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Long-term variability in the water budget and its controls in an oak-dominated temperate forest

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Abstract:

Water availability is one of the key environmental factors that control ecosystem functions in temperate forests. Changing climate is likely to alter the ecohydrology and other ecosystem processes, which affect forest structures and functions. We constructed a multi-year water budget (2004–2010) and quantified environmental controls on an evapotranspiration (ET) in a 70-year-old mixed-oak woodland forest in northwest Ohio, USA. ET was measured using the eddy-covariance technique along with precipitation (P), soil volumetric water content (VWC), and shallow groundwater table fluctuation. Three biophysical models were constructed and validated to calculate potential ET (PET) for developing predictive monthly ET models. We found that the annual variability in ET was relatively stable and ranged from 578 mm in 2009 to 670 mm in 2010. In contrast, ET/P was more variable and ranged from 0.60 in 2006 to 0.96 in 2010. Mean annual ET/PET_FAO was 0.64, whereas the mean annual PET_FAO/P was 1.15. Annual ET/PET_FAO was relatively stable and ranged from 0.60 in 2005 to 0.72 in 2004. Soil water storage and shallow groundwater recharge during the non-growing season were essential in supplying ET during the growing season when ET exceeded P. Spring leaf area index (LAI), summer photosynthetically active radiation, and autumn and winter air temperatures (T_a) were the most significant controls of monthly ET. Moreover, LAI regulated ET during the whole growing season and higher temperatures increased ET even during dry periods. Our empirical modelling showed that the interaction of LAI and PET explained >90% of the variability in measured ET. Altogether, we found that increases in T_a and shifts in P distribution are likely to impact forest hydrology by altering shallow groundwater fluctuations, soil water storage, and ET and, consequently, alter the ecosystem functions of temperate forests. Copyright © 2013 John Wiley & Sons, Ltd.

KEY WORDS evapotranspiration; water budget; interannual and seasonal variability; climatic warming; eddy-covariance; temperate deciduous forest

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INTRODUCTION

Temperate forests are particularly vulnerable to climatic change, which alters soil water dynamics and ecosystem structure and functions (IPCC, 2007). Global climate models predict an above average increase in air temperature (T_a) in northern mid-latitudes. An understanding of how temperate forests respond to climatic changes and variability is important because these forests play an important roles in the global C budget, climate moderation, and quality and quantity of freshwater supplies (Cox *et al.*, 2000; Welp *et al.*, 2007; Bonan, 2008; Sun *et al.*, 2011a).

Recent ecohydrology studies have focused on evapotranspiration (ET), a key hydrologic process that is critical to understanding the hydrologic response to climate change (Bonan, 2008; Jung *et al.*, 2010; Rao *et al.*, 2011; Jones *et al.*, 2012; Williams *et al.*, 2012). ET rates are an integral component of the water cycle that affects vegetation distribution and productivity, climate, and water resources across multiple spatial-temporal scales (Brutsaert, 1982; Bonan, 2008; Sun *et al.*, 2011b). Globally, terrestrial plant ET returns approximately 60% of annual precipitation (P) to the atmosphere (Oki and Kanae, 2006), and the rates are even higher in forested watersheds (Lu *et al.*, 2003; Sun *et al.*, 2011a, b). Small

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watershed-scale manipulation experiments have shown that changes in forest covers impact streamflow amounts and periodicity by altering ET processes (Swank and Douglass, 1974; Zhang *et al*., 2001; Ford *et al*., 2005). Global eddy-flux synthesis studies suggest that forests have very different ET rates compared with other ecosystems within the same climatic regimes (Williams *et al*., 2012). In spite of the importance of ET, quantifying temporal variability in ET and its environmental controls remain challenging because ET processes of forests are complex and costly to quantify on moderate temporal scales (Chen *et al*., 2004; Sun *et al*., 2011a).

Seasonal and interannual climatic changes directly alter the regional hydrological cycle by altering ET processes in temperate forests (Restrepo and Arain, 2005; Sun *et al*., 2008a). Responses of ET to a changing climate are uncertain. Bates *et al*. (2008) found that evaporative demand was expected to increase almost everywhere because the water holding capacity of the atmosphere increased with higher T_a. Yet, Roderick and Farquhar (2002) reported a long-term decrease in pan evaporation globally. Keenan *et al*. (2013) also found a long-term decrease in ET through eddy-covariance (EC) techniques. Moreover, ET responds differently to climatic and biophysical variables between forest types (Stoy *et al*., 2006; Chen *et al*., 2008). Granier *et al*. (2007) suggested that ET in coniferous forests appeared to be less impacted by the warming than ET in broadleaved forests. Yet, Gholz and Clark (2002) found that ET was more controlled by the climatic variability rather than forest type or age. Evidently, the majority of the literature supports a conclusion that multiple environmental factors regulate ET simultaneously (Ewers *et al*., 2002; Barker *et al*., 2009).

Existing hydrologic models in studying climate change are often unreliable due to ET model deficiency (Sun *et al*., 2011b). The ET models that are often embedded in hydrological models are rarely validated due to the lack of long-term ET measurements (Shuttleworth, 2008). The main techniques for estimating ET include soil water budget (Eastham *et al*., 1988), catchment water balance (Luxmoore and Huff, 1989), plant and soil weighing lysimeters (Daamen *et al*., 1993), Bowen ratio (Denmead *et al*., 1993), plant chamber (Cienciala and Lindroth, 1995), sap flow (Wullschleger *et al*., 1998), chemical tracing (Kalma *et al*., 1998), biophysical mathematical models (*e.g*. Penman–Monteith equation; McCarthy *et al*., 1992), and biophysical models (Zhou *et al*., 2008; Liu and Yang, 2010). However, these techniques often yield considerably different results due to differences in the scales of measurements (Wilson *et al*., 2001; Shuttleworth, 2008). Recent advancements in the EC technique allow for direct and continuous measurements of ET at an ecosystem scale (*e.g*. Chen *et al*., 2002; Sun *et al*., 2011a). The EC technique also offers an advanced method for developing the water budget of forest ecosystems on fine temporal scales (from 30 min to years) (Wilson *et al*., 2001).

Little is known about the long-term impacts of variations in both external (*i.e*. climatic) and internal (*i.e*. biophysical) drivers on variability in the ET of natural temperate forests on a seasonal and interannual basis (Barr *et al*., 2007; Campbell *et al*., 2011; Sun *et al*., 2011a). With this interest, we have maintained an EC tower in a 70-year-old mixed oak woodland forest, from which continuous measurements of water vapour fluxes were collected over a 7-year period from 2004 through 2010. The study site is within a large oak forest in Oak Openings Preserve Metropark of northwest Ohio, USA. The study region has a remarkable number of rare and endangered species, including 145 plants as potentially threatened or endangered in Ohio (Brewer and Vankat, 2004; Deforest *et al*., 2006). The study site is one of the US-China Carbon Consortium (USCCC) sites (Sun *et al*., 2009), which has been incorporated into the Ameriflux network of sites (http://public.ornl.gov/ameriflux/Site_Info/siteInfo.cfm?KEYID=us.ohio_oak.01).

The objectives of this study were the following: (1) to construct a 7-year water budget and quantify seasonal and interannual variability in water fluxes for an oak-dominated forest; (2) to determine internal (*i.e*. leaf area index, LAI) and external (*i.e*. P, temperature, solar energy, vapour pressure, and soil moisture) factors regulating ET; and (3) to develop empirical ET models that can be readily used to estimate monthly ET for similar temperate deciduous forests. We hypothesized that monthly ET can be adequately predicted with PET and leaf area dynamics.

MATERIAL AND METHODS

Site characteristics

The study site is located in an oak-dominated forest (N 41.55°, W 83.84°) near the city of Toledo, Ohio, USA (Figure 1). The mean annual T_a was 9.2 °C and P was 840 mm during 1971–2000 (http://www.ncdc.noaa.gov/oa/ncdc.html). Snow was considered and it made up an average of 10% of the annual P. Year 2004 had the highest value at 20% and 2007 had the lowest value at 5%. The research site is characterized by flat topography with elevations ranging from 200 to 205 m. The site sits on a band of sandy soil deposits along an ancient lakeshore that was created during the last glacial retreat, approximately 11 000 years ago. Sandy soil lies above a layer of clay that opposes the penetration of water and causes the perched groundwater table to be close to the surface. High exposed spots are xeric, providing conditions suitable for dry prairie communities, whereas

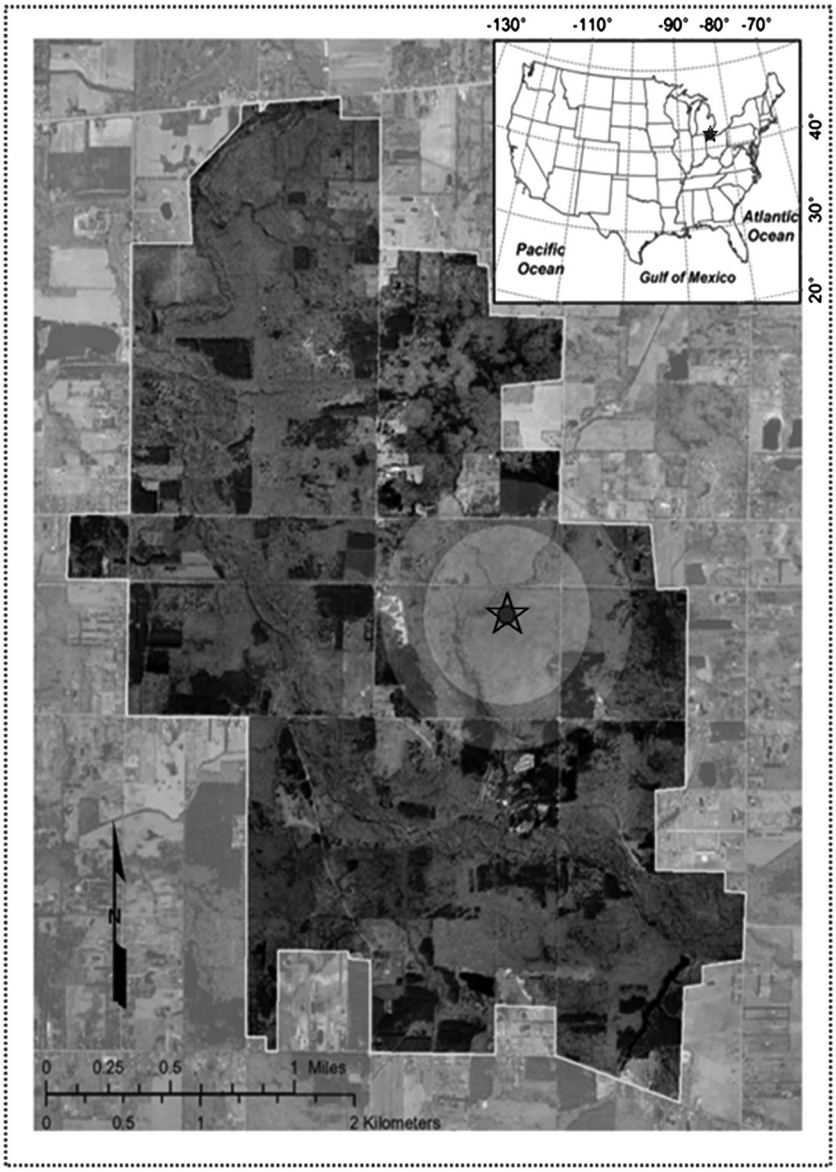

Figure 1. A long-term eddy-covariance flux tower and ancilary biophysical and physical measurements of microclimate, groundwater, and vegetation were installed in November, 2003 at the Oak Openings Preserve Metropark in northwest Ohio, USA.

low spots are moist to wet throughout the entire year, with standing water appearing in spring and winter. The landscape provides diverse habitats for different plant communities. The heights of dominant trees were ~24 m, with an average canopy height of ~20 m. As of 2012, the 70-year-old mixed oak woodland forest is dominated by *Quercus rubra* (red oak; 31%), *Quercus alba* (white oak; 26%), *Quercus velutina* (black oak; 14%), *Quercus macrocarpa* (Bur Oak; 8%), and other species including *Acer rubrum* (red maple; 10%), *Prunus serotina* (black cherry; 5%), *Sassafras albidum* (sassafras; 2%), and *Carya spp.* (Hickory; <1%) (Brewer and Vankat, 2004; Deforest *et al.*, 2006). We defined April through May as spring, June through September as summer, October through November as autumn, and the first 3 months and the last month of the year as winter. The growing season length was defined as the period between the first and the last occurrence of 3 continuous days when daytime carbon uptake exceeded 5% of the summer maximum carbon uptake and the growing season started in mid-May.

Eddy flux and meteorological measurements

The 34-metre-tall tower was surrounded in all directions by a uniform canopy of similar tree species and ages, extending to approximately 600 m of unbroken fetch. Following AmeriFlux protocols, turbulent fluxes of water vapour between the forest canopy and atmosphere were

measured using EC instruments placed at the tower tops. The open-path EC system consisted of a LI-7500 infrared gas analyzer (IRGA; Li-Cor Biosciences, Lincoln, NE, USA) and a 3-dimensional sonic anemometer [CSAT3; Campbell Scientific, Inc. (CSI), Logan, UT, USA]. The IRGA was calibrated every 4 to 6 months in the laboratory using zero grade nitrogen, a dew point generator (LI-610, Li-Cor, Inc., Lincoln, NE, USA), and NOAA/CMDL-traceable primary CO_2 standards.

The 30-min mean flux value was calculated as the covariance of vertical wind speed, T_a, and water vapour densities using the Webb–Pearman–Leuning correction (Webb *et al*., 1980; Massman and Lee, 2002) and the EC_processor software (http://research.eeescience.utoledo.edu/lees/ECP/ECP.html; Noormets *et al*., 2008). Wind coordinates to mean streamline plane were rotated (Wilczak *et al*., 2001), which were calculated from mean wind data over an entire year. Sonic temperature was corrected for changes in atmospheric humidity and pressure (Schotanus *et al*., 1983). Raw data spikes (> 6 standard deviations) were removed and 30-min fluxes for the warming of the IRGA above T_a were corrected (Burba *et al*., 2006; Grelle and Burba, 2007). Latent heat (W m^{-2}) was calculated from the difference between measured water vapour flux and storage change in water vapour flux in the canopy air space. The water storage was estimated as the mean rate of 30-min change in water vapour concentrations, measured at four heights (*i.e*. 1.5, 5, 16, and 22 m) using a LI-800 analyzer (Li-Cor) within the canopy (Noormets *et al*., 2007; Yang *et al*., 2007).

A number of meteorological factors at multiple vertical levels were measured and reported as 30 min means. Both T_a and relative humidity (RH, %) were measured by HMP45AC (Vaisala, Finland) above the canopy at the same height as the IRGA and below the canopy. Photosynthetically active radiation (PAR, μmol m^{-2} s^{-1}) was measured by a LI-190SB (Li-Cor) above and below the canopy. A CNR-1 sensor (Kipp and Zonen, Delft, The Netherlands) measured downward and upward shortwave and longwave radiation. Net radiation (R_n, W m^{-2}) was derived by summing up net shortwave and longwave radiation measurements using all four measured radiation components. P (mm) was measured using tipping-bucket-type raingauges, TE-525WS-L (Texas Electronics, Dallas, TX, USA). Soil temperature (T_s, °C) was monitored with CS107 temperature probes (CSI) at 5 (T_{s5}) and 25 cm (T_{s25}) soil depths. Soil heat flux (G, W m^{-2}) was measured at three locations using HFT3 flux plates (REBS, Seattle, WA, USA) buried 5 cm below the ground. Data were recorded at 30-min intervals along with other meteorological variables. LAI (m^2 m^{-2}) was obtained from 1-km resolution MODIS LAI/FPAR Collection 5 (http://daac.ornl.gov/MODIS/modis.shtml; Shabanov *et al*., 2005) with an online subset output of a 3 × 3 km pixel subset centered on the flux tower, which provided a time series of LAI from 2004 through 2010. There were large uncertainties in MODIS LAI estimates. For example, the uncertainties for the red and near infrared (NIR) bands were set to 30% and 15%, respectively, for deciduous and evergreen broadleaf forests (Shabanov *et al*., 2005). The estimated uncertainty on *in situ* LAI can be 15% and MODIS LAI values are slightly overestimated with values of 0.03–0.2 above *in situ* measurements (Fensholt *et al*., 2004).

Water budget

Constructing a monthly water budget is an effective way to examine how environmental factors interact with ecosystem processes (Sun *et al*., 2010). The key water fluxes for the water budget considered in this study included P (mm), ET (mm), deep seepage (Q, mm), changes in soil water storage (ΔS, mm), and measurement errors (ε) [Equation 1]. Soil water storage in the top 3 m of the soil includes both saturated and unsaturated zones in the soil profile. We installed a 3-m-deep shallow groundwater well to track the groundwater table fluctuations on an hourly basis to fully estimate ΔS, in addition to soil moisture measurements. The water table position was estimated based using a pressure transducer that measures water pressure above a sensor head (Infinities USA, Port Orange, Florida, USA). The water budget for a 3 m soil profile at the study site was expressed as (Sun *et al*., 2010; Lu *et al*., 2011):

$$P = ET + Q + \Delta S, \quad (1)$$

where ET is consistent with plant transpiration (*i.e*. dry canopy transpiration) and evaporation from soil and plant surfaces. Daily, seasonal, and annual ET were calculated by summing the total corresponding 30-min values converted from the latent heat of EC measurements. Q (mm) was deep seepage (*i.e*. water flowing out of the 3-m soil profile and water recharges to groundwater). After ΔS was estimated using Equations 2 and 3, Q was estimated as the residual of P, ET, and ΔS in Equation 1. In terms of ΔS in the saturated zone, porosity was 0.4, field capacity was 0.25, and drainable porosity was their difference, being 0.15. We assumed that surface runoff was negligible because the site was flat and infiltration rates of the sandy soil were high. The ΔS term was estimated for unsaturated and saturated zones using Equation 2 (Sun *et al*., 2010; Lu *et al*., 2011):

$$\Delta S = \Delta WT \times \theta_d + \Delta VWC \times 1000, \quad (2)$$

where ΔWT (mm) was the change in the depth of the groundwater table, and θ_d was soil drainable porosity that was estimated as the difference between soil porosity (*i.e*. saturated VWC) and field capacity. Soil volumetric water content (VWC, %) was averaged across the top 30 cm using vertically inserted CS616 Time Domain Reflectometer probes (CSI). ΔVWC (%) was changes in VWC for the

unsaturated zone in the top 1000 mm. We assumed that the changes in soil moisture in the unsaturated zone between 1 m and the water table level were small and negligible in the overall water budget. The following formula was used to estimate ΔS:

$$\Delta S = \left[\sum_{i=1}^{n} \theta_i D_i(t+1) + \theta_s D_s(t+1) \right] - \left[\sum_{i=1}^{m} \theta_i D_i(t) + \theta_s D_s(t) \right] \quad (3)$$

where θ_i is the soil moisture content at layer *i* that varies from *m* to *n* depending on the water table level from 1 month (or year) *t* to next month (or year) *t* + 1. D_i is the soil thickness interval (mm) corresponding to layer *i*. D_s is the saturated soil thickness (mm) above an arbitrary reference level and θ_s is the saturated soil moisture content or soil porosity.

Potential evapotranspiration

Because actual ET is rarely measured in the field, PET is widely used to approximate actual ET. In this study, we examined three PET models: the Food and Agriculture Organization (FAO) of the United Nations grass reference ET (PET_FAO; Allen *et al*., 1994) model, the Priestley—Taylor PET (PET_PT; Priestley and Taylor, 1972) model, and the Hamon PET (PET_Hamon; Hamon, 1961) model for practically estimating actual ET on a monthly scale. Daily PET was summed to derive monthly, seasonal, and annual values.

The Penman–Monteith model (Monteith, 1965) estimates ET as a function of T_a, VPD, available energy, atmospheric pressure, aerodynamic resistance (a function of wind speed and plant-canopy height and roughness), and canopy resistance (a measure of resistance to vapour transport from plants). The process-based Penman–Monteith ET model requires several climatic variables that are often not available, nor can the parameters be derived for large areas (Sun *et al*., 2011a). The PET_FAO model is a simplified version of the Penman–Monteith model (Monteith, 1965) and is widely used in agriculture as a standard way to represent energy conditions for a particular region. The PET_FAO model needs substantial corrections to provide ET estimates for certain landscapes (*e.g*. forests) at a daily or monthly scale (Sun *et al*., 2011a). Using the process-based Penman–Monteith ET equation, actual daily ET of a hypothetical well-watered grass (*i.e*. PET_FAO) that has a 0.12-m canopy height, a leaf area of 4.8, a bulk surface resistance of 70 s m^{-1}, and an albedo of 0.23 is estimated as follows

$$PET_{FAO} = \frac{0.408\,\Delta\,(R_n - G) + \gamma \frac{900}{T_a+273} u_2 (e_s - e_a)}{\Delta + \gamma(1 + 0.34u_2)}, \quad (4)$$

where Δ (kPa °C^{-1}) was the slope of saturation water vapour pressure versus temperature curve, R_n (MJ m^{-2}) was net radiation, G (MJ m^{-2}) was soil heat flux, γ (kPa ° C^{-1}) was the psychometric constant modified by the ratio of canopy resistance to atmospheric resistance, T_a (°C) was mean air temperature, e_s (kPa) was saturation vapour pressure, e_a (kPa) was actual vapour pressure, u_2 (m s^{-1}) was mean wind speed at 2 m height, and 900 was the unit conversion factor.

The Priestley–Taylor equation (PT; Priestley and Taylor, 1972) is a radiation-based semi-empirical model that was derived from the Penman–Monteith model. In the PT model, the atmosphere is assumed to be saturated, in which case the aerodynamic term is zero. Daily PET_PT (mm) was estimated by

$$\text{PET\_PT} = 1.26 R_n \frac{\Delta}{(\Delta + \gamma)\lambda}, \quad (5)$$

where λ (MJ kg^{-1}) was latent heat of vaporization (=2.501–0.002 × T_a), and R_n, Δ, γ, and T_a were the same as in Equation (4).

The PET_Hamon model (Hamon, 1961) has been widely used in regional hydrologic modelling in climate change due to its simplicity and few data requirements for future prediction (Vörösmarty *et al*., 1998; Lu *et al*., 2003; Rao *et al*., 2011). The model was based on T_a and theoretical daytime length and daily PET_Hamon (mm) was estimated by

$$\text{PET}_{\text{Hamon}} = 0.1651 \times L_{\text{day}} \times \frac{216.7 \times e_s}{\text{T}_\text{a} + 273.3}, \quad (6)$$

where L_{day} was the ratio of daytime length to 12 and e_s (mbar) was saturation vapour pressure at a given T_a.

Data quality control and Gap-filling method

All 30-min flux data from the EC tower were quality checked, including stationarity, integral turbulence characteristic, and friction velocity thresholds (Noormets *et al*., 2007, 2008). The periods with poor turbulent developments were filtered out and treated as gaps. Gaps were more frequent during the night than during the day and during winter than the other seasons. The mean percentage of gaps in latent heat was 33%, which was lower than the network-wide average of 35% (Falge *et al*., 2001). The longest gap was a 12-day gap during January in 2007. Short gaps (<1.5 h) were filled by linear interpolation and longer gaps (≥1.5 h) were filled using the mean diurnal variation method (Falge *et al*., 2001) for missing 30-min ET, R_n, and VWC. For the mean diurnal variation method, a missing value for any given 30-min gap was replaced with an average of values of the adjacent 7-day or 14-day windows, and this average was at exactly the same time of day.

Missing PAR gaps were filled by using the PAR-R_n regression models by month for our site. Missing T_a and RH gaps were filled with regression models between T_a

and RH above and below the canopy from existing 30-min data by month. T_{s5} gaps were filled using T_{s5}–T_{s25} regression models by month. Missing P gaps were replaced by the *P*-values from the Toledo Express Airport weather station (N 41.5886°, W 83.8014°), which was ~3 km northeast of the flux site. Saturation vapour pressure (e_s, kPa) and actual vapour pressure (e_a, kPa) were calculated using filled T_a and RH above the canopy. The vapour pressure deficit (VPD) was the difference between e_s and e_a. All data processing and statistical analyses were conducted using SAS 9.2 software (SAS, Institute Inc., Cary, NC, USA). Variables T_s, T_a, P, VWC, PAR, VPD, and LAI were identified as potential drivers (*i.e.* independent variables). We determined the contribution of each variable on C fluxes on a monthly scale with the Pearson correlation coefficient.

RESULTS

Microclimate and vegetation dynamics

Mean annual T_a was 9.9 °C from 2004 through 2010 (Figure 2a; Table I), and annual T_a during this period was always higher than the 1971–2000 mean of 9.2 °C. The highest T_a was 10.7 °C in 2006 and the lowest was 9.3 °C in 2008. The mean summer T_a was lower than that of past 30 years but mean T_a of winter, spring, and autumn were higher than those of past 30 years. The highest daily R_n ranged from 215 W m^{-2} in 2006 to 230 W m^{-2} in 2010, with mean daily values being 36 W m^{-2} in winter, 123 W m^{-2} in spring, 142 W m^{-2} in summer, and 30 W m^{-2} in autumn (Figure 2b). The highest mean LAI was 4.2 ± 0.37 (m^2 m^{-2}) in July (Figure 3). Mean LAI was 1.73 in spring, 3.73 in summer, and 0.79 in autumn (Figure 4). Annual VWC decreased while annual T_{s5}, T_a, PAR, R_n, VPD, and LAI increased from 2004 through 2010 (Table I).

Changes in water budget over time

There were contrasting annual changes in ET/P (*i.e.* ET index) and P. Year 2004 had the lowest cumulative annual P (at 668 mm), whereas year 2006 had the highest P (at 1019 mm) (Figure 2d; Table I). Although the mean annual P was lower than that of past 30 years, the mean spring and summer P were higher than the means of past

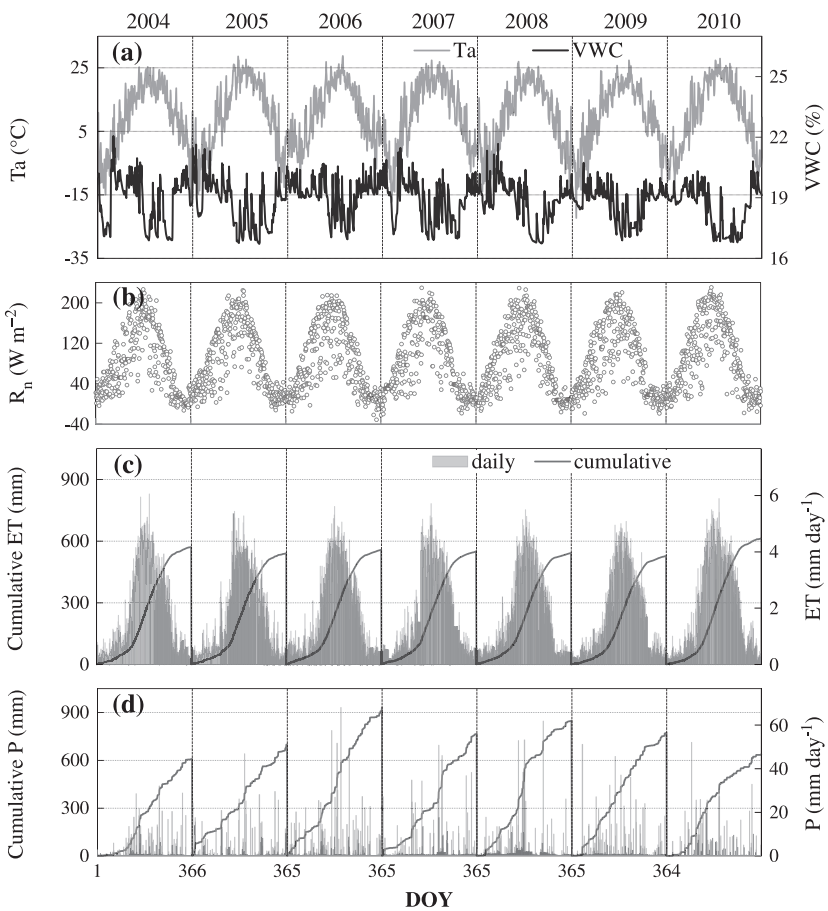

Figure 2. Daily (a) air temperature (T_a, °C) and soil volumetric water content (VWC, %) above the canopy, (b) net radiation (R_n, W m^{-2}), (c) daily and cumulative actual evapotranspiration (ET, mm day^{-1}), and (d) precipitation (P, mm day^{-1}) from 2004 through 2010.

Table I. Annual totals of evapotranspiration and precipitation and annual means of T_a, T_{s5}, VWC, PAR, VPD, R_n, and LAI from 2004 through 2010. ET (mm yr^{-1}) is evapotranspiration; P (mm yr^{-1}) is precipitation; T_a (°C) is air temperature above the canopy; T_{s5} (°C) is soil temperature at 5 cm; VWC (%) is soil volumetric water content in the top 30 cm; PAR (μmol m^{-2} s^{-1}) is photosynthetically active radiation; VPD (kPa) is vapour pressure deficit; R_n (W m^{-2}) is net radiation; and LAI (m^2 m^{-2}) is leaf area index. Values in parentheses indicate monthly standard error

Year	ET	P	T_a	T_{s5}	VWC	PAR	VPD	R_n	LAI
2004	626	668	9.5	10.0	18.88	186	0.41	80.7	1.67
	(12)	*(12)*	*(2.7)*	*(2.1)*	*(0.25)*	*(26)*	*(0.07)*	*(16)*	*(0.4)*
2005	593	766	10.0	10.1	18.68	210	0.54	86.5	1.76
	(12)	*(10)*	*(3.0)*	*(2.2)*	*(0.26)*	*(30)*	*(0.10)*	*(17)*	*(0.5)*
2006	611	1019	10.7	10.5	19.14	197	0.50	82.8	1.78
	(12)	*(14)*	*(2.4)*	*(2.0)*	*(0.19)*	*(29)*	*(0.08)*	*(17)*	*(0.57)*
2007	603	840	10.0	10.2	18.80	207	0.54	85.0	1.94
	(12)	*(15)*	*(2.9)*	*(2.1)*	*(0.23)*	*(31)*	*(0.10)*	*(17)*	*(0.5)*
2008	594	930	9.3	10.6	18.82	208	0.49	85.8	1.82
	(12)	*(19)*	*(2.8)*	*(2.1)*	*(0.25)*	*(30)*	*(0.09)*	*(18)*	*(0.5)*
2009	578	847	9.4	11.1	18.76	204	0.48	83.0	1.55
	(11.5)	*(10)*	*(2.7)*	*(1.7)*	*(0.19)*	*(28)*	*(0.08)*	*(17)*	*(0.4)*
2010	670	700	10.5	12.1	18.51	211	0.55	90.1	1.83
	(13.2)	*(11)*	*(3.0)*	*(1.9)*	*(0.27)*	*(29)*	*(0.09)*	*(18)*	*(0.5)*
Annual mean	*611*	*824*	*9.9*	*10.7*	*18.80*	*203*	*0.50*	*84.8*	*1.76*
(±SE)	*(30)*	*(125)*	*(0.6)*	*(0.7)*	*(0.19)*	*(9)*	*(0.05)*	*(3.1)*	*(0.1)*

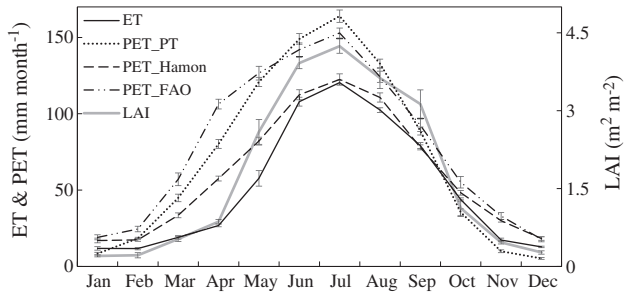

Figure 3. Comparisons among mean monthly values (with standard error) of actual evapotranspiration (ET, mm), potential evapotranspiration (PET, mm; *i.e.*, PET_FAO, PET_PT, and PET_Hamon), and leaf area index (LAI, m^2 m^{-2}) over 7 years.

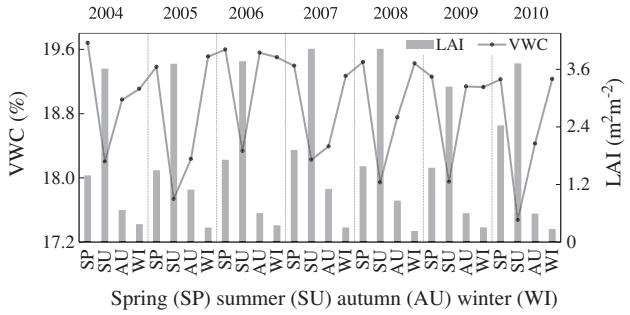

Figure 4. Seasonal (spring SP, summer SU, autumn AU, and winter WI) soil volumetric water content (VWC, %) in the top 30 cm and leaf area index (LAI, m2 m-2) from 2004 through 2010.

30 years (168 *vs* 162 mm and 377 *vs* 321 mm, respectively). Year 2010 had the lowest summer P (at 265 mm), whereas 2008 had the highest value (at 531 mm), which were 18% lower and 65% higher than mean summer P of the 30-year mean, respectively. Although P was highly variable seasonally and interannually, ET remained relatively stable (Figure 2c and d; Table I). Annual ET varied from 578 mm in 2009 ($P = 847$ mm in 2009) to 670 mm in 2010 ($P = 700$ in 2010). In contrast, mean annual ET/P was 0.74, ranging from 0.60 in 2006 to 0.96 in 2010 (Figure 5a). The largest difference in annual ET was 92 mm (15% of 7-year mean annual ET), whereas that of P was 351 mm (43% of 7-year mean annual P) among the 7 years. Mean annual ET/PET_FAO (*i.e.* ET efficiency) was 0.64, ranging from 0.60 in 2005 to 0.72 in 2004, whereas the mean annual PET_FAO/P (*i.e.* dryness index) was 1.15, ranging from 0.94 in 2006 to 1.47 in 2010 (Figure 5a). Total ET was lower than P across all study years (Table I). Annual difference between P and ET varied widely from 30 mm in 2010 (a dry year; $P = 700$ mm) to 408 mm in 2006 (a wet year; $P = 1019$ mm).

The mean ET was lower than that of P in winter, spring, and autumn but not in summer (Figures 6 and 7b). Seasonal ET was significantly correlated with seasonal P and LAI ($R^2 = 0.70$ and 0.87, respectively, with $P = 0.0001$, $N = 28$). The VWC was at a maximum level in the spring and a minimum level in the summer (Figure 2a). Spring VWC significantly decreased during the 7 years. VWC significantly decreased during April through September in 2004, 2008, 2009, and 2010. Mean annual ET, ΔS, and Q accounted for 78%, −3%, and 25% of the mean annual P, respectively. Approximately 39% of seasonal ΔS and all summer ΔS were negative and Q had relatively small negative values with the exception of winters (Figure 7).

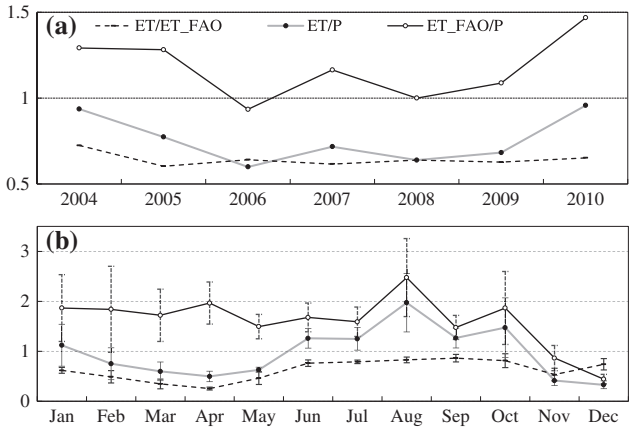

Figure 5. (a) Annual evapotranspiration efficiency (ET/PET_FAO), ET index (ET/P), and dryness index (PET_FAO/P) and (b) their mean monthly values (with standard error) from 2004 through 2010.

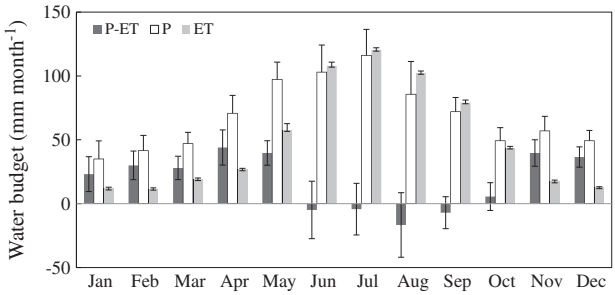

Figure 6. Mean monthly values (with standard error) of precipitation (P, mm), actual evapotranspiration (ET, mm), and the difference between precipitation and evapotranspiration from 2004 through 2010.

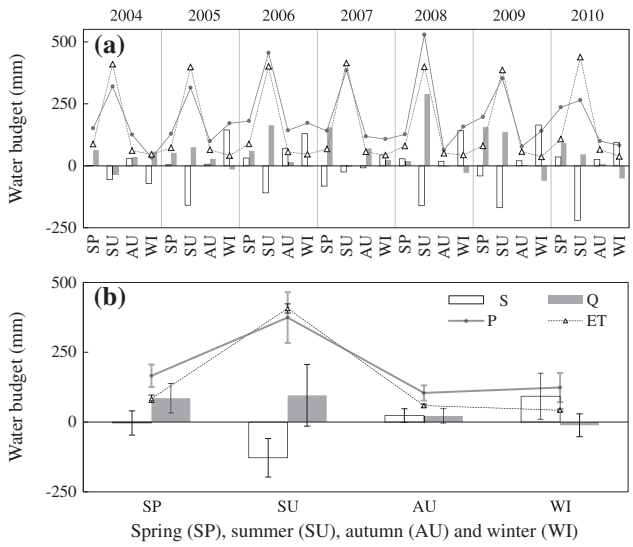

Figure 7. (a) Seasonal (spring SP, summer SU, autumn AU, and winter WI) and (b) 7-year means (with standard error) of the water budget, including changes in soil water storage (ΔS, mm), estimated deep seepage (Q, mm), precipitation (P, mm), and actual evapotranspiration (ET, mm) from 2004 through 2010.

Climatic and biophysical regulations

Monthly ET was significantly correlated with the following variables, in the significance order from highest to lowest: vegetation biomass (LAI) > energy inputs ($T_{s5} > T_a >$ PAR) > atmospheric demand (VPD) > water inputs (VWC > P) on an annual basis (Table II). The two most important drivers on monthly actual ET were LAI (Figure 8a) and T_{s5} for the spring, PAR (Figure 8b) and T_a for the summer, T_a (Figure 8c) and T_{s5} for the autumn, and T_a (Figure 8d) and VWC for the winter, respectively (Table II). Only T_a and PAR were significantly correlated with ET in all four seasons of study years. LAI explained >90% of the variability in monthly ET on an annual basis.

Modelling seasonal evapotranspiration

Basically, different combinations of environmental variables against PET by the three methods [*i.e.* Equations (4), (5), and (6)] were tested to derive the best-fit model for estimating ET on monthly scale. The models were evaluated to ensure independence among all variables (*i.e.* no multicollinearity) with the following statistics:

$$\mathrm{ET} = 16.96 + 0.16 \times \mathrm{PET_{FAO}} \times \mathrm{LAI}, \quad (7)$$

Adjusted R^2=0.91, p <0.001, root mean square error (*RMSE*) = 12.10 mm month^{-1}, and coefficient variance (CV) = 23.8.

$$\mathrm{ET} = 19.01 + 0.15 \times \mathrm{PET_{PT}} \times \mathrm{LAI}, \quad (8)$$

Adjusted R^2=0.91, p <0.001, *RMSE* = 11.8 mm month^{-1}, and CV = 23.2.

$$\mathrm{ET} = 17.51 + 0.20 \times \mathrm{PET_{Hamon}} \times \mathrm{LAI}, \quad (9)$$

Adjusted R^2=0.93, p <0.001, *RMSE* = 10.4 mm month^{-1}, and CV = 20.4.

Surprisingly, the three models produced similar forms and accuracies. Both LAI and PET were important variables and the combination of the two was sufficient to predict monthly forest ET.

DISCUSSION

Evapotranspiration and water budget variability

The annual site ET (578–670 mm) and the highest daily ET (4.6–5.5 mm) were moderate over the 7-year study period compared with data reported for other temperate deciduous forests. Mean annual ET measured by the EC method was reported to be 579 mm, and the highest daily ET was 4–5 mm in a broadleaved deciduous forest in the Appalachian Mountains that had a warmer and wetter climate (Wilson and Baldocchi, 2000). Our results were similar to the annual ET rates of 515 mm to 644 mm

Table II. Pearson correlation coefficient (PCC) values of the most significant environmental variables on monthly evapotranspiration on a seasonal and annual basis over 7 years (the significance of p values <0.0001, <0.001, and <0.01 *vs* ***, **, and *)

	Variables	PCC values
Spring	LAI^{***}, T_{s5}^{***}, T_a^{**}, PAR^{*}	0.88, 0.85, 0.81, 0.63
Summer	PAR^{***}, T_a^{***}, LAI^{***}, VPD^{**}, T_{s5}^{**}	0.86, 0.80, 0.66, 0.63, 0.61
Autumn	T_a^{***}, T_{s5}^{***}, PAR^{***}, VPD^{**}, LAI^{**}, VWC^{*}	0.92, 0.91, 0.90, 0.74, 0.73, −0.64
Winter	T_a^{***}, VWC^{***}, VPD^{***}, P^{**}, PAR^{**}	0.88, 0.75, 0.73, 0.51, 0.49
Annual	LAI^{***}, T_s^{***}, T_a^{***}, PAR^{***}, VPD^{***}, VWC^{***}, P^{***}	0.97, 0.92, 0.92, 0.84, 0.82, −0.68, 0.51

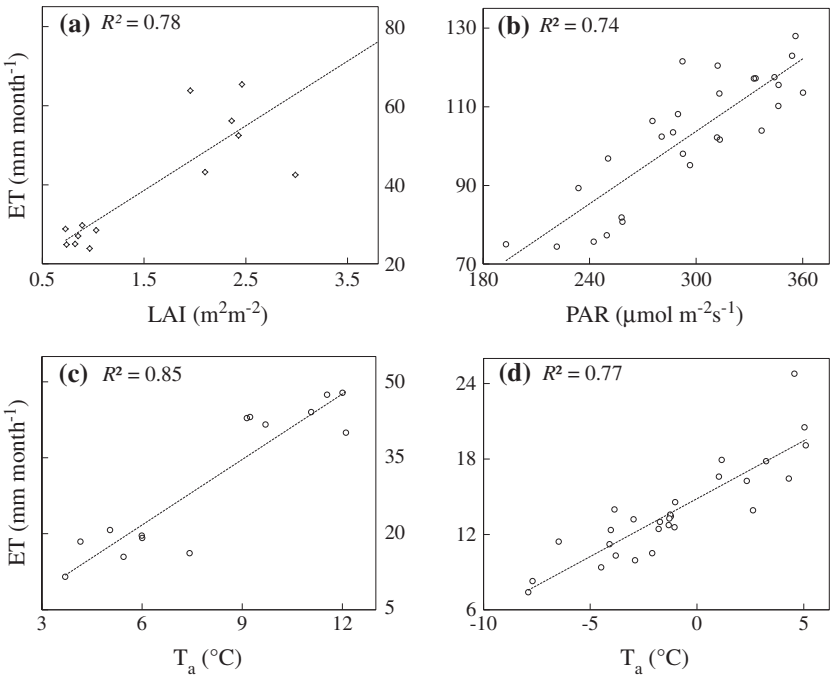

Figure 8. Representative scatter plots to show linear relationships between monthly actual ET (mm) and (a) LAI ($m^2 m^{-2}$) and (b)T_{s5} (°C) in spring, (c) PAR (μmol $m^{-2} s^{-1}$) and (d) T_a (°C) in summer, (e) T_a in autumn, and (f) T_a in winter over the 7-year study period. See Table I for abbreviations.

reported by Hanson *et al.* (2004) for an upland, oak-dominated forest of eastern Tennessee (N 35.95°, W 84.28°; 250 ± 330 m elevation), with long-term (50 years) mean annual P being 1352 mm, mean annual T_a being 14.2 °C, and the soils being primarily typical Paleudults. The interannual variability in ET is comparatively smaller than that of P, which is consistent with other long-term EC measurements (Ohta *et al.*, 2001; Grünwald and Bernhofer, 2007; Granier *et al.*, 2008). It was interesting to note that the wettest year of 2006 (P = 1019 mm, ET = 611 mm) did not have the highest ET. Year 2009 had the lowest annual ET (ET = 578 mm, P = 847 mm) but it was not the driest year, this is because its groundwater and P supported its annual ET. The groundwater table of the driest year (*i.e.* 2004) dropped to its lowest recorded level (>2.5 m) from 2004 through 2010 (Figure 9) and this year had the lowest P (P = 668 mm, ET = 626 mm).

Availability of both water and energy and biomass dynamics affect actual water loss (Williams *et al.*, 2012). Although annual ET/P was highly variable (Figure 5a), especially during the summer months (Figure 5b), the annual variability in ET and ET/PET_FAO remained stable (Table I; Figure 5). On an average, according to annual ET/P (mean = 0.74) and PET_FAO/P (=1.15) found at our site, our site fell between the mean theoretical Budyko (1974) curve and the best-fit curve for all eddy flux sites as presented in Figure 4 in Williams *et al.* (2012). Our study suggested that the general Budyko model produced reasonable estimates of long-term annual ET for the study site. Mean ET and PET reached the maxima in July (Figure 3). PET rates were higher than those of ET on monthly and annual scales, suggesting that the forest used less water than well-watered grass (*i.e.* PET_FAO). The order of mean PET

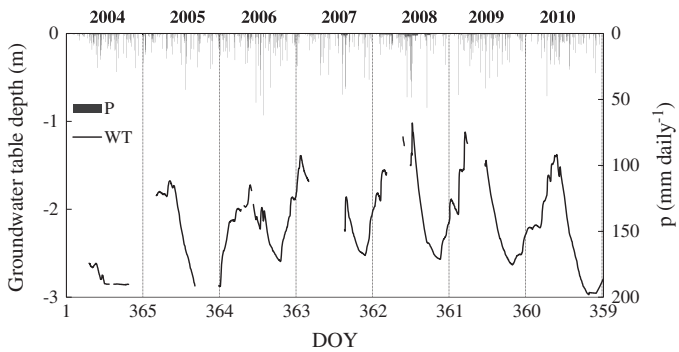

Figure 9. Daily dynamics of groundwater table depth (WT, m) and precipitation (P, mm day^{-1}) from 2004 through 2010. Groundwater fluctuations were given hourly, and the measurement limit was approximately at a depth of 3 m.

and actual ET from highest to lowest were: PET_FAO PET_PT > PET_Hamon > ET on the monthly and annual bases (Figure 3). Moreover, PET_FAO was much higher than ET from March through June than in other months when LAI increased relatively dramatically, which suggests the roles of LAI in regulating the variations in ET.

The water budget model [Equation 1] appeared to produce reasonable estimates of the water budget on a monthly scale and over a long period (Figure 7). Seasonal patterns of Q and ΔS were dictated by those of ET. However, in some cases, the method produced negative Q during winters (Figure 7). Thus, ΔS might be overestimated during winter months. Q varied from 96 mm in 2010 to 274 mm in 2008, representing 12% and 33% of their mean annual P, respectively. These values were higher than those of 1 to 10 mm yr^{-1} of a woodland (Knight *et al*., 2002) and that of 9 to 44 mm yr^{-1} of a broadleaved forest on sandy soils under a much drier climate (Zeppel *et al*., 2006). However, the estimates are somewhat lower than the modelled water yield (~300 mm yr^{-1}) for the study region by Sun *et al*. (2011b). High infiltration rates and low water-holding capacity of sandy soils, flat topography, and a cool climate all contributed to the high deep seepage. High ET of oak forests reduced deep seepage and streamflow in the growing season.

Soil water storage and shallow groundwater at this study site played an important roles in meeting ET demands in the growing season. Although annual P exceeded ET by 30%, mean summer P was lower than that of ET by 9% (Figure 6). An increase in T_a generally meant an increase in water loss and thus, a decrease in the groundwater table (Sun *et al*., 2008b). Although P significantly increased in the spring during the 7 years at this study site, annual higher T_a and associated increasing ET negated the positive impacts of the proportionally highest amounts of summer P and spring VWC, as indicated by a clearly decreasing groundwater table (Figures 4 and 9) and VWC (Figure 2a) in the growing season. Therefore, soil water storage of the growing season was an important water supplement for the increasing ET. Moreover, soil water and shallow groundwater, which were recharged from non-growing season P, were essential in supporting increasing ET (Figure 6). Significant controls of soil moisture on the water budget are also found at European forest sites (Baldocchi *et al*., 2001). Similarly, Zha *et al*. (2010) found that a shallow groundwater table allowed trees to maintain high levels of ET even under dry summer conditions. Teuling *et al*. (2010) also found that increasing forest ET due to climatic warming might rely on soil water accessed by a deep root system.

Climatic and biophysical controls on the trend of hydrologic fluxes

Long-term monitoring of the water budget allows for the quantification of temporal variability in major hydrologic fluxes and determines the environmental drivers of ET on a seasonal and interannual basis (Table II). Our results indicated that both internal (*i.e*. LAI) and external (*i.e*. T_a, P, PAR, VPD, VWC, and T_{s5}) drivers had significant contributions to monthly ET over 7 years. Annual LAI and T_{s5} were positively correlated with the observed annual ET. Global synthesis studies using the EC measurements also suggest that LAI is the most important variable controlling ET for deciduous forests (Zha *et al*., 2010; Sun *et al*., 2011a). Higher T_{s5}, T_a, PAR, and LAI contributed to the increased ET in the spring. Early leaf emergence caused by the spring warming enhanced ET. Higher T_{s5}, T_a, PAR, VPD, and LAI explained increasing ET in the summer. Jung *et al*. (2010) concluded that increasing ET caused by a warmer climate eventually depleted soil moisture. A previous study also suggested that high temperatures led to water stress and increasing ET rates under certain conditions. Higher T_{s5}, T_a, PAR, VPD, and LAI explained an increasing ET trend in the autumn.

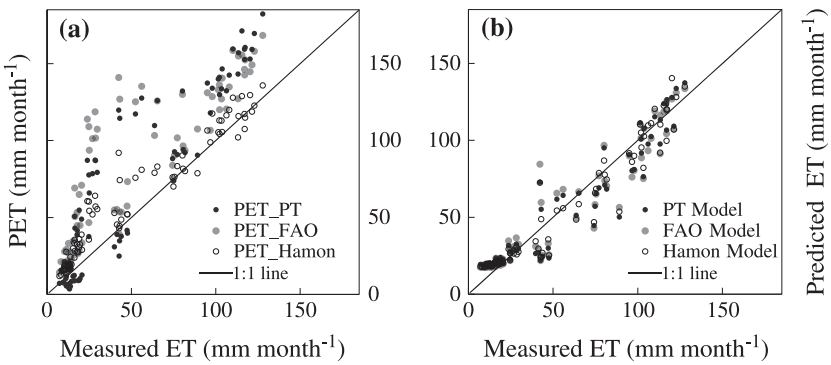

Figure 10. Monthly linear relationships between (a) actual evapotranspiration (mm) by measured by the EC technique and potential evapotranspiration (PET, mm; *i.e.* PET_FAO, PET_PT, and PET_Hamon) and (b) between measured ET and predicted ET by FAO, Priestley-Taylor, and Hamon PET models.

Monthly evapotranspiration model

Practical models have been widely used to estimate ET due to the high cost of measuring ET on large time and space scales (Sun *et al*., 2011a, b). We parameterized three ET models that use three different PETs containing different data input requirements so that users have three options for estimating ET depending on the availability in the input data. It appears that all three models produced similar accuracy, and users may have complex [PET_FAO in Equation 7], medium complex [PET_PT in Equation 8], or simple [PET_Hamon in Equation 9]. Equation 9 is the most practical model for this site because PET_Hamon produces values closest to measured ET (Figure 10b). Equation 9 indicated that the interaction of PET_Hamon and LAI explained the majority (93%) of the variability in ET for the site in which it was developed. Actual ET was more correlated with predicted ET than with PET (Figure 10 b and a). These monthly modelling results suggested that ET was controlled by interactions of PET and plant biomass. A scatter plot suggests that these models can accurately predict monthly ET (Figure 10b). Three PETs are widely used for calculating the PET for ecosystems and our ET models can be used for estimating actual ET in oak forests if meteorological and basic parameters are available. Our ET models offer new insights into basic external and internal environmental controls on ET. These models can assist in quantifying ET under normal climatic conditions and for estimating the mean monthly ET controls of similar temperate forests.

CONCLUSIONS

The 7-year continuous ET and hydrologic measurements from this study using EC and other monitoring techniques provided a good opportunity to examine the long-term responses of the water budget to environmental variabilities in an oak-dominated temperate forest. The seasonal and interannual variabilities in ET are significantly controlled by the external (*i.e.* T_a, P, PAR, VPD, VWC, and T_{s5}) and the internal (*i.e.* LAI) drivers. ET is a major driver of the forest-water budget (soil moisture, groundwater table, and deep seepage) on multiple temporal scales. Three ET models developed from this study offer convenient ways to empirically estimate monthly water loss from similar ecosystems. The occurrences of temperature rises during the long period enabled the analysis of warming impacts on water fluxes. Contrasting seasonal ET responses to the warmer conditions reflected the adaptation of natural temperate forests to the climatic changes and the dynamics of soil moisture and the groundwater table. We found that the growing season water loss exceeded the amount of natural precipitation. Soil water storage and shallow groundwater recharged from non-growing season precipitation were important in maintaining ecosystem functions in the growing season when water use exceeded precipitation. When climatic variability and droughts increase in the study region, maintaining groundwater is essential to the sustainability of the oak forest ecosystem.

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