3 Forest Evapotranspiration: Measurement and Modelling at Multiple Scales

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3.1 Introduction

Compared with traditional engineering hydrology, forest hydrology has a relatively long history of studying the effects of vegetation in regulating streamflow through evapotranspiration (Hewlett, 1982; Swank and Crossley, 1988; Andreassian, 2004; Brown *et al*., 2005; Amatya *et al*., 2011, 2015, 2016; Sun *et al*., 2011b; Vose *et al*., 2011). It is estimated that more than half of the solar energy absorbed by land surfaces is used to evaporate water (Trenberth *et al*., 2009). Evapotranspiration (ET), the sum of evaporation from soil (E), canopy and litter interception (I), and plant surface and plant transpiration (T), is critical to understanding the energy, water and biogeochemical cycles in forests (Baldocchi *et al*., 2001; Levia *et al*., 2011).

The linkage among energy, water and carbon balances at a forest-stand level over a long time period (Fig. 3.1), in which ET plays a key role, can be described conceptually in the following interlinked formulae (Sun *et al*., 2010, 2011a).

Water balance:

$$P = ET + Q \tag{3.1}$$

Energy balance:

$$R_n = L_E + H = ET \times L + H \tag{3.2}$$

Carbon balance:

$$\mathrm{NEP} = \mathrm{GPP} - R_e - L_C = \mathrm{ET} \times \mathrm{WUE} - \mathrm{R}_e - L_C \tag{3.3}$$

In the above, P is precipitation (mm), Q is runoff (mm), R_n is net radiation (W/m^2), L_E is latent heat (W/m^2) that represents the energy used to evaporate the amount of water by ET assuming a constant conversion factor called the latent heat of vaporization of water (L = 539 cal/g H_2O = 2256 kJ/kg H_2O), H is sensible heat that is consumed to heat the air near the forest canopy. The net ecosystem productivity (NEP; g C/m^2) is the carbon balance between carbon gain by gross ecosystem productivity (i.e. plant photosynthesis) and carbon loss by ecosystem respiration (R_e; g C/m^2) and lateral export in stream runoff (L_C; g C/m^2). The magnitude of both gross primer productivity (GPP; g C/m^2) and R_e is much larger than that of NEP and L_C, and all four variables are influenced by soil moisture and the hydrology. In many cases, ET explains the majority of the seasonal variability of GPP for all ecosystems

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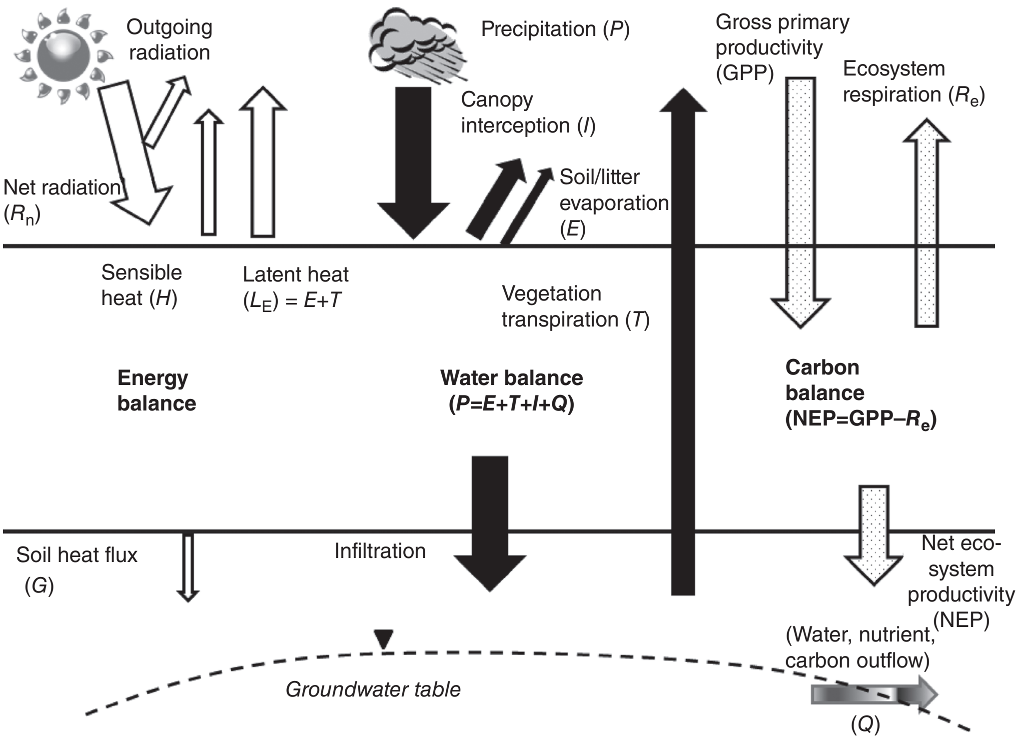

Fig. 3.1. Linkages among energy, water and carbon cycles in a forest ecosystem on the lower coastal plain of North Carolina in the USA. Note that net radiation (R_n) is a result of total incoming minus reflected shortwave radiation, along with the absorbed minus emitted longwave radiation.

(Law *et al.*, 2002; Sun *et al.*, 2011b). The ratio GPP/ET is termed water-use efficiency (WUE) and has been used as an important variable to understand the linkages of water–carbon coupling (Law *et al.*, 2002; Gao *et al.*, 2014; Frank *et al.*, 2015).

3.1.1 Understanding ecosystem processes

ET is a key variable linking meteorology, hydrology and ecosystem sciences (Baldocchi *et al.*, 2000; Oishi *et al.*, 2010; Sun *et al.*, 2011b). Plant transpiration T is a key variable directly coupled with ecosystem productivity (Rosenzweig, 1968) and carbon sequestration (Aber and Federer, 1992). This is easy to understand by the simple fact that CO_2 intake during plant photosynthesis uses the same pores, stomata, as the water loss, transpiration, uses (Canny, 1998). However, although E and T are both driven by atmospheric demand, T is actively controlled by stomatal regulation. ET is the only variable that links hydrology and biological processes in many ecosystem models (Aber and Federer, 1992). ET is also highly linked to ecosystem productivity and net ecosystem exchange of CO_2 because both photosynthesis and ecosystem respiration are controlled by soil water availability (Law *et al.*, 2002; Jackson *et al.*, 2005; Huang *et al.*, 2015).

3.1.2 Constructing water balances

ET is a large component of the water budget. Worldwide, mean annual ET rates are estimated to be about 600 mm (Jung *et al.*, 2010; Zeng *et al.*, 2014), or 60–70% of precipitation (Oki and Kanae, 2006; Teuling *et al.*, 2009). In the USA, more than 70% of the annual precipitation returns to the atmosphere as ET (Sanford and Selnick, 2013). Annual forest ET can exceed precipitation in the humid southern USA (Sun *et al.*, 2002, 2010) in dry years and it is not uncommon that

ET exceeds precipitation during the growing season in forests. Vegetation affects watershed hydrology and water balances through ET (Zhang *et al*., 2001; Oudin *et al*., 2008; Ukkola and Prentice, 2013; Jayakaran *et al*., 2014). Land-use conversion (i.e. bioenergy crop expansion) can dramatically change plant cover and biomass, affecting transpiration and evaporation rates, and therefore site water balances (King *et al*., 2013; Albaugh *et al*., 2014; Amatya *et al*., 2015; Christopher *et al*., 2015), including streamflow quantity (Ford *et al*., 2007; Palmroth *et al*., 2010; Amatya *et al*., 2015) and quality such as total sediment loading (Boggs *et al*., 2015).

3.1.3 Understanding climate change, variability and feedbacks

The ET processes are closely linked to energy partitioning, water balances and climate systems (Betts, 2000; Bonan, 2008). ET is tightly coupled to land-surface energy balance and thus influences vegetation–climate feedbacks (Bonan, 2008; Cheng *et al*., 2011). Changes in ET directly affect runoff, soil water storage, and local precipitation and temperature at the regional scale (Liu, 2011). The cooling or warming effects of reforestation are due to the increase in ET by planted trees or altered surface albedo (Peng *et al*., 2014). ET may be considered an 'air conditioner'.

Global climate change, in turn, directly affects the local water resources through ET (Sun *et al*., 2000, 2008). An increase in air temperature generally means an increase in vapour pressure deficit and evaporative demand or potential ET, resulting in an increase in water loss by ET, and thus a decrease in groundwater recharge and soil water availability to ecosystems and human water supply. Regions that are experiencing more warming would see more severe hydrological droughts regardless of changes in precipitation (Mann and Gleick, 2015).

3.1.4 Modelling regional ecosystem biodiversity

ET has long been regarded as an index to represent the available environmental energies and ecosystem productivity by bioclimatologists. Thus, ET has been used to explain the large regional variations in plant and animal species' richness and biodiversity. For example, the variability in species richness in vertebrate classes could be statistically explained by a monotonically increasing function of a single variable, potential evapotranspiration (PET) (Currie, 1991). In contrast, regional tree richness was more closely related to actual ET (Currie, 1991; Hawkins *et al*., 2003).

3.2 Evapotranspiration Processes

Forest ET processes are inherently complex due to the many ecohydrological interactions within a forest ecosystem that often consists of multiple plant species with heterogeneous spatial distribution and variable microclimate over space and time (Canny, 1998). Both the physiological (e.g. stomata control) and physical processes (e.g. water potential control) influence the water vapour movements from plant organs of roots, xylem and leaf, to stands and landscapes (i.e. watersheds). Since soil evaporation can be minor in closed-canopy forests (McCarthy *et al*., 1992; Domec *et al*., 2012b), this chapter focuses on the processes that control canopy and litter interception (I) and transpiration (T), and methods to quantify these two major components of ET.

3.2.1 Canopy and litter interception

The quantity of canopy and litter interception (I) in forests can be a large component of the ET and water balances, depending on forest structure characteristics such as leaf area index (LAI) and canopy holding capacity, and the amount of litter and litter water-holding capacity, respectively (Gash, 1979; Deguchi *et al*., 2006). In addition, the frequency of storms and the drying and wetting cycles affect total canopy and litter interception. Although interception can be 20–50% of the precipitation, most hydrological models do not simulate this process explicitly (Gerrits *et al*., 2007).

The earliest studies by Horton (1919) showed highly variable interception rates between and across species, with the spruce–fir–hemlock forest type the highest, followed by pines and then hardwoods. Helvey (1974) reported annual canopy interception as

17% for red pine (*Pinus resinosa* Ait.), 16% for ponderosa pine (*Pinus ponderosa* Dougl. ex. Laws.), 19% for eastern white pine and 28% for the spruce–fir–hemlock forest type. The difference in canopy interception rates between hardwood and conifer forests partially explained the observed difference in streamflow (Swank and Miner, 1968). Summer interception rates of deciduous forests in the south-eastern USA ranged from 8 to 33%, with a mean of 17%, and winter rates ranged from 5 to 22%, with a mean of 12% (Helvey and Patric, 1965). Annual canopy interception rate was 18% for wetland sites, 20% for hardwood sites and a longleaf pine (*Pinus palustus* Mill.) plantation and 23% for pine-dominated forests in the south-eastern USA (Bryant *et al*., 2005). Thinning of a loblolly pine (*Pinus taeda* L.) plantation forest reduces basal area and subsequent leaf area, resulting in a decrease in canopy interception (McCarthy *et al*., 1992). Interception rates vary between 10–35% and 5–25% for un-thinned versus thinned loblolly pine stands, respectively (Gavazzi *et al*., 2015). Forests in tropical and subtropical regions could intercept 6 to 42% of precipitation (Bryant *et al*., 2005). In the USA, reported annual values of litter precipitation interception rate for eastern forests vary by about 2–5%, generally less than 50 mm per year (Helvey and Patric, 1965). However, litter interception may be higher than canopy interception in other forest ecosystems (Gerrits *et al*., 2007).

3.2.2 Transpiration

The transpiration process (T) represents water loss through leaf stomata, the tiny openings found on one side or both sides of the tree leaves (Canny, 1998). Because T is an inevitable consequence of CO_2 assimilation by plants through photosynthesis, maintaining of leaf tissue turgidity and plant nutrient uptake, together with soil evaporation, T represents an ecosystem water loss and thus is a 'necessary evil' for net ecosystem productivity.

A global synthesis study indicates that T accounts for 61 ± 15% of total ET and returns approximately 39 ± 10% of incident precipitation to the atmosphere, playing a great role in the global water cycle (Schlesinger and Jasechko, 2014). The T/ET ratios are highest in tropical rainforests (70 ± 14%) and lowest in steppes, shrublands and deserts (51 ± 15%). Transpiration is the major component of the total evapotranspiration in global hydrological cycles and ET is highly dependent upon biophysical parameters like stomatal conductance (Jasechko *et al*., 2013). Therefore, changes in transpiration due to increasing CO_2 concentrations, land-use changes, shifting ecozones, air pollution and climate warming may have significant impacts on water resources (Schlesinger and Jasechko, 2014). An increase in CO_2 concentrations may reduce plant leaf stomata conductance and increase WUE, but T can arise from increased leaf area in addition to lengthened growing seasons and enhanced evaporative demand in a warming climate with increased CO_2 concentration (Frank *et al*., 2015).

Carbon and water fluxes are coupled through the stomata activities: water vapour exits the stomata along with oxygen; carbon dioxide flows into the stomata and is absorbed by the photosynthesis process to produce carbohydrate (Crétaz and Barten, 2007). Transpiration is an active water translocation process that occurs only when water exists continuously along the soil–root–stem–branch–leaf–stomata flow pathway (Kumagai, 2011). However, transpiration rates differ tremendously among different tree species and ages (Plate 2). For example, a *Quercus rubra* tree with a 50 cm trunk diameter transpires an average of 30 kg H_2O/day, but *Betula lenta* can transpire as high as 110 kg H_2O/day under the same climate in the southern Appalachians in the south-eastern USA (Vose *et al*., 2011). A review of 52 whole-tree water use studies for 67 tree species worldwide using different techniques concluded that maximum daily water use rates for trees averaging 21 m in height were within 10–200 kg/day (Wullschleger *et al*., 1998).

The transpiration rates are controlled by numerous biophysical factors such as microclimatic characteristics, atmospheric CO_2 concentration, soil water potential, stand characteristics (e.g. leaf area, species compositions, tree density) and hydraulic transport properties of plant tissues (Domec *et al*., 2009, 2010, 2012a). The species compositions of forests change over space and time due to natural regeneration or in response to climatic change and/or human

Table 3.1. A comparison of major methods for estimating evapotranspiration (ET) at multiple scales.

	Method	Strength	Weakness	Source
Direct field-based	Porometer and cuvette	Leaf-level physiological process	Difficult to scale up due to uncertainty on the influence of boundary layers and variability of leaf age, radiation and humidity	Olbrich (1991)
	Weighing lysimeter	Single whole-tree water use	High cost	
	Heat balance/heat dissipation sapflow	Allows routine unsupervised measurement accurately at single plant scale	Large-scale measurement errors are determined by sample size and the variability of samples	Granier (1987)
	Eddy covariance	Measuring fluxes continuously, offering data with high temporal resolution	High cost in instrumentation, gap filling required, energy imbalance problems	Baldocchi *et al*. (2001)
	Bowen ratio	Works for both crops and natural vegetation	Relies on several assumptions, errors associated with low gradients	Irmak *et al*. (2014)
	Catchment water balance	Easy to measure	Only long-term average is reliable	Ukkola and Prentice (2013)
Remote sensing	MODIS	Provides high-resolution spatial, continuous and temporal data	Uncertainties due to errors generated by measurement of sparse canopies, data mostly from clear-sky conditions	Kalma *et al*. (2008)
Mathematical modelling for ET alone or the full hydrological cycle	Theoretical models (e.g. Penman–Monteith equation)	Widely tested, including all conditions, low cost	Requires site-specific parameters, not easy to apply to data-poor regions	McMahon *et al*. (2013)
Isotopes	Stable isotope H and O	Process-based understanding of water source of ET; partitioning of evaporation and transpiration	Cost and scaling up to stand level	Good *et al*. (2015)

activities such as silviculture (i.e. reforestation, afforestation). In addition, forest ecosystem structure changes in both above-ground characteristics, including leaf (i.e. leaf biomass) and stem (i.e. sapwood area) (Domec *et al*., 2012a; Komatsu and Kume, 2015), and below ground (i.e. root biomass) over time. Little is known about water pathways between soil water and roots and the water uptake mechanism of deep roots in response to drought (Meinzer *et al*., 2004; Warren *et al*., 2007).

Different from croplands, forests have multiple canopies and the understorey vegetation is an important component of a forest stand by intercepting and transpiring a significant amount of water. For example, over 20% of the total ET for a 17-year-old pine plantation was from understories (Domec *et al*., 2012b). Emergent understorey vegetation soon after harvest in the humid coastal plain was shown to have a substantial LAI, potentially affecting water balance for 4–5 years until the planted pine seedlings dominated the understorey (Sampson *et al*., 2011).

3.2.3 Hydraulic redistribution by roots: exchange of water at the soil–root interface

Plants can reduce water stress by extracting water from deeper and moist soil layers through plant roots and storing it in the upper, drier soil layers for use by shallow roots. The bidirectional (upward and downward) processes are termed ‘hydraulic redistribution’ (HR) (Burgess *et al*., 1998). The HR process occurs widely in all water-limited vegetated environments (Meinzer *et al*., 2004; Neumann and Cardon, 2012). HR is a passive process that depends on the soil suction head (soil water potential) and the root distribution within the soil column. HR by roots acts as a large water capacitor, increasing the efficiency of whole-plant water transport, buffering the seasonality of ET against water stress during seasonal water deficits, and representing 20–40% of whole-stand water use (Domec *et al*., 2010). Even when HR represents only a relatively small amount of ecosystem water use (e.g. <0.5 mm/day) and just a fraction (e.g. 5–10%) of total ET during the dry period, the partial recharge of upper soil moisture by HR is important to slow down the decline of soil water content and thus maintain water availability in topsoil layers (Warren *et al*., 2007). The influx of soil water maintains root water-uptake capacity and extends root functioning later into the drought period (Domec *et al*., 2004), influencing forest productivity (Domec *et al*., 2010).

3.2.4 Total evapotranspiration

The total ET rates at the ecosystem or watershed landscape level are controlled mainly by regional energy and water availability (Douglass, 1983; Zhang *et al*., 2001), but also are influenced by other anthropogenic management factors such as site fertilization (CO_2 effects and N deposition) (Tian, H.Q., *et al*., 2012; Frank *et al*., 2015), tree genetic improvement, species conversion (Swank and Douglass, 1974), artificial drainage (Amatya *et al*., 2000) and irrigation (Amatya *et al*., 2011). During the course of the forest stand development, site-level energy and water availability also vary, resulting in dramatic seasonal changes in total ET and its partitioning into sensible heat and other energy balance variables (Sun *et al*., 2010).

Forested watershed ET generally decreases soon after removal of the canopy by either harvesting or other natural disturbances (hurricanes, invasive species, fires, wind and snow storms, etc.) as a result of reduced canopy interception and transpiration (Sun *et al*., 2010; Tian, S.Y., *et al*., 2012; Jayakaran *et al*., 2014; Boggs *et al*., 2015). However, ET generally tends to increase soon after plantation (afforestation/reforestation) and after natural regeneration (Sun *et al*., 2010; Jayakaran *et al*., 2014). Figures 3.2 and 3.3 present an example of increase in annual ET after planting a harvested watershed (Amatya *et al*., 2000; Amatya and Skaggs, 2001, 2011; Tian, S.Y., *et al*., 2012) and after natural regeneration of a watershed (Jayakaran *et al*., 2014) substantially impacted by hurricane force winds. The inter-annual variability of ET was a result of precipitation variability at both the sites, consistent with other studies (Sun *et al*., 2002, 2010; Ukkola and Prentice, 2013).

Forest ET rates also vary dramatically across space and time on a heterogeneous terrain.

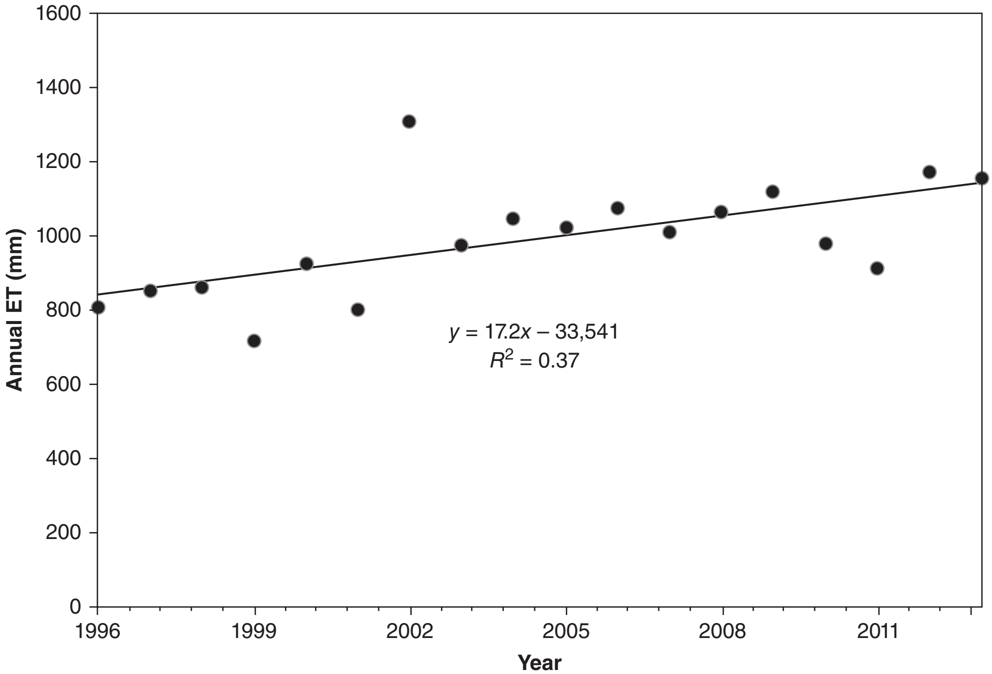

Fig. 3.2. Annual forest ecosystem evapotranspiration (ET) calculated as the differences between measured precipitation and measured streamflow for an experimental watershed. The ET rate increases gradually following tree/forest harvest in 1995 and replanting with loblolly pine in 1997 in Carteret County, coastal North Carolina, USA.

For example, ET rates of a forest stand are higher in the sunny side or/and near the ridges in a mountain watershed due to more solar radiation available (Douglass, 1983; Emanuel *et al*., 2010). Forest thinning practices reduce forest biomass, thus canopy interception and transpiration from remaining trees (Boggs *et al*., 2015), but do not necessarily reduce total ET (Sun *et al*., 2015).

3.3 Direct Measurement of Evapotranspiration

Forest ET processes have been quantified at multiple temporal and spatial scales from leaf to watershed, and even to global scale, using various methods from the hand-held cuvette method to the remote sensing approach (Table 3.1). The porometer method has been used to understand the environmental control on gas (CO_2 and H_2O) exchange at the leaf level (Olbrich, 1991). Other methods to measure T include ventilated chambers (Denmead *et al*., 1993), complex models parameterized by leaf-scale physiological traits and three-dimensional tree architecture (Kumagai *et al*., 2014), or sap flux density based on thermal dissipation and heat transport theories (Granier *et al*., 1996; Granier, 1987).

The sapflow technique has the advantage of not being limited by landform heterogeneity (Granier, 1987). The sapflow method measures water use by a single plant or tree, and thus answers questions on water use at the species and whole-stand levels. Components of forest water loss may be determined by measuring differences between total ET and tree sapflow, providing insights in terms of the response of water use by plants to climatic variability and stand development (Domec *et al*., 2012a). Sapflow measurements provide a powerful tool for quantifying plant water use and physiological responses of plants to environmental conditions (Domec *et al*., 2009).

In contrast, the eddy covariance technique measures forest ET by calculating the covariance

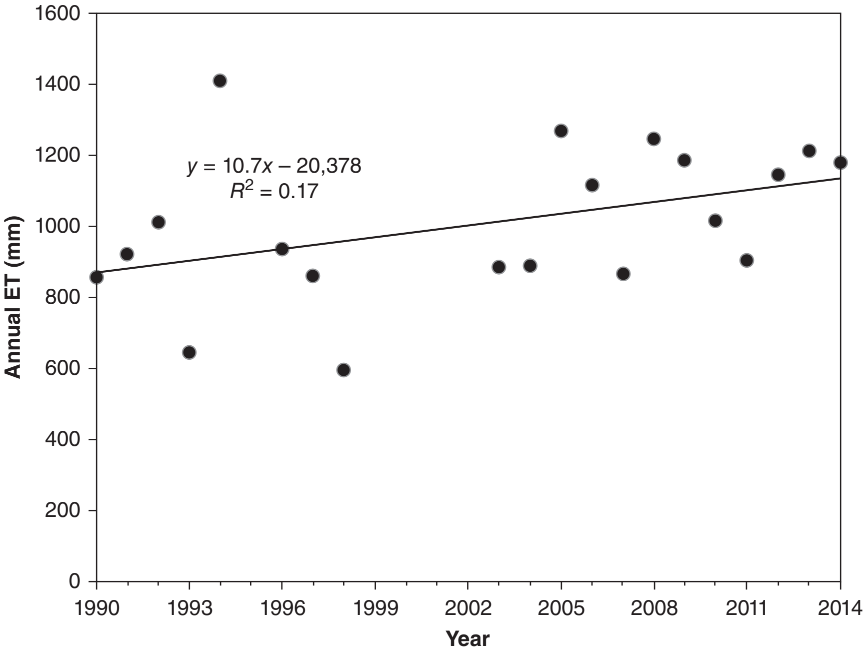

Fig. 3.3. Recovery of annual evapotranspiration (ET) calculated as the differences between measured precipitation and streamflow for a forested watershed that was naturally regenerated after the impact of Hurricane Hugo in 1989 at Santee Experimental Forest in coastal South Carolina, USA.

between fluctuations in vertical eddy velocity and the specific water vapour content above forest canopies (Baldocchi and Ryu, 2011). The method is designed to understand the gas exchange at the boundary layer between vegetation and the atmosphere, and answers questions at the landscape scale (the footprint of the flux tower) (Baldocchi *et al.*, 1988). The method relies on several assumptions such as an extensive fetch over a homogeneous surface.

Global participation in flux measurements through the FLUXNET (over 500 sites) (http://www.fluxnet.ornl.gov) since the 1990s has been a major driving force for advancing ET science (Baldocchi *et al.*, 2001).

The Bowen ratio methods have been used in quantifying ET in croplands under various soil (tillage), crop and irrigation management (sprinklers, subsurface drip, gravity irrigation, etc.) practices through the NEBFLUX project (Irmak, 2010) and have similar accuracy to the eddy flux methods (Irmak *et al.*, 2014). The method estimates ET from the ratio of sensible heat to latent heat, using air temperature and humidity gradients measured above the canopy, net radiation and soil heat flux. The fetch requirements for the Bowen ratio method are less than those for the eddy covariance method.

In addition to micrometeorological methods, stable isotopes have been used as tracers for identifying the sources of water uptake in ecosystems and evaluating quantitatively the relationships among water, energy and isotopic budgets. For example, tree-ring ^{13}C is used to identify changes in WUE and soil water stress (McNulty and Swank, 1995), and ^{18}O assists in determining whether those changes in WUE are due to changes in photosynthetic rate or stomatal conductance. Vegetation affects water/energy balance and isotopic budget through transpiration. Recently, using the D/H isotope ratios of continental runoff and evapotranspiration, independent of terrestrial hydrological partitioning, Good *et al.* (2015) demonstrated that globally the transpired fraction of evapotranspiration is estimated to be 56 to 74% (25th to 75th percentile), with a median of 65% and mean of 64%. Furthermore, studies across an ecosystem gradient in the USA and Mexico provided evidence of ecohydrological separation, whereby different subsurface compartmentalized pools of water supply either plant transpiration fluxes or the combined fluxes of groundwater and streamflow (Evaristo *et al*., 2015).

Estimating regional ET using satellite remote sensing data has emerged since the 1980s when there was an increasing interest in spatial dynamics in water use at the landscape scale (Kalma *et al*., 2008). Remote sensing ET products such as MODIS (Moderate Resolution Image Spectroradiometer) (Mu *et al*., 2011) have provided spatially and temporally continuous ET estimates at a 1 km resolution for understanding regional hydrology and environmental controls. However, uncertainties in modelling effective surface emissivity and effective aerodynamic exchange resistance, and sparse canopies and cloud conditions may make the remote sensing methods less reliable (Shuttleworth, 2012). Coupling energy balance models with remotely sensed land-surface temperature retrieved from thermal infrared imagery provides proxy information regarding the surface moisture and vegetation growth status (Anderson *et al*., 2012). Models such as the regional Atmosphere–Land Exchange Inverse (ALEXI) and the associated flux disaggregation model (DisALEXI) are based on the Two Source Energy Balance (TSEB) land-surface representations (Kustas and Norman, 1996). These modelling systems have recently been applied in a lower coastal plain in North Carolina and show promise to map high-resolution ET (e.g. daily, 30 m) for a landscape with mixed land uses with natural wetland forests, drained pine forest with multiple stand ages, and croplands (see also Chapter 9, this volume).

Long-term and annual watershed water balance ET are generally estimated using a simple water balance as the difference between measured precipitation and streamflow, assuming a negligible change in storage (Wilson *et al*., 2001; Sun *et al*., 2005; Amatya and Skaggs, 2011; Ukkola and Prentice, 2013). Watershed-scale ET is also dependent upon its land use or the area covered by vegetation (Amatya *et al*., 2015) in addition to the broader controls of precipitation and potential ET. Using observed data from 109 river basins during 1961–1999, Ukkola and Prentice (2013) showed strong control by precipitation followed by vegetation processes on ET trends and variability.

A few studies comparing multiple ET methods found that each method has its own limitations (Wilson *et al*., 2001; Ford *et al*., 2007; Domec *et al*., 2012b). The eddy covariance method measures fluxes continuously, offering time series data with high temporal resolution, but data availability is limited by costly site instrumentation, gap filling issues and extensive data corrections issues. In addition, the eddy covariance method may underestimate ET by as much as 30% due to a lack of energy balance closure (Wilson *et al*., 2002). The eddy covariance technique has also been shown to be problematic to underestimate ET on wet days because the sonic anemometer and infrared gas analyser must be dry to function properly (Wilson *et al*., 2001).

3.4 Indirect Estimates of Evapotranspiration

3.4.1 Methods based on potential evapotranspiration

Due to the high cost of trained personnel requirements for measuring ET directly at field and larger scales, mathematical modelling has been widely used to estimate ET (McMahon *et al*., 2013). ET models can be roughly divided into two groups: biophysical (theoretical) and empirical models. The former type of models refers to those developed based on physical and physiological principles describing energy and water transport in the soil–plant–atmosphere continuum (SPAC). Many theoretical models have evolved from the famous Penman (1948) and later from the Penman–Monteith model (Monteith, 1965) that represents the most advanced process-based ET model. The Penman–Monteith model estimates ET as a function of available energy, vapour pressure deficit, air temperature and pressure, and aerodynamic and canopy resistance. In contrast, empirical ET models are models developed using empirical observed ET data, land cover type, biophysical variables of plant characteristics such as LAI, soil moisture and atmospheric conditions. Empirical ET models do not intend to describe the processes of vaporization, but can give a reasonable estimate with limited environmental information.

In practice, it is often rather difficult to parameterize the process-based ET models to estimate actual ET. To simplify calculations, the concept of potential ET (PET) was introduced in the 1940s. For any ecosystem, PET represents the potential maximum water loss when soil water is not limiting. Actual ET then can be scaled down from the hypothetical PET by limiting canopy conductance and soil moisture, and correlates to pan evaporation (Grismer *et al.*, 2002). Such PET models are often embedded in hydrological models that can simulate the dynamics of soil moisture, a major control on soil evaporation and transpiration (Sun *et al.*, 1998; Tian, S.Y., *et al.*, 2012). McMahon *et al.* (2013) provide a comprehensive review on conceptual PET models and the techniques to estimate actual ET from open-surface waters, landscapes, catchments, deep lakes, shallow lakes, farm dams, lakes covered with vegetation, irrigation areas and bare soils.

Existing PET models can be classified into five groups (Lu *et al.*, 2005): (i) water budget; (ii) mass transfer; (iii) combination; (iv) radiation; and (v) temperature-based. There are approximately 50 models available to estimate PET that are developed considering input data availability and regional climate characteristics. The models give inconsistent values due to their different assumptions and input data requirements, or because they were often developed for specific climatic regions.

Numerous studies have suggested that different PET methods may give significantly different results (Amatya *et al.*, 1995; Lu *et al.*, 2005; McMahon *et al.*, 2013), so the standardized grass-reference PET method (Allen *et al.*, 2005), ET_o, is recommended to achieve comparable results across sites. Details of the computation procedures for ET_o are found in Allen *et al.* (1994). A computer program is available for public use (http://www.agr.kuleuven.ac.be/lbh/lsw/iupware/downloads/elearning/software/EtoCalculator.pdf). Once ET_o is calculated, actual ET for a particular ecosystem type can be estimated by simply multiplying by a 'crop coefficient, K_c' developed for that crop using ET measured by lysimeter or some other method (Allen *et al.*, 2005; Irmak, 2010). The K_c method works well in irrigation agriculture for various croplands that have uniform phenology. However, for forests, this method can be problematic given the large variability of species composition of a forest, leaf biomass dynamics throughout the season, and the age and density effects on tree biomass and water transport properties (canopy conductance, sapwood area). In addition, the reference ET concept may be misleading, because actual forest ET rates in humid climates often exceed the ET_o (Sun *et al.*, 2010). A casual use of ET_o as the maximum ET in a hydrological model may result in underestimation of actual ET (Amatya and Harrison, 2016). A recent study suggests that K_c for any forest type may vary tremendously and latitude, precipitation and LAI are the best predictors of K_c (Liu *et al.*, 2015). Forests generally have higher K_c values than other ecosystem types (Fig. 3.4).

3.4.2 Empirical evapotranspiration models

Empirical ET models are derived from direct ET measurements at the ecosystem scale. Empirical models may be best used as a first-order approximation of mean climatic conditions. The following model was derived from field data collected at 13 sites using a variety of methods (Sun *et al.*, 2011a). The model estimates monthly ET as a function of LAI, ET_o (mm/month) and precipitation P (mm/month) (see equation 3.4 at the bottom of the page), where ET_o is the FAO (Food and Agriculture Organization) reference ET as discussed above.

Other forms of the ET model use Hamon's potential ET (*PET*) instead of the more data-demanding FAO reference ET method (Sun *et al.*, 2011b) (see equation 3.5 at the bottom of the page).

$$ET = 11.94 + 4.76 \times \text{LAI} + ET_o\left(0.032 \times \text{LAI} + 0.0026 \times P + 0.15\right) \quad (3.4)$$

$$ET = 0.174 \times P + 0.502 \times PET + 5.31 \times \text{LAI} + 0.0222 \times PET \times \text{LAI} \quad (3.5)$$

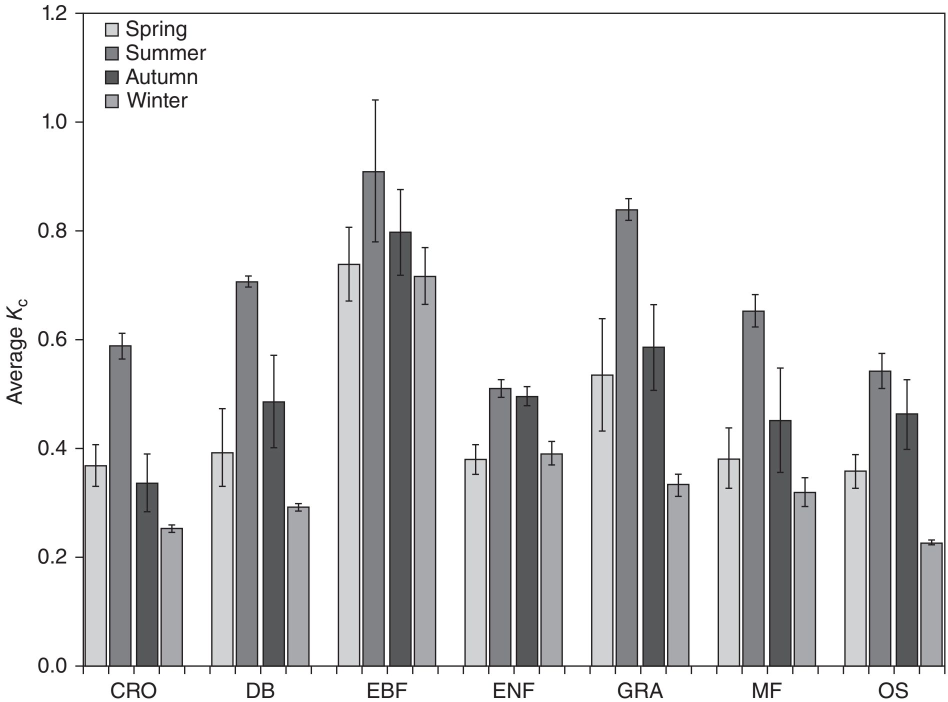

Fig. 3.4. A comparison of seasonal mean crop coefficient (K_c) calculated from global eddy flux measurements for cropland (CRO), deciduous broadleaf forest (DB), evergreen broadleaf forest (EBF), evergreen needle-leaf forest (ENF), grassland (GRA), mixed forest (MF) and open shrub land (OS). K_c is estimated as measured ET divided by the grass reference ET (ET_o) calculated by the standardized FAO-56 method; error bars represent standard deviation.

Using a similar concept and a 250 FLUXNET synthesis data set, Fang *et al.* (2015) developed the two monthly ET models (Eqns 3.6 and 3.7) that require different input variables (see equation 3.6 at the bottom of the page), where *PET* is monthly potential ET (mm) calculated by Hamon's method, VPD is vapour pressure deficit (hundreds of Pascals) that can be estimated from relative air humidity, R^2 is the coefficient of determination and RMSE is root-mean-squared error. Since R_n is rarely available at the regional scale, another model that uses more commonly available data was developed (see equation 3.7 at the bottom of the page).

A series of ecosystem-specific monthly scale ET models was also developed using the global eddy flux data (Fang *et al.*, 2015) (Table 3.2). An empirical annual ET model was developed by combining a water balance method with a climate and land cover regression equation to estimate mean annual ET across the conterminous USA (Sanford and Selnick, 2013). The climate variables included mean annual daily maximum and daily minimum air temperature and mean annual precipitation. The land cover types included developed, forest, shrubland, grassland, agriculture and marsh.

$$ET = 0.42 + 0.74 \times PET - 2.73 \times \mathrm{VPD} + 0.10 \times R_n \qquad \left(R^2 = 0.73, \mathrm{RMSE} = 17.0\ \mathrm{mm/month}\right) \tag{3.6}$$

$$ET = -4.79 + 0.75 \times PET + 3.92 \times \mathrm{LAI} + 0.04 \times P \qquad \left(R^2 = 0.68, \mathrm{RMSE} = 18.1\ \mathrm{mm/month}\right) \tag{3.7}$$

Table 3.2. Empirical models by land cover type developed using three commonly measured biophysical variables.

Land cover type	Model	RMSE	R^2	*n*
Shrubland	$ET = -3.11 + 0.39 \times PET + 0.09 \times P + 11.127 \times \text{LAI}$	12.5	0.80	193
Cropland	$ET = -8.15 + 0.86 \times PET + 0.01 \times P + 9.54 \times \text{LAI}$	20.9	0.70	653
Grassland	$ET = -1.36 + 0.70 \times PET + 0.04 \times P + 6.56 \times \text{LAI}$	16.8	0.66	803
Deciduous forest	$ET = -14.82 + 0.98 \times PET + 2.72 \times \text{LAI}$	23.7	0.74	754
Evergreen needle-leaf forest	$ET = 0.10 + 0.64 \times PET + 0.04 \times P + 3.53 \times \text{LAI}$	17.8	0.68	1382
Evergreen broadleaf forest	$ET = 7.71 + 0.74 \times PET + 1.85 \times \text{LAI}$	16.8	0.76	233
Mixed forest	$ET = -8.763 + 0.95 \times PET$	13.1	0.79	259
Savannah	$ET = -5.66 + 0.18 \times PET + 0.10 \times P + 44.63 \times \text{LAI}$	11.1	0.68	36

ET = evapotranspiration (mm/month); *P* = precipitation (mm/month); *PET* = potential ET estimated by Hamon's method (mm/month); LAI = leaf area index; RMSE = root-mean-squared error; R^2 = coefficient of determination; *n* = sample size.

The long-term mean ET in a region is controlled mainly by water availability (precipitation) and atmosphere demand (potential ET), and this relationship is well described in the Budyko framework (Budyko *et al.*, 1962; Zhang *et al.*, 2001; Zhou *et al.*, 2015). Using the same concept, Zhang *et al.* (2001) analysed watershed balances data for over 250 catchments worldwide and developed a simple two-parameter ET model. The model offers a practical tool that can be readily used for assessing the long-term average effect of vegetation changes on catchment evapotranspiration:

$$ET = P \times \frac{1 + w(PET/P)}{1 + w(PET/P) + (P/PET)} \quad (3.8)$$

where w is the plant-available water coefficient which represents the relative difference in plant water use for transpiration. *PET* can be estimated by the Priestley and Taylor (1972) model. *P* is annual precipitation. The best fitted value of w for forest and grassland is 2.0 and 0.5, respectively, when PET is estimated using the Priestley and Taylor (1972) model (Zhang *et al.*, 2001). Sun *et al.* (2005) suggested that w can be as high as 2.8 when the Hamon PET method is used in applying the model for the humid south-eastern USA, consistent with a study for a managed pine forest in the Atlantic coastal plain (Amatya *et al.*, 2002). Kumagai *et al.* (Chapter 6, this volume) modified the above equation to obtain ET for tropical forests.

By combining remote sensing and climate data for 299 large river basins, Zeng *et al.* (2014) developed an annual ET model that has been used to estimate global ET (see equation 3.9 at the bottom of the page), where *ET* is basin-averaged annual evapotranspiration (mm/year), *P*, *T* and NDVI are annual precipitation (mm/year), mean annual temperature (°C) and annual normalized difference vegetation index, respectively. Similarly, an empirical model was developed using only mean annual temperature from 43 catchment water balance data sets in Japan (Komatsu *et al.*, 2008).

3.5 Future Directions

3.5.1 Response to climate change

Climate change is the largest environmental threat to forest ecosystems in the 21st century (Vose *et al.*, 2012). Climate warming and the

$$ET = 0.4(\pm 0.02) \times P + 10.62(\pm 0.39) \times T + 9.63(\pm 2.27) \times \text{NDVI} + 31.58(\pm 7.89) \qquad (R^2 = 0.85) \quad (3.9)$$

increased variability of precipitation form, amount and timing are expected to have rippling effects on forest ecosystem structure and functions through directly or indirectly altering ET processes. However, because precipitation, a key environmental control of tree transpiration and soil evaporation, is uncertain and difficult to predict, we have little capacity to project ET changes at the local scale.

3.5.2 Managing evapotranspiration in a water-shortage world

Accurate quantification of watershed water budgets including water use by trees and shrubs is becoming increasingly important given the growing competition for water resources among all users, from agricultural irrigation and bioenergy development to domestic water withdrawals by cities, in the Anthropocene (Sun *et al*., 2008). We need better simulation models to reliably account for the role of forest ET in regulating streamflow and other ecosystem services (carbon fluxes) in large basins. Land managers have long asked the question: is it practical to manage upland headwater forests to meet future water supply demand in an urbanizing world (Douglass, 1983)? We know a lot of the basic relationships among forest cover, ET and water yield, but applying the knowledge to management remains a challenge (Vose and Klepzig, 2014). The services provided by forests in regulating local and regional climate (e.g. urban heat island, or cooling effects) through influencing the local energy balances, ET and precipitation patterns have been studied using computer simulation models (Liu, 2011), but these regional climate models need further parameterization, validation and refinement to enhance their prediction accuracy.

3.5.3 Measuring evapotranspiration everywhere all the time

Although large progress has been made in the past two decades towards measuring ET 'everywhere all the time' (Baldocchi *et al*., 2001; Baldocchi and Ryu, 2011), the study of ET is still regarded as an imprecise science (Shuttleworth, 2012). Research is needed to scale up or scale down among plot, watershed, regional and global scales to integrate methods and data (Amatya *et al*., 2014). In recent years remote sensing and radar technologies have advanced rapidly and enhanced our capability to accurately quantify water use and irrigation scheduling for croplands. However, the remote sensing applications in forest water management and water supply monitoring are rare. In fact, few studies have examined the accuracy of remote sensing-based ET products for forested areas. Forest ET measurements on the ground for calibrating remote sensing models are costly and the remote sensing techniques are often hampered by cloud cover and the complexity of multilayered tree canopies that vary spatially and temporally. For example, leaf clustering and light saturation problems are often problematic in estimating LAI for forests. Although images with high spatial and temporal resolution obtained from unmanned aerial vehicles may potentially play a role for precision agriculture and irrigation scheduling in the future, the validity of this method in estimating forest ET requires a significant amount of research (Amatya *et al*., 2014). The best approach to estimate ET for large watersheds is achieved by combining field hydrological measurements with high-resolution remote sensing and energy balance-based land-surface modelling (Wang *et al*., 2015).

3.5.4 New generation of ecohydrological models

Field measurements of ET at the leaf, tree, stand and landscape scale are essential to parameterize process-based hydrological models that have often not been validated with spatial and temporal distribution of various ET components (Sun *et al*., 2011b). The so-called 'equifinality' in hydrological models is common, partially due to the lack of understanding of ET processes or the lack of ET data for model verification. To develop reliable predictive models, there is a great need for better understanding of the interactions and feedback mechanisms of ET and other ecohydrological processes (Evaristo *et al*., 2015), including the canopy resistance factor used in the Penman–Monteith based ET models. More information is needed about how forest ET may be affected by species, density, stand age and management (managed versus natural forests, fertilization, thinning) in various ecoregions. Budyko's framework has been widely used to explain the mean spatial patterns of ET under land cover change (Zhou *et al*., 2015) and climate change (Creed *et al*., 2014). However, the model needs to be extended to finer temporal scale such as daily or seasonal to fully capture the dynamics of ET over time (Zhang *et al*., 2008; Wang *et al*., 2011). A new generation of ecohydrological models that combine the effects of CO_2 on ET processes and couple the physical and biological processes such as soil moisture redistribution, hydraulic distribution, photosynthesis, canopy conductance and tree growth is needed to fully understand the atmosphere–vegetation–soil processes mechanistically (Cheng *et al*., 2014). Such models can provide better information to regional land-surface and climate models for quantifying the feedbacks of forest cover change to regional and global climate systems. Oversimplified model designs in the ET processes likely contain errors in the computation of dry-season water balances and the associated heat fluxes, and thus in the possible feedbacks between soil moisture and climate (Bonetti *et al*., 2015).

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