HYDROLOGICAL PROCESSES
Hydrol. Process. **24**, 2405–2416 (2010)
Published online 24 March 2010 in Wiley InterScience
(www.interscience.wiley.com) DOI: 10.1002/hyp.7643

Evapotranspiration of a tropical rain forest in Xishuangbanna, southwest China

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Abstract:

The eddy covariance flux measurements from 2003 to 2006 at a tropical rain forest site in Xishuangbanna, southwest China, were used to study the ecosystem evapotranspiration (ET). During the four study years, the mean annual precipitation and air temperature were slightly lower than their long-term averages (1322 vs 1487 mm and 20·1 vs 21·7 °C, respectively). Mean annual ET in the four study years was estimated at 1029 mm, accounted for 78% of the corresponding annual precipitation. Mean daily ecosystem ET was 2·6 mm day^{-1} during the dry season from November to April and was 3·1 mm day^{-1} during the rainy season from May to October. The ET diurnal variations in the cool-dry season (from November to February), the hot-dry season (March and April) and the rainy season had similar maximum values, although the soil water content, leaf area index (LAI) and climate conditions differed greatly. ET was mainly controlled by soil water availability in the hot-dry season, by LAI in the early rainy season (May and June) and by atmospheric conditions in the mid-to-late rainy season (from July to October) and the cool-dry season. The total ET was substantially higher than the corresponding precipitation in the dry season. The extra amount of water evapotranspirated in the dry season was mainly due to the depletion of soil water stored in the previous rainy season. Fog deposition during the dry season also played a role in providing water for ET. Our results indicated the importance of interannual interactions of water balance in the seasonal distribution of ecosystem ET at this site.

KEY WORDS tropical rain forest; evapotranspiration; eddy covariance measurement; water balance

Received 9 July 2009; Accepted 1 February 2010

INTRODUCTION

Tropical forests comprise 60% of the total global forest area (Food and Agriculture Organization, 2001). These forests are a major source of land surface evapotranspiration (ET) and have profound impacts on regional and global climate and hydrological cycles (Nobre *et al.*, 1991; Choudhury *et al.*, 1998; Malhi *et al.*, 2002). Field investigations on hydrological cycles of tropical forests have been largely conducted in the Amazon region (Shuttleworth, 1988; Jipp *et al.*, 1998; Carswell *et al.*, 2002; Malhi *et al.*, 2002; da Rocha *et al.* 2004; Negrón Juárez *et al.*, 2007). These studies showed large spatial variability in ecosystem ET. The proportion of ET in precipitation over the region ranged from 50% to 90%. The seasonal variation of ET in Amazonian tropical forests also varied remarkably with locations. For example, ET was found lower in the dry season than that in the wet season in central Amazon (Malhi *et al.*, 2002), but higher in eastern Amazon (Carswell *et al.*, 2002; da Rocha *et al.*, 2004). Some studies suggested that ET was mainly determined by atmospheric evaporative demand followed by the seasonal variations in leaf area index (LAI) (Carswell *et al.*, 2002; da Rocha *et al.*, 2004), whereas other studies showed the importance of available soil water (Burgess *et al.*, 1998), rooting-depth (Nepstad *et al.*, 1994) or the upward flow of soil water such as hydraulic lift (da Rocha *et al.*, 2004) in determining ET. ET studies in the tropical forests of southeast Asia are limited and mainly conducted in Malaysia (Tani *et al.*, 2003; Kumagai *et al.*, 2005), Indonesia (Calder *et al.*, 1986) and Thailand (Pinker *et al.*, 1980; Tanaka *et al.*, 2008). Results from these studies showed that the ET at the tropical hill evergreen forest in Thailand was mainly determined by the atmospheric evaporative demand (Tanaka *et al.*, 2008), whereas the ET at the Malaysian tropical forests, which are located near the equator with a indistinct dry period, was relatively constant throughout the year (Tani *et al.*, 2003; Kumagai *et al.*, 2005). Tanaka *et al.* (2008) reviewed ET studies in southeast Asia and highlighted its importance in the regional and global climate. The controlling mechanisms of ET over this region are highly complex due to the diverse ecosystem types and climate regimes.

Xishuangbanna in southwest China is located at the northern edge of the tropical zone in southeast Asia (Figure 1). Despite its relatively high latitude, Xishuangbanna has a tropical moist climate due to the impact

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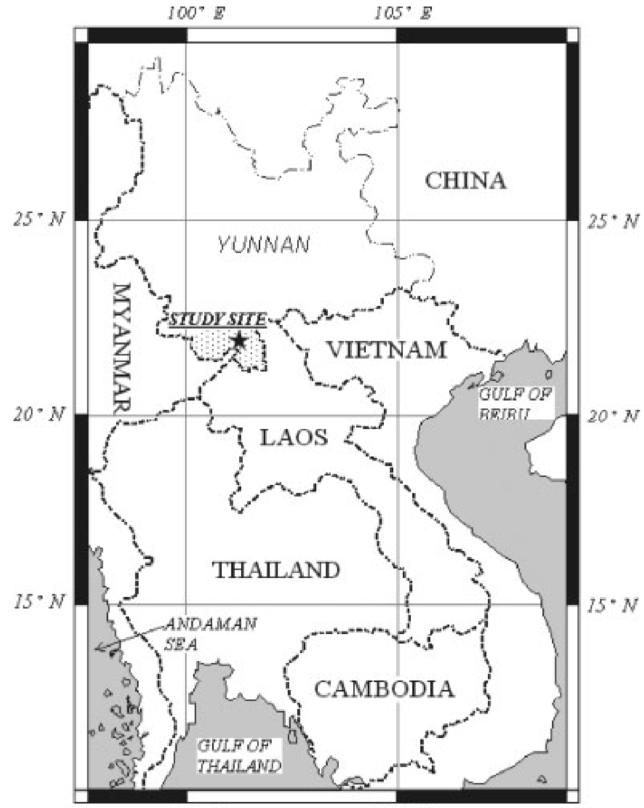

Figure 1. Location of the study site (indicated by a solid star; 21°55′39″N; 101°15′55″E) in Xishuangbanna, Yunnan Province, southwest China

of the Himalayas Mountains. The Himalayas Mountains lead to the penetration of warm and moist tropical air from the Indian Ocean to Xishuangbanna in summer, and blockade cold air from the north to the region in winter (Cao *et al.*, 2006). The Xishuangbanna tropical rain forests are unique in forest types compared with those in the equatorial region of southeast Asia. They have highly diverse and mixed types of floristic compositions due to the unique geographical location, which is in an ecotone with tropical zone to the south and subtropical zone to the north. Xishuangbanna is thereby very special in biogeography in southeast Asia. In the past 30 years, the Xishuangbanna tropical rain forests have been considerably replaced by rubber tree plantations and agriculture crops due to the developing economy in the region. This land cover change would potentially alter the regional hydrology and climate although the details have not been adequately studied. Consequently, it is highly desirable to understand the processes that control the ecosystem energy, water and carbon cycles over the region.

To fill this knowledge gap, the Chinese Terrestrial Ecosystem Flux Research Network (CTEFRN, ChinaFlux) initiated a tower flux measurement station at the end of 2002, aiming at the long-term measurements of energy, water vapour and CO_2 fluxes in the tropical rain forest in Xishuangbanna. In this study, we present the ecosystem ET estimates using 4 years' flux measurements (2003–2006) at this site. The main objectives of this article are: (1) to quantify the ecosystem ET and its temporal variations, and (2) to analyse the major environmental controls on ET in this tropical rain forest.

MATERIALS AND METHODS

Study site

The measurement site (21°55′39″ N, 101°15′55″ E, 750 m asl) is located at a National Natural Reserve in Xishuangbanna, Yunnan Province, China (Figure 1). It is a long-term site for ecosystem research and managed by the Xishuangbanna Station for Tropical Rain Forest Ecosystem Studies (XSTRFES), Chinese Academy Sciences (CAS). The terrain over the region is complex and the measurement site has a slope of about 20°. The vegetation at the site is dominated by tropical seasonal rain forests. The canopy structure is complicated and can be generally separated into three vertical layers (Zhu *et al.*, 2006). The upper layer is more than 30 m tall (up to 60 m) with about 30% crown coverage. The dominant species are *Pometia tomentosa* and *Tenninnalia myriocarpa*. The middle layer contains the main canopy, which is up to 30 m tall with a high crown closure (70–80% coverage) and stand density [796 ha^{-1}; >5 cm diameter at breast height (dbh)]. The dominant species are *Gironniera subaequalis* and *Barringtonia macrostachya*. The lower layer is about 5–18 m tall with a crown coverage of about 40%. The dominant species are *Milleuia laptobotrya, Mezzettiopsis creaghii* and *Myristica yunnanensis*. The stands are mainly broadleaf evergreen although there are some deciduous trees. Total canopy LAI is about 5·58 $m^2\ m^{-2}$, with variations between 3·56 $m^2\ m^{-2}$ in the hot-dry season of March and April and 6·34 $m^2\ m^{-2}$ in the rainy season of May through October (Zhang *et al.*, 2006). More detailed descriptions of the vegetation at this site can be found in Cao *et al.* (1996, 2006) and Zhu *et al.* (2006). The soil at the site is sandy clay loam (latosol), developed from siliceous rocks. The sand, clay and silt contents are about 47·8–57·4%, 19·5–29·5% and 22·8–23·7%, respectively. Soil bulk density ranged from 1410 to 1620 kg m^{-3} within the top 1 m of soil layer. Water table depends on precipitation and varies from 1·8 to 2·2 m (data provided by XSTRFES, CAS).

With the predominant effect of the Asian Monsoon, the climate in the study region is strongly seasonal. The long-term (1959–2002) climate records show that the mean annual total precipitation is 1487 mm, and mean annual air temperature is 21·7 °C (data provided by XSTRFES, CAS). There is a distinctive dry season and wet season throughout the year. The wet season (rainy season) is from May to October, with 87% (1294 mm) of annual total precipitation occurring during this period. Heavy precipitation events in the rainy season are mainly brought by the southwest summer monsoon from the Indian Ocean and the Bay of Bengal which are more than 800 km away from the study site. The rest of the year has little precipitation and is known as the dry season. The dry season can be further divided into a cool-dry season from November to February and a hot-dry season from

March to April. The cool-dry season is characterized by low temperature (seasonal mean value of 15·8 °C) and high frequency of radiation fogs at night and in the morning. The hot-dry season is characterized by high temperature (seasonal mean value of 21·3 °C, daytime value can exceed 38 °C) in the afternoon and radiation fogs in the morning only (Liu *et al.*, 2005).

Field measurements

Eddy covariance (EC) technique was used to measure latent heat (LE), sensible heat (H) and CO_2 fluxes at the study site. The EC wind speed and virtual temperature were measured using a three-dimensional sonic anemometer (Model CSAT-3, Campbell Scientific Inc., Logan, UT, USA). The EC water vapour and CO_2 concentrations were measured using an open-path infrared gas analyser (IRGA; Model LI-7500, Li-cor Inc., Lincoln, NE, USA). All these sensors were installed at the height of 48·8 m above the ground surface on a tower. To minimize the potential disturbance of the tower on measurements, these sensors were oriented in the direction of 210° from north (the dominant wind direction). These measurements were made at a frequency of 10 Hz and data were recorded by a data logger. After removing unreasonable spikes, data caused by sensor and logger malfunctions and discontinuities in the raw 10-Hz data, means and standard deviations (SDs) of wind (U_x, U_y, U_z), air temperature (T_{sonic}), water vapour (H_2O) and carbon dioxide (CO_2) concentrations, and air pressure were computed at a 30-min period. To eliminate horizontal and vertical advections, the coordinate system rotation for the three-dimensional wind speed was applied according to Tanner and Thurtell (1969). The LE, H and CO_2 fluxes were then calculated using the covariance of the rotated vertical wind speed and the water vapour concentration, air temperature and CO_2 concentration, respectively. Fluxes were also corrected for the effects of variation in air density after Webb *et al.* (1980). Errors in flux estimation under stable conditions were examined using friction velocity (u^*) following Saleska *et al.* (2003) and Miller *et al.* (2004). We observed decreasing trends in CO_2 flux when u^* was less than 0·2 m s^{-1}. In this study, all the flux data obtained at $u^* < 0{\cdot}2$ m s^{-1} were filtered and gap-filled.

Additional air temperature, relative humidity (RH) and wind speed were measured at the heights of 4·2, 16·3, 26·2, 36·5, 42·0, 48·8 and 69·8 m above the ground surface using platinum resistance thermometers, capacitive RH chips (Model HMP45C, Vaisala Inc., Helsinki, Finland) and anemometers (Model A100R, Vector Instrument, RHYL, North Wales, UK, UK), respectively. Soil temperature measurements were made at depths of 0·0, 0·05, 0·1, 0·15, 0·2, 0·4, 0·6, 0·8 and 1·0 m using thermocouple probes (Model 105T, Campbell Scientific Inc., Logan, UT, USA). Soil volumetric water content was recorded at depths of 0·05, 0·2 and 0·4 m using water content reflectometers (Model CS-616-L, Campbell Scientific Inc., Logan, UT, USA). Two soil heat flux plates (Model HFT-3, Campbell Scientific Inc., Logan, UT, USA) were placed at a depth of 0·05 m below the ground surface. Precipitation was recorded by a rain gauge (Model 52203, R.M. Young Inc., Traverse City, MI, USA) installed on the top of the tower which was about 70·2 m above the ground.

Net radiation (R_n) was measured on the tower at 41·6 m above the ground using a four-component net radiometer (Model CNR-1, Kipp & Zonen Inc., Delft, the Netherlands) which measures downward and upward short- and long-wave radiations. Measurements of photosynthetically active radiation (PAR) were recorded above and below the canopy using quantum sensors (Model LQS70-10SUN, Apogee Instruments, Logan, UT, USA).

Measurement of LAI began in December 2003 and was conducted around the 15th of each month using a plant canopy analyzer (Model LAI-2000, Li-cor Inc., Lincoln, NE, USA). Leaf transpiration measurements were made using a portable photosynthesis system (Model LI-6400, Li-cor Inc., Lincoln, NE, USA) on a clear-sky day (31 March 2004) at the height of 30 m (lower-canopy layer), 33 m (mid-canopy layer) and 36 m (upper-canopy layer) above the ground. Nine leaves of the dominant species (*P. tomentosa*) were measured at each height every 2 h from 08 : 00 to 16 : 00 (local time).

The missing data in flux and micrometeorological measurements were filled using different methods depending on the length of the gap. Gaps of less than 2 h were interpolated linearly from adjacent values. Gaps of longer than 2 h were filled using the mean diurnal variation method (15-day moving window; Falge *et al.*, 2001). Gaps in precipitation were filled using data from a nearby weather station that is 5 km away from the study site. Elevation correction was made to the precipitation gaps after Chen *et al.* (2007).

Data analysis

Surface energy closure of the measurement was evaluated first. If all components of the energy fluxes were measured accurately, the energy budget should close such that:

$$\mathrm{LE} + H = R_n - G - S - Q \tag{1}$$

where G is the ground surface heat flux, S is the sum of changes of heat storage in the air (S_{air}) and biomass (S_{biomass}) between the ground surface and the EC instruments and Q is the energy used for photosynthesis. Compared with other items, Q was small in magnitude and neglected in this study. G (W m^{-2}) was estimated as the sum of heat flux plate measurements and the soil heat storage change in the soil layer above the plates:

$$S_{\text{soil}} = (\rho_b C_s + \rho_w \theta C_w)\frac{\Delta T_{\text{soil}}}{\Delta t} z \tag{2}$$

where S_{soil} (W m^{-2}) is the soil heat storage change for the soil above the heat flux plate, ρ_b is soil bulk density (kg m^{-3}), C_s is the specific heat for mineral soil (J kg^{-1} K^{-1}), ρ_w is the density of water, θ is the volumetric soil water content (m^3 m^{-3}), C_w is the specific heat for water (J kg^{-1} K^{-1}), z is the thickness of soil layer above the heat flux plate (= 0·05 m) and ΔT_{soil} is the change of soil temperature of the top 0·05 m soil layer during the time interval of Δt (= 30 min).

Energy storage of the air S_{air} (W m^{-2}) between the ground surface and the EC instruments was estimated from the air temperature and humidity profiles:

$$S_{\mathrm{air}} = \rho_{\mathrm{a}} C_{\mathrm{p}} \sum_{i=1}^{n} \left(\frac{(\Delta T_{\mathrm{air}})_i}{\Delta t} \Delta z_i \right) + \rho_{\mathrm{a}} L \sum_{i=1}^{n} \left(\frac{\Delta q_i}{\Delta t} \Delta z_i \right) \tag{3}$$

where ρ_{a} (kg m^{-3}) is the air density, C_{p} (J kg^{-1} K^{-1}) is the specific heat of air, L (J kg^{-1}) is the LE of water vapourization, i is the layer of air temperature and humidity measurements (4·2, 16·3, 26·2, 36·5, 42·0 and 48·8 m), $(\Delta T_{\mathrm{air}})_i$ (K) and Δq_i (kg kg^{-1}) are the changes in air temperature and specific humidity of layer i during the time interval of Δt (= 30 min), and Δz_i (m) is the thickness of each layer (4·2, 12·1, 9·9, 10·3, 5·5 and 6·8 m).

Energy storage in the biomass S_{biomass} (W m^{-2}) was estimated as:

$$S_{\mathrm{biomass}} = \sum_{i=1}^{n} \left(\frac{(\Delta T_{\mathrm{biomass}})_i}{\Delta t} (C_i \mathrm{BM}_i) \right) \tag{4}$$

where i represents the different plant elements (stems, branches, leaves and litter biomass) above ground surface, $(\Delta T_{\mathrm{biomass}})_i$ (K) is the change in temperature during the time interval of Δt, C_i is a combined specific heat for woody biomass and water (3340 J kg^{-1} K^{-1} was used in this study, based on Wilson and Baldocchi, 2000), and BM$_i$ (kg m^{-2}) is the wet biomass of element i above ground surface. As no temperature measurements were made for stems, branches and leaves at the study site, we used the changes in mean air temperature below the canopy to represent the changes of stem temperature, and the changes of air temperature near the canopy to represent the changes in branch and leaf temperatures. Oliphant *et al.* (2004) evaluated the errors from this simplification and found that the S_{biomass} based on the air temperature was close to that based on stem and branch temperatures. The temperature of litter was assumed equal to that measured on the ground surface. The biomass of stems, branches and leaves at the study site was determined using a survey of dbh in 100 plots (plot size = 10 × 10 m^2) and the allometric equations of Zheng *et al.* (2006). Litter biomass was determined from the average mass of 25 surface litter samples (sample size = 1 m^2). More details of these measurements are available in Zheng *et al.* (2006) and Lu *et al.* (2007).

RESULTS

Site climate during the study period

The mean annual precipitation in the 4-year study period of 2003–2006 was slightly lower than the long-term average (1322 mm for the 4-year mean vs 1487 mm for the long-term average). The difference was mainly caused by the lower precipitation in the rainy season (1125 vs 1294 mm); the dry season precipitation was very close to the normal value (197 vs 193 mm). The mean annual air temperature during the four study years was 20·1 °C, approximately 1·5 °C lower than the normal. Figure 2 shows the temporal variations of daily total precipitation (P) and solar radiation, and daily mean air temperature (T_{air}), soil temperature (T_{soil}), water vapour pressure deficit (VPD) and soil moisture (SM) at three depths during the 4-year study period. The precipitation showed strong seasonal variations. The solar radiation also changed remarkably throughout the year due to the changes in solar zenith angle and the cloudy or rainy conditions. The maximum air temperature occurred in March or April instead of June or July was mainly due to the large precipitation events during the later months. Soil temperature varied seasonally following the air temperature, with 1–2 months delay in phase. Daily VPD peaked in the hot-dry season, and had lowest values in the cool-dry season in December. The high values of VPD in hot-dry season corresponded to low precipitation and high air temperature during this period. Soil moisture in the three soil layers stayed at high values during the rainy season as the soil water was recharged by frequent rainfall, and declined gradually to as low as 0·1 m^3 m^{-3} during the dry season as the soil water was depleted by drainage and root uptake. The soil moisture in 2003 did not reach a minimum in hot-dry season, which was mainly due to the abnormal high precipitation during that period.

Energy closure of the measurements

The ET can be directly determined by the EC measurements. To ensure the accuracy of the measurements, we first evaluated the energy closure before calculating ET. Figure 3 shows the relationship between daily turbulent energy flux (LE + H) measured by the EC system and the available energy ($R_{\mathrm{n}} - G - S$) calculated using the methods described in the section on Data analysis. There were a total of 1219 days available for this analysis in the four study years. Our results indicated that on average the measured turbulent fluxes (LE + H) only accounted for 64% of the available energy ($R_{\mathrm{n}} - G - S$). Energy closure was further examined separately in the cool-dry season, hot-dry season and rainy season. We found that large energy imbalance occurred in cool-dry season, in which the overall proportion of turbulent energy fluxes in the available energy was as low as 55%. This was mainly due to the impact of fog on EC measurements which caused water droplets on the sensor surface and the stable atmospheric conditions prevailing during fog events. We also compared the energy closure in daytime and nightime. Nocturnal energy budgets were usually less closed than that during daytime, although fluxes obtained at $u^* < 0.2$ m s^{-1} were rejected in the energy closure analysis. This indicated the overall lower quality of data measured at night than those measured during daytime.

The energy imbalance at our study site was unlikely caused by the R_{n} measurement. The radiometers for R_{n}

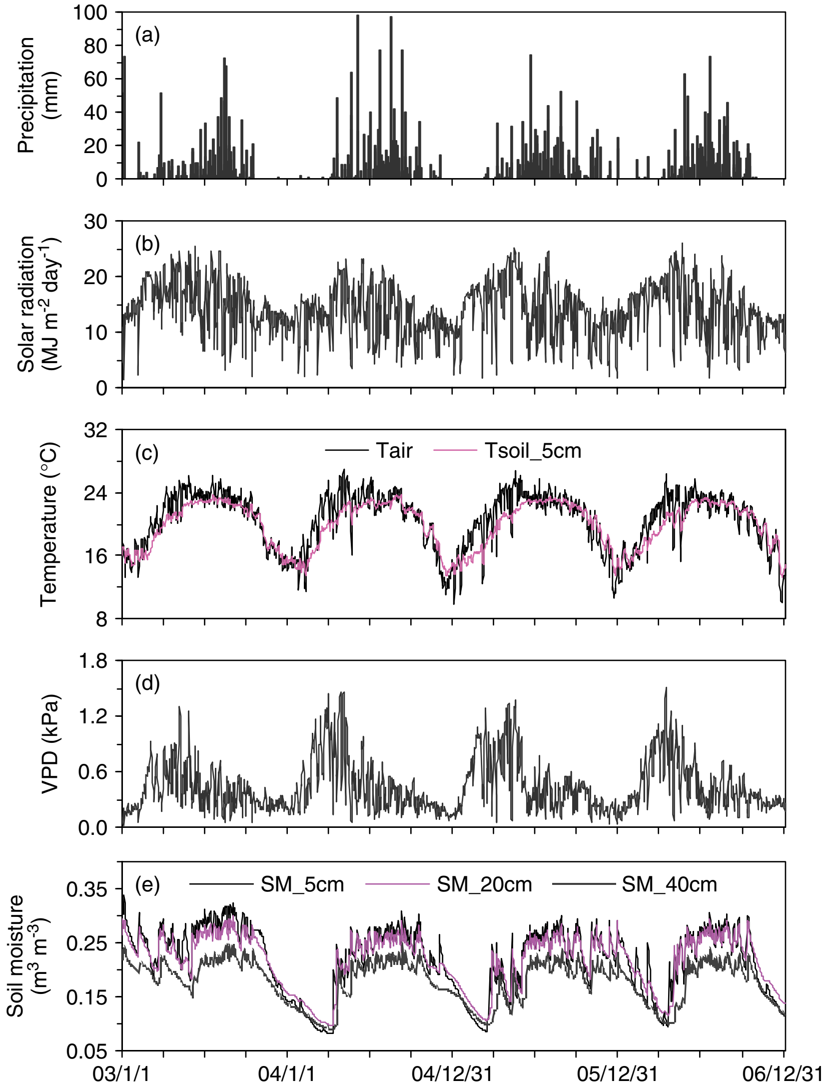

Figure 2. Time series of daily totals of precipitation (P) and solar radiation, and daily means of air temperature (T_{air}), soil temperature (T_{soil}), water vapour pressure deficit (VPD) and soil moisture (SM)

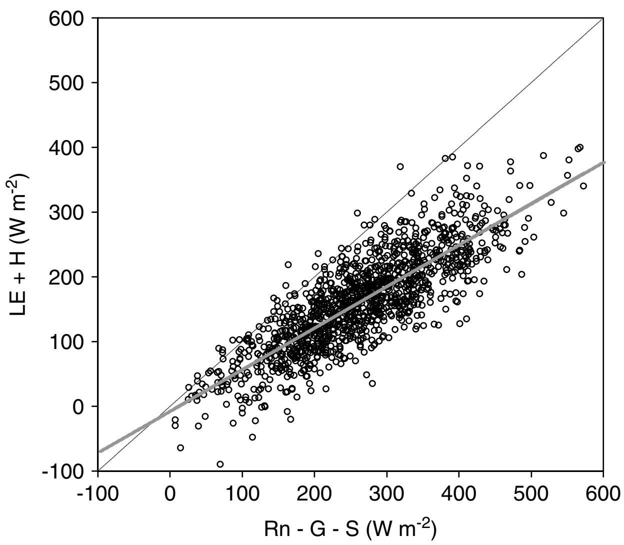

Figure 3. Relationship between the daily turbulent fluxes ($LE + H$) measured by the eddy covariance system and available energy ($R_n - G - S$) during the study period from 2003 to 2006 ($y = 0{\cdot}64x - 7{\cdot}02$, $r = 0{\cdot}83$, $n = 1219$)

measurement were well maintained during the observation period (levelled and cleaned at least once a month, calibrated once a year). The CNR-1 radiometers had an accuracy of 94–99%. Daily ground surface energy flux (G) in 2004 is shown in Figure 4a. It had a small seasonal variation, with a slight increase in hot-dry season and decrease in rainy season. Twine *et al*. (2000) investigated the uncertainties in G due to the sensor calibration and soil water content uncertainty and spatial variability. They suggested an error of 15%. G was particularly small at our site due to the low penetration of radiation through the dense canopy to the ground surface. It only accounted for 2–3% of net radiation. Therefore, G was unlikely to cause the large gap in the surface energy closure.

Figure 4b–d shows the variations of daily mean energy storage (S_{air}, $S_{biomass}$ and S) throughout the year of 2004. It could be seen that the energy storage in the air and biomass usually changed between -5 and 5 W m^2. The total energy storage S ($= S_{air} + S_{biomass}$) varied between -10 and 10 W m^{-2}. Energy was stored in the air and biomass during the daytime and released from them at

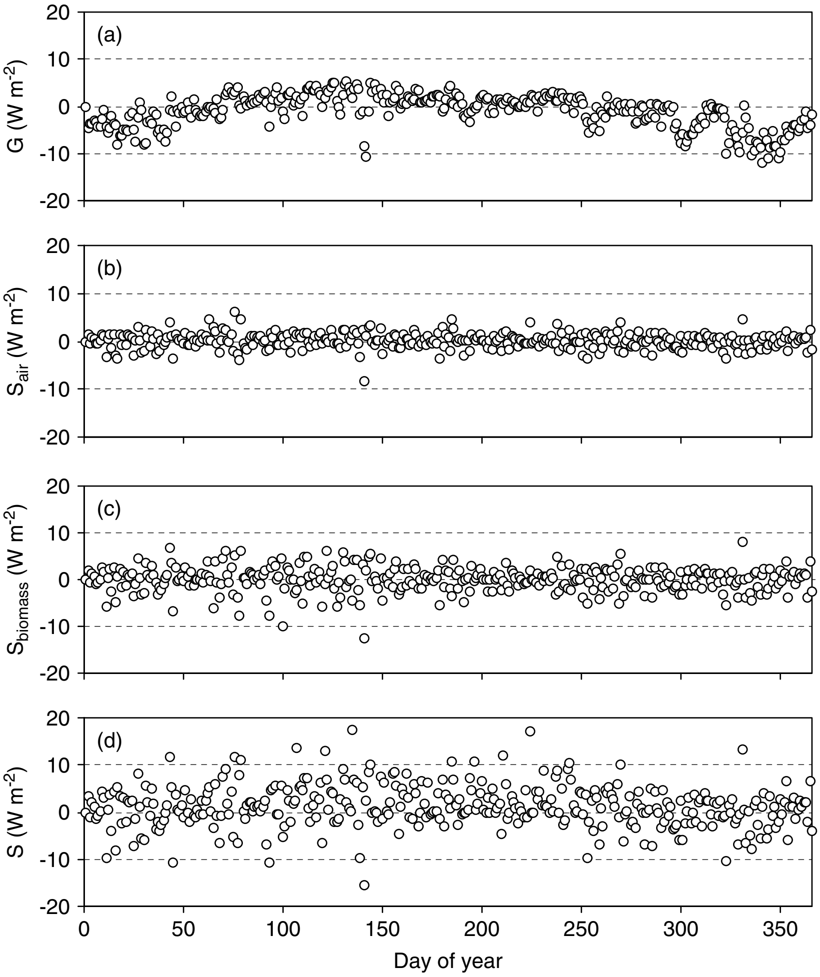

Figure 4. Daily mean ground surface heat flux (G) and energy storage changes in the canopy air (S_{air}), in the biomass (S_{biomass}) and their total ($S = S_{\mathrm{air}} + S_{\mathrm{biomass}}$) at the study site in 2004

night, and hence the daily energy storage changes in the air and biomass were small. Therefore, the energy storage changes were also unlikely to cause the surface energy imbalance.

Our analyses suggested that the energy closure problem was most likely due to the underestimation of turbulent fluxes of LE and H. The underestimates of LE and H by EC measurements that caused energy imbalance have also been found in many other study sites. Wilson *et al.* (2002) evaluated the energy closure at 22 sites with a total of 50 site-years in the FLUXNET across Europe and North America, and indicated the underestimation of 1–47%. Some investigators recommended that the energy closure should be forced by adjusting the EC fluxes to account for the missing energy (Blanken *et al.*, 1998; Twine *et al.*, 2000; Oliphant *et al.*, 2004; Barr *et al.*, 2006; Kosugi *et al.*, 2007). Through comparing different methods for forcing energy closure, Twine *et al.* (2000) suggested that the 'Bowen-ratio closure' method is the most appropriate approach. This approach assumes that the Bowen ratio ($\beta = H/\mathrm{LE}$) is correctly measured although LE and H are underestimated. In this study, we used this approach to adjust the LE and sensible heat fluxes at a 30-min interval. It is worth noting that this approach may lead to unrealistic results when the total turbulent flux ($\mathrm{LE} + H$) is close to 0 (e.g. in the early morning and at night when the magnitudes of LE and H are extremely small or when LE and H have similar magnitudes but with opposite sign). Under these extreme conditions, Oliphant *et al.* (2004) partitioned the residual energy equally between LE and H, whereas Kosugi *et al.* (2007) attributed the residual energy to the sensible heat flux. To keep the sign of the measured LE unchanged, we used the latter approach to force energy closure when the above conditions were present. This treatment produced annual total ET or LE 3·5%, 5·5%, 5·6% and 2·6% smaller than those using the former approach in 2003, 2004, 2005 and 2006, respectively. Overall, the difference of annual ET between the two treatment methods was small.

Diurnal variations in ecosystem ET

The turbulent energy fluxes including ET discussed hereafter referred to the values after energy closure correction. The mean diurnal cycles of ecosystem ET in the cool-dry season, hot-dry season and rainy season, calculated using all the available 30-min data in each of the three seasons, were shown in Figure 5. The

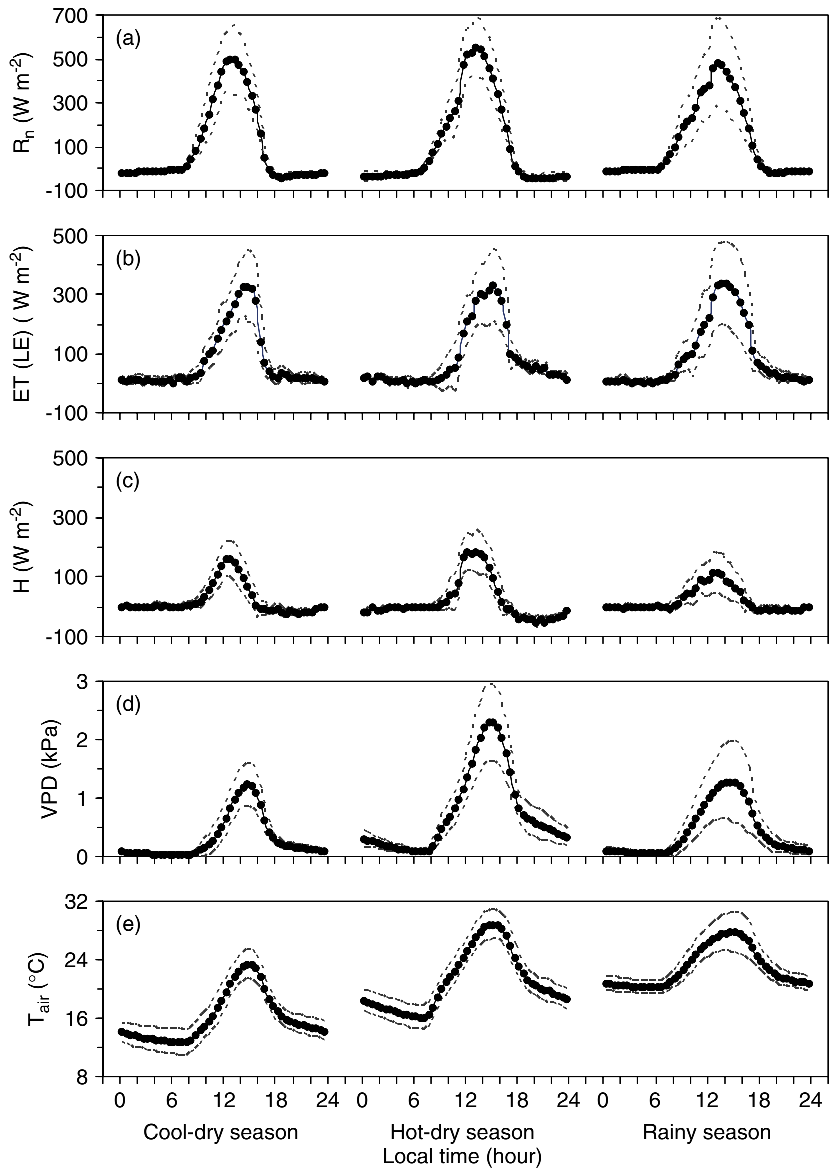

Figure 5. Diurnal variations of ecosystem ET (represented by latent heat flux), sensible heat flux (H), net radiation (R_n), water vapour pressure deficit (VPD) and air temperature (T_{air}) in the cool-dry season, hot-dry season and rainy season. Solid circles are the means of all available 30-min data in each of the three seasons. Dash lines represent the SD of the 30-min data

corresponding values of H, R_n, air temperature and VPD were also shown. In the cool-dry season, the ecosystem ET reached a maximum of 326 ± 101 W m^{-2} (0·48 ± 0·15 mm h^{-1}, ±SD; Figure 5b) for 2 h after R_n reached its maximum around noon (Figure 5a). On the contrary, the diurnal pattern of H (Figure 5c) corresponded closely to that of R_n (Figure 5a). The time delay of maximum ET was mainly due to the further increase in VPD (Figure 5d) and air temperature (Figure 5e) after R_n reached its maximum value (Figure 5a).

In the hot-dry season, the atmosphere evaporative demand increased substantially due to the increase in VPD (Figure 5d) and air temperature (Figure 5e). However, the maximum ET remained at values similar to that in the cool-dry season (Figure 5b). This was mainly due to the low soil water content (Figure 2e) in the late dry period. In addition, the high VPD in the hot-dry season, which could reach 3 kPa, was likely to cause stomatal closure and further constrain ET. This was observed through leaf-level measurements as shown in Figure 6 which gave the daytime variations of leaf transpiration (T_r) and stomatal conductance (g) on a clear-sky day (31 March 2004) in different canopy layers of the dominant species (*P. tomentosa*). The observed soil water content on this day was as low as 0·08 m^3 m^{-3}. Leaf transpiration and stomatal conductance showed relatively low values around noon in all the three canopy layers. The results implied stomatal closure in response to the high atmospheric evaporative demand under dry soil conditions. This was in consistent with the results of da Rocha *et al.* (2004), Pejam *et al.* (2006) and Giambelluca *et al.* (2009) where stomatal resistance increased under high evaporative demand and constrained transpiration around noon time.

In the rainy season, the soil water content was high and the forest experienced rapid growth at this study

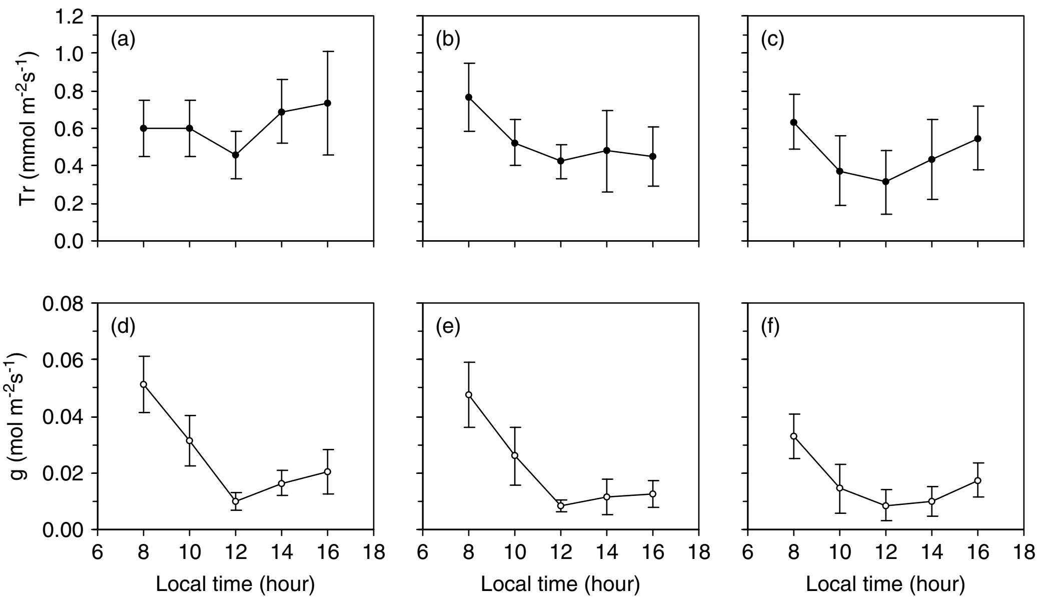

Figure 6. Leaf transpiration (T_r) and stomatal conductance (g) measured during the hot-dry season (31 March 2004) on the canopy leaves of the dominant species (*Pometia tomentosa*) in the upper-canopy layer (a and d), mid-canopy layer (b and e) and lower-canopy layer (c and f). Vertical lines represent the SD of the nine measured leaves

site. Most of the available energy was partitioned to ET. The diurnal distribution of ecosystem ET generally followed the variations of net radiation and H (Figure 5a–c). The maximum ET was 335 ± 136 W m^{-2} ($0{\cdot}50 \pm 0{\cdot}20$ mm h^{-1}, $\pm$SD; Figure 5b) and it was much higher than the sensible heat flux H (Figure 5c). The large variations (large SD) in ET (Figure 5b), R_n (Figure 5a) and VPD (Figure 5d) indicated the frequent rainy conditions in this season.

Seasonal pattern and annual total of ecosystem ET

The seasonal variations of the mean daily ecosystem ET for each month during the 4-year study period (2003–2006) are shown in Figure 7. The ET generally had a decline from November to January (early cool-dry season). This was mainly due to the gradual decrease in air temperature (Figure 2c) which reached its minimum (about 10 °C) in late December and early January. VPD in this period was also small (Figure 2d). With the increase in air temperature and VPD in February (late cool-dry season), the ecosystem ET had a slight increase. Air temperature and VPD further increased in March (early hot-dry season) which led to the increase in the atmospheric evaporative demand. On the other hand, the upper soil water content had decreased substantially during this period which constrained root water uptake. As the result, the actual ecosystem ET in the hot-dry season did not change significantly. This also indicated the role of deep soil water in the canopy transpiration during this dry period. In the rainy season of May through October, the ecosystem ET had high values but with large variations (large SD; Figure 7). This was mainly due to the large changes in weather conditions. The maximum daily ET in the year occurred in the rainy season and it was close to 5 mm day^{-1}.

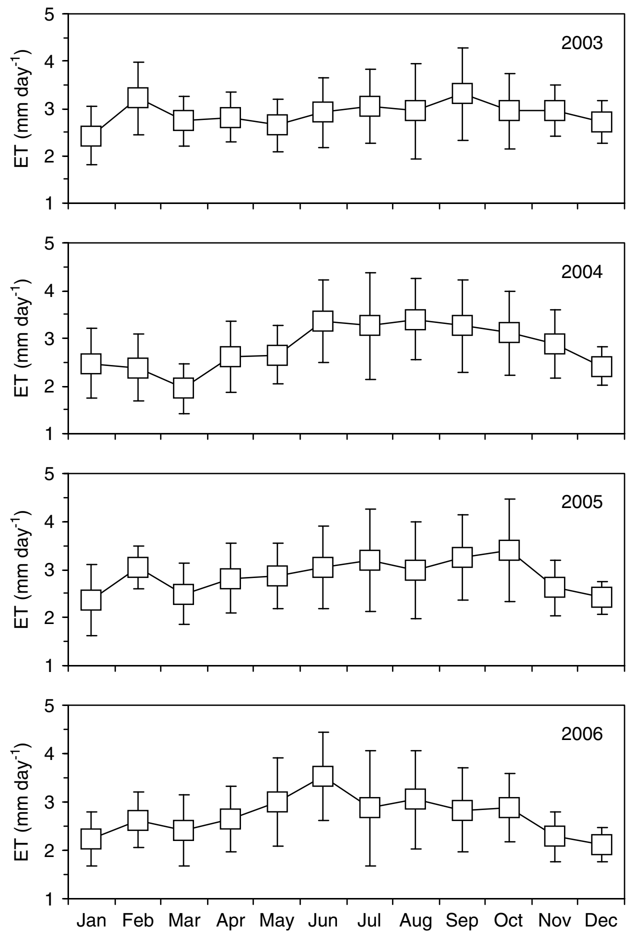

Figure 7. Means and SDs of daily ET by month during the four-year study period (2003–2006)

The Priestley–Taylor parameter α ($\alpha = \mathrm{ET}/\mathrm{ET}_{eq}$, where ET_{eq} is the equilibrium ET; Priestley and Taylor, 1972) was calculated to analyse the seasonal variations of controlling factors on ET. When α is greater than or close to 1, ET is mainly constrained by atmospheric demand. When α is less than 1, ET is mainly constrained by water supply (e.g. dry soil or low LAI). In contrast to ET, the Priestley–Taylor parameter α at the site had significant seasonal changes (Figure 8a), suggesting the changes in the main controlling factors on ET. In the hot-dry season, the α values were low and corresponded to the low soil water content and LAI (Figure 8b and c), implying the dominant constraint of water supply on ET. Further analyses on the relationships between ET and LAI and soil moisture (Figure 9) indicated that soil moisture was the main controlling factor on ET in this season (Figure 9e). LAI showed no evident effect on ET (Figure 9a). In the early rainy season, the α values were also low especially in May. However, ET was found to be more significantly related to LAI (Figure 9b) than soil moisture (Figure 9f). In the mid-to-late rainy season and the cool-dry season, the α values were generally high and the relationships of ET versus LAI and ET versus soil moisture were statistically insignificant (Figure 9c, d, g and h), indicating that atmospheric conditions played the major role in controlling ET. Indeed, correlation analysis between VPD and ET showed that ET was significantly related to VPD in the two seasons ($p = 0{\cdot}02$ and 0·04 in the mid-to-late rainy season and cool-dry season, respectively; p is the significance level). On the contrary, the corresponding p was 0·06 and 0·10 in the hot-dry season and early rainy season, respectively.

Interannual differences of daily ET were obvious at the site (Figure 7). For example, the daily ET in March (the early hot-dry season) was much lower in 2004 (monthly mean 1·8 mm day^{-1}) than in 2005 (monthly mean 2·3 mm day^{-1}). This was mainly due to the differences of deep soil moisture between the 2 years (note that the soil moisture in the upper soil layers was similar between the 2 years; Figure 2e). Deep soil moisture was mainly determined by the precipitation amount in the rainy season of the previous year. The total precipitation

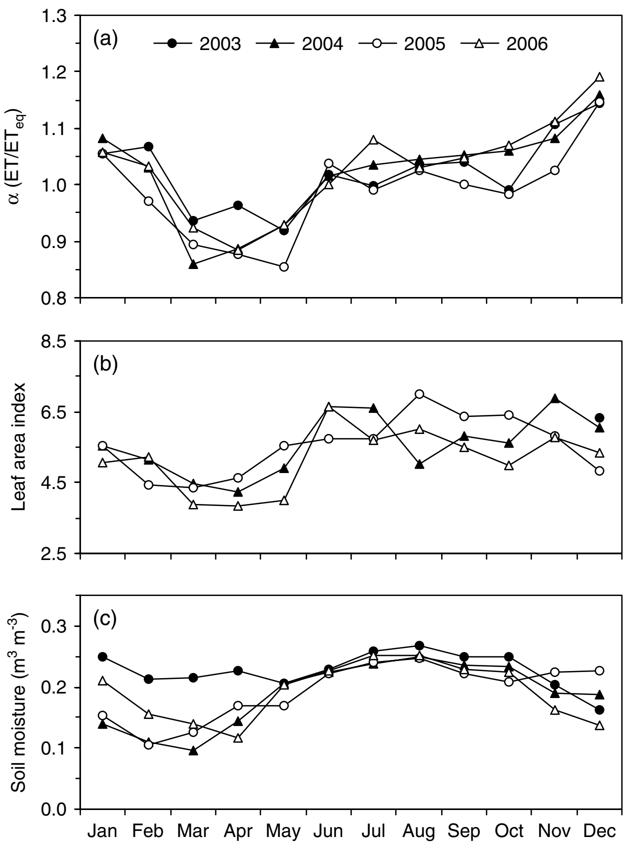

Figure 8. Monthly mean Priestley–Taylor parameter α (a), leaf area index (LAI; b) and soil moisture (SM; c) during the 4-year study period (2003–2006)

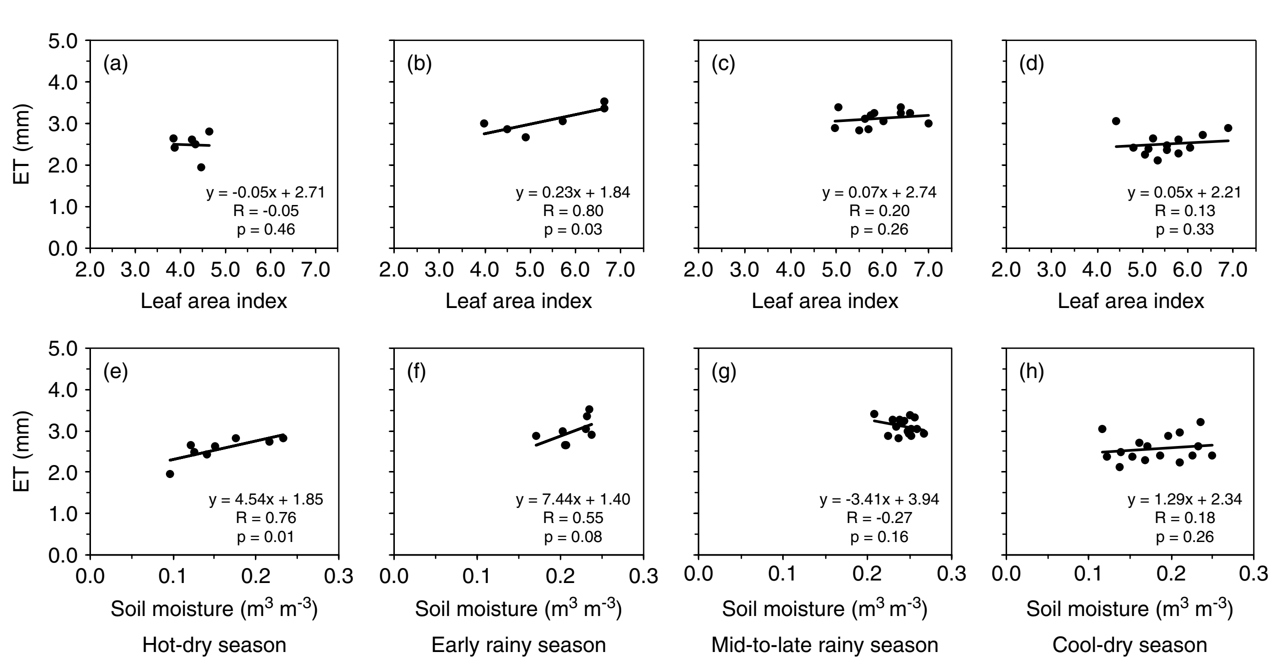

Figure 9. Relationships between mean daily ET by month and monthly mean leaf area index (LAI) and soil moisture (SM) in the hot-dry season (March and April), early rainy season (May and June), mid-to-late rainy season (from July to October) and cool-dry season (from November to February). p is the significance level

in rainy season was as low as 930 mm in 2003, whereas in 2004 it was 1283 mm (Table I). This caused large differences in soil water storage between the 2 years, which led to the difference of ET in the subsequent hot-dry season in 2004 and 2005.

The annual total ET measured directly by the EC system (without energy closure correction) was 730 ± 58 mm (±SD) in the 4 years. After adjusted for energy closure using the Bowen-ratio approach, the total annual ET was 1029 ± 29 mm (±SD). It accounted for about 78% of the corresponding annual total precipitation. Table I summarizes the annual and seasonal totals of ET in the four study years. The interannual variations of the total ET among the 4 years were generally small (SD = 29 mm). The mean total ET in rainy season, cool-dry season and hot-dry season was 565 ± 17, 308 ± 25, and 156 ± 13 mm, respectively, which was approximately 0·5, 3·41 and 1·46 of its corresponding seasonal total precipitation. Liu *et al.* (2004) reported that in the cool-dry season and hot-dry season, the deposition of radiation fog onto the same studied ecosystem could amount to 0·49 and 0·33 of the corresponding precipitation. After including the fog deposition in the precipitation, the ET was still well above *P* in the cool-dry season. It reflected the fact of dry-season depletion of soil water stored from previous rainy season. This can be seen from the large decline of soil water content during the dry period as shown in Figure 2e.

DISCUSSION AND CONCLUSIONS

Based on the EC flux measurements in Xishuangbanna tropical rain forest in the 4 years of 2003–2006, we estimated the ecosystem ET and its temporal variations. Our results showed that the diurnal variations of 30-min ET in the three seasons had similar maximum values although the soil water content, LAI and climate conditions differed greatly. The main controlling factors on ET varied with season. ET was mainly controlled by soil water availability in the hot-dry season, LAI in the early rainy season, and atmospheric conditions in the mid-to-late rainy season and the cool-dry season. The seasonal total ET was lower than the corresponding precipitation in the rainy season, but substantially higher in the cool-dry season. The extra amount of water evapotranspirated in the dry season was mainly from the depletion of soil water stored in the previous rainy season. The fog deposition also played a moderate role in providing water for ET during dry seasons. Overall, the mean annual total ET in the four study years was estimated at 1029 mm at our study site, accounted for 78% of the mean annual total precipitation.

Our results were in consistent with that of Negrón Juárez *et al.* (2007) where it was found that the recharge of deep soil moisture storage from the previous wet season could normally provide sufficient water to maintain ET in the subsequent dry season, mitigating the impact of the dry-season rainfall deficit. Our annual ET after energy closure correction was comparable with the results from other studies at XSTRFES–CAS. Liu *et al.* (2006) estimated the ET at the same site using the isotopic technique and water balance method. Their results showed that the ecosystem ET in 2002 and 2003 was 1186 and 987 mm, respectively. It accounted for about 70% of the mean annual precipitation (1579 mm) in the 2 years.

Compared with the other study sites in southeast Asia (Table II), the annual ET at our site was close to that of the tropical evergreen forest in central Cambodia (1140 mm), which had a similar annual total precipitation (1565 mm) and dry–wet seasonal cycle (the rainy season was from May to October and the dry season was from November to April) (Nobuhiro *et al.*, 2007). The ET values at our site was significantly lower than that at Sarawak (Kumagai *et al.*, 2005) and Pasoh (Tani *et al.*, 2003) in Malaysia and Java (Calder *et al.*, 1986) in Indonesia which are located in the equatorial region of southeast Asia and have high precipitation. Tanaka *et al.* (2008) observed a slightly lower annual ET in a tropical hill evergreen forest in Kog-Ma, although the site had high-annual precipitation. They found that the ET at their site was mainly constrained by the atmospheric evaporative demand. The VPD values at their site were generally smaller than those at our site, which contributed to its lower annual ET. Compared with the sites in South America (Amazonian tropics), the annual ET at our site was mostly lower (Table II). The South America sites mostly had similar annual dry season with our site, but they had much higher precipitation. A low-annual ET (1026 mm) was observed in a *terra firme* tropical forest in Rebio Jarú, Brazil (Negrón Juárez *et al.*, 2007),

Table I. Annual and seasonal totals of precipitation (*P*, mm) and evapotranspiration (ET, mm) during the 4-year study period of 2003–2006[a]

Year	Annual total			Hot-dry season			Rainy season			Cool-dry season		
	P	ET	ET/*P*	*P*	ET	ET/*P*	*P*	ET	ET/*P*	*P*	ET	ET/*P*
2003	1247	1052	0·84	147	169	1·15	930	546	0·59	171	338	1·98
2004	1428	1027	0·72	105	139	1·32	1283	584	0·46	40	305	7·58
2005	1284	1048	0·82	87	162	1·86	1063	574	0·54	134	312	2·33
2006	1328	988	0·74	88	154	1·75	1223	556	0·45	17	277	16·80
4-year mean (σ)	1322 (78)	1029 (29)	0·78	107 (28)	156 (13)	1·46	1125 (160)	565 (17)	0·50	90 (74)	308 (25)	3·41

[a] σ is the 4-year SD. Calendar year (January to December) was used in the Table. The cool-dry season in a year included January to February and November to December of the same year.

Table II. The annual precipitation (*P*, mm) and evapotranspiration (ET, mm) estimated in different tropical rainforest ecosystems

Site	*P*	ET	ET/*P*	Source
Southeast Asian tropics				
Xishuangbanna	1322	1029	0·78	This study
Kog-Ma, Thailand	1768	812	0·46	Tanaka *et al*. (2008)
Kampong Thom, Cambodia	1565	1140	0·73	Nobuhiro *et al*. (2007)
Sarawak, Malaysia	2151	1545	0·72	Kumagai *et al*. (2005)
Pasoh Forest Reserve, Malaysia	1571	1548	0·99	Tani *et al*. (2003)
Java, Indonesia	2892	1481	0·51	Calder *et al*. (1986)
South American tropics (Amazonian tropics)				
Tapajós, Brazil	2167	1300	0·60	da Rocha *et al*. (2004)
Cuieiras, Brazil	2089	1124	0·54	Malhi *et al*. (2002)
Rebio Jarú, Brazil	2291	1026	0·45	Negrón Juárez *et al*. (2007)
Pará, Brazil	1550	1516	0·98	Jipp *et al*. (1998)
Rondô nia, Brazil	2200	1359	0·62	von Randow *et al*. (2004)
Puerto Rico	1473	1219	0·83	Van der Molen (2002)
La selva, Costa Rica	3732	2139	0·57	Loescher *et al*. (2005)

although the site had high-annual total precipitation (2291 mm). Negrón Juárez *et al*. (2007) attributed this to the impacts of La Niña and El Niño events during their study period (2000–2003).

With more data becoming available at our study site, our knowledge on the ecosystem energy and water cycles over this tropical rain forest is expected to be further improved. Additional measurements of deep soil moisture, deep root system and soil water hydraulic lifting and redistribution are necessary at this site to further analyse the details of ET controlling mechanisms. It is worth mentioning that a nearby companion study site at a rubber tree forest was established in 2008. Comparison studies of the two sites will help us to assess the effects of land cover change on the ecosystem energy, water and CO_2 fluxes over the region.

ACKNOWLEDGEMENTS

We thank Mr Mengnan Liu, Mr Qinghai Song and other investigators in the Xishuangbanna Station for Tropical Rain Forest Ecosystem Studies, Chinese Academy of Sciences, for their data contribution to this study. This study was conducted during Zhiheng Li's visit to the Canada Centre for Remote Sensing, Natural Resources Canada. Financial support was from the Knowledge Innovation Program of the Chinese Academy of Sciences (KZCX2-YW-Q1-05-04, KJCX2-YW-432-1), National Science Foundation of China (40801035, 40571163) and the Development Program in Basic Science of China (2002CB412501).

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