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# **RESEARCH ARTICLE**

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#### **Kev Points:**

- A 7 year evapotranspiration (ET) in a tropical deciduous forest was measured
- · Higher-order Taylor series expansions were needed to explain ET controls
- · Rainfall variations exert physical and biological controls on ET variation

#### **Supporting Information:**

Supporting Information S1

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# Separating physical and biological controls on long-term evapotranspiration fluctuations in a tropical deciduous forest subjected to monsoonal rainfall

JGR

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Abstract Evapotranspiration (ET), especially in the mainland of the Indochina Peninsula, can impact and is impacted by the Asian monsoonal (AM) system, thereby prompting interest in its long-term variability. To separate the physical and biological factors controlling ET variability in a tropical deciduous forest under the AM influence, 7 year eddy covariance and ancillary measurements were collected and analyzed. The 7 year mean rainfall ( $P_r$ ) and ET along with their standard deviations were 1335 ± 256 and 977 ± 108 mm (about 73% of  $P_r$ ), respectively, suggesting close coupling between these two hydrologic fluxes. However, other physical and biological drivers decouple seasonal and annual variations of ET from Pr. To explore them, a big-leaf model complemented by perturbation analysis was employed. The big-leaf model agreed well with the measured ET at daily to multiyear time scales, lending confidence in its ability to separate biological and physical controls on ET. Using this formulation, both first-order and second-order Taylor series expansions of the total ET derivatives were applied to the big-leaf model and compared with measured changes in ET (dET). Higher-order and joint terms in the second-order expansion were necessary for matching measured and analyzed dET. Vapor pressure deficit (D) was the primary external physical controlling driver of ET. Leaf area index (LAI) and bulk stomatal conductance  $(q_s)$  were shown to be the main significant biological drivers of the transpiration component of ET. It can be surmised that rainfall variability controls long-term ET through physical (mainly D) and biological (mainly LAI and  $q_s$ ) factors in this ecosystem.

# 1. Introduction

The Asian monsoon, a large-scale circulation system driven by seasonal differential heating between the Eurasian continent and surrounding ocean, controls the seasonal and interannual rainfall variability in southeast (SE) Asia [e.g., Yasunari et al., 2006; Yasunari, 2007]. There is now significant interest in how the Asian monsoon maintains its wet state and to what extent ongoing rapid (and projected) land cover changes impact the onset, strength, and duration of the Asian monsoon. Numerical modeling can provide information about the interplay between land surface processes and the Asian monsoon such as the effects of deforestation on monsoonal rainfall. Using a general circulation model (GCM), earlier studies reported that the deforestation in SE Asia has less of an impact on regional monsoonal precipitation than the deforestation in Amazonian rainforests [Henderson-Sellers et al., 1993; Zhang et al., 1996]. However, recent numerical experiments using finer-resolution models and realistic land surface schemes have revealed that the deforestation in SE Asia can appreciably alter Asian monsoon circulation and precipitation [Mabuchi et al., 2005], and the concomitant regional-scale precipitation [Kanae et al., 2001]. Yasunari et al. [2006] also showed, from GCM experiments, that water vapor supplied from surface vegetation plays a critical role in the formation of the Asian monsoon circulation, and subsequent precipitation over the Asian monsoon region, suggesting the significance of whole ecosystem evapotranspiration (ET). Using satellite-derived precipitation and atmospheric reanalysis data, a recent observational study revealed earlier onset of the Asian monsoon in recent decades. The earlier arrival of the Asian monsoon appears to be connected to the enhanced atmospheric warming over land trend in May (mainly through sensible heat), as

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this warming trend provides the key necessary environmental conditions for forming the Asian monsoonal circulation [*Kajikawa et al.*, 2012].

Given the tight coupling between sensible heat flux and evapotranspiration over land, imposed by the surface energy balance, it follows that low-frequency variability in evapotranspiration arising from possible changes in physical or biological factors can directly contribute to the fluctuation in sensible heat flux. Sensible heat flux changes amplify the land-ocean heating contrast, thereby altering the atmospheric convergence responsible for the formation of the Asia monsoon in this region. It is precisely these findings that motivated this investigation into what controls low-frequency variability in evapotranspiration. At these low frequencies, biological and physical processes can interact to reshape the functional traits of ecosystems, particularly with regard to water and energy balances.

At the local spatial scale, fluctuations in precipitation and other climate variables alter evapotranspiration through multiple physical and biological pathways. For example, leaf phenology including timing of the leaf-on/off cycle is an important factor that determines seasonal and interannual variations of evapotranspiration, as evidenced from field studies of a tropical monsoonal deciduous forest in northern Thailand [*Yoshifuji et al.*, 2006; *Tanaka et al.*, 2009; *Yoshifuji et al.*, 2014]. To explore how these pathways separately and jointly impact variability in evapotranspiration, a 7 year record of half-hourly evapotranspiration measured by the eddy covariance method is analyzed, complemented by a perturbation analysis conducted on the Penman-Monteith equation [e.g., *Wilson and Baldocchi*, 2000; *Stoy et al.*, 2006]. The primary objective is to isolate the low-frequency contributions of physical and biological factors controlling evapotranspiration in a tropical deciduous forest subjected to the Asian monsoon climate. It is envisaged that such an analysis can "pinpoint" some of the slowly evolving physical and biological variables that impact evapotranspiration on long time scales thereby assisting climate models in projecting shifts in the Asian monsoon due to present and possible future evapotranspiration variations.

### 2. Materials and Methods

#### 2.1. Site Description

The experiments were carried out in an even-aged teak (*Tectona grandis* Linn. f.) plantation located in the upper part of the Chao Phraya river basin, in Mae Mo district, Lampang Province, northern Thailand (18°25'N, 99°43'E, 380 m above sea level). Teak plantations have been widely established throughout the tropics, especially in Indochina and India, due to their high market value—a trend that is likely to continue in the foreseeable future [*Kollert and Cherubini*, 2012]. A case in point is Thailand, where significant decreases in forest area were documented until the 1980s. Forest rehabilitation and plantation have since been vigorously promoted, resulting in the establishment of large-scale teak plantations by the Forest Industry Organization, one of which is used here as a case study.

The Mae Mo teak plantation was established in 1968. Tree density around the study site was 440 trees ha<sup>-1</sup> in 2005 and, due to tree mortality, dropped to 433 trees ha<sup>-1</sup> in 2012. The mean stem diameter at breast height (DBH) was 22.4 cm (standard deviation (SD) = 7.3 cm, n = 308) in November 2005. This DBH increased to 24.2 cm (SD = 8.3 cm, n = 303) by February 2012. The mean tree height was 19.9 m (SD = 3.5 m, n = 308) in November 2005 and 23.8 m (SD = 5.4 m, n = 303) in February 2012. Various herbaceous plants and shrubs occupy the understory throughout the year, but their density is low. The study site is situated on flat land where the stand structure is almost homogeneous. The soil is Loamy Paleustults (Thai classification) or Ultisol (U.S. Department of Agriculture, soil taxonomic classification) with 35.5% sand, 43.5% silt, and 21.0% clay, with a total porosity of 35–40%. Most fine and coarse roots were detected in the top 0.4 m below the soil surface [*Kume et al.*, 2013].

The climate is governed by the Asian monsoon and is characterized by two distinct seasons: a wet season (approximately May–October) and a dry (approximately November–April) season (Figure 1). This region occasionally suffers from severe flood events. For example, in 2011, the Chao Phraya river basin experienced annual precipitation of 1439 mm, which was 143% more than the long-term average rainy season precipitation (1982–2002) [*Komori et al.*, 2012]. In fact, a record-level annual precipitation of 1760 mm was measured in 2011 at this site. The mean annual temperature in the period of 2001 to 2010



**Figure 1.** Year-long climatic data of (a) daily precipitation ( $P_r$ ), (b) relative extractable soil water in the 0–60 cm soil layer ( $\Theta$ ), (c) daily total solar radiation ( $R_s$ ), (d) daily mean air temperature ( $T_a$ ), (e) daily mean vapor pressure deficit (D), (f) leaf area index (LAI), and (g) midday-averaged above-canopy water vapor flux under dry canopy condition (E). Shaded columns represent periods of rainy season. Mean yearly cycles of (h)  $P_r$ , (i)  $\Theta$ , (j)  $R_s$ , (k)  $T_a$ , (l) D, (m) LAI, and (n) E. For reference, one standard deviation of the binned variables is also shown (gray zone).

was 25.4°C, with a minimum mean monthly value of 21–22°C (in December–January) and a maximum mean monthly value of 29–30°C (in April).

#### 2.2. Micrometeorological Measurements

A 40 m tall scaffold tower was constructed in the study area for conducting micrometeorological and eddy covariance flux measurements. Incident solar radiation above the canopy (R<sub>s</sub>; Wm<sup>-2</sup>) and downward longwave radiation were measured with a pyranometer (CMP21, Kipp & Zonen, Delft, Netherlands) and a pyrgeometer (Model PIR, Epply, Newport, RI and CG4, Kipp & Zonen before and after February 2007), respectively, at a height of 41 m above the forest floor. Upward short-wave and long-wave radiations were measured using a pyranometer (CMP21) and a pyrgeometer (Model PIR), respectively, installed upside down at a height of 35 m above the forest floor, respectively. The sensors were mounted away from the tower using a 3 m horizontal boom. The boom was fixed on its south corner so as to minimize the effects of the tower. These four radiation components were used to calculate net radiation ( $R_{n}$ ; W m<sup>-2</sup>). Air temperature ( $T_{a}$ ; °C) and air relative humidity above the canopy were measured at 39 m above the forest floor with an air temperature and humidity sensor (HMP45A and HMP155, Vaisala, Helsinki, Finland before and after June 2012, respectively). Solar radiation ( $S_{b}$ ; W m<sup>-2</sup>) was also measured near the forest floor at a height of ~0.5 m, some 30 m from the tower, using a pyranometer (MS401, EKO, Tokyo, Japan and CMP21 before and after August 2008, respectively). Wind speed was measured using a cup anemometer (AC750, Makino Instruments, Tokyo, Japan), which was installed on the tower at 39 m above the forest floor. All sensors were sampled every 5 s, and their 10 min averages were recorded by data loggers (CR10X, CR23X, and CR1000, Campbell Scientific, Logan, UT).

Rainfall ( $P_r$ ; mm) was measured with a tipping bucket rain gauge (No. 34T, Ohta Keiki, Tokyo, Japan) connected to a data logger (Hobo Event, Onset Computer, Bourne, MA) in an open site about 500 m from the main meteorological tower.

Volumetric soil water content ( $\theta$ ; m<sup>3</sup> m<sup>-3</sup>) was measured about 40 m away from the meteorological tower at depths of 10, 20, 40, and 60 cm using time domain reflectometers (CS-615, Campbell Scientific) below the soil surface. Soil heat flux (G; W m<sup>-2</sup>) was also measured adjacent to the  $\theta$  measurement array at a depth of 3 cm using three soil heat flux plates (MF-180, EKO, Tokyo, Japan), respectively. The data were recorded with a data logger (CR10X) at 10 min intervals. The weighted average  $\theta$  in the 0–60 cm soil layer was calculated as  $\theta_{0-60} = (10\theta_{10} + 20\theta_{20} + 20\theta_{40} + 10\theta_{60})/60$ , where  $\theta_{10}$ ,  $\theta_{20}$ ,  $\theta_{40}$ , and  $\theta_{60}$  are  $\theta$  at depths of 10, 20, 40, and 60 cm, respectively. The relative extractable water in the soil ( $\Theta$ ) was calculated using  $\theta_{0-60}$  as  $\Theta = (\theta_{0-60} - \theta_r)/(\theta_s - \theta_r)$  [cf. *Kumagai et al.*, 2005: *Yoshifuji et al.*, 2006], where  $\theta_s$  and  $\theta_r$  are the saturated water content and the residual water content, respectively, averaged in the 0–60 cm layer. The values of  $\theta_s$  and  $\theta_r$  were determined as the maximum and minimum measured  $\theta$  within the 0–60 cm soil column during the entire experiment period (January 2006 to March 2013).

#### 2.3. Eddy Covariance Measurements

The sensible (H; W m<sup>-2</sup>) and latent ( $L_E$ ; W m<sup>-2</sup>) heat fluxes were measured using the eddy covariance technique. A three-dimensional sonic anemometer (USA-1, METEK, Elmshorn, Germany) and an open-path CO<sub>2</sub>/H<sub>2</sub>O infrared gas analyzer (LI-7500, Li-Cor, Lincoln, NE) were installed at 28.0 m above the forest floor. Wind speeds and gas concentration time series were all sampled and stored at 10 Hz (CR5000 data logger, Campbell Scientific, and CR1000). Calibrations were routinely performed on the gas analyzer every 3–4 months using LI620 (Li-Cor), zero gas, and CO<sub>2</sub> span gas (Thai Industrial Gases Public Company Limited, Bangkok, Thailand). Two large data gaps (March 2007 to January 2008 and April–November 2009) occurred in the 7 year observation campaign due to equipment failures following lightning strikes and electric power supply problems.

Postprocessing of the data series was conducted as follows, Spikes, outliers, possible sensor/logger malfunctions, and data discontinuities in the raw 10 Hz data were first detected and removed [e.g., Moncrieff et al., 2005]. A double rotation method was then applied to the sonic anemometer velocities [McMillen, 1988; Kaimal and Finnigan, 1994]. Then, sensible and latent heat flux were computed using the 30 min averaged covariances of the rotated 10 Hz vertical velocity, air temperature, and water vapor concentration time series, respectively. Corrections for density effects for water vapor were also applied prior to computing covariances with the vertical velocity [Webb et al., 1980]. The influence of the minimum fetch, which was about 500 m, on the estimated scalar fluxes was examined using a 2-D analytic solution to the diffusion equation proposed by Schuepp et al. [1990]. The 80% cumulative footprint value [Schuepp et al., 1990] decreased by  $218 \pm 98$  m from  $228 \pm 92$  m (mean  $\pm$  SD) with increasing vegetation high during the study period (from February-August 2006 to February-August 2012). However, in both periods, the main footprints (source area) of the flux measurement were determined to be within the site border. The in-stationarity test [Foken and Wichura, 1996] was employed, and if there was a difference of more than 60% between the 30 min averaged covariance and an average of six 5 min averaged covariances within the 30 min run, the flux calculated in the 30 min run was rejected. To minimize the effects of tower distortions and other anomalies, sensible and latent heat flux values measured during rain events and when the friction velocity was  $< 0.20 \text{ m s}^{-1}$  [see Saleska et al., 2003; Miller et al., 2004] were discarded, as well as those runs where the mean wind direction originated from a sector bounded by 270-360° (when the wind flow was distorted by the tower).

Using the postprocessed eddy covariance fluxes with such filters, the energy balance was checked. It was found that  $H + L_E = 0.69$  ( $R_n - G$ ) reasonably represented ( $R^2 = 0.70$ ) the half-hourly energy balance closure. Then, a high-frequency spectral correction was applied using standard surface layer transfer functions [*Moncrieff et al.*, 1997]. As a result of applying this procedure, the lack of an energy balance closure was improved from 69% to 72%.

We did not find a significant hysteresis relation between  $R_n - G$  and  $H + L_E$  (data not shown) without any adjustment by storage terms. This implies that the contribution of the storage flux to the available energy and the scalar fluxes might be small [*Leuning et al.*, 2012]. Storage fluxes are not expected to be significant in a canopy with low stem density, moderate leaf area index, and moderate canopy height conditions [see *Ohkubo and Kosugi*, 2008], especially when the friction velocity exceeds 0.2 m s<sup>-1</sup>.

Here the median of imbalance ratio  $((H + L_E)/(R_n - G))$  in the morning (from 8:00 A.M. to 10:30 A.M., weakly unstable condition) and midafternoon period (from 1:30 P.M. to 4:00 P.M., unstable condition) were 0.68 and 0.70, respectively, in the dry season (January–April). The difference in measured Bowen ratio between the morning and the midafternoon period was also small (0.96 in the morning and 0.95 in the midafternoon period were 0.74 and 0.68, respectively, in the wet season (June–September). The median of the Bowen ratio in the morning and midafternoon period were 0.74 and 0.68, respectively, in the wet season (June–September). The median of the Bowen ratio in the morning and midafternoon period were large (0.40 and 0.24, respectively). These results suggest that entrainment processes connected to boundary layer height and its temporal dynamics (leading to entrainment flux differences during these two periods) are not the main culprit for the lack of energy balance closure as earlier suggested [*Lee et al.*, 2004]. Therefore, we did not employ the Bowen ratio correction for the fluxes. However, further analyses will be needed to resolve all the reasons for the imbalance and the underestimation of the sensible and latent heat fluxes measured by eddy covariance, the degree of closure is well within the range reported for other forest sites. For example, mesoscale circulation, which is driven in part by landscape level (approximately 20 km × 20 km) heterogeneity, might affect the energy balance closure [*Foken*, 2008; *Stoy et al.*, 2013].

#### 2.4. Leaf Area Index Estimate

According to Beer's law, the silhouette area index of leaves plus stems and branches (plant area index, PAI;  $m^2 m^{-2}$ ) can be expressed as [Holst et al., 2004]

$$\mathsf{PAI} = \mathsf{LAI} + b = -\frac{1}{k}\mathsf{ln}(R_t),\tag{1}$$

where LAI is leaf area index (m<sup>2</sup> m<sup>-2</sup>), b is an empirical constant representing the fixed contribution of stems and branches to radiation extinction, k is the extinction coefficient, and  $R_t$  is the radiative transmittance defined as  $S_b/R_{sr}$ , where  $S_b$  and  $R_s$  are separately measured as described earlier every 30 min. Once the parameters b and k are determined, the temporal variation in leaf area index can be produced by equation (1) with continuous monitoring of radiative transmittance [see Maass et al., 1995; Holst et al., 2004; Wang et al., 2004]. The parameters b = 1.03 and k = 0.52 were determined by regressing periodic data of leaf area index estimated from litterfall collection and the  $R_t$  observation [Yoshifuji et al., 2014]. The value of k is known to vary with zenith angle and cloudiness [Baldocchi et al., 1984; Holst et al., 2004]. To reduce dayby-day variation in the cloudiness or direct sunlight condition, we selected radiative transmittance values measured during the period where the zenith angle was  $> 75^{\circ}$  and used the 15 day moving average of the daily radiative transmittance as an input into equation (1) [Yoshifuji et al., 2014]. On the other hand, the periodic leaf area index data were calculated for dates in which litterfall measurements were made, during the period from the day of the maximum radiative transmittance in the dry season to the day of the minimum radiative transmittance in the previous rainy season. Since the studied teak canopy was completely dormant (i.e., LAI = 0) on the day of the maximum radiative transmittance, the leaf area index for each day in which litterfall measurements were taken was calculated by multiplying the total dry weight of the litter lost from the teak canopy during the period between this day and the day of the maximum radiative transmittance with the specific leaf area. Here the specific leaf area was estimated at  $87.6 \text{ cm}^2 \text{g}^{-1}$  using measurements of 40 sample leaves obtained at the site. Using a plant canopy analyzer (LAI-2000, Li-Cor), periodic checking was conducted to determine whether the leaf area index estimates from equation (1) with b = 1.03 and k = 0.52 were reasonable. These checks suggest no large discrepancies between the two leaf area index estimates throughout the studied period [Yoshifuji et al., 2014].

#### 2.5. Models for Estimating Evapotranspiration

Daily evapotranspiration (ET;  $mm d^{-1}$ ) was expressed as

$$ET = E + E_i$$
  
=  $E_{tc} + E_{ts} + E_i$ , (2)

where  $E_{tc}$  is canopy transpiration (mm d<sup>-1</sup>),  $E_{ts}$  is vegetation floor evaporation (mm d<sup>-1</sup>), and  $E_i$  is canopy interception (mm d<sup>-1</sup>). Evapotranspiration was calculated with a big-leaf model [e.g., *Kelliher et al.*, 1995]. We note that canopy transpiration and vegetation floor evaporation were calculated using the Penman-Monteith equation [e.g., *Monteith and Unsworth*, 1990] and a modified Priestley and Taylor's equation [*Priestley and Taylor*, 1972] with half-hourly forcing data and aggregated to daily.

Abbreviation	Units	Definition
А	$Wm^{-2}$	Daily total available energy
Ac	$W m^{-2}$	Daily total absorbed energy by the canopy
As	$W m^{-2}$	Daily total absorbed available energy by the vegetation floor
CA	-	Extinction coefficient determined
C <sub>p</sub>	$J kg^{-1} K^{-1}$	Specific heat of air at a constant pressure
D	Pa	Vapor pressure deficit
d	m	Zero plane displacement
d <sub>m</sub>	-	Model parameter (equation (7))
Ε	$mm d^{-1}$	Daily above-canopy water vapor flux under dry canopy conditions
Ei	$mm d^{-1}$	Canopy interception
ET	$\operatorname{mm} \operatorname{d}^{-1}$	Total ecosystem evapotranspiration
<i>E</i> <sub>tc</sub>	$mm d^{-1}$	Canopy transpiration
Ets	mm $d^{-1}$	Vegetation floor evaporation
G	$W m^{-2}$	Soil heat flux
Ga	m s <sup>-1</sup>	Aerodynamic conductance
G <sub>c</sub>	m s <sup>-1</sup>	Canopy conductance
g <sub>s</sub>	$m s^{-1}$	Bulk stomatal conductance
g <sub>sref</sub>	m s <sup>-1</sup>	Model parameter (equation (6))
Н	$W m^{-2}$	Sensible heat fluxes
<i>k</i> 1	$W m^{-2}$	Model parameter (equation (6))
k2	$\ln(\Theta)^{-1}$	Model parameter (equation (6))
LAI	$m^2 m^{-2}$	Leaf area index
LE	$W m^{-2}$	Latent heat fluxe
m	$\ln(D)^{-1}$	Model parameter (equation (6))
Pr	mm $d_2$	Rainfall
Q <sub>max</sub>	$W m_{2}^{2}$	Model parameter (equation (6))
R <sub>n</sub>	$Wm^{-2}$	Net radiation
R <sub>s</sub>	W m <sup>-2</sup>	Incident solar radiation above the canopy
R <sub>t</sub>		Radiative transmittance
Sb	W m <sup>-2</sup>	Solar radiation near the forest floor
Ta	°C1	Air temperature
U	m s <sup>-1</sup>	Wind speed
z <sub>0m</sub>	m	Momentum roughness length
z <sub>0v</sub>	m	Roughness length
α	-	Interception loss coefficient
α <sub>PT</sub>	- 	Priestley-laylor coefficient
γ	PaK	Psychometric constant
Δ	$PaK^{-1}$	Ratio of charge of saturation water vapor pressure with temperature
3	mmm s kg $30$ min $3 -3$	Unit adjustment constant
θ	m <sup>-</sup> m <sup>-</sup>	Volumetric soil water content
Θ	3	Relative extractable water in the soil
θ <sub>0-60</sub>	m m -	Weighted average $\theta$ in the 0–60 cm soil layer
ĸ		von Karman constant
λ	J Kg	Latent heat of water vaporization
ρ	kg m	Density of dry air

 Table 1. List of Abbreviations

Canopy transpiration was calculated as follows:

$$E_{\rm tc} = \frac{\varepsilon}{\lambda} \frac{\Delta A_{\rm c} + \rho C_{\rm p} D G_{\rm a}}{\Delta + \gamma [1 + G_{\rm a} / (g_{\rm s} {\rm LAI})]},\tag{3}$$

where  $\Delta$  is the ratio of charge of saturation water vapor pressure with temperature (Pa K<sup>-1</sup>),  $\rho$  is the mean density of dry air (kg m<sup>-3</sup>),  $C_p$  is the specific heat of dry air at constant pressure (J kg<sup>-1</sup> K<sup>-1</sup>), D is the vapor pressure deficit above the canopy (Pa),  $\varepsilon$  is the unit adjustment constant (1800 mm m<sup>2</sup> s kg<sup>-1</sup> 30 min<sup>-1</sup>),  $\gamma$  is the psychometric constant (66.5 Pa K<sup>-1</sup>), and  $\lambda$  is the latent heat of water vaporization (2.5 × 10<sup>6</sup> J kg<sup>-1</sup>). The  $A_c$  is the absorbed energy by the canopy (W m<sup>-2</sup>) given as

$$A_{\rm c} = [1 - \exp(-c_{\rm A} {\rm LAI})]A, \tag{4}$$

where  $c_A$  is an extinction coefficient determined to be 0.7 from net radiation measurements above and below the canopy [*Igarashi et al.*, 2015], and A is the available energy (W m<sup>-2</sup>).  $G_a$  is the aerodynamic conductance

 $(m s^{-1})$  in the absence of any thermal stratification (a plausible assumption given the small sensible heat flux, the large friction velocity, and the small measurement heights above the canopy) as

$$G_{a} = \frac{\kappa^{2}U}{\ln[(z-d)/z_{0m}]\ln[(z-d)/z_{0v}]},$$
(5)

where  $\kappa$  is the von Karman constant (= 0.41), U is the wind speed (m s<sup>-1</sup>) at height z (= 39 m), where humidity was also measured in this study, d is the zero plane displacement (= 20.7 m),  $z_{0m}$  is the momentum roughness length (= 1.93 m) [*Igarashi et al.*, 2015]. Here  $z_{0v}$  is the roughness length (m) governing the transfer of vapor assumed to be the same as heat and 1/10 of  $z_{0m}$ . The  $g_s$  is the stomatal conductance (m s<sup>-1</sup>), expressed by series of multiplicative functions [e.g., *Jarvis*, 1976]:

$$g_{\rm s} = g_{\rm sref} \left[ \frac{R_{\rm s}(Q_{\rm max} + k_1)}{Q_{\rm max}(R_{\rm s} + k_1)} \right] (1 - k_2 \ln\Theta) (1 - m \ln D), \tag{6}$$

where  $R_s$  (W m<sup>-2</sup>),  $\Theta$ , and D denote values of  $R_s$ ,  $\Theta$ , and D, respectively, and  $g_{sref}$  (m s<sup>-1</sup>),  $Q_{max}$  (W m<sup>-2</sup>),  $k_1$  (W m<sup>-2</sup>),  $k_2$  (In( $\Theta$ )<sup>-1</sup>), and m (In(D)<sup>-1</sup>) are fitting parameters determined so as to minimize the root-mean-square error (RMSE) between predicted and actual stomatal conductance. The derivation of the actual stomatal conductance is reviewed in Appendix A.

A modified Priestley and Taylor's equation [*Priestley and Taylor*, 1972] was used to compute vegetation floor evaporation, given by

$$E_{\rm ts} = \alpha_{\rm PT} \frac{\varepsilon}{\lambda \Delta + \gamma} A_{\rm s} (1 + d_m \ln \Theta), \tag{7}$$

in which

$$A_{\rm s} = A - A_{\rm c},\tag{8}$$

where  $A_s$  is the available energy absorbed by the vegetation floor (W m<sup>-2</sup>). Priestley-Taylor coefficient  $\alpha_{PT}$  and  $d_m$  are the parameter estimated using nonlinear least squares regression between vegetation floor evaporation calculated from equation (7), and measured latent heat flux, during the leafless (i.e., LAI = 0) conditions, where measured latent heat flux can be assumed to be vegetation floor evaporation.  $\alpha_{PT}$  and  $d_m$  were validated as 0.76 and 0.23, respectively. It should be noted that forest floor evaporation contains not only soil evaporation but also transpiration from understory vegetation. It is assumed that herbaceous plants and shrubs on the forest floor are active even in the dry season because of their green leaves.

From the throughfall and stemflow measurements at the study site, the canopy interception was determined from daily rainfall ( $P_r$ ; mm d<sup>-1</sup>) as

Ei

$$= \alpha P_r,$$
 (9)

where  $\alpha$  is an interception loss coefficient. It should be noted that throughfall ratios, which ranged from 0.89 to 0.93, showed no clear seasonality, despite different canopy phenophases (from leafless to leaf-on) [*Tanaka et al.*, 2014]. Here  $\alpha$  is estimated to be 0.075 (N. Tanaka et al., unpublished data, 2006). List of abbreviations data are available, see Table 1.

#### 3. Results

#### 3.1. Variability in Physiological and Biological Drivers of Measured Latent Heat Flux

Daily measurements of environmental drivers, as well as leaf area index and measured latent heat flux during the study period (January 2006 to February 2013) are shown in Figures 1a–1g. The mean annual rainfall and the SD during 2006–2012 were 1326±268 mm. There was a clear seasonal variation in rainfall, rainy, and dry seasons. The delineation of these seasons was determined from a 5 day mean rainfall criteria [*Matsumoto*, 1997], which is commonly used in Asian monsoon region. The relative extractable water exceeded 0.5 in the wet seasons and dropped to about 0.1 in the dry seasons. The annual average of daily cumulative incident solar radiation and their SD were  $17.6 \pm 4.2 \text{ MJ m}^{-2} \text{ d}^{-1}$ , but clear seasonal variations were also noted. A decrease in incident solar radiation caused by increased cloud cover in the wet season was compensated for by the maximal solar elevation. Hence, the mean incident solar radiation values in the wet and dry seasons were similar. However, the maximum incident solar radiation occurred in the wet season.



**Figure 2.** Comparison between eddy covariance-measured (observed *E*) and big-leaf modeled daily above-canopy water vapor flux for dry canopy conditions (modeled *E*). The 95% upper and lower prediction interval (dashed lines) and the 1:1 line (solid line) are also shown.

The maximum daytime mean air temperature surpassed 30°C in the late dry seasons, and the minimum air temperature was around 20°C in the early dry seasons (November-December). The value of vapor pressure deficit showed a distinct seasonal variation with the highest daily mean vapor pressure deficit (around 3.0-4.0 kPa) occurring in the late dry season (January-March) because of the high air temperature and the lowest vapor pressure deficit values (≤1.0 kPa) predictably occurring throughout the wet seasons. Seasonal variation in leaf area index and the canopy duration (leaf-on) can be explained by seasonal variation in rainfall and soil moisture availability, and the leaf area index values reached a plateau early in the wet seasons. It should be noted that the timing of leaf-out and leaf-fall, i.e., the canopy duration, differed considerably between study years. Measured latent heat flux,

which is averaged in midday (10:00–15:00), also showed clear seasonal variations. The increased measured latent heat flux and the timing of leaf-out corresponded to the start of the rainy season, but measured latent heat flux and leaf area index were not perfectly in phase, because of seasonal variation in physiological and biological factors.

Interannual fluctuation in environmental factors, leaf area index, and measured latent heat flux are shown in Figures 1h–1n. Daily rainfall showed clear seasonality and reached almost zero in the dry season. The value of the SD of rainfall increased as rainfall increased and was large in the rainy season. Some rainfall events (premonsoon rainfall) during the late dry season also contributed to the rainfall SD in the dry season. These rainfall events in the rainy season and premonsoon events were important in explaining the interannual fluctuations of other variables. The SD in the late dry season was larger than that in the early dry season. This means that the beginning of the rainy season was quite variable, and the end time was less so at this site. Relative extractable water also showed clear seasonality, reflecting the seasonality in rainfall. The clear seasonality in incident solar radiation was also evident; the incident solar radiation peaked just before the rainy season due to the absence of clouds. The fluctuations in incident solar radiation were large in the rainy season and late dry season (January-March) due to the occurrence of rainfall. The air temperature also showed seasonal trends, high in the late dry season (January–March) and low in December-January. However, seasonal and interannual variations in air temperature were smaller than those in other variables. The SD of vapor pressure deficit was large in the dry-rainy season transition period due to the on-off nature of rainfall. The SD of the leaf area index was at maximum during the period from day of year (DOY) 140-150 (from end of May to beginning of June). The SD of daily measured latent heat flux increased with increasing daily measured latent heat flux in the dry-rainy season transition period but remained high in the rainy season. In contrast, some fluctuations in daily measured latent heat flux were measured in the late dry season (February-March) due to the occurrence of premonsoon rainfall.

#### 3.2. Estimating Canopy Transpiration and Vegetation Floor Evaporation

Figure 2 compares half-hourly measured latent heat flux from the eddy covariance method with modeled *E* (sum of canopy transpiration and vegetation floor evaporation). The parameters in equation (6) were determined as  $g_{sref} = 10.8 \times 10^{-3}$  (m s<sup>-1</sup>),  $Q_{max} = 660$  (W m<sup>-2</sup>),  $k_1 = 450$  (W m<sup>-2</sup>),  $k_2 = 0.29$  (ln( $\Theta$ )<sup>-1</sup>), and m = 0.81 (ln(D)<sup>-1</sup>), and the model reproduced the measurements well. It should



**Figure 3.** Seasonal and interannual variations in (a) daily rainfall  $(P_r)$ ,

transpiration ( $E_{tc}$ ), and vegetation floor evapotranspiration ( $E_{ts}$ ).

(b) leaf area index (LAI), and (c) estimated canopy interception  $(E_i)$ , canopy

be noted that these empirical parameters (equation (6)) showed insignificant differences between years and were not affected by biases associated with specific periods [Igarashi et al., 2015]. Hence, regardless of some uncertainty introduced by measurement errors, the big-leaf model is sufficiently reliable to link physical and biological drivers of evapotranspiration for the purposes of this work (at least at these space and time scales). The mean annual rainfall and evapotranspiration, with their SD, during March 2006 to February 2013 were 1335 ± 256 mm and 977 ± 108 mm, respectively.

Daily rainfall, leaf area index, and evapotranspiration components computed using the big-leaf model are shown in Figure 3. Because the leaf area index dropped to near zero during the period from DOY 60–70 (beginning of March), increasing of canopy transpiration started in concurrence with rainfall in the rainy season. Vegetation floor evaporation decreased with increasing leaf area index as expected. Hence, leaf area index (a biological factor here) played a significant role in partitioning evapotranspiration into canopy transpiration and vegetation floor evaporation. The timing for increases in canopy transpiration and decreases in vegetation floor evaporation in the leaf-out season corresponded to the start of the rainy season, though did not perfectly correspond to the initial increase in leaf area index because of seasonal and interannual variations in physical (e.g., rainfall, vapor pressure deficit, and aerodynamic conductance) and biological (e.g., stomatal conductance, and relative extractable water) factors.

#### 4. Discussion

To address the study objectives, the discussion is presented in three sections. The first reports the annual evapotranspiration and the categorization of the record of each of the 7 years record based on the annual rainfall and length of rainy season. The second presents the outcome of a multivariate higher-order perturbation analysis separating the physical from biological drivers. The third considers physical and biological factors affecting year-to-year differences in evapotranspiration for each of the categories, thereby completing the study objectives.



**Figure 4.** Relation between annual rainfall ( $P_r$ ) and annual evapotranspiration (ET) during the study period 2006–2012. The 1:1 and the regression ( $R^2 = 0.83$ ) lines are also shown. The observation year are also listed along with the reference year (2012) (see Figure 5).

#### 4.1. Annual Evapotranspiration and Yearly Classification

The relation between annual rainfall and evapotranspiration is shown in Figure 4. When annual rainfall was < 1400 mm, evapotranspiration increased with annual rainfall. On the other hand, when annual rainfall > 1400 mm, evapotranspiration did not exhibit any significant dependence on rainfall. This result compares favorably with earlier metaanalytic reviews that suggested that evapotranspiration depends on rainfall when rainfall < 1000 mm (due to water supply limitations) and plateaus or scatters when rainfall > 2000 mm [e.g., *Zhang et al.*, 2001; *Komatsu et al.*, 2012].

Hence, one measure of "cycling" of water is given by the mean evapotranspiration/rainfall  $(ET/P_r)$ , which was 0.73 for the entire study duration. This ratio was reported to vary from 0.3 to 0.9 across some



**Figure 5.** Relation between rainy season length and annual rainfall ( $P_r$ ) for each year. The observation year is also shown. The observation years were categorized into three rainfall regimes, i.e., "Large annual rainfall & Long rainy season," "Small annual rainfall & Long rainy season," and "Small annual rainfall & Short rainy season." Note that 2012 is the reference year, and vertical and horizontal dashed lines represent mean rainy season length and mean  $P_r$  during the study duration, respectively.

138 tropical forests that included seasonal forests [Kume et al., 2011]. The value of ET/Pr was also reported as 0.54 in an evergreen forest in northern Thailand [Tanaka et al., 2003], 0.67 at lowland dry evergreen forest in central Cambodia [Nobuhiro et al., 2009], and 0.66-0.71 at a Eucalyptus plantation in southeast China [Lane et al., 2004]. The  $ET/P_r$  value at this site was toward the upper end of the range compared to other tropical forestssuggesting that here a large part of the incident rainfall consumed was by evapotranspiration. Unsurprisingly, expanding teak plantations might play a prominent role in moisture supply to the atmosphere within the mainland of the Indochina Peninsula thereby impacting the Asian monsoon. As shown earlier [Igarashi et al., 2015], the effect of interseasonal and interannual variations in A available energy and  $R_s$  incident solar radiation to evapotranspiration was minor.

Based on the annual rainfall and rainy season length, each year was classified into one of three classes with the year 2012 as a baseline reference (Figure 5). This base reference was selected because 2012 closely approximates the long-term average annual rainfall and rainy season length. In fact, the annual evapotranspiration in 2012 (984 mm) was close to the mean evapotranspiration during the 7 year study period. Compared to 2012, years 2006 and 2011 were characterized by a "Large annual rainfall & Long rainy season," years 2008 and 2009 "Small annual rainfall & Long rainy season," and years 2007 and 2010 "Small annual rainfall & Long, respectively). As noted earlier, a major flood event occurred in 2011 resulting in the highest rainfall on record.

#### 4.2. Perturbation Analysis

The big-leaf model, which compared favorably with the eddy covariance-measured evapotranspiration at daily time scales, is now used to analytically link evapotranspiration variations to variations in physical and biological factors. Because of the analytical tractability of this model, it is ideal for conducting perturbation analyses intended to isolate the relative contributions of physical and biological factors that give rise to seasonal and annual changes in evapotranspiration. Because of the multiple variables involved, a brief



**Figure 6.** Averaging time versus root-mean-square error (RMSE) between observation and total change in evapotranspiration (dET). Closed and open circles represent computations from the first-order and the first-order with second-order Taylor series expansions, respectively. Vertical lines denote 30 days.

review of perturbation analysis with the simplest case of a two-variable function is provided for illustration.

Consider a dependent variable u (e.g., representing evapotranspiration) and two independent variables  $x_1$  and  $x_2$  (e.g., representing light and vapor pressure deficit) linked to u by a function f (e.g., represented by the big-leaf model) so that

$$u = f(x_1, x_2).$$
 (10)

Total changes in u = du can be determined from the total derivative of equation (10), which can be minimally represented by a firstorder Taylor series expansion given as

$$du = \frac{\partial u}{\partial x_1} dx_1 + \frac{\partial u}{\partial x_2} dx_2.$$
(11)

Here the  $\partial u/\partial xi$  (*i* = 1, 2) defines the overall sensitivity of the dependent variable *u* on each of the two independent variables and d*xi* defines the changes (or perturbations) in the independent variables.

The total derivative of equation (10) can be generalized using higher-order Taylor series expansions. For example, in the case of a second-order Taylor series expansion, du is given as

$$du = \underbrace{\left[\frac{\partial u}{\partial x_1}dx_1 + \frac{\partial u}{\partial x_2}dx_2\right]}_{\text{First Order}} + \frac{1}{2!}\underbrace{\left[\frac{\partial^2 u}{\partial x_1^2}dx_1^2 + \frac{\partial^2 u}{\partial x_2^2}dx_2^2 + 2\frac{\partial^2 u}{\partial x_1\partial x_2}dx_1dx_2\right]}_{\text{Higher Order and Mixed Term}}.$$
(12)

Here  $(\partial^2 u/\partial x_i^2) dx_i^2$  and  $(\partial^2 u/\partial x_1 \partial x_2) dx_1 dx_2$  are higher-order and mixed terms, respectively. The mixed terms signify "joint effects"—measuring how perturbations in x1 and x2 jointly (or in a multiplicative manner) impact du above and beyond the linear sum characterizing the first-order term. Higher-order and mixed terms are commonly neglected when small perturbation analysis is assumed [e.g., *Wilson and Baldocchi*, 2000; *Stoy et al.*, 2006]. However, there is no a priori reason to neglect them as they may be significant (especially when extreme events such as the 2011 flood exist in the record). Given the large seasonal and year-to-year variability expected here (see Figures 1h–1n), higher-order and mixed terms are likely not to be so small.

Averaging times also determine the size of these increments (i.e., dx1, dx2, dx12, dx22, and dx1dx2). Longer averaging times (e.g., yearly and half yearly) may yield smaller increments when compared to shorter averaging times (e.g., hourly). Defining these increments at time scales where the function *f* accurately links *u* to *x*1 and *x*2 is also imperative.

Using equations (2), (3), (7), and (9), the changes in evapotranspiration (= dET) can be expressed via a first-order linear Taylor series expansion as

$$dET = dE_{tc} + dE_{ts} + dE_{i}$$

$$= \frac{\partial E_{tc}}{\partial A} dA + \frac{\partial E_{tc}}{\partial D} dD + \frac{\partial E_{tc}}{\partial G_{a}} dG_{a} + \frac{\partial E_{tc}}{\partial LAI} dLAI + \frac{\partial E_{tc}}{\partial g_{s}} dg_{s}$$

$$+ \frac{\partial E_{ts}}{\partial A} dA + \frac{\partial E_{ts}}{\partial LAI} dLAI + \frac{\partial E_{ts}}{\partial \Theta} d\Theta + \frac{\partial E_{i}}{\partial P_{r}} dP_{r}.$$
(13)

Here rainfall, available energy, vapor pressure deficit, and aerodynamic conductance were defined as physical factors, while stomatal conductance, leaf area index, and relative extractable water were defined as biological factors, respectively. We note that perturbation analysis applied for a daily summed evapotranspiration data and averaged forcing for daytime. While relative extractable water is generally a physical factor, largely derived by precipitation characterizing the volume of water per unit volume of soil available for evapotranspiration, it is categorized here as a biological factor. Because of the bimodality of the relative extractable water (equation (7)) in wet and dry seasons, it may be considered to be a biological factor because it dictates the plant hydraulic capacity to uptake water on the time scales of interest (daily to seasonal). This categorization is based on the premise that during the dry season, precipitation is small (equation (7)). The dET also can be expressed using the second-order Taylor series expansion (equation (S11) in Text S1 in the supporting information).

To determine the optimal averaging times needed for calculating dET, both first- and second-order expansions were applied to the big-leaf model and compared with measured dET for several averaging times (Figure 6). Root-mean-square error (RMSE) of dET between 2012 and other years was then calculated for each averaging time and for both expansions. The higher-order expansion suggested minor reductions in RMSE for averaging times > 30 days (<10 mm yr<sup>-1</sup>). However, the RMSE in the first-order analysis remained large even at the annual time scale (= 39 mm yr<sup>-1</sup>). These findings suggest that a second-order expansion is necessary when explaining how biological and physical drivers explain long-term variability in evapotranspiration (>1 month). These results indicate that (i) the performance of the big-leaf model in explaining evapotranspiration from physical and biological factors and (ii) the increments needed in the perturbation analysis depend on the averaging time with monthly time scales (or longer) are necessary.



**Figure 7.** Contribution of (a) physical, (b) biological, and (c) mixed physical and biological drivers to total change in evapotranspiration (dET), categorized into three rainfall regimes, i.e., "Large annual rainfall & Long rainy season," "Small annual rainfall & Long rainy season," and "Small annual rainfall & Short rainy season." Variables in legend can be seen in text.

transpiration by saturating the leaf surface. At the same time, small rainfall events are associated with higher vapor pressure deficit when compared to intense rainfall periods. As a result, vapor pressure deficit physically drives evapotranspiration. This impact becomes drastic on the Small & Short condition (Figure 7a), resulting in higher positive contribution of physical effect to dET in 2007, 2009, and 2010 (Figure 8b). Although year 2008



■:Physical effect, ■:Biological effect, □:Physical & Biological effect

**Figure 8.** (a) Total change in evapotranspiration (dET) and (b) contribution of total physical, biological, and mixed physical and biological effects to dET, categorized into three rainfall regimes, i.e., "Large annual rainfall & Long rainy season," "Small annual rainfall & Long rainy season," and "Small annual rainfall & Short rainy season."

Unsurprisingly, truncating the dET expansion at second order does not perfectly agree with the measured dET partly due to the fact that third- and higher-order terms are ignored and partly due to the fact that the big-leaf model equation does not perfectly describe measured evapotranspiration. Nonetheless, what is important is the improvement gained when expanding dET and retaining second-order terms (including the joint or interactive terms).

#### 4.3. Physical Controls on Evapotranspiration

Second-order perturbation analysis revealed the relative importance of each variable on interannual dET on the monthly time scale. In general, the significance of each variable for dET were in the order vapor pressure deficit rainfall > vapor pressure deficit & aerodynamic conductance > available energy (for vegetation floor evaporation) > available energy, aerodynamic conductance, available energy & aerodynamic conductance (Figure 7a) (mix term effects are represented by "&"). Of cause, a small rainfall event causes a decrease in evapo-

is classified as Small & Long, changes in vapor pressure deficit were not so large to increase evapotranspiration; thus, small changes in rainfall could not alter evapotranspiration as well (Figure 7a). With the other small but negative effects, the physical effect on dET in 2008 was overall negative (Figure 8b).

Large annual rainfall, i.e., in 2006 and 2011, provided a positive effect on evapotranspiration via the canopy interception process (Figure 7a). Here it should be noted that rainfall and vapor pressure deficit are not always coupled. For example, for the same Large & Long condition, vapor pressure deficits in 2006 and 2011 were high and low, respectively (see Figure 7a). These induced positive and negative physical effects on dET in 2006 and 2011, respectively (Figure 8b).

As Wilson and Baldocchi [2000] reported in temperate deciduous forest, the atmospheric

evaporative demand was the primary external driver on evapotranspiration rather than rainfall and/or available energy. Other forests are shown not to be as sensitive to atmospheric water vapor demand alone [e.g., *Kumagai et al.*, 2005; *Stoy et al.*, 2006].

#### 4.4. Biological Controls on Evapotranspiration

The dET was, in general, influenced by biological factors in the order of LAI > stomatal conductance > leaf area index (for vegetation floor evaporation), relative extractable water (for vegetation floor evaporation) > leaf area index & stomatal conductance, leaf area index & relative extractable water (for vegetation floor evaporation) (Figure 7b). A large amount of rainfall induced positive effect on LAI and increased evapotranspiration (Figure 7b). The leaf area index effect was significant during the Large & Long conditions, but insignificant in 2008 in the Small & Long condition.

Interestingly, the stomatal conductance effect decreased evapotranspiration throughout the study period with the exception of 2011 (Figure 7b). This suggested that low humidity and soil moisture deficits in the studied forest lower evapotranspiration. In the case of 2011, the Large & Long condition reduced the atmospheric demand and supplied sufficient ecosystem water resources, resulting in an increase in stomatal conductance (equation (6) and Figure 7b). Note that also, in 2011, both stomatal conductance and LAI effects on dET were positive and thus raised the importance of the biological effect on evapotranspiration (Figure 8b).

Small amounts of rainfall, especially during the Small &Short conditions, represented insufficient ecosystem water resources and reduced  $g_s$  (Figure 7b). This caused negative biological effects on evapotranspiration during such Small & Short conditions (Figure 8b).

The effects of leaf area index (for vegetation floor evaporation) and relative extractable water (for vegetation floor evaporation) on evapotranspiration were not small (Figure 7b). During the Large & Long condition, their negative effects compensated for the positive effect of leaf area index on dET, and thus, the biological effect in 2006 was almost zero (Figure 8b) though individually the changes in leaf area index and relative extractable water.

Fluctuations in rainfall and other climatic variables alter evapotranspiration through multiple physical and biological pathways. For example, leaf phenology including timing of the leaf-on/off cycle is an important factor that determines the seasonal and interannual variations in evapotranspiration, as evidenced by field studies of a tropical monsoonal deciduous forest in northern Thailand [*Yoshifuji et al.*, 2006; *Tanaka et al.*, 2009; *Yoshifuji et al.*, 2014]. The interannual variations in the growing season length spanned several months, even in the 40+ year old forest that did not experience any major disturbance during the study period [*Yoshifuji et al.*, 2006]. As a reference, these variations are considerably larger than interannual variations previously reported in temperate deciduous forests [e.g., *Black et al.*, 2000; *Wilson and Baldocchi*, 2000; *Barr et al.*, 2004; *Hirata et al.*, 2007].

*Yoshifuji et al.* [2014] reported that the canopy is nearly closed (i.e.,  $LAI = ~ 3 m^2 m^{-2}$ ) during the wet season; therefore, the effect of dLAI to dET can be small during this period due to leaf-shading effects [see *Baldocchi et al.*, 1984; *Kelliher et al.*, 1995]. The teak plantation has much larger interannual variations in the leafy period that significantly impact the length of the growing season [*Yoshifuji et al.*, 2006] when compared to temperate and boreal deciduous forests [*Black et al.*, 2000; *Wilson and Baldocchi*, 2000; *Granier et al.*, 2002; *Barr et al.*, 2004], due to the large interannual variation in rainfall. Therefore, the contribution of the interannual variability of leaf-out and leaf-off timing explains much of the interannual variability of dLAI. In this context, estimation of leaf-out, lead-off timing and the length of growing season is needed for computing evapotranspiration in large-scale global climate and vegetation dynamics models [see *Sato et al.*, 2007], especially for the Asian monsoon climate.

#### 4.5. Effects of the Mixed Term on Evapotranspiration

The contributions of the mixed terms to dET with each of the physical and biological factors are shown in Figure 7c. Each of the mixed term did not significantly impact evapotranspiration when compared to its physical or biological counterparts in the second-order expansion. However, since almost all the mixed term effects are negative, their sum total on evapotranspiration for each year, especially in 2011, 2006, 2007, and 2010, was not so small (Figure 7c).

#### 4.6. Summary: Contribution of Physical and Biological Factors and Their Mix Term Controls on dET

It is evident that during the Large & Long and Small & Short conditions, the total derivatives of equation (13), dET (i.e., annual change in evapotranspiration), were determined to be positive and negative, respectively, while during the Small & Long period, no clear sign direction was evident (Figure 8a). Here the negative dET during the Small & Short conditions was mainly due to large negative contributions to dET from biological and mixed physical and biological effects. However, there was no dominant physical or biological factor to explain the positive dET during the Large & Long conditions. Also, there were no dominant factors for dET during the Small & Long periods.

Given the well-known logarithmic decrease in stomatal conductance with increasing vapor pressure deficit [e.g., *Oren et al.*, 1999] and the decrease in stomatal conductance with soil moisture in this studied forest [*Igarashi et al.*, 2015]. It can be surmised that the negative, high stomatal conductance effect in 2007 and 2010 during the Small & Short conditions was caused by a decrease in stomatal conductance with a simultaneous increase in vapor pressure deficit and decrease in soil moisture. This is due to small interseasonal and interannual variations in incident solar radiation and stomatal conductance being correlated with vapor pressure deficit and soil moisture at the site [*Igarashi et al.*, 2015]. The small contribution of dA to dET (Figure 7) also means that incident solar radiation changes do not appreciably impact stomatal conductance at these time scale. The other conditions for the remaining years (in 2006 and 2009) can also be explained by the same two pathways (Figures 7b and 8b), where vapor pressure deficit impacted both physical and biological drivers of evapotranspiration.

Thus, changes in vapor pressure deficit cause opposing changes in physical and biological drivers: the increase in vapor pressure deficit increases evapotranspiration directly as a physical driver and reduces evapotranspiration indirectly via a decrease in stomatal conductance as a biological driver, and *vice versa*. Cancelation was actually demonstrated in the analysis here between physical (i.e., the vapor pressure deficit effect) and biological (i.e., the stomatal conductance effect) effects on ET in 2007 and 2010 during the Small & Short conditions.

## 5. Conclusion

The low-frequency physical and biological controls on evapotranspiration were examined in a teak plantation in northern Thailand subjected to Asia monsoon rainfall. A 7 year record of half-hourly eddy latent heat flux was complemented by a perturbation analysis conducted on the Penman-Monteith equation that "pinpointed" the key physical and biological controls impacting long-term variability in measured evapotranspiration. The novelty of the analysis is the use of higher-order Taylor series expansions that retain higher-order and joint effects among physical and biological drivers of evapotranspiration. The analyzed big-leaf model suggested that higher-order terms were necessary for matching measured and analyzed variation in evapotranspiration, when physical and biological factors exhibit large seasonal and interannual variability. Fluctuations in atmospheric demand (through vapor pressure deficit) were shown to be the primary external physical driver of evapotranspiration perturbations, more significant than available energy or rainfall perturbations. Leaf area index and stomatal conductance were the main biological drivers of canopy transpiration though evapotranspiration, though relative extractable water played an important role in vegetation floor evaporation. Our analysis also suggested how water limitation explains some of the controls on evapotranspiration through the vapor pressure deficit, leaf area index, and stomatal conductance in this ecosystem where evapotranspiration and rainfall are in near balance.

Moving forward into the future, the consensus from global climate models thus far is that this region will experience longer rainy seasons and increased annual rainfall [e.g., *Ueda et al.*, 2006; *Kajikawa et al.*, 2012; *Kitoh et al.*, 2013]. Thus, the dET and its machinery (physical drivers) are likely to be shifting toward the Large & Long conditions observed here (see Figures 7 and 8), which are quite different from the current norms described by 2012. How this evapotranspiration shift might impact or "feed back" to the formation of the Asian monsoon is an interesting question for the future.

## **Appendix A: Estimation of Stomatal Conductance**

The canopy conductance ( $G_c$ ; m s<sup>-1</sup>), representing the efficiency of the latent heat flux from the canopy trees ( $L_{Ec}$ ; W m<sup>-2</sup>), was calculated using the actual data of latent heat flux above the canopy ( $L_E$ ; W m<sup>-2</sup>) and the

inverted Penman-Monteith equation [Dolman et al., 1991] implemented in the simple two-layer big-leaf model [e.g., Kelliher et al., 1995] with some modifications:

$$G_{c}^{-1} = \left[\frac{\Delta}{\gamma} \left(\frac{A_{c}}{L_{Ec}} - 1\right) - 1\right] G_{a}^{-1} + \frac{\rho C_{p} D}{L_{Ec} \gamma},$$
(A1)

in which

$$A_{c} = [1 - \exp(-c_{A}LAI)]A, \tag{A2}$$

where  $\Delta$  is the ratio of charge of saturation water vapor pressure with temperature (Pa K<sup>-1</sup>),  $\gamma$  is the psychometric constant (Pa K<sup>-1</sup>),  $\rho$  is the density of dry air (kg m<sup>-3</sup>),  $C_p$  is the specific heat of air at a constant pressure (J kg<sup>-1</sup> K<sup>-1</sup>), and *D* is the vapor pressure deficit above the canopy (Pa).  $G_a$  is the aerodynamic conductance (m s<sup>-1</sup>) and was estimated as  $u_*^2/U$ , where  $u_*$  is the friction velocity (m s<sup>-1</sup>) and *U* is the vanopy (M m<sup>-2</sup>),  $A_c$  is the *A* absorbed by the canopy (W m<sup>-2</sup>), and  $c_A$  is an extinction coefficient determined to be 0.7 from the net radiation observations above and below the canopy [*Igarashi et al.*, 2015].

Latent heat flux from the canopy trees can be expressed as a contribution from the canopy to the total ecosystem latent heat flux:

$$L_{\rm Ec} = L_{\rm E} - L_{\rm Es},\tag{A3}$$

in which

$$L_{\rm Es} = \alpha_{\rm PT} \frac{\Delta}{\Delta + \gamma} A_{\rm s} (1 + d_m \ln \Theta), \tag{A4}$$

$$A_{\rm s} = A - A_{\rm c},\tag{A5}$$

where  $L_{Es}$  and  $A_s$  are latent heat flux (W m<sup>-2</sup>) and available energy (W m<sup>-2</sup>), respectively, at the vegetation floor.  $a_{PT}$  and  $d_m$  are Priestley-Taylor coefficient and an empirical parameter representing the effect of the soil moisture deficit on the latent heat flux vegetation floor, respectively. The parameter  $a_{PT}$  and  $d_m$  was estimated using nonlinear least squares regression between latent heat flux vegetation floor calculated from equation (A4) and the latent heat flux above the canopy observed under the leafless (i.e., LAI = 0) conditions, where latent heat flux above the canopy can be assumed to be equal to latent heat flux vegetation floor  $a_{PT}$  and  $d_m$  were validated as 0.76 and 0.23, respectively.

We assumed that vapor pressure deficit is well mixed within the canopy and can represent the driving force of evapotranspiration from the entire canopy and that canopy conductance can be regarded as the parallel sum of the individual leaves' stomatal conductances [e.g., *Raupach and Finnigan*, 1988; *Raupach*, 1995], resulting in

$$g_{\rm s} = \frac{G_{\rm c}}{{\sf LAI}},$$
 (A6)

where  $g_s$  is bulk stomatal conductance per unit leaf area (m s<sup>-1</sup>, bulk stomatal conductance: see *Pataki and Oren* [2003]).

Note that for the quality control, we removed the  $g_s$  values obtained on rainy days and when  $D \le 0.6$  kPa [cf., *Ewers and Oren*, 2000] and LAI < 1.0 [cf., *Kelliher et al.*, 1995] from the parameter determination to avoid near singularities and obvious uncertainties in small D measurements.

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