

Ten-year evapotranspiration estimates in a Bornean tropical rainforest

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ABSTRACT

This study was undertaken to quantify interannual variations of total evaporation (ET) in a tropical rainforest in Sarawak, Malaysia. To this end, we conducted 10-year meteorological measurements and formulated a simplified big-leaf model that reproduces transpiration (Et), rainfall interception (Ei), and ET as the sum of these. The model was validated independently using eddy covariance fluxes, rainfall interception based on throughfall and stemflow measurements, and sap flow measurements conducted for more than 2 years. Using the model, Et , Ei , and ET were estimated for the period 2000–2009. Annual Et , Ei , and ET averaged over 10 years were estimated as 1114, 209, and 1323 mm, respectively, with small seasonal fluctuations. The derived estimates showed conservative year-to-year variations in annual Et , Ei , and ET ($CV = 5–7\%$) in contrast to considerable year-to-year variations in annual rainfall ($CV = 11\%$). Specific rainfall characteristics (e.g. intense short duration storms) at this site can explain the conservative year-to-year Ei variations. Small interannual variations in meteorological conditions and absence of severe drought during the study period can explain the small year-to-year Et variations. To characterize ET at our site, we also compared Ei and ET at our site with those of other tropical forests. Based on the derived ET characteristics, we discuss possible ET changes in response to changes in rainfall regime at this site.

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1. Introduction

Tropical rainforests are a major latent energy source, and influence regional and global water cycling (e.g., Lean and Warrilow, 1989; Mabuchi et al., 2005). In Southeast Asian tropical forests where relative deforestation rates are among the highest (e.g., Houghton and Hackler, 1999), clarifying water cycling is critical for understanding the consequences of land-use and/or climate changes (e.g., van der Molen et al., 2006). In tropical ecosystem water cycling, total evaporation (ET) as the sum of wet and dry canopy evaporation is a hydrologic component of major importance in determining the water budget of forest areas due to the significant volumes involved (e.g., Calder et al., 1986; Shuttleworth, 1988; Malhi et al., 2002).

The eddy covariance method is a widely used technique for assessing long-term ET , and has advanced our understanding of ecosystem water balance processes (e.g., Fisher et al., 2009; Giambelluca et al., 2009; Li et al., 2010). Some studies conducted in temperate forests showed that forest ET had large interannual variations and that soil drought resulting from seasonal and interannual rainfall variations could be factors determining interannual ET variations, which reduced ET due to limitations on transpiration (Wilson and Baldocchi, 2000; Lafleur et al., 2005; Granier et al., 2007). In some cases, annual ET increased with increasing rainfall due to more frequent occurrence of rainfall interception (e.g., Loescher et al., 2005). These results suggest that limited measurement periods can occasionally lead to erroneous water balance estimations in forested ecosystems. On the other hand, interannual variations in tropical rainforest ET are less certain, and long-term field monitoring studies have been poorly documented compared to temperate forest studies (e.g., Loescher et al., 2005; Takanashi et al., 2010). Furthermore, significant interannual variations in rainfall coinciding with El Niño have been observed in Southeast Asian tropical rain-

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forests (Nakagawa et al., 2000; Kishimoto-Yamada and Itioka, 2008).

Tropical rainforests on Borneo Island are characterized normally by high rainfall under maritime conditions distributed continuously through the year, although there is distinct seasonality of rainfall with phase-locked dry seasons in the South American and the Central African tropics under continental conditions (e.g., Walsh, 1996; Kumagai et al., 2005). Here, eddy covariance techniques tend to provide unreliable measurements during and after rainfall resulting from the physical impact of rain droplets on probes (Mizutani et al., 1997; Granier et al., 2000), suggesting limitations in the applicability of eddy covariance techniques in humid regions. Furthermore, ET derived from eddy covariance techniques cannot examine transpiration (E_t) and rainfall interception (E_i) components separately. E_t and E_i are different processes depending on dry and wet canopy. Thus, E_t and E_i should be estimated separately using component measurements, in addition to eddy covariance techniques, for examining impacts of possible changes in rainfall regime in this region.

This study was undertaken to quantify 10-year ET , E_t , and E_i in a Bornean tropical rainforest. To this end we conducted 10-year meteorological measurements and developed a simplified big-leaf model. The model was validated using water vapor fluxes under dry canopy conditions derived from eddy covariance techniques, rainfall interception based on throughfall and stemflow measurements, and sap flow measurements that were conducted for more than 2 years at this site. The model was applied to the 10-year continuous hourly meteorological measurements above the canopy. We then examined seasonal and interannual variations in E_i , E_t and ET for 10 years. The estimates were compared with those observed in previous studies conducted in other tropical forests to provide a global context for ET characterized at this site.

2. Materials and methods

2.1. Site description

This study was conducted in a lowland, mixed dipterocarp forest in Lambir Hills National Park, Sarawak, Malaysia ($4^{\circ}20' N$, $113^{\circ}50' E$), situated ~ 30 km southwest of Miri, Sarawak. A continuous canopy layer occurs ~ 40 m above the ground, and the height of emergent trees reaches 50 m. The leaf area index (LAI) ranges spatially between $4.8 \text{ m}^2 \text{ m}^{-2}$ and $6.8 \text{ m}^2 \text{ m}^{-2}$, with a mean of $6.2 \text{ m}^2 \text{ m}^{-2}$ (Kumagai et al., 2006). Litterfall was distributed evenly through the year (Nakagawa, unpublished), suggesting few variations in LAI . Soils at this site are classified as Red-Yellow Podzolic Soils (Malaysian classification) or Ultisols (United States Department of Agriculture Soil Taxonomy), with high sand content (62–72%) and high porosity (54–68%).

In Lambir Hills National Park, a 93-m-tall crane facility and a 49.5-m tree tower have been constructed for long-term ecological and hydro-meteorological studies (Ozanne et al., 2003). Our measurements in this study were conducted mainly in the area surrounding the crane, a 4-ha study plot comprising 400 $10 \text{ m} \times 10 \text{ m}$ subplots. Some measurements were also conducted in the tree tower site located ~ 540 m southwest of the crane facility.

2.2. Meteorological measurements and characteristics of the elements

A solar radiometer (MS401, EKO, Tokyo, Japan), an infrared radiometer (Model PIR, Epply, Newport, RI, USA), and a tipping bucket rain gauge (RS102, Ogasawara Keiki, Tokyo, Japan) with data logger (HOBO event, Onset Computer, USA) were installed at the top of the crane, 85.8 m above the ground. Upward short-

wave and long-wave radiation were measured temporally at tree tower located 100 m south of the crane, using an upside-down solar radiometer (CM06E, Kipp and Zonen, Delft, Netherland) and infrared radiometer (Model PIR, Epply, Newport, RI, USA), as the tree canopy gap was formed around the crane due to crane construction with its largest radius (~ 25 m). Radiations were sampled every 5 s, and an average value from each 10-min period was used (CR10X, Campbell Scientific). These four radiation measurements were used to calculate net radiation (R_n). When upward short-wave and long-wave radiation were not available, R_n was calculated using the relationships between solar radiation (R_s) and R_n ($R_n = 0.87R_s - 35.0$, $R^2 = 0.99$). A thermohygrograph (HMP35A, Visala Co., Helsinki, Finland) and cup-anemometer (AC750, Makino Instruments, Tokyo, Japan) were installed on the crane at 76 m height. Cup-anemometers were also installed at heights of 93 m and 60 m, and wind speeds were calculated using cumulative pulses from the cup-anemometers every 10 min (CR10X, Campbell Scientific). Air temperature and relative humidity were sampled every 10 min and recorded by a data logger (SQ1250, Grants Instruments, Cambridge, UK). At the top of the 49.5-m tree tower, the same meteorological characteristics were measured simultaneously. When data from the crane were not available, data from the tree tower were used instead. When this occurred, wind speed at 76 m was estimated using known relationships between height and wind speed.

Fig. 1 shows seasonal and interannual variations in meteorological elements over the last 10 years at this site. Mean annual temperature was $25.8^{\circ}C$. Rainfall was distributed through the year (Fig. 1a), and mean annual rainfall was 2600 mm. Although there are no distinctive phase-locked dry seasons (Fig. 1a), rainfall power spectra showed a major peak at the 1-year cycle (data not shown). Rainfall was relatively higher between October and January. The power spectra of R_n , air temperature, vapor pressure deficit (VPD) and wind speed showed a major peak at the 1-year cycle (data not shown). R_n , air temperature and VPD were slightly higher between July and September in most of cases (Fig. 1b and c). Wind speed was higher between January and March under northeast monsoon influences (Fig. 1d).

2.3. Eddy covariance method

A three-dimensional sonic anemometer (USA-1, METEK, Elmshorn, Germany) was installed at 60.2 m on the crane facility, 20 m above the mean canopy height. An open-path $\text{CO}_2/\text{H}_2\text{O}$ analyzer (LI-7500, Li-Cor, Lincoln, NE, USA) was placed ~ 30 cm from the sonic anemometer in an inclined orientation. A ventilated psychrometer (MS020S, EKO, Tokyo, Japan) was installed adjacent to the sonic anemometer and open-path analyzer to calibrate and correct for density effects (Webb et al., 1980).

Wind speeds and gas concentrations were sampled at 10 Hz, and the averages were computed at 30-min intervals. Although the effect of the low-frequency contribution to eddy fluxes was insignificant at this site, the results were detrended, as an example, using a moving-average time constant of 200 s and the wind field (u, v, w) (m s^{-1}) coordinates were rotated so that mean v and w values were zero over 10-min periods. The percentage corrections for the coordinate rotations were less than 1%, probably because of the gentle slope. We applied the density fluctuation correction method described by Webb et al., 1980. In this study, flux data measured during rain events were abandoned because of unreliability of the data. Calibrations of the open path analyzer were performed periodically using a N_2 standard gas and a dew point generator (LI-610, Li-Cor, Lincoln, NE, USA), and the zero and span were checked at intervals of 2–3 months.

When we compared the sum of latent (IE) and sensible (H) heat fluxes to net radiation (R_n) at the site, the relation $IE + H = 0.80R_n$

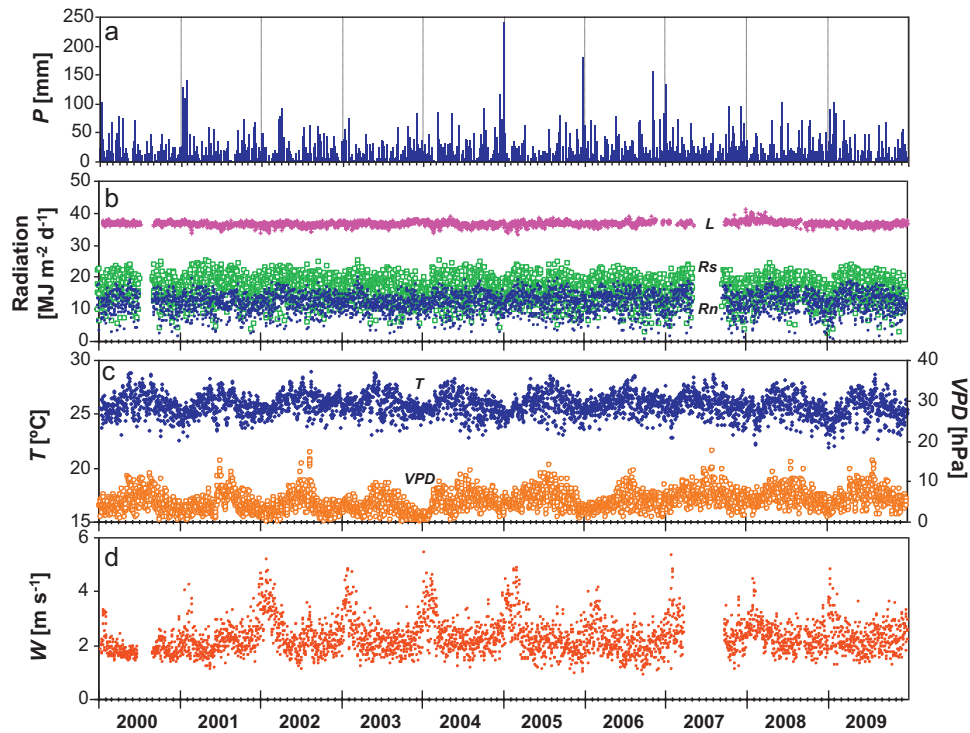


Fig. 1. Temporal variations in (a) rainfall (P), (b) downward short-wave (R_s), long-wave (L), and net radiation (R_n), (c) temperature (T), and vapor pressure deficit (VPD), (d) wind speed (W) in the period from January 2000 to December 2009.

($R^2 = 0.81$) was derived. Non-closure of the energy balance is a common feature of eddy covariance measurements above forest canopies, and the closure deficit of $\sim 20\%$ is comparable to that obtained at most forest sites (Goulden et al., 1996; Aubinet et al., 2001; Kosugi and Katsuyama, 2007; Takashi et al., 2010). The degree of closure is within the range reported for other forested sites. In this study, we assumed that the eddy fluxes had been underestimated consistently, and we added 20% to the flux. The measurements were conducted in two periods: June 2001–November 2002, and August 2004–January 2005. A more complete description of the flux measurements is available in Kumagai et al. (2004, 2005), and Saitoh et al. (2005).

2.4. Throughfall and stemflow measurements

Throughfall and stemflow measurements were conducted in the 4-ha study plot. In one fixed subplot of the 400 subplots, stemflow generated by all trees (78 trees) with DBH > 1 cm and throughfall at 20 points were observed continuously on a daily basis for 3 years from July 2001 to July 2004. In addition, daily measurements of throughfall in 22 other subplots, relocated periodically in the 4-ha plot, were conducted in the same 3-year study period using a set of 20 throughfall collectors. From the relationship between estimated net rainfall (throughfall + stemflow) and gross rainfall on an event basis, rainfall interception at the 4-ha scale was estimated for the 3 years. A more complete description of the long-term intensive measurements of throughfall and stemflow at the 4-ha plot is available in Manfroï et al. (2004, 2006).

2.5. Sap flow measurements

Sap flow measurements were conducted using the thermal heat-pulse method (Closs, 1958; Marshall, 1958) in four upper canopy trees: *Dryobalanops aromatica* (53 m) and *Shorea beccariana* (51 m) in the area surrounding the crane, and *Dipterocarpus*

pachyphyllus (48 m) and *Ficus crassiramea* (46 m) adjacent to the tree tower. A heat pulse sensor (HP-1, Hayasi Denko, Tokyo, Japan) was inserted radially into the trunk just above the buttress in the emergent trees at the crane site, and at 35 m above ground level in the emergent trees adjacent to the tree tower. A more complete description of the sap flow measurements is available in Kume et al. (2008b, 2008c).

3. Analysis

3.1. Big-leaf model structure

The big-leaf model used in this study was formulated based on the Penman–Monteith equation (e.g., Monteith and Unsworth, 1990). The canopy was assumed to have canopy water storage capacity (S_c), and current water storage (S) was charged by rainfall (P) and discharged by rainfall interception (E_i), throughfall (T_f), and stemflow (S_f). Canopy water balance was calculated as follows:

$$\frac{dS}{dt} = \int (P - E_i - T_f - S_f) dt. \quad (1)$$

Here, we did not distinguish between T_f and S_f , which were calculated as over-flow from the canopy water storage capacity.

When $S = S_c$, both expressed as millimeters of water, evaporation from the wet canopy (E_i) was calculated from simplified Rutter's model (Rutter et al., 1975) as follows:

$$E_i = E_p = \frac{\Delta (R_n - G) + \rho C_p VPD / Ra}{(\Delta + \gamma) \lambda}, \quad (2)$$

in which

$$Ra = \frac{\ln^2 \{(z - d) / z_0\}}{k^2 u(z)}, \quad (3)$$

where E_p is the potential evaporation rate (mm h^{-1}), Δ is the slope of the saturation vapor pressure function (hPa K^{-1}), R_n is

net radiation ($W m^{-2}$), G is soil heat flux ($W m^{-2}$), ρ is air density ($kg m^{-3}$), C_p is the specific heat of air ($J kg^{-1} K^{-1}$), VPD is the vapor pressure deficit of air (hPa), R_a is the aerodynamic resistance for heat and water from vegetation ($s m^{-1}$), γ is the psychrometer constant (hPa K^{-1}), and λ is the latent heat of water vaporization ($J g^{-1}$), z is the reference height (m), d is the zero plane displacement (m), z_0 is the roughness length (m), k is the von Karman's constant ($=0.41$) and $u(z)$ is the wind speed at level z ($m s^{-1}$). Here, G was negligible at this study site (Sato et al., 2005). The free throughfall coefficient in Rutter's original model was assumed to be zero because of the dense canopy in this site, and the drainage function in the original model was neglected because it has less impact on the reproducibility of the model.

When the canopy is partially dry ($0 < S < S_c$), E_i and dry canopy transpiration (E_t) were calculated as follows:

$$E_i = E_p \frac{S}{S_c} \quad (4)$$

$$E_t = E_{pt} \left(1 - \frac{S}{S_c}\right) \quad (5)$$

in which

$$E_{pt} = \frac{\Delta(R_n - G) + \rho C_p VPD / R_a}{\Delta + \gamma(1 + R_c / R_a)} \quad (6)$$

$$R_c = \frac{1}{G_c} = \frac{1}{(aR_s + b)(1 - m \ln VPD)}, \quad (7)$$

where E_{pt} is the E_t assuming a fully dry canopy, R_c is the canopy resistance ($s m^{-1}$) using a multiplicative-type function, G_c is the canopy conductance ($m s^{-1}$), R_s is solar radiation ($W m^{-2}$), and a , b , and m are the parameters to be estimated through non-linear regression analysis.

Note that dry canopy evaporation ($S=0$) is denoted by E_t , and the E_t was expressed as follows:

$$E_t = E_{tp}. \quad (8)$$

Finally, total evaporation (ET) was expressed as follows:

$$ET = E_i + E_t. \quad (9)$$

3.2. Parameterization

Aerodynamic parameters, zero plane displacement (d) and roughness length (z_0) for R_a were determined by wind profile analysis using a standard least-squares approach (e.g., Lafleur, 1992). At the 93-m crane, wind speed profiles measured using the three cup anemometers mounted at 60 m, 76 m and 93 m above ground level were selected such that atmospheric conditions were near neutral stability. The bulk Richardson number (Ri_B) was used to select wind speed profiles for near-adiabatic conditions ($0.0 < Ri_B < 0.1$) according to Komatsu et al., 2005. Values of $d = 37.8$ m and $z_0 = 1.73$ m were computed previously (Kume et al., 2008a).

Canopy conductance (G_c) parameters a , b and m were computed previously from the eddy covariance method conducted in the period between June and December 2001, including wet and dry soil conditions (Kumagai et al., 2004). Consequently, G_c parameters were determined as $a = 1.43 \times 10^{-5}$, $b = 6.71 \times 10^{-3}$, $m = 0.658$ under wet soil conditions, and $a = 1.25 \times 10^{-5}$, $b = 2.99 \times 10^{-3}$, $m = 0.254$ under dry soil conditions. Fortunately, the period in the previous study included the driest period (i.e., minimum extractable soil water) in the last 10 years (Fig. 8), suggesting that the stomatal response in this study period can be described using the two parameter sets, and that the total amount of E_t at this site is within the ranges of the two E_t calculated from the two parameter sets. In this study, E_t was described using the average of the maximum E_t estimated using the wet soil condition G_c parameter (E_{t1})

and minimum E_t estimated using the dry soil condition G_c parameter (E_{t2}). Here, differences between calculated E_{t1} and E_{t2} were less than 20%, indicating the impacts of variations in G_c on E_t were relatively insignificant at this site because of a larger decoupling coefficient ($\Omega \geq 0.3$) (Kumagai et al., 2004, 2005).

Canopy water storage capacity (S_c) was determined previously by a mean method based on the throughfall and stemflow measurements (e.g., Klaassen et al., 1998) conducted in the 4-ha study plot. A value of $S_c = 0.7$ – 1.0 mm was derived from the relationship between net rainfall (throughfall + stemflow) and rainfall on an event basis (Manfroi et al., 2006). We also independently estimated the S_c value from comparisons between modeled wet canopy duration after rainfall (i.e. the period with $S > 0$ just after rainfall) and estimated wet canopy duration after rainfall determined by sap flow measurements (Kume et al., 2006, 2008a). Consequently, the derived S_c value ($=0.7$ mm) corresponded well to the S_c value from throughfall and stemflow measurements. Overall, we assumed that real E_i at this site is in the E_i range calculated from $S_c = 0.7$ – 1.0 mm. In this study, E_i was described using the average of the minimum E_i estimated using $S_c = 0.7$ mm (E_{i1}) and maximum E_i estimated using $S_c = 1.0$ mm (E_{i2}).

3.3. Analysis method

Model input data were hourly rainfall, net radiation (R_n), vapor pressure deficit (VPD), air temperature, wind speed, and solar radiation (R_s) measured above the canopy in the period January 2000 to December 2010. The model outputs were transpiration (E_t), rainfall interception (E_i), and total evaporation (ET) as the sum of the two. In this study E_t and E_i estimated from the derived parameters were validated independently as described below.

Under dry canopy conditions, accuracy of the modeled E_t was validated by comparing the total amount of daily E_t from eddy covariance methods. Under wet canopy conditions, accuracy of the modeled E_i was validated by comparing the total amount of E_i from rainfall interception measurements, based on throughfall and stemflow measurements for 3 years (Manfroi et al., 2006). Further, to validate accuracy of our model under wet canopy conditions, we compared modeled wet canopy duration (i.e. the period with $S > 0$) with diurnal variations in sap flow measured during and after day-time rainfall (Kume et al., 2008b). Because transpiration decreases during canopy wetting, the decrease in transpiration can be monitored continuously by sap flow measurements (Bosveld and Bouten, 2003).

In the meteorological input data, rainfall, air temperature, and VPD had no data gaps over the last 10 years, while R_n and wind speed had some data gaps (Fig. 1). To estimate the total amount of annual ET , E_i and E_t , annual values were calculated from $E = \frac{365}{n} \sum_{i=1}^n E(n)$, where E is the total annual value, n is the number of measurement days, and $E(n)$ is the values measured at n -day. The percent coverage of ET , E_i and E_t in this study period excluding 2007 ranged between 80% and 100%, and the data gaps were never longer than 2 months per year. This suggests that data gaps have a negligible effect on reliability of annual values. The percent coverage of data in 2007 was only 51%, and we excluded the 2007 data for ET analysis.

To examine characteristics of temporal change in ET , we defined the water deficit (WD) as monthly ET minus monthly rainfall (P) ($WD = ET - P$, but $WD = 0$ when $ET < P$). The WD for 10 years was compared with temporal changes in soil water content at the site. Soil water content at the crane site was measured at the 10-cm, 20-cm and 50-cm depth, and at the tree tower site at 10-cm, 30-cm, and 60-cm depth. Soil measurements at the crane and tree tower sites were conducted at 10-min intervals (CR23X and CR10X data

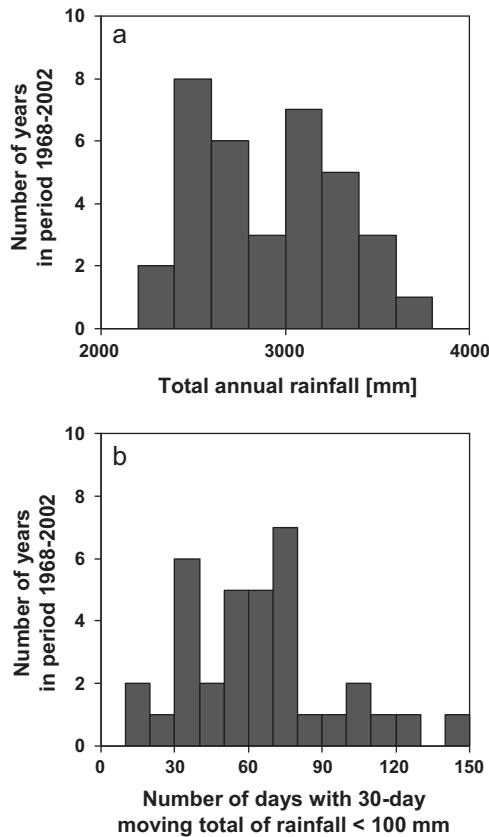


Fig. 2. Frequency distributions of (a) rainfall and (b) number of dry periods at Miri Airport between 1968 and 2002. A dry period is defined as days having a 30-day moving total of <100 mm rainfall.

logger, Campbell Scientific). A time-domain reflector (TDR; CS615, Campbell Scientific) was used to measure time series of volumetric soil water content (θ ; $\text{m}^3 \text{m}^{-3}$). Extractable soil water (mm) in the 0–60-cm zone (ESW_{0-60}) was calculated using the cumulative θ

measured in the three layers: $ESW_{0-60} = \sum_{i=1}^3 (\theta_i - \theta_r) \times SL_i$, where

θ_r is residual water content in each layer, SL is the soil layer thickness (=200 mm), and i is the order of the sensors from the top. θ_r was estimated as the minimum θ over the last 10 years.

4. Results

4.1. Rainfall characteristics

It is important to understand the representativeness of rainfall data from the last 10 years in the context of past climate. Fig. 2 shows rainfall records from Miri Airport, 20 km from the study site, for the period 1968–2002. Although the rainfall data showed some discrepancy with those of the study site on a daily basis, seasonal rainfall distribution and annual rainfall at Miri Airport corresponded to those of the study site (data not shown). At Miri Airport, the total annual rainfall over the period ranged between 2124 mm and 3499 mm, and showed a bimodal frequency distribution that peaked at 2400 mm and 3000 mm (Fig. 2a). The period mean was 2723 mm, and the median 2645 mm. Annual rainfall over the last 10 years ranged between 2191 mm and 2944 mm, and mean and median annual rainfall were 2600 mm and 2631 mm, respectively. Although rainfall over the last 10 years did not exceed the annual rainfall of >3000 mm, the average for the last 10 years approximated the 34-year average.

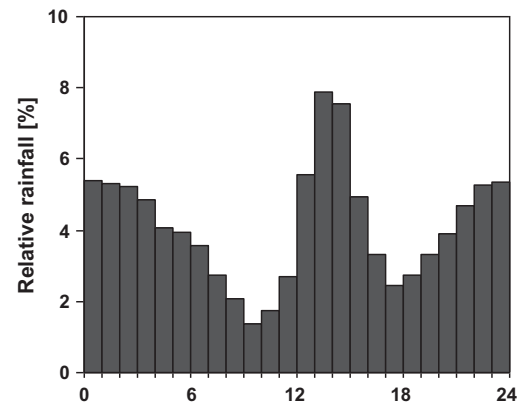


Fig. 3. Diurnal rainfall cycle. Relative rainfall represents total rainfall at each time divided by total rainfall in the study period from January of 2000 and December of 2009.

The number of months with rainfall <100 mm can be a useful indicator of water stress in tropical forests (e.g., Walsh, 1996; Malhi et al., 2002). Further, a dry period, which is defined as days having a 30-day moving rainfall total of <100 mm, is a more useful indicator than monthly rainfall in Southeast Asian tropical rainforests where intra-annual dry spells rarely exceed a month (e.g., Becker, 1996; Sakai et al., 2006). The annual sum of the dry periods during 1968–2002 ranged between 7 and 140 days, with a mean of 55 and median of 52 (Fig. 2b). The longest continuous dry period of 119 days occurred in the period 25 December 1986 to 23 April 1987. Our study period had 3–70 days of dry periods in each year, with a mean of 38 and median of 37. The longest continuous dry periods were 24 days in September 2001, May 2005 and August 2009 (data not shown). This indicates that the dry periods in the study period were shorter than those of the 34-year, and that our study period did not include an unusually severe drought year.

Fig. 3 shows the rainfall diurnal cycle at the study site. It shows two peaks: one occurs between 14:00 and 15:00, and the other occurs at midnight. In the study period, each rain event at the site had a short duration with a mean of 3.2 h, and never exceeded 24 h. In this study, to separate each rain event, we used a non-rain period of 3 h. Note that the threshold (3 h) corresponded to the canopy drying time after day-time rainfall using sap flow measurements, which could be used as surrogates of wetness sensors (Kume et al., 2006).

Fig. 4 shows interannual variations in rainfall in the study period. Interannual differences in annual rainfall ranged 753 mm, while cumulative rainfall values with events of <10 mm were consistently ~500 mm each year (Fig. 4b). Also, cumulative rainfall values from events <60 mm were consistently ~2000 mm during the study period (Fig. 4b). Although most events were <60 mm and events ≥ 60 mm occurred only four to 10 times each year (Fig. 4a), annual rainfall had a moderate relationship with total events >60 mm ($R^2 = 0.36$, $p = 0.08$) and frequency of events >60 mm ($R^2 = 0.33$, $p = 0.10$), respectively.

4.2. Model validation

Comparisons between the total modeled E_t and measured E_t on a daily basis without rain showed mostly differences of <1.0 mm day^{-1} (Fig. 5). The data in Fig. 5 were differentiated using soil moisture conditions. Although there were some model over-estimations (Fig. 5), there was no clear bias resulting from soil moisture conditions. The mean difference between modeled and measured E_t was 0.35 mm d^{-1} , or 11% of the population mean. These agreements indicated accuracy of our model under dry canopy conditions.

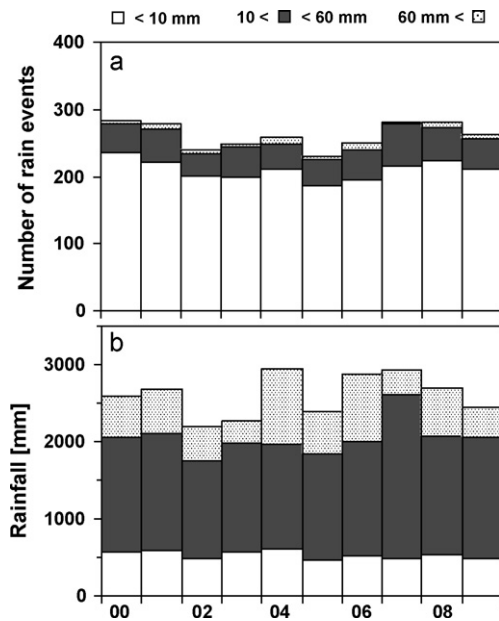


Fig. 4. (a) Number of rain events, and (b) total amount of annual rainfall between 2000 and 2009. Both are classified using specific rainfall depth. White bars show cumulative rainfall when the rain event is <10 mm, black bars show cumulative rainfall when 10 mm < rain event < 60 mm, and the dotted gray bars show cumulative rainfall when the rain event is >60 mm.

Fig. 6 shows typical time series of modeled E_t and modeled wet canopy duration, with diurnal cycles of sap flow velocities on rainy days. The modeled wet canopy duration corresponded to the period when sap flow velocities were low or increasing abruptly after rainfall cessation. Diurnal variations in modeled E_t corresponded well to those of sap flow velocities. Here, the diurnal patterns of sap flow velocities measured in the basal stems showed remarkable concordance with canopy-scale modeled E_t , with short time lags of <20 min ($R=0.90-0.94$; Kume et al., 2008b). This suggests diurnal patterns of sap flow velocities in emergent trees at this site are highly spatially representative. These correspondences between modeled E_t and sap flow velocities during and after rainfall suggest that our model

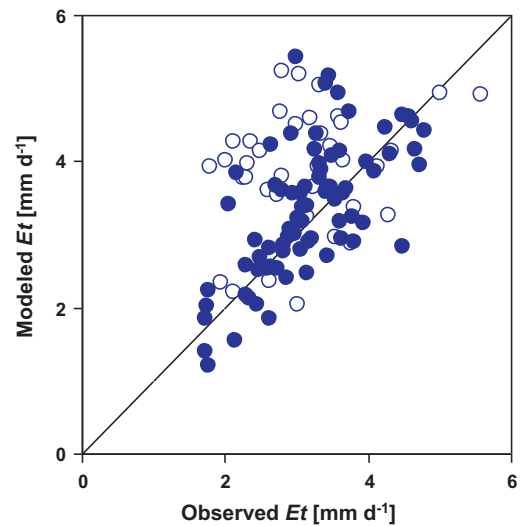


Fig. 5. Relationship between observed transpiration (E_t) and modeled results under dry canopy conditions. The modeled E_t is the average of maximum and minimum estimates. The solid line indicates observed $E_t =$ modeled E_t . The data were differentiated with soil moisture condition; ● >50 mm, ○ <50 mm.

also reproduced switching of E_t and E_i during and after rainfall.

The mean annual E_i measured for 3 years at the crane site was 210.2 ± 10 mm (Manfroi et al., 2006). The mean E_i for the 3 years estimated from our model ranged between 164.7 ± 1.5 mm and 230 ± 5.0 mm using $0.7 \text{ mm} < S_c < 1.0 \text{ mm}$, and the average (197.3 mm) approximated the measured value (210.2 mm). Further, the measured 3-year E_i showed almost constant annual E_i ($CV = 5\%$), which corresponded to the year-to-year variations in calculated annual E_i ($CV = 2\%$). Consequently, our model is sufficiently accurate to reproduce the amount of annual rainfall interception.

4.3. E_t , E_i , and ET estimates

The temporal variations in rainfall, E_t , and E_i and their cumulative values for each year are shown in Fig. 7. Monthly E_{t1} and

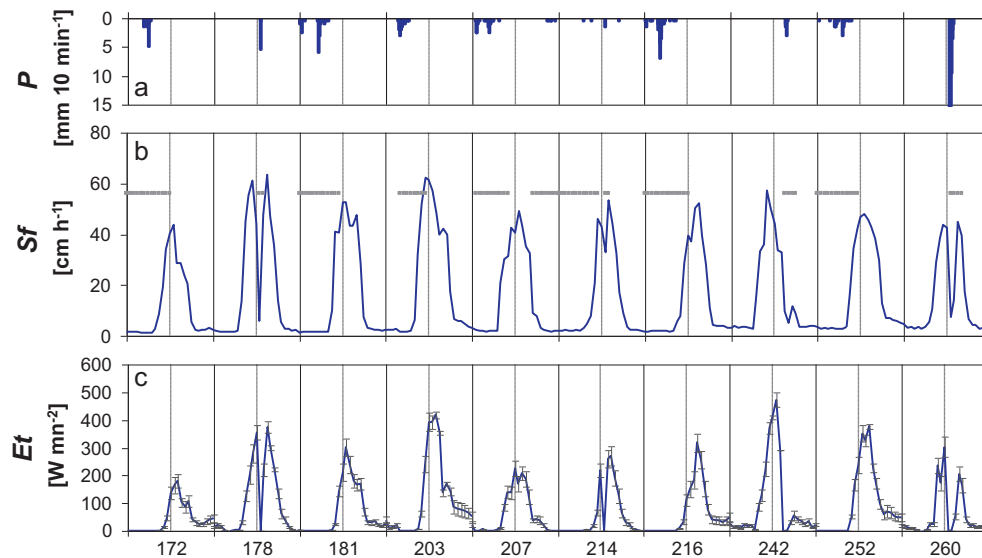


Fig. 6. Measured time series of (a) rainfall (P), (b) sap flow velocities (F_d) measured in *Dipterocarpus pachyphyllus* from DOY 172 (21 June) to 260 (17 September), 2001, and (c) modeled transpiration (E_t) under wet and dry canopy conditions. The bold gray line in Fig. 6b indicates modeled wet canopy duration (i.e., the time when $S > 0$). The vertical lines in Fig. 6c represent upper and lower limits of estimations, and the solid lines show the average value.

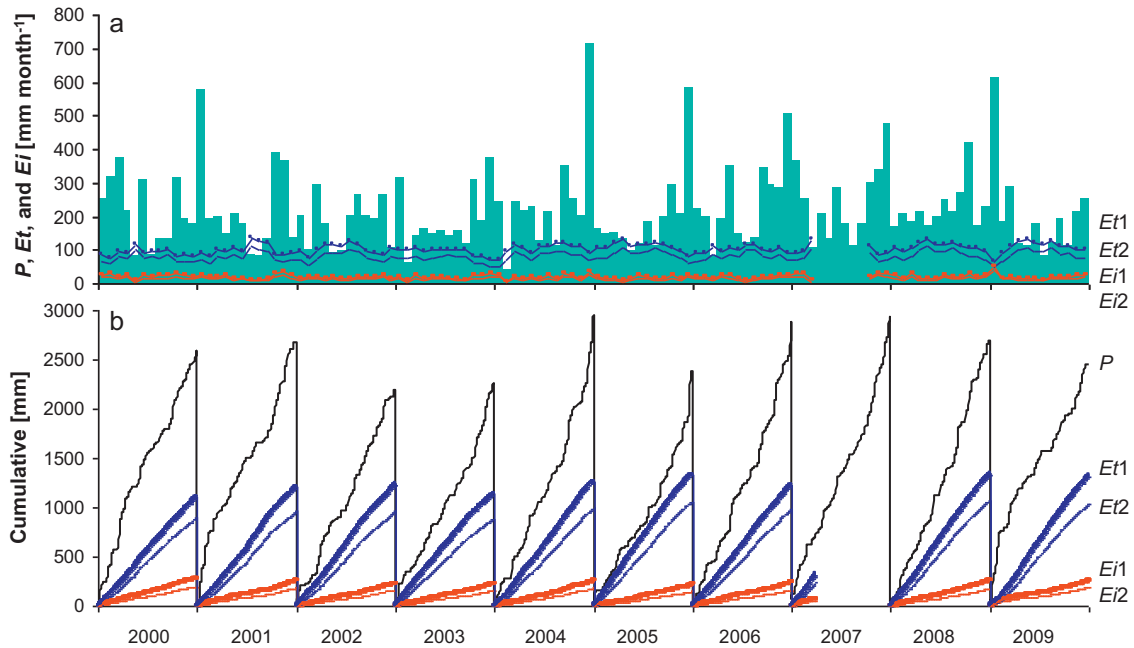


Fig. 7. (a) Temporal changes in monthly rainfall (P), transpiration (Et), and rainfall interception (Ei) in the period between 2000 and 2009. (b) Cumulative results for each year. Et_1 and Et_2 represent upper and lower limit of Et calculated using wet and dry soil G_c parameter sets, respectively. Ei_1 and Ei_2 represent the upper and lower limit of Ei estimated from $Sc = 0.7$ mm and 1.0 mm, respectively.

Et_2 were consistently ~ 100 mm through the years (Fig. 7a). The monthly Ei_1 and Ei_2 ranged between 10 mm and 30 mm through the years (Fig. 7a). The cumulative Et , Ei and ET values showed less significant year-to-year variations in the study period ($CV = 5\text{--}7\%$) (Fig. 7b; Table 1), although annual rainfall showed considerable year-to-year differences ($CV = 11\%$) (Fig. 7b; Table 1). Note that Et and Ei are the average of the maximum and the minimum estimates, and that ET is the sum of Et and Ei .

Annual Et was related to the mean annual VPD ($R^2 = 0.93$, $p < 0.001$) rather than mean annual Rn ($R^2 = 0.52$, $p < 0.05$), mean annual temperature ($R^2 = 0.25$, $p = 0.21$) or mean annual wind speed ($R^2 = 0.03$, $p = 0.78$). Annual Ei was related moderately to annual rainfall ($R^2 = 0.33$, $p = 0.10$) and the mean annual temperature ($R^2 = 0.34$, $p = 0.04$) rather than mean annual Rn ($R^2 = 0.02$, $p = 0.70$), mean annual VPD ($R^2 = 0.11$, $p = 0.42$) or mean annual wind speed ($R^2 = 0.05$, $p = 0.60$).

The annual Et and Ei averaged over 10 years were estimated to be 1114 mm and 209 mm, respectively (Table 1). The ratio of mean annual Ei to mean annual rainfall in the 10-year period was $\sim 8\%$, which approximated that of our previous study based on 3-year 4-ha-scale measurements of throughfall and stemflow

($\sim 8.5\%$; Manfroli et al., 2006). Annual ET averaged over 10 years was 1323 mm, which accounted for $\sim 51\%$ of the 10-year mean annual rainfall. This was smaller than that of our previous study ($\sim 70\%$) estimated from 1-year meteorological measurements and the small spatial-scale Ei measurements at this site (Kumagai et al., 2005).

4.4. Water deficit and soil moisture

Temporal changes in extractable soil water content in the 0–60-cm soil layer (ESW_{0-60}) and water deficit (WD) are shown in Fig. 8. There were no regular seasonal variations in ESW_{0-60} in the study period. Positive water deficit (WD) values were found in 23 months, and were distributed unpredictably in the last 10 years. The largest WD was measured in July 2001 and February 2004, with a value of 63 mm. In 8 of the 23 months, the WD was larger than ESW_{0-60} . The differences between WD and ESW_{0-60} (i.e., $WD - ESW_{0-60}$) were most significant in July–August 2001, and were $\sim 20\text{--}30$ mm. This result indicates that our forest could also suffer temporarily from water stress if the extractable soil water layer was restricted to 60-cm thickness.

Table 1

Summary for the period 2000–2009 of annual total rainfall (P), mean daily net radiation (Rn), mean daily downward short-wave radiation (Sd), mean daily downward long-wave radiation (Ld), mean vapor pressure deficit (VPD), mean temperature (T), total evaporation (ET) as the sum of total transpiration (Et) and total rainfall interception (Ei). Et and Ei were calculated using the average of the maximum and minimum estimates.

Elements	Year	2000	2001	2002	2003	2004	2005	2006	2007	2008	2009	Avg \pm S.D.
P [mm]		2582	2681	2191	2263	2944	2387	2875	2932	2689	2455	2600 \pm 272
Rn [$MJ m^{-2} d^{-1}$]		11.5	11.2	11.8	11.1	11.6	11.8	11.7	–	11.9	11.6	11.6 \pm 0.3
Sd [$MJ m^{-2} d^{-1}$]		16.2	16.7	16.8	16.0	16.3	16.1	16.0	–	16.1	15.7	12.2 \pm 0.4
Ld [$MJ m^{-2} d^{-1}$]		36.9	36.5	36.4	36.7	36.6	36.7	33.9	–	37.3	36.5	36.4 \pm 1.0
VPD [hPa]		4.8	4.4	4.0	3.5	4.7	5.6	4.5	6.2	6.0	5.8	5.0 \pm 0.9
T [$^{\circ}C$]		25.9	25.7	26.1	25.9	25.8	25.9	26.0	25.9	25.7	25.4	25.8 \pm 0.2
W [ms^{-1}]		1.8	2.1	2.4	2.3	2.4	2.3	2.1	–	2.3	2.3	2.2 \pm 0.2
ET [mm]		1229	1306	1293	1210	1351	1382	1310	–	1425	1402	1323 \pm 74
Et [mm]		1005	1087	1100	1012	1130	1197	1108	–	1207	1182	1114 \pm 74
Ei [mm]		224	219	192	198	221	184	202	–	219	221	209 \pm 15

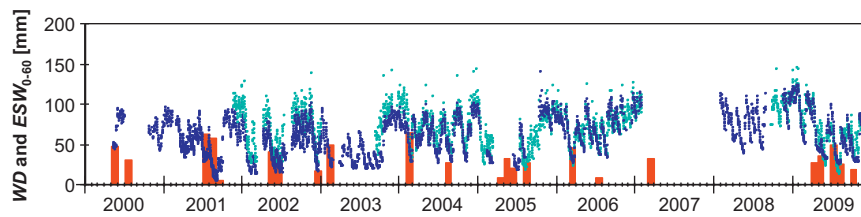


Fig. 8. Time series of extractable soil water at 0–60-cm depth (ESW_{0-60}) and water deficit (WD) calculated from monthly total evaporation (ET) minus monthly rainfall. Red bars represent WD . Light blue and navy dots represent ESW_{0-60} measured at the crane site. Blue dots represent ESW_{0-60} measured at the tower site (for reference to the color information in the figure legend refer to the on-line version of the article).

4.5. Comparison with other tropical forests

We related annual E_i measured in 40 tropical forests to annual rainfall (Fig. 9a). The 40 tropical rainforests included both seasonal forests and rainforests (Supplementary data). Most E_i values measured in tropical forests accounted for 10–20% of annual rainfall with a mean of 17% ($n=40$), although there was extremely large E_i relative to annual rainfall at some research sites. On the other hand, rainfall interception by our forest with an E_i ratio (i.e., annual E_i /rainfall) of 8% was at the lower limit of the data.

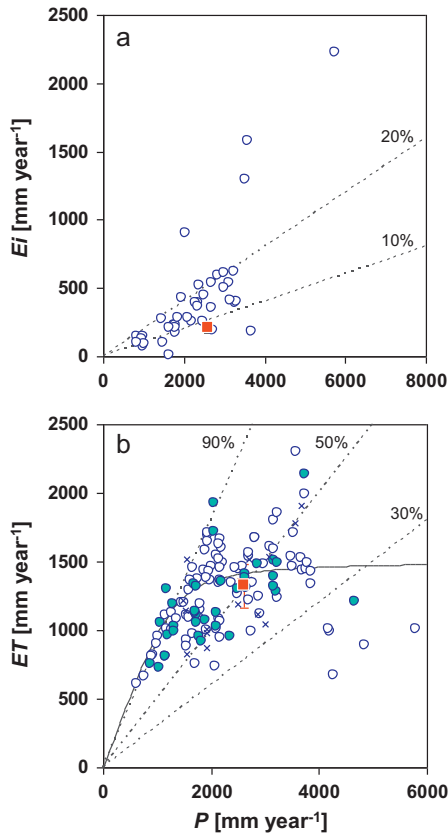


Fig. 9. Comparison between rainfall and (a) rainfall interception (E_i), and (b) total evaporation (ET). Dotted lines in Fig. 9a represent 10% and 20% of annual rainfall. Dotted lines in Fig. 9b represent 30%, 50%, and 90% of annual rainfall. In Fig. 9a, the red box represents our datum, and the open circles represent data from other tropical forests (Supplementary data). In Fig. 9b, the red box represents our datum. The open circles represent data obtained in other tropical forests using the watershed water balance method (WB). Filled circles are data from other forests using methods such as sap flux measurements, the meteorological method, model calculation, soil water balance method, and rainfall interception method based on canopy water balance. Crosses represent data derived from unknown methods (Supplementary data). The data were fitted using the non-rectangular hyperbola (Kume et al., 2007) ($R^2 = 0.35$) (for reference to the color information in the figure legend refer to the on-line version of the article).

We also related annual ET measured in 138 tropical forests to annual rainfall (Fig. 9b). The 138 tropical forests include both seasonal forests and various types of rainforest (Supplementary data). Most ET ratios (i.e., annual ET /rainfall) ranged between 30% and 90%. When rainfall was <2000 mm, ET increased with increases in annual rainfall. On the other hand, ET under annual rainfall of >2000 mm plateaued at ~ 1500 mm, although 22 forests with annual rainfall of >2000 mm had $ET > 1500$ mm. Note that the watershed water balance method could have possibly overestimated ET resulting from deep percolation (e.g., Oda et al., 2009), and that ET in 17 of the 22 forests was measured using the watershed water balance method. ET in our forest (1323 mm) was slightly lower than the global standard maximum value (1500 mm) under annual rainfall of >2000 .

5. Discussion and conclusions

To clarify the interannual variations in ET in the Bornean tropical rainforest, this study conducted 10-year meteorological measurements and developed an ET big-leaf model that successfully reproduced E_i and E_t at our study site (Figs. 5 and 6). The annual ET estimated from the model for a 10-year period showed less significant year-to-year variation than that of annual rainfall (Fig. 7b, Table 1). Consequently, the ratio of ET to annual rainfall ranged between 46% and 59%. Thus, this study suggests that the assumption of constant ET ratio based on a short-term measurement period could occasionally provide erroneous interpretations of water balance in tropical rainforests.

Fairly constant annual E_i was found in this study (Table 1). The fairly constant annual E_i was related to rainfall characteristics at this site. According to previous studies (e.g., Gash, 1979), the total number of rain events (n) and total rainfall duration (d) determine total E_i when the evaporation rate is assumed to be almost constant in a given period. Loescher et al., 2005 reported significant interannual variations in annual E_i in accordance with large interannual variations in annual rainfall. In Costa Rican tropical forest, wet canopy duration (i.e., total d in 1 year) was more than 30% of the year under mean annual rainfall of 3700 mm (Loescher et al., 2005), while at our site wet canopy duration ranged between 7% and 11% of the year under mean annual rainfall of 2600 mm. This suggests less intense longer duration storms in the Costa Rican tropical forests and more intense shorter duration storms at our Bornean site. In regions with more-intense shorter duration storms, interannual variations in annual rainfall might have less impact on year-to-year fluctuation of E_i , with smaller interannual variations in n and d than those of regions with less-intense longer duration storms. Note that E_i ratios on a rain event basis decreased with increases in storm size at our site (Manfroi et al., 2006), suggesting that beyond the canopy water storage capacity (Sc) considerable amounts of rainwater in intense storms is shed quickly from the canopy. Thus, contribution of E_i in intensive storms (i.e., ≥ 60 mm) to the total annual E_i could be less significant, although the contribution of intensive storms of ≥ 60 mm to total annual rainfall was signifi-

cant (Fig. 4b). To clarify whether the discrepancy between this and previous studies could be caused by regional-specific rainfall characteristics, cross-sites comparisons of *Ei* and rainfall characteristics in tropical forests could be useful.

Annual *Et* ranged between 1005 mm and 1207 mm, and the small fluctuations corresponded to year-to-year *VPD* differences. Normally, daily *Et* at our site is controlled by *Rn* rather than *VPD* due to the decoupled canopy to the atmosphere by boundary layer (Kumagai et al., 2004; 2005). However, our 10-year measurements showed conservative *Rn* (CV=2%) and significant year-to-year fluctuations in *VPD* (CV=18%). Consequently, at our site the year-to-year fluctuations in *Et* could be caused by *VPD*.

In previous temperate forest studies, soil drought due to seasonal and interannual variations in rainfall could be a determining factor for interannual variations in *Et* due to stomatal closure or leaf-fall (Wilson and Baldocchi, 2000; Lafleur et al., 2005; Scott et al., 2004; Granier et al., 2007; Thomas et al., 2009). At our site such severe drought did not occur because of the continuously distributed rainfall through the study period (Figs. 2 and 7a). The potential water deficit ($WD = ET - P$) could be replenished by the soil water content in the soil layer of 0–60-cm depth in most cases (Fig. 8). While, in the most severe dry period in the last 10 years (September 2001), the water deficit was larger than soil water content at 0–60-cm depth (Fig. 8). Thus, there is a possibility that soil water at >60-cm depth could be also utilized by trees in severe dry periods. Some previous studies noted the importance of deep roots in tropical forested ecosystems to maintain their water use and carbon assimilation in dry periods (e.g., Rocha et al., 2004; Kume et al., 2007). Further research including rooting depth measurements and stable isotope composition analysis of water in tree xylem and soil (Liu et al., 2010; Meinzer et al., 1999) under soil drought conditions would enable us to understand the contribution of deep soil water to total water use in unusually severe dry periods at this Bornean site.

Annual *ET* averaged over 10 years (1323 mm) was smaller than the standard global *ET* maximum in tropical forests (~1500 mm) (Fig. 9b). The smaller *Ei* (ratio of 8%) at our site compared to other tropical forests may explain the smaller *ET* (Fig. 9a). Most tropical forests show *Ei* ratios of 10–20% with a mean of 17% ($n = 40$). If we assume that our forest has the global mean *Ei* ratio (=17%) and that *Et* is constant, *ET* in our site would be 1557 mm, which approximates the global standard maximum of *ET* in tropical forests. The specific rainfall characteristics and the small water storage capacity of the canopy (*Sc*) could explain the smaller *Ei* at this site. Note that *Sc* at our site (=0.7–1.0 mm) was smaller than that of the Costa Rican tropical rainforest (=1.5 mm) (Loescher et al., 2005), that of an Indonesian forest (=1.1–4.5 mm) (Calder et al., 1986), and that of a Puerto Rican forest (=1.2–1.7 mm) (Schellekens et al., 1999).

Long-term trend analysis at this site using the rainfall record in the period 1968–2002 showed that rainfall intensity between February and August had a significant rising trend (Kumagai et al., 2009). On the other hand, a significant trend was not detected in the total rainfall and rainfall event duration through the year (Kumagai et al., 2009). These results suggest that *ET* in our forest is conservative in response to possible changes in rainfall regime, because intensive storm with short duration have little impact on interannual variations in annual *Ei*, *Et* and hence *ET*. On the other hand, this study did not include unusually severe drought (Fig. 2b), which tends to occur in Southeast Asia corresponding with El Niño cycles (Walsh and Newbery, 1999; Wang and Weisberg, 2000). An unusually severe drought event was observed in the Bornean forest during an El Niño period in 1987 and in 1998. Such unusually severe drought led to high tropical rainforest tree mortality (Nakagawa et al., 2000). This implies that more severe drought conditions than those of this study period could cause significant decrease in *Et* or tree mortality due to hydraulic failure (McDowell et al., 2008). Thus,

such severe droughts can have considerable impacts on *ET* due to limitation of *Et* and decrease of *Ei* frequency and duration. To clarify the impact of interannual rainfall variability on *ET* in the context of past and current climate, further long-term monitoring is required in this region.

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Appendix A. Supplementary data

Supplementary data associated with this article can be found, in the online version, at doi:10.1016/j.agrformet.2011.04.005.

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