eddy covariance. And today eddy covariance measurements of evaporation continue worldwide through various regional networks associated with the FLUXNET project (Baldocchi et al. 2001; Baldocchi 2008).

5.2 Forest Evaporation by the Eddy Covariance Method

The eddy covariance technique measures evaporation by assessing the covariance between fluctuations in vertical velocity (w) and the specific water vapor content ($q = \rho_v/\rho_a$where ρ_a is dry air density and ρ_v is H_2O density):

$$E = \overline{\rho_a} \cdot \overline{w'q'}. \tag{5.1}$$

In (5.1), the overbars denote time-averaging (e.g., 30–60 min) and primes represent fluctuations from the mean (e.g., $q' = q - \overline{q}$). A positively signed covariance represents net H_2O transfer into the atmosphere and a negative value denotes the reverse.

Many issues remain about the applicability and accuracy of eddy covariance measurements over forests. Of most concern are the many circumstances where investigators fail to close the surface energy balance (Twine et al. 2000; Wilson et al. 2002), which is used as an independent data quality check. Lack of energy balance closure, on the other hand, should not always indict the quality or the accuracy of the evaporation measurements. Mitigating factors include: (1) nonrepresentative measurements of the net radiation balance across the flux footprint; (2) biases in net radiation measurements via improper mounting of the sensor close to a tower; and (3) insufficient sampling of soil heat flux and canopy heat storage across the flux footprint (Meyers and Hollinger 2004; Lindroth et al. 2010). In fact, there is growing body of evidence showing good agreement between long-term evaporation measurements by eddy covariance with independent hydrologically based methods. Three studies report that annual sums of evaporation, based on eddy covariance, agree within 6% of independent assessments of forest evaporation; these have been produced by water budgets from catchments (Wilson et al. 2001; Scott 2010), deep groundwater piezometers (Barr et al. 2000), and changes in soil moisture profiles (Baldocchi et al. 2004; Yaseef et al. 2010). Furthermore, daily and annual integrations of eddy covariance water fluxes do not suffer from the night-time systematic biases that plague CO_2 flux measurements (Moncrieff et al. 1996).

5.3 Evaporation from Forests, Magnitudes, and Variations

Over the past decade, several hundred research teams commenced measuring fluxes of water, carbon dioxide, and energy continuously with the eddy covariance method. So, today, many forest evaporation datasets exist, with measurements accumulated over years to decades. Ironically, a small fraction of these research teams have found the time or inclination to publish their long-term evaporation measurements, compared to the several hundred papers that have been published on ecosystem CO_2 exchange (Baldocchi 2008). Nevertheless, there exists a substantial and growing body of literature on long-term forest evaporation, which we have compiled, that merits scrutiny. For this chapter we compiled and evaluated 185 site-years of forest evaporation measurements, derived with the eddy covariance method. These studies are associated with over 40 forests and include data from tropical evergreen broadleaved, temperate evergreen conifer, deciduous broadleaved forests, savanna woodlands, and shrublands. Below, we draw upon this compiled database and address the following questions relating to forest evaporation: what is the range of annual evaporation from the World's forests and woodlands? Is there a ranking among annual evaporation rates for different forest types, canopy structure, and climates? And how is the amount of annual evaporation constrained or related to annual precipitation and available energy?

Figure 5.1 summarizes the evaporation database, of 185 site-years, by presenting the probability density distribution of annual evaporation. The mean annual evaporation rate of forests from across the globe (and its standard deviation) is 503 ± 338 mm year^{-1}.

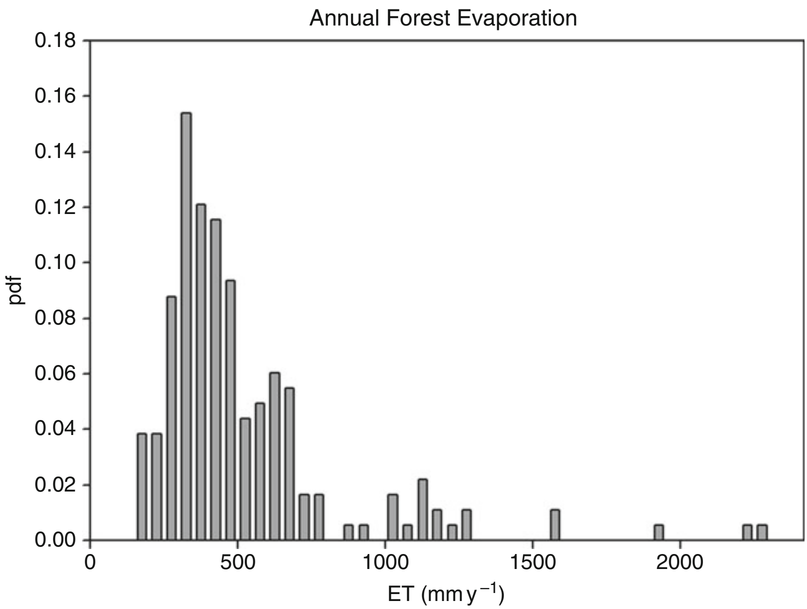

Fig. 5.1 Probability distribution of annual evaporation from forested sites. The probability density function (pdf) is derived from 185 site-years of eddy covariance flux measurements. The mean is 503 ± 338 mm year^{-1} and the median is 408 mm year^{-1}

The probability distribution is positively skewed towards high sums, as annual evaporation from tropical forests can exceed 2,000 mm year^{-1} (Loescher et al. 2005; Fisher et al. 2009). For perspective, these statistical values fall within the range of estimates on land evaporation that are being produced by a new generation of global evaporation models that are being generated using a combination of climate, eddy flux, and remote sensing information; 550 mm year^{-1} (Jung et al. 2010); 539 mm year^{-1} (Zhang et al. 2010); 655 mm year^{-1} (Fisher et al. 2008).

The seasonal pattern of daily evaporation is very distinct for different forest types, distributed across the globe (Fig. 5.2). Tropical forests experience little seasonality in maximum evaporation, which approaches 5 mm day^{-1} (Fig. 5.2a). Most temporal variation occurs on a day-by-day basis and is modulated by cloud cover and daily changes in solar radiation (Vourlitis et al. 2001; Araujo et al. 2002; da Rocha et al. 2004; Loescher et al. 2005; Giambelluca et al. 2009). Deciduous broadleaved forest, in temperate and boreal climates, experience much seasonal variation in daily evaporation (Black et al. 1996; Moore et al. 1996; Lee et al. 1999; Wilson and Baldocchi 2000; Blanken et al. 2001; Oliphant et al. 2004; Barr et al. 2007). During the winter leafless period, daily evaporation fluxes are below 1 mm day^{-1} (Fig. 5.2b). By summer, peak evaporation rates are not dissimilar from those observed over tropical forests. Daily evaporation from savanna woodlands experiences much seasonality (Baldocchi et al. 2004; David et al. 2007; Paço et al. 2009). Highest evaporation rates (~4 mm day^{-1}) occur during the spring,

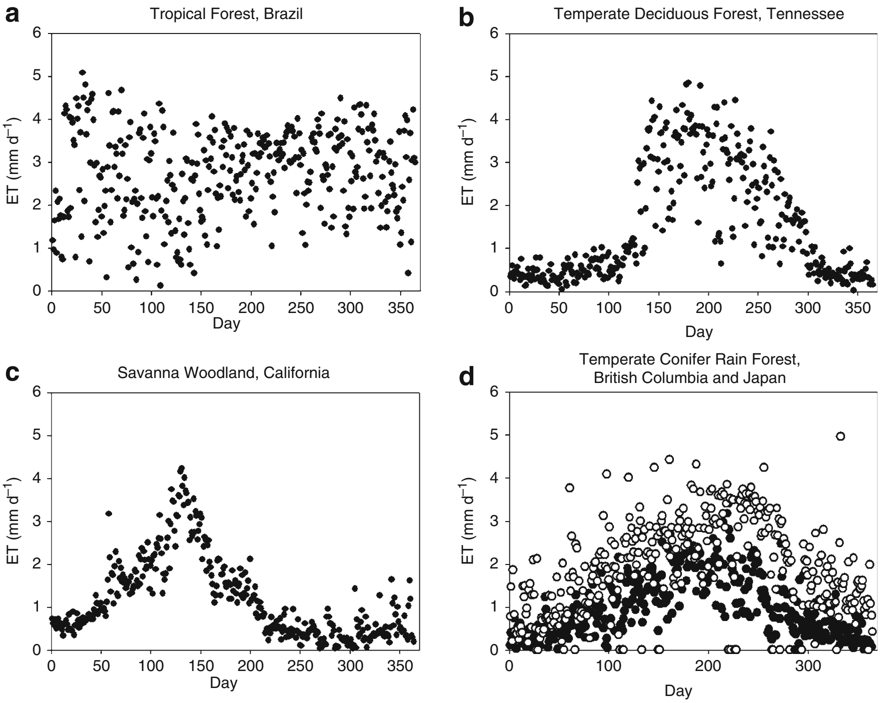

Fig. 5.2 Seasonal variation in daily integrated evaporation from: (**a**) tropical rain forest in Brazil (Araujo et al. 2002); (**b**) a deciduous broadleaved forest in Tennessee (Wilson and Baldocchi 2000); (**c**) a savanna woodland in California (Baldocchi et al. 2004); and (**d**) a temperate conifer rain forest in British Columbia (Humphreys et al. 2003) and a cypress forest in Japan (*open dot*) (Kosugi et al. 2005)

after winter rains have replenished the soil profile. Lack of rain during the summer depletes the soil moisture reservoir, causing stomata to close, transpiration to be restricted, and soil evaporation to be nil (Fig. 5.2c). Evaporation from temperate evergreen, conifer forests occurs year-round and their daily magnitude follows the seasonal course of the sun (Anthoni et al. 2002; Humphreys et al. 2003; Kosugi et al. 2005; Grunwald and Bernhoffer 2007). The maximum rates of daily evaporation from an evergreen, conifer, Douglas fir forest in British Columbia tend to be much smaller (less than 3 mm day^{-1}) than that from a temperate evergreen Cypress forest in Japan (Fig. 5.2d). In contrast, daily evaporation from evergreen boreal conifer forests is highly seasonal and is nil during the winter freezing and snow period (Amiro et al. 2006) (not shown).

Very few longer term evaporation records – a decade and longer – have been collected and reported in the literature based on either the eddy covariance (Grunwald and Bernhoffer 2007; Granier et al. 2008) or flux-gradient (Jaeger and Kessler 1997) methods. In general, no trends in evaporation have been detected or attributed to climate, forest function, and structure or successional stage in these few studies.

The lack of a long-term trend in evaporation contrasts with trends associated with CO_2 exchange and forest age (Urbanski et al. 2007; Stoy et al. 2008); net carbon exchange of a closed canopy is a strong function of stand age. This small sampling of the very long evaporation literature also contrasts with findings reported in a 30-year catchment study in the United Kingdom (Marc and Robinson 2007). Younger forests evaporate more than grasslands, while the opposite is true for older forests (Marc and Robinson 2007). On the other hand, Jung et al. (2010) found no trend in terrestrial evaporation at the global scale between 1998 and 2008; they used a machine learning algorithm based on the flux tower network and remote sensing.

5.4 Forest Evaporation and Hydrology

Forests cannot evaporate more water than is available from precipitation. But the questions we ask here include: What fraction of annual precipitation is lost by annual evaporation? Is the evaporation to precipitation ratio conservative? Or does it vary with climate and forest type? Figure 5.3 shows there is a strong linear relationship between precipitation and annual evaporation from forests distributed across the globe. On average, forest evaporation increases 46 mm for every 100 mm increase in rainfall. Furthermore, variations in annual precipitation explain over 75% of the variance in annual evaporation, a remarkably high r^2 value in our opinion. Surprisingly, few workers have examined or reported a correlation

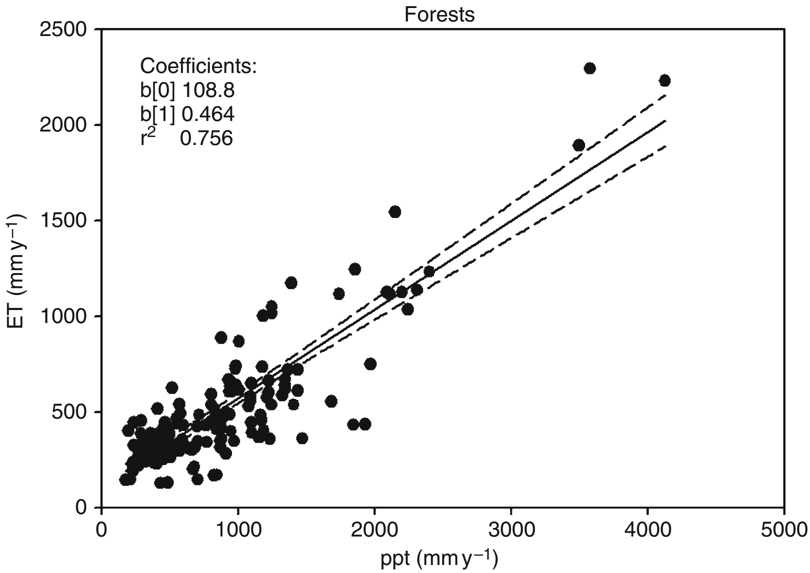

Fig. 5.3 Correlation between annual precipitation (ppt) and annual evaporation (ET) from forests. $N = 165$

between direct measurements of forest evaporation and precipitation at annual timescales, to our knowledge. Albeit indirect correlations are inferred from the Budyko function (Budyko 1974; Donohue et al. 2007). And recently, Alton et al. (2009) evaluated the Marconi version of the FLUXNET database and reported that an ensemble of ecosystems evaporates about 58% of monthly precipitation.

We do not claim that the slope between annual precipitation and forest evaporation (0.46) holds within specific climate spaces, just across the global climate space. For example, in semiarid regions forests tend to evaporate all precipitation (Anthoni et al. 1999; Scott 2010; Yaseef et al. 2010), so the local evaporation–precipitation ratio is near one. In Mediterranean climates, evaporation from evergreen and deciduous oak woodlands is capped below 500 mm year^{-1}, despite interannual variations in rainfall that can range between 600 and 1,200 mm (Joffre and Rambal 1993; Baldocchi et al. 2010).

To investigate the residual sources of variance shown in Fig. 5.3, we plotted annual evaporation (ET) as a function of annual precipitation and net radiation (Fig. 5.4). This surface plot indicates that adding net radiation increases the coefficient of determination (r^2) only slightly, to 0.82. In general, the highest sums of annual evaporation occur where both precipitation and net radiation are high (tropical forests), and the lowest evaporation occurs where annual sums of precipitation and net radiation are low (e.g., boreal forests).

On longer timescales, eco-hydrological factors conspire with one another to limit actual evaporation rates in semiarid regions from meeting extremely high potential evaporation levels (>1,200 mm year^{-1}) that are conducive of this region (Rambal 1984; Baldocchi and Xu 2007; Yaseef et al. 2010). In effect, limitations in precipitation and soil moisture limit stand recruitment by modulating sapling and

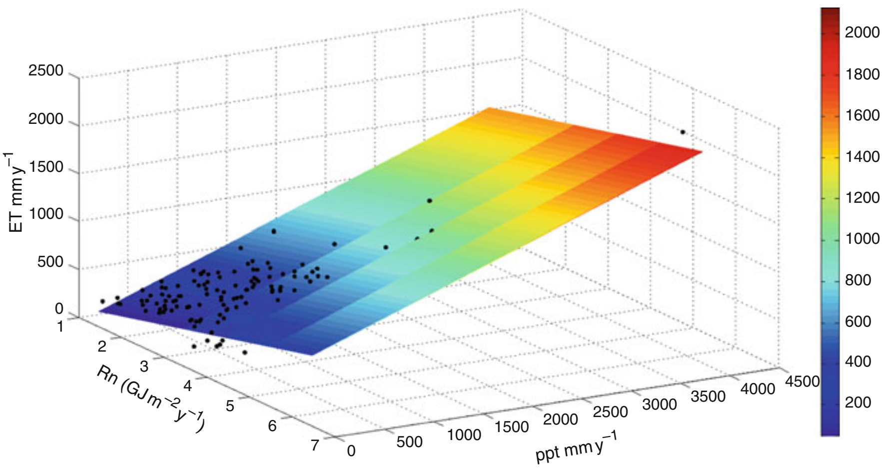

Fig. 5.4 Three-dimensional plot between annual evaporation (ET), net radiation (Rn), and precipitation (ppt). A linear additive model has the following statistics: ET = −141 + 116*Rn + 0.378*ppt, r^2 = 0.819. The *color bar* refers to annual ET

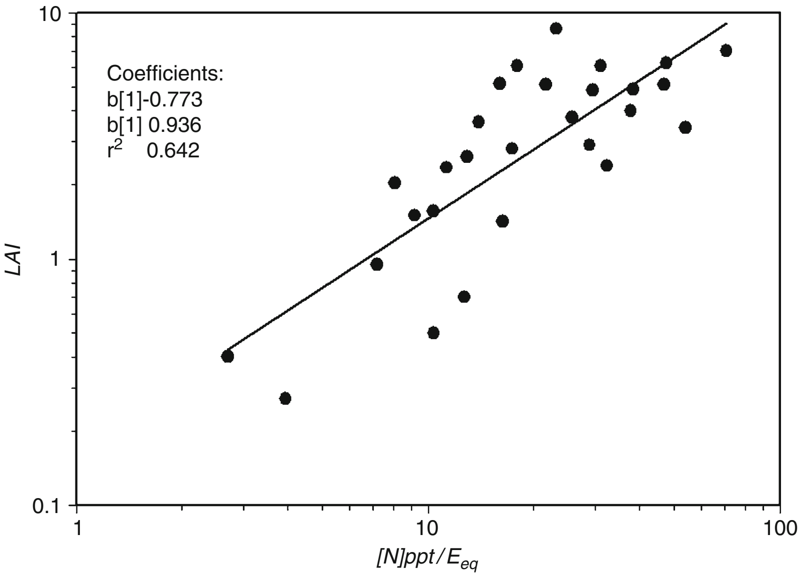

Fig. 5.5 Relation between leaf area index of forests vs. a nondimensional index defined as the product of leaf nitrogen times annual precipitation divided by the annual sum of equilibrium evaporation

seedling mortality. An equilibrium between water use and availability is eventually met to form a canopy with a relatively low tree density and low leaf area index (Eamus and Prior 2001; Baldocchi and Xu 2007; Joffre et al. 2007) (Fig. 5.5).

Another ecological question pertaining to forest evaporation relates to the functional role of broadleaved vs. needleleaved and deciduous vs. evergreen trees on evaporation of forests growing in similar climates. This question was addressed by Komatsu et al (2007) for a number of forest catchments across Japan. They found that evaporation from broadleaved forests was approximately the same for young conifer stands and it was higher than evaporation from old conifer stands. Swank and Douglass (1974), on other hand, found stream flow was reduced by 20% by converting a deciduous forest to conifer forest due to more interception losses by pine. In Mediterranean climates, annual evaporation from deciduous oaks is significantly greater than evaporation from evergreen oaks, by 110 mm year^{-1}, after difference in leaf area index and local climate is considered (Baldocchi et al. 2010).

5.5 Biophysical Controls on Evaporation

The Penman–Monteith equation provides a theoretical framework for quantifying forest evaporation, in terms of latent heat exchange (Monteith and Unsworth 1990), and for diagnosing how evaporation rates will respond to changes in weather and plant variables. For improved diagnostic reasons, the Penman–Monteith equation has been redefined as the additive combination of equilibrium (E_{eq}) and imposed (E_{imp}) evaporation (Jarvis and McNaughton 1986):

$$E = \Omega \cdot E_{eq} + (1 - \Omega)E_{imp}. \tag{5.2}$$

Equilibrium evaporation is a function of available energy and computed as:

$$E_{eq} = \frac{s}{s + \gamma} \frac{(R_{net} - G - S)}{\lambda}. \tag{5.3}$$

Imposed evaporation is a function of atmospheric demand and physiological supply and is expressed as:

$$E_{imp} = \frac{G_s D}{P}. \tag{5.4}$$

In (5.2) through (5.4), Ω denotes the coupling factor and ranges between zero and one, s is the slope of the saturation vapor pressure–temperature relationship, R_n is the net radiation balance, S is canopy heat storage, G is soil heat storage, D is vapor pressure deficit, γ is the psychrometric constant, ρ is air density, C_p is specific heat of air, λ is the latent heat of evaporation, and G_h and G_s are the canopy-scale conductances for boundary layer and surface.

Forests are aerodynamically rough and tend to be better coupled with their environment, hence they tend to have low omega values (Jarvis and McNaughton 1986; Verma et al. 1986). Despite the theoretical association between forest evaporation with imposed evaporation, a large number of investigators have interpreted their forest evaporation rates as a multiplicative fraction of equilibrium evaporation, α:

$$E = \alpha \frac{s}{s + \gamma} (R_{net} - G - S). \tag{5.5}$$

For short vegetation with adequate soil moisture and extensive fetch, α is about 1.26. But a number of studies show that α deviates from 1.26 for dry (Baldocchi et al. 1997; Komatsu 2005; Baldocchi and Xu 2007) and wet forest canopies (Shuttleworth and Calder 1979). Table 5.1 reproduces a survey of α values for a range of forest types. In general, α values for forests range between 1.09 and 0.53 and rank according to the following: broadleaved deciduous > tropical broadleaved evergreen > temperate conifer > boreal conifer > boreal deciduous conifer. Empirical evidence adds that α increases with decreasing canopy height and increasing leaf area index (Kelliher et al. 1993; Komatsu et al. 2007) and it decreases with progressively drying soils (Kelliher et al. 1993; Baldocchi et al. 2004; Chen et al. 2008). Using a theoretical model, we determined that this ranking depends on leaf area index, photosynthetic capacity, and soil moisture (Baldocchi and Meyers 1998); highest α values are produced by forests with high leaf area

Table 5.1 The ratio between actual and equilibrium evaporation for a number of forest categories, otherwise denoted as the Priestley–Taylor coefficient, α (Komatsu 2005)

Forest type	Mean	Std dev.
Boreal broadleaved, deciduous	1.09	–
Temperate broadleaved deciduous	0.851	0.147
Tropical broadleaved, evergreen	0.824	0.115
Temperate broadleaved evergreen	0.764	0.181
Temperate conifer	0.652	0.249
Boreal conifer, evergreen	0.55	0.102
Boreal conifer, deciduous	0.53	0.084

indices, ample soil water, and high photosynthetic capacity. Conversely, lowest α values are associated with sparse forest canopies with low photosynthetic capacity, low leaf area index, low hydraulic conductivity, and/or soil moisture deficits.

The role of biodiversity on forest evaporation remains unclear. On one hand, Currie and Paquin (1987) reported that tree species richness was positively correlated with an inferred estimate of annual evaporation. It was proposed that forest productivity scales positively with biodiversity, so it was expected that evaporation would scale with increased productivity. On the other hand, one of us (Baldocchi 2005) reported that the alpha coefficient in (5.5) decreased with increasing number of the dominant tree species in deciduous broadleaved forests. In the latter study, it was hypothesized that greater biodiversity increases the diversity of xylem architecture (e.g., ring vs. diffuse porous) and xylem conductivity. Hence, more diverse forests would have a mixture of trees with higher and lower xylem conductivities (Cochard et al. 1996), causing the area integrated hydraulic conductivity of a diverse forest to be less than a less diverse forest. The idea remains contentious and merits further scrutiny as it was derived from a relatively small cross-section of the FLUXNET database.

5.6 Understory Evaporation

Not all water is lost by tree transpiration. Some water is lost by re-evaporation of intercepted rainfall, a second fraction is lost by soil evaporation, and the remainder is transpired by understory vegetation. But until the advent of eddy covariance systems, it was difficult to measure understory evaporation directly, albeit lysimeters and chambers provide useful information on soil evaporation (Black and Kelliher 1989; Yaseef et al. 2010). The tall nature of forests enables investigators to deploy eddy covariance systems in the understory (Baldocchi and Meyers 1991) to measure soil evaporation directly. In general, the fraction of evaporation under forests is significant. It ranges between 10 and 50% of total evaporation and the evaporation ratio tends to increase with decreasing leaf area index as the net radiation flux density at the soil increases (Table 5.2).

Table 5.2 Survey on the fraction of understory evaporation based on eddy covariance measurements made in the forest understory and overstory

Location	$ET_{understory}/ET$ (%)	LAI	References
Deciduous forest, Tennessee	10	6	Wilson et al. (2000)
Boreal pine forest, Saskatchewan	20–40	~2	Baldocchi et al. (1997)
Oak savanna, California	<20	0.7	Baldocchi et al. (2004)
Temperate pine, Metolius, OR	<20	1.5	Baldocchi et al. (2000)
Boreal deciduous, broadleaved, Prince Albert, Sask	25	5.6	Blanken et al. (2001)
Semiarid pine, Israel	36	1.5	Yaseef et al. (2010)
Semiarid, woodland, Arizona	30–40		Scott et al. (2003)
Larch, Siberia	51	2.0	Iida et al. (2009)
Larch, Siberia	50		Kelliher et al. (1997)
Boreal pine forest, Sweden	10–15		Constantin et al. (1999)
Larch, Siberia	35	3.7	Ohta et al. (2001)

By being nonintrusive, eddy covariance measurements of soil evaporation have produced an alternative interpretation on the controls of soil evaporation. The timescale for turbulent exchange inside a deep forest is on the order of 2–5 min. This periodic re-flushing of the canopy air space inhibits soil evaporation rates from attaining equilibrium with the net radiation budget, and instead forces soil evaporation to be more closely coupled to their vapor pressure deficit (Baldocchi and Meyers 1991; Baldocchi et al. 2000). A consequence of this finding is the potential to restrict or inhibit soil evaporation by using soil chambers.

5.7 Final Comments and Future Directions

Forests play pivotal, and at times contrarian, roles on the water balance of catchments (Salati and Vose 1984) and the climate system (Bonan 2008; Jackson et al. 2008). If one is managing watersheds to maximize water yield, forested catchments tend to provide less runoff than cleared catchments (Bosch and Hewlett 1982; Marc and Robinson 2007). This finding has important implications on the hydrological cost of sequestering carbon by afforestation and reforestation. In semiarid regions, replacing herbaceous vegetation with forests will increase evaporation because forests are aerodynamically rougher and radiatively darker than herbaceous vegetation (Kelliher et al. 1993; Baldocchi et al. 2004). On the other hand, if one is trying to sustain large-scale precipitation in tropical and temperate humid regions, the presence of forests can generate a positive feedback on the hydrological cycle and promote runoff (Salati and Vose 1984; Bonan 2008; Jackson et al. 2008); trees are effective conduits for transferring soil moisture into the atmosphere, which in turn condenses, forms clouds and rain. Conversely, large-scale tropical deforestation has the potential to break this hydrological cycle because C_4 pastures (that typically replace tropical forests) evaporate less water due to their lower stomatal and surface conductances (Dickinson and Henderson-Sellers 1988; Vourlitis et al. 2002; Sakai et al. 2004).

We are now at the dawn of a new era with the potential to produce decade long, and longer, data records of direct evaporation rates for a wide range of forests in a changing world. The application of eddy covariance, however, is still restricted to rather ideal and flat terrain. Hence, there is still value to continue studying forest evaporation with gauged watersheds in complex terrain. Research questions that need continued attention include the roles of annual precipitation and biodiversity on forest evaporation and how to upscale tower fluxes across complex landscapes to regional and global scales using remote sensing.

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