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Interannual Variation in Transpiration Peak of a Hill Evergreen Forest in Northern Thailand in the Late Dry Season: Simulation of Evapotranspiration with a Soil-Plant-Air Continuum Model

Tanaka K.¹, Wakahara T.², Shiraki K.², Yoshifuji N.³ and Suzuki M.⁴

¹*Japan Agency for Marine-Earth Science and Technology*

²*Tokyo University of Agriculture and Technology*

³*Kyoto University*

⁴*The University of Tokyo
Japan*

1. Introduction

Northern Thailand, which experiences rainy and dry seasons under an Asia monsoon climate, is characterized by hilly and mountainous landscapes. The rainfall tends to increase with altitude (Kuraji et al., 2001; Dairaku et al., 2004). Forests in northern Thailand at 1000 m above sea level (a.s.l.) are classified as lower montane rain forests (Santisuk, 1988). These areas receive high amounts of precipitation and provide a stable supply of high-quality water that is crucial for irrigation and drinking water (Bruijnzeel et al., 2011). Generally, water resources or stream flow are estimated by the difference between precipitation and evapotranspiration (i.e., the sum of canopy interception, soil evaporation, and transpiration). Thus, it is important to examine how forests consume rainwater as evapotranspiration, in conjunction with hydrological and meteorological variables. Such modeling is also essential for water resource management.

This study is a continuation of previous studies of transpiration peaks in an evergreen forest in northern Thailand (18°48'N, 98°54'E, 1265–1450 m a.s.l.) in the late dry season (Tanaka et al., 2003, 2004). Tanaka et al. (2003) concluded that transpiration in evergreen forests peaked in the late dry season. They suggested that reduced canopy wetness lowered evaporation; however, stomatal conductance declined only slightly, even under the driest conditions and highest net radiation. These results counter previous reports of an evapotranspiration decline in Thailand's dry season in evergreen forests (Pinker et al., 1980) and other vegetation (Toda et al., 2002). Tanaka et al. (2004) examined the impact of rooting depth and soil hydraulic properties on forest transpiration using a newly developed soil-plant-air (SPAC) multilayer model. They found that a rooting depth of 4–5 m was needed to effectively simulate heat-pulse velocity variations corresponding to dry-season transpiration and annual discharge or stream flow. Moreover, a penetration test showed that the soil

became harder at depths of 4–5 m, supporting the estimated rooting depths. Numerical simulations indicated that a late dry season transpiration peak is theoretically possible on the basis of rooting depth limitations on soil water use because the rooting depth was within the reported maximum for trees. Canadell et al. (2006) reviewed numerous reports of maximum rooting depth and calculated the average and standard deviation as 7.0 ± 1.2 m. Although these studies emphasized the late-dry-season transpiration peak and its mechanism, using combinations of modeling and observation for 2–3 years periods, the impact of interannual variation in rainfall or dry season period length on the peak was not sufficiently examined.

Our objective was to clarify the interannual variation in the late-dry-season transpiration peak in a hill evergreen forest in northern Thailand. A numerical simulation of the seasonal variation in evapotranspiration was performed using a SPAC multilayer model with hydrometeorological variables for the period 1999–2005. The heat pulse velocity corresponding to water use by individual trees was monitored and evapotranspiration was estimated from the water budget (i.e., the difference between rainfall and stream flow) for the 7-year period. These values were compared with the simulated temporal transpiration and annual evapotranspiration.

2. Materials and methods

2.1 Site

Since February 1997, the hydrological and meteorological parameters of a sub-watershed of the Kog-Ma Experimental Watershed have been measured. The sub-watershed has an area of 8.63 ha and is situated 1265–1420 m a.s.l. on Mount Pui (18°48' N, 98°54' E) near Chiang Mai (18°47' N, 98°58' E, 310 m abs.) in northern Thailand. A 50-m meteorological tower was built in the sub-watershed and equipped with instruments for measuring meteorological parameters, such as radiation, wind velocity, and air temperature. Evergreen forest covers the hills of the experimental watershed. *Fagaceae* dominates, with species including *Lithocarpus*, *Quercus*, and *Castanopsis* (Bhumibhamon & Wasuwanich, 1970). Northern Thailand is characterized by hilly and mountainous landscapes, and rainfall tends to increase with altitude (Kuraji, 2001). Dairaku et al. (2004) reported that the larger amount of high-altitude rainfall was due to duration and frequency rather than intensity, which implies that the appearance of clouds is more frequent at higher altitudes. The forest is lower montane rain forest according to the classification of Santisuk (1988). This area receives more rainfall than areas at lower altitudes such as Chiang Mai, with annual precipitation of 1183 mm for 1960–1990 (Thai Meteorological Department, 2011). It is cloudier, particularly in the rainy season, and occasionally experiences fog (N. Tanaka et al., 2011). Average annual rainfall and air temperature for a 7-year period (1999–2005) were 1881 mm and 19.8°C, respectively. The leaf area index (LAI) is approximately 4.5, with a seasonal range from 3.5 to 5.5 (Tanaka et al., 2003). Forest floor soils derive from granitic materials and are classified as Reddish Brown Lateritic (Tangtham, 1974).

Stream flow in the sub-watershed was measured at a concrete weir with a 90° triangular notch. Data were missing for 164 days in 2002, 2004, and 2005, representing 6.41% of the total days in the period 1999–2005. Stream flow on these days was estimated by data assimilation with a river flow model (Fukushima, 1988). The rainfall data and measured

stream flow were used as input and output data, respectively, and the data for 10 (Sep. in 2001–Jun. in 2002), 3 (Oct.–Dec. in 2002), 8 (Jan.–Aug. in 2004), and 12 months (Jan.–Dec. in 2005) around the missing data were used to assimilate the stream flow data.

The heat pulse velocity corresponding to water use by an individual tree was monitored in three tree trunks (No. 1: *Phoebe paniculata*. Nos. 2 and 3: *Lithocarpus elegans*). The height and diameter of the three trees at 1.2 m were 28.0 m and 0.51 m, 23.0 m and 0.29 m, and 15.5 m and 0.20 m, respectively, in 1999–2005 (Tanaka et al., 2003). The observation of heat pulse velocity near a ridge, where no water table seemed to form, showed that the water use (or sap flow) of individual trees had a seasonal trend similar to those of the three trees (Tanaka et al., 2004). These trees belonged to the uppermost or second story. Therefore, the water use by these trees should reflect the transpiration over the forest as a whole (e.g., Kelliher et al., 1992; Tanaka et al., 2003) because transpiration from the upper layers is thought to contribute most of the total transpiration. Here, measured seasonal and interannual changes were used to validate the simulated transpiration.

2.2 A one-dimensional SPAC multilayer model for evapotranspiration

We used a one-dimensional SPAC multilayer model (Tanaka & Hashimoto, 2006) consisting of a soil multilayer model (Kondo & Xu, 1997) and a canopy multilayer model (Tanaka et al., 2003; Fig. 1). The soil multilayer model considers the variation in albedo and evaporation efficiency with changes in soil moisture at the top of the soil column (Kondo & Xu, 1997). The canopy multilayer model (Tanaka et al., 2003) for sensible and latent heat and CO₂ gas exchange consists of a second-order closure model for atmospheric diffusion coupled with a radiation transfer model (Tanaka, 2002), a rainfall interception model (Tanaka, 2002), a Farquhar-type photosynthesis model (Farquhar et al., 1980), and a stomatal conductance model (Ball, 1988). The rainfall interception model assumes that rainfall does not wet the lower sides of leaves with stomata, only the upper sides without stomata, while condensation wets both sides. In the photosynthesis model, the maximum potential rate of electron transport and dark respiration at 25°C (J_{MAX25} and R_{d25}) were scaled to V_{cMAX} at 25°C (V_{cMAX25}); that is, $J_{MAX25} = 2.14 V_{cMAX25}$ after Tanaka et al. (2002) and $R_{d25} = 0.015 V_{cMAX25}$ after Collatz et al. (1991). The assumptions, functions, and procedures in the calculations of the gross CO₂ assimilation rate A , dark respiration R_d , and stomatal conductance g_s were described by Tanaka et al. (2002).

Combined, the two multilayer models by Kondo and Xu (1997) and Tanaka et al. (2003) consider the loss of soil moisture by water uptake (or transpiration) and the effect of soil water content on stomatal closure (Tanaka et al., 2004). The water uptake at depth z was assumed to be proportional to the ratio of the extractable to the entire extractable soilwater content (W_e ; Tanaka et al., 2006). The sum of the water uptakes corresponds to the temporal canopy transpiration. When canopy transpiration can be supplied by the entire extractable soilwater content at 0–1m depths (see Case A in Fig. 1), where the major plant nutrients C, N, P, and K appear to be concentrated (Jackson et al., 2000), W_e is calculated between the depths of 0–1m from which the water uptake is supplied. In the other case, W_e is calculated as the extractable soilwater content between 1m and the maximum rooting depth Z_{ROOT} (here, = 4 m) (see Case B in Fig. 1), and the water uptake is supplied from soil layers at 1m to Z_{ROOT} . The water uptake at depth z , regardless of the vertical root distribution, is expressed as

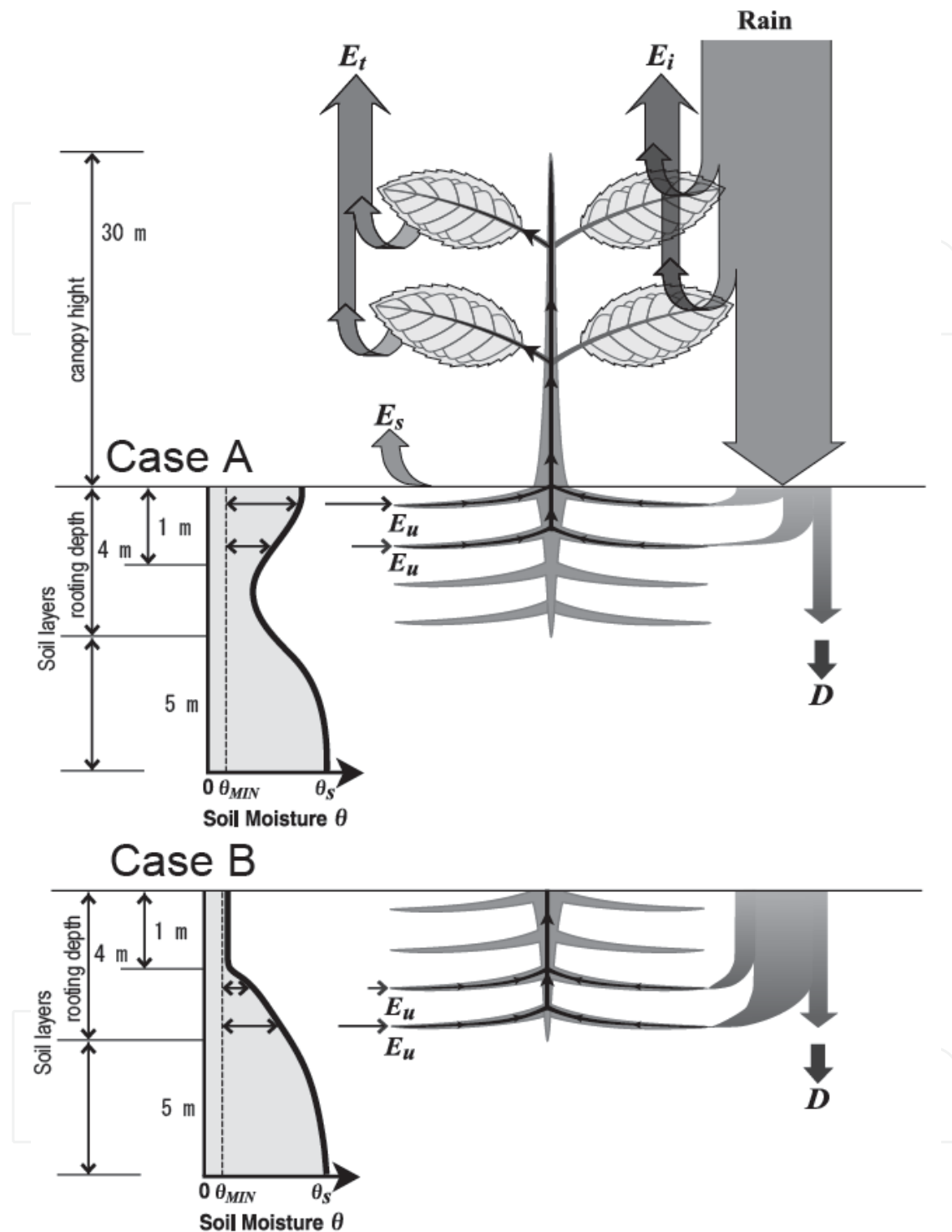


Fig. 1. A one-dimensional soil-plant-air continuum multilayer model for evapotranspiration (i.e., the sum of canopy transpiration E_t , canopy interception E_i , and soil evaporation E_s). Discharge D was calculated as the downward water flux passing through the rooting depth (4 m), and volumetric soil moisture at the bottom of the soil layer (9 m) was set to the saturated volumetric soil moisture θ_s . E_u and θ_{MIN} are the water uptake by roots and soil moisture, respectively, at the lower limit of soil water potential ψ_{LL} (-100 m), where trees cannot take up water. Case A shows water uptake when canopy transpiration can be

supplied by the entire extractable soilwater content at 0–1m depths, while Case B shows water uptake in the other case. E_u at depth z was assumed to be proportional to the ratio of the extractable soil water content (i.e., $\theta(z) - \theta_{MIN}$) to the entire extractable soil water content W_e at 0–1m depths (Case A) or from 1m to the maximum rooting depth Z_{Root} ($= 4$ m) (Case B) (Tanaka et al., 2006). The canopy height was set to 30 m (Tanaka et al., 2003).

$$E_u(z) = E_t \frac{(\theta(z) - \theta_{MIN})}{W_e} \tag{1}$$

Here θ_{MIN} is the volumetric soil water content at which trees cannot take up water. This corresponds to the value at the upper limit of the soil water potential ($\psi_{UL} = -100$ m). W_e in Case A or B is expressed as

$$W_e = \int_0^1 (\theta(z) - \theta_{MIN}) dz, \text{ or } W_e = \int_1^{Z_{ROOT}} (\theta(z) - \theta_{MIN}) dz \tag{2}$$

This assumption of water uptake is simple compared to another frequently used weighting scheme (e.g., Dickinson et al., 1993; Desborough, 1997) based on the assumption that the root length density distribution is proportional to water extraction throughout the profile. Radersma and Ong (2004) did not find a clear relationship between root length density and water extraction. Other researchers have questioned the various proposed relationships between root length density and water uptake (Dardanelli et al., 2004). These findings suggest that the process of water uptake by roots is not entirely clear. Therefore, we used a simpler assumption.

Stomatal conductance was assumed to decrease with the ratio R_{We} of integrated extractable water content W_e from the surface to the rooting depth (i.e., $W_e = \int_0^{Z_{ROOT}} (\theta(z) - \theta_{MIN}) dz$) to the integrated extractable water content at saturation from the surface to the rooting depth $W_{es} = (\theta_s - \theta_{MIN})Z_{ROOT}$, given as

$$g_s = f(R_{We}) \cdot g_{sW} \tag{3}$$

where g_{sW} is the stomatal conductance in well-watered soil and $f(R_{We})$ is a function of the ratio $R_{We} = W_e/W_{es}$ ranging from 0 to 1. The function $f(R_{We})$ was calculated as

$$f(R_{We}) = \min[1.6R_{We} + 0.2; 1] \text{ at } R_{We} > 0 \tag{4}$$

$$f(R_{We}) = 0 \text{ at } R_{We} = 0 \tag{5}$$

Equation (4), including the values of the slope and intercept, is close to the relationship between the extractable water content and stomatal conductance shown by Gollan et al. (1985).

In the canopy multilayer model, the evapotranspiration depends on the LAI, the slope m in Ball’s stomatal conductance model (Ball, 1988), and V_{cmax} at 25°C (V_{cmax25}) in a Farquhar-type photosynthesis model (Farquhar et al., 1980). These parameters are based on the estimated LAI and determined by referring to the measured net photosynthesis rate and stomatal conductance for single leaves (Tanaka et al., 2003). The values were set at 4.5, 10, and 25 $\mu\text{mol m}^{-2} \text{s}^{-1}$, respectively. The vertical profile of the LAI is also a required parameter. It was

assumed to obey a beta distribution, with the greatest leaf area density at 0.7 times the canopy height (B-type canopy; see Figure 6 of Tanaka et al. 2003). Tanaka et al. (2003) investigated the impact of each parameter on evapotranspiration. Kondo and Xu (1997) verified the method by comparing observed and calculated results for four soil textures (i.e., volcanic ash, clay loam, silty sand, and sand). Silty sand was selected as the sub-watershed soil texture, whose observed relationship between volumetric soil water content θ and soil water potential ψ was close to that in the model (Tanaka et al., 2004). The soil and rooting depth were set at 9 and 4 m, respectively (Tanaka et al., 2004). Kondo and Xu (1997) and Tanaka et al. (2003) detailed the other parameter values used in the simulation.

The canopy (height = 30 m) was divided into 50 layers. Each soil layer was 0.1 m thick. The time interval was set at 3 min in the soil multilayer model because of the thin soil layers (0.1 m), but it was set at 15 min in the canopy multilayer model. The model simulated soil evaporation E_s , canopy interception (i.e., evaporation from a wet canopy) E_i , transpiration E_t , discharge D , and soil moisture. The profiles of all the meteorological elements were calculated repeatedly among all sub-models until the differences between the previous and new values of leaf temperature, air temperature, humidity, ambient CO₂ concentration, downward and upward longwave radiation, and water storage on both upper and lower sides of the leaves were within 1% (Tanaka, 2002). The maximum number of repetitions was set at 100 (Tanaka, 2002). Here, D was calculated as the downward water flux passing through the rooting depth (Fig. 1). The initial soil moisture condition at the beginning of 1998 calculated by Tanaka et al. (2004) was used here. The initial soil moisture condition was calculated repeatedly until it corresponded to the value at the end of 2000 in the study by Tanaka et al. (2004). This implies that the total rainfall was used as E_t , E_i , and E_s , and discharged without changing into stored soil water between the beginning and end of the 3-year period (1998–2000). Soil moisture at the bottom of the soil layer (= 9 m; Fig. 1) was set to the saturated soil moisture θ_s . This initial condition did not consider the impact of the decrease in rainfall in the rainy season in 1997 caused by the 1997–1998 El Niño (Wang & Weisberg, 2000). The initial soil moisture appeared greater because of the impact of more rainfall in 2001. Heat pulse velocity was not monitored in the late dry season in 1998. Therefore, simulation results for 1999–2005 are shown here. The initial soil moisture at the beginning of 1999 was calculated using hydrometeorological variables in 1998.

3. Results

Figure 2 shows seasonal and interannual temporal variations in hydrometeorological variables in 1999–2005. The study area has three seasons in terms of air temperature and rainfall changes: a rainy season and early and late dry seasons (Tanaka et al., 2003). The light gray, gray, and black bars in Fig. 2a indicate the point 30 days before the day when the rainfall amount exceeded 150 mm (i.e., the wet period; WP) (Fig. 2b), the days excluding those in the WP whose following 5 days had mean air temperatures below 21°C (i.e., the cool dry period; CDP), and the days excluding those in the WP whose following 5 days had average values of air temperature over 21°C (i.e., hot dry period; HDP), respectively. The horizontal bars in Fig. 2b show the points at which the 90 previous days had less than 50 mm of total rainfall (i.e., a drought condition; DC). The CDP was concentrated in the early dry season, while the HDP was concentrated in the late dry season. The HDP occasionally appeared in the early dry seasons, in much shorter periods than in the late dry seasons. The

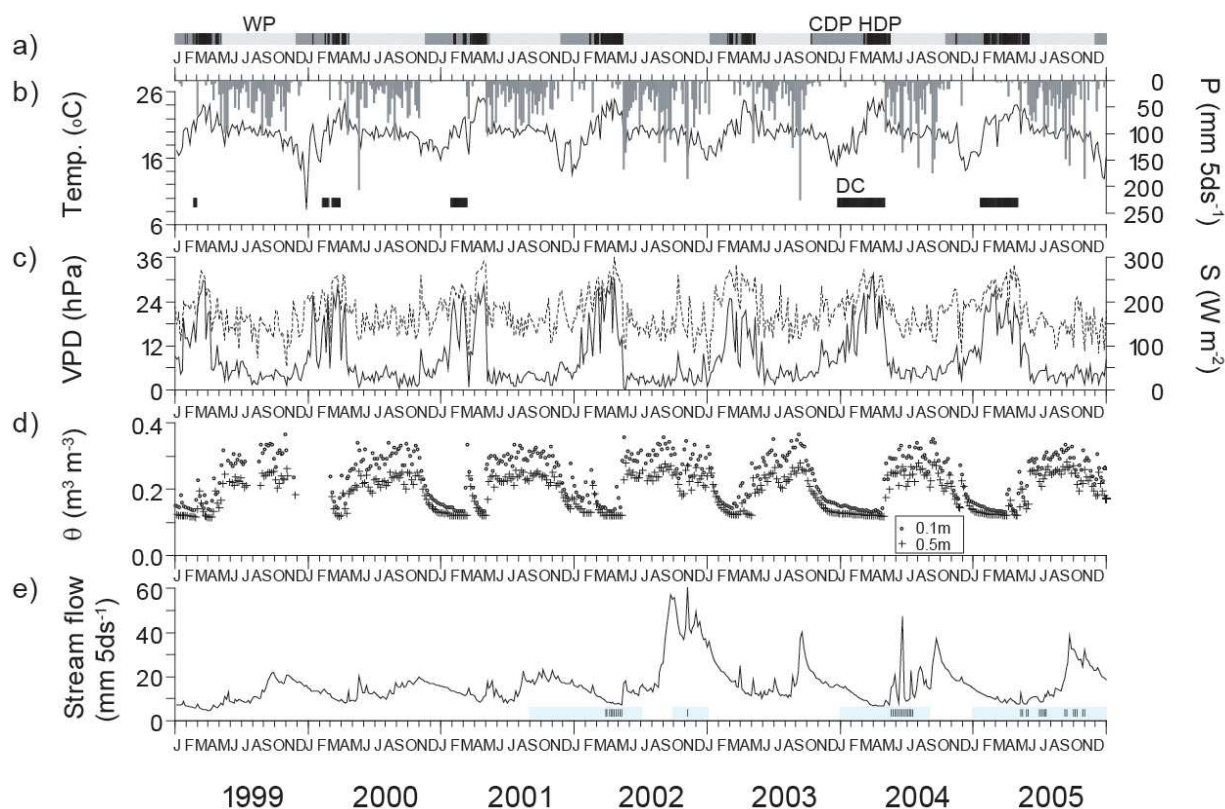


Fig. 2. (a) Hydrometeorological periods in 1999–2005. Light gray, gray, and black bars show the days on which the amount of rainfall 30 days previously was over 150 mm (i.e., the wet period; WP), the days excluding those in the WP for which the following 5 days had average air temperature values below 21°C (i.e., the cool dry period; CDP), and the days excluding those in the WP for which the following 5 days had average air temperature values over 21°C (i.e., the hot dry period; HDP), respectively. Seasonal changes in air temperature and rainfall P (b), vapor pressure deficit and downward solar radiation S (c), volumetric soil moisture θ at depths of 0.1 and 0.5 m (d), and stream flow (e) are shown for 1999–2005. The horizontal bars in (b) show the days for which the previous 90 days had less than 50 mm of total rainfall (i.e., drought conditions; DC). The horizontal bars in (e) show the days without measurements of stream flow. During the days with missing data, stream flows were estimated by data assimilation using a river flow model (Fukushima, 1988). The shaded areas in (e) correspond to the data assimilation periods.

lengths of both the CDP and HDP changed interannually, being longest in 2004–2005 and shortest in 2002–2003. In the longest dry season, the DC period was longest. The DC period did not appear in the 2001–2002 and 2002–2003 dry seasons. The annual amount of rainfall was smallest in 2003 (= 1504 mm) and largest in 2002 (= 2458 mm). The vapor pressure deficit (VPD) and downward solar radiation S peaked in the HDP (Fig. 2c), indicating the strongest atmospheric evaporative demand. These quantities were lower in the WP. The VPD peak was the lowest in the HDP of 2003 because the shortest dry season was in 2002–2003. The solar elevation at noon and the day length were higher and longer, respectively, in the rainy season, but the frequent appearance of clouds modified the less intense solar radiation. The volumetric soil moisture values at 0.1 and 0.5 m were also lowest in the late dry season (Fig. 2d). The duration of the driest conditions increased with the DC period,

particularly in 2004. The stream flow was never interrupted, even in the late dry season (Fig. 2e). This implies that the deeper soil portion was still moist even though the soil surface layer at depths of 0.1–0.5 m was dry. The peak stream flow appeared in the late rainy season or at the end of WP (e.g., September–November). The peak value was the largest in 2002 due to the rainfall amount. The horizontal bars in Fig. 2e show the days without stream flow measurements. For days with missing data, stream flows were complemented by data assimilation with a model for river flow (Fukushima, 1988). The shaded areas in Fig. 2e correspond to the data assimilation periods. The total measured and modeled stream flows were 3025 mm and 3040 mm, respectively, for 835 days during the periods with measured stream flow. The total amount of complemented stream flow was 498 mm for the 164 days with missing data, and the estimated error was extremely small (several millimeters).

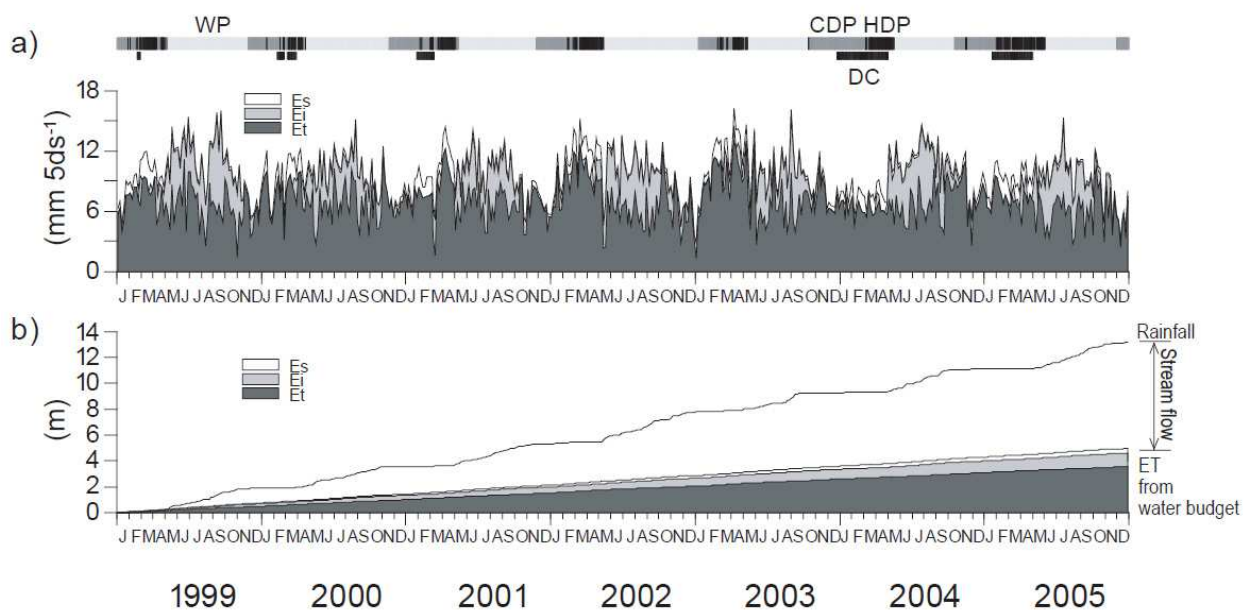


Fig. 3. (a) Simulation of the seasonal variation in evapotranspiration (i.e., the sum of soil evaporation E_s , canopy interception or evaporation from the wet canopy E_i , and transpiration E_t) in 1999–2005. (b) The cumulative result. Light gray (wet period; WP), gray (cool dry period; CDP), and black bars (hot dry period; HDP) in (a) correspond to those in Fig. 2a. The shorter black bar (drought condition; DC) also corresponds to that in Fig. 2b.

Figure 3a shows the simulation results of evapotranspiration ET (i.e., the sum of soil evaporation E_s , canopy interception E_i , and transpiration E_t) in 1999–2005. The E_i appeared to increase with rainfall in the rainy seasons. In particular, the ratio of E_i to ET was close to half of the ET in 2002, the year with the largest rainfall. The E_t peaked in the HDP in 1999–2003, particularly in 2002 and 2003, but the E_t in the HDP almost equaled that in the WP in 2004 and 2005. The simulated ET maintained large values during the HDP and WP in 1993–2003, although the ET was smaller in the HDP in 2004 and 2005. The simulated ET values were smaller in the CDP due to weaker atmospheric conditions (lower temperature, lower VPD, and less intensive solar radiation due to the decline in solar elevation), even though the soil was wetter. The simulated E_s was small under humid conditions within a canopy due to both the lower VPD and the wetter soil, while E_s increased in the HDP and/or the

DC. Figure 3b shows the cumulative results from Fig. 3a. The annual amounts of simulated ET , E_s , E_i , and E_t were 707, 49, 151, and 507 mm yr⁻¹, respectively, for the 7-year period. The closed canopy reduced E_s , and almost all E_i disappeared outside the rainy season. Thus, the percentages of E_s , E_i , and E_t within ET were 7, 21, and 72% in the 7-year period, respectively. Assuming a negligible difference in the storage of soil moisture between the beginning of 1999 and the end of 2005, the annual ET was 694 mm yr⁻¹, the difference between rainfall (1881 mm yr⁻¹) and stream flow (1187 mm yr⁻¹). The ET value from the water budget was very close to the simulated ET . The error in stream flow estimated by data assimilation appeared to be negligible (several millimeters over 7 years), as the above-mentioned.

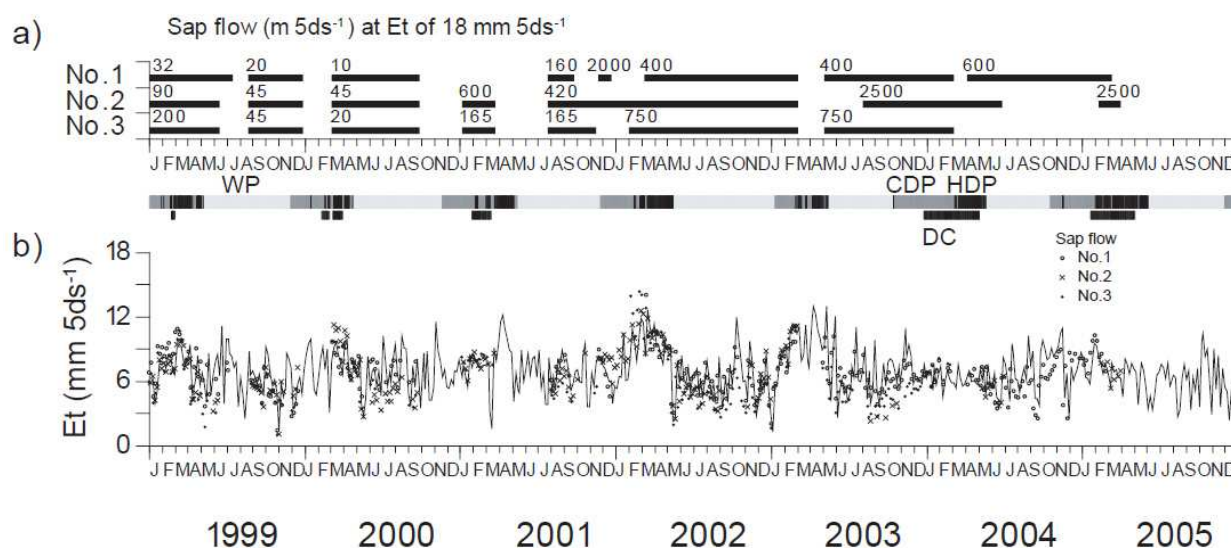


Fig. 4. (a) Continuous period of measurement of heat pulse velocity of three trees (number 1, *Phoebe paniculata*; numbers 2 and 3, *Lithocarpus elegans*) using the same sensor. The values in the figure correspond to the heat pulse velocity (m 5ds⁻¹) at 18 mm per 5 days of transpiration. (b) Simulation of the seasonal variation in transpiration E_t and heat pulse velocity of the three trees in 1999–2005. Light gray (wet period; WP), gray (cool dry period; CDP), and black bars (hot dry period; HDP) in (b) correspond to those in Fig. 2a. The shorter black bar (drought condition; DC) also corresponds to that in Fig. 2b.

Figure 4 shows seasonal and interannual variations in the simulated E_t and heat pulse velocity of the three trees (number 1, *Phoebe paniculata*; numbers 2 and 3, *Lithocarpus elegans*). The horizontal bars in Fig. 4a indicate the duration of the measurement of heat pulse velocity for each tree with the same sensor and position. The values above the bars correspond to the heat pulse velocity (m s⁻¹) at 18 mm per 5 days of transpiration. These changes depended on where and how deep the probe and sensor were placed in the trunks (Phillips et al., 1996; Takizawa et al., 1996; James et al., 2002). Probes and sensors were inserted in the trunks. The simulated E_t captured the variation in heat pulse velocity. Both E_t and the heat pulse velocities exhibited late-dry-season transpiration peaks in 1999, 2000, and particularly 2002. However, the peaks were smaller in 2003 and 2004 and almost the same as E_t in the rainy season.

Figure 5 shows the simulated mean±1 standard deviation values of daily transpiration during the HDP (see Fig. 4b) in the dry season from 1999–2000 to 2004–2005. The gray,

black, and narrower black bars denote the number of days in the CDP, HDP, and DC, respectively, in the dry season from 1999–2000 to 2004–2005. Daily transpiration in the HDP was the smallest and second smallest in the 2003–2004 and 2004–2005 periods, respectively. These periods had longer dry seasons; in particular, the longest CDP and HDP were in 2004–2005. There were also 54 and 78 days of DC in the 2003–2004 and 2004–2005 periods, respectively. The longer period of atmospheric strong evaporative demand and insufficient soil moisture decreased the transpiration in the HDP. The transpiration in the HDP in 2004–2005 was larger than that in 2003–2004, although both the HDP and DC were longer in 2004–2005 than in 2003–2004. In the shortest dry season without DC (for which the sum of CDP and HDP was 118 days; 2002–2003), the daily transpiration was the largest.

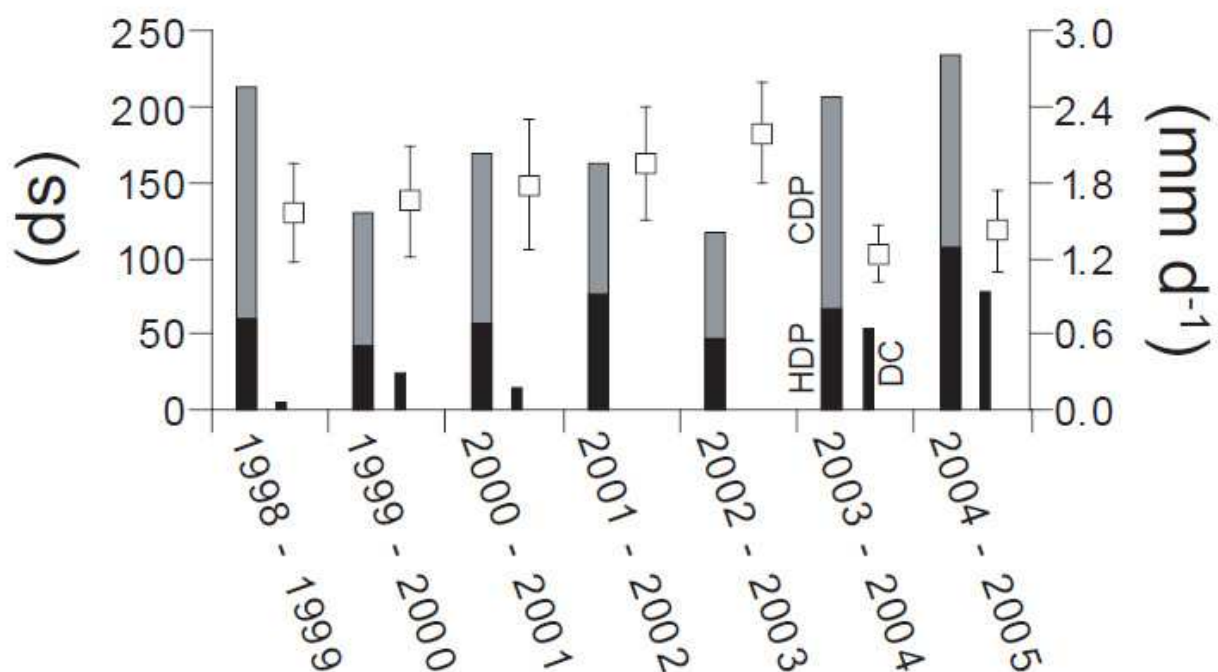


Fig. 5. Simulated mean \pm 1 standard deviation values of daily transpiration E_t during the hot dry period (HDP, see Fig. 4b) in the dry season from 1999–2000 to 2004–2005. Gray, black, and narrow black bars show the number of days of cool dry period (CDP), HDP, and drought condition (DC), respectively, in the dry season from 1999–2000 to 2004–2005.

4. Discussion

The SPAC multilayer model (Tanaka & Hashimoto, 2006) was used to simulate the interannual variation in ET at a hill evergreen forest in northern Thailand between 1999 and 2005. The simulated annual ET was close to the difference between rainfall and stream flow (i.e., the ET from the water budget) during the 7-year period. The simulated transpiration E_t captured the measured heat pulse velocity corresponding to water use by an individual tree, particularly the decrease in the late-dry-season E_t in 2004 and 2005. The assumption that the decrease in extractable soil moisture had reduced impact on stomatal closure (i.e., equation 4) thus appeared to be appropriate for the estimation of transpiration at the forest canopy

level. To confirm this assumption, the model should be applied to data for different vegetation under drought conditions. Tanaka et al. (2009) found that this assumption was reasonable for modeling deciduous teak plantations with shallower rooting depth under drought conditions in northern Thailand. Although the simulated canopy interception E_i was not validated here, Tanaka et al. (2003) simulated the E_i of the forest in 1998–1999 using a canopy multilayer model (Tanaka, 2002) but no soil multilayer model (Kondo & Xu, 1997). They compared the simulated E_i with that estimated from the difference between rainfall and the sum of throughfall and stemflow and found that the annual values were close. Although the simulated soil evaporation E_s was also not validated, E_s likely occupied a small portion of the ET due to the decline in downward solar radiation and wind velocity because of the closed canopy, as in the simulation (Fig. 3; the ratio of E_i to ET is annually 7%). Thus, this study is the first to reveal features of the interannual variation in ET (i.e., the sum of E_s , E_i , and E_t).

The simulated annual ET ($= 707 \text{ mm yr}^{-1}$) was smaller than that estimated from the water budget ($= 694 \text{ mm yr}^{-1}$) by 13 mm. The smaller LAI decreased E_t and E_i and increased E_s (Tanaka et al., 2009). Smaller values of V_{cMAX25} or shallower rooting decreased E_t and changed little both E_i and E_s within the model (Tanaka et al., 2004, 2009). These impacts on ET were larger in the late dry season and the HDP, when atmospheric evaporation demand was stronger than in the rainy season because of the higher temperatures and VPD and the intensive solar radiation (Tanaka et al., 2003, 2004). Therefore smaller values of V_{cMAX25} or shallower rooting depths are required for simulated ET to become closer to that estimated from measurements. Then the seasonal and interannual variation in ET would show slightly smaller E_t values in the HDP, and neither E_s nor E_i would change significantly. In tuning with smaller LAI values, the decrease in E_i and E_t and increase in E_s must be considered.

The simulated E_t and monitored heat pulse velocity showed that the late-dry season E_t peak was lower in 2004 and 2005 (Fig. 4). The rooting depth was set at 4 m. This is deeper than the 1 m depth often used in land surface models within general circulation models (Kleidon & Heimann, 1998). Tanaka et al. (2004) noted that, during dry periods, rainwater remaining from the previous rainy season may be sufficient for trees with greater water use capacities. The longest DC periods, however, as well as the longest dry season lengths in 2003–2004 and 2004–2005, largely limited E_t in the late dry season beyond this larger water use capacity. Kume et al. (2007) compared heat pulse velocities (or sap flow) among two large trees corresponding to Nos. 1 and 2 in this study (Fig. 4b) and two smaller trees (4.8 m and 1.4 m height) in the study forest in 2003 and 2004. They found that the reduced impact of soil drought on sap flow was clearer in the smaller trees in the late dry season in 2004, likely due to their shallower rooting depths. They suggested that the larger trees might avoid water uptake limitations with their deeper roots. Furthermore, transpiration over the whole forest canopy could also have been limited in the late dry season in 2004 and 2005 (Figs. 4b and 5).

The annual ET ($= 694 \text{ mm}$) was small compared with values reported for other tropical and sub-tropical forests (e.g., Doley 1981); in the latter, values often exceed 1000 mm, with maxima of 1750 mm, and the ratio of annual discharge to annual rainfall exceeds that of the annual ET . Air temperature decreases with altitude, while rainfall tends to increase with altitude (Kuraji, 2001; Dairaku et al., 2004) in northern Thailand, and the downward solar

radiation decreases due to the frequent appearance of clouds during the rainy season (Fig. 2c). Under the weaker evaporative demand in the rainy season, most rainwater probably infiltrates the soil layers with relatively smaller ET values. This rainwater is likely used by evergreen trees with deeper roots, even in the following dry season. Such trees can continuously transpire, control their leaf temperature, and assimilate carbon, although late-dry-season transpiration peaks likely decreased in 2004 and 2005. Hence the trees can maintain leaves all year round. The population of deciduous trees increases as altitude decreases below 1000 m a.s.l. in northern Thailand (Santisuk, 1998). Tanaka et al. (2009) numerically simulated canopy net assimilation A_n , ET , and soil moisture in a deciduous teak plantation with shallow rooting depth (< 1 m) in a dry tropical climate in northern Thailand ($18^{\circ}25'N$, $99^{\circ}43'$; 380 m a.s.l.) using the SPAC multilayer model. That site had annual rainfall of 1361 mm for 2001–2008 and higher annual temperature of $25.4^{\circ}C$ (K. Tanaka et al., 2011). The first experiment in that study involved seasonally varying LAI estimates based on time-series of radiative transmittance through the canopy, and the second experiment applied an annually constant LAI. The first simulation captured the measured seasonal changes in soil surface moisture; the simulated transpiration agreed with seasonal changes in heat pulse velocity, corresponding to the water use of individual trees. The simulated A_n almost always became positive during leafy seasons. The simulated annual ET was ~ 900 – 1200 mm. However, in the second simulation in the dry season, A_n and E_t became negative and small, respectively, because the decline in stomatal conductance due to severe soil drought limited the assimilation. The simultaneous increase in leaf temperature increased dark respiration. These experiments revealed that leaflessness in the dry season is reasonable for carbon gain, and trees cannot maintain leaves year round at the site. Therefore, it may be more difficult (easier) for trees to maintain leaves in the late dry season as altitudes decrease (increase) in northern Thailand.

5. Conclusion

The E_t simulated with the SPAC multilayer model and heat pulse velocities indicated that the late-dry-season transpiration peak weakened in 2004 and 2005, even with an assumed rooting depth of 4 m. The 2003–2004 and 2004–2005 dry seasons were relatively longer, and they had the second longest (= 67 days) and longest (= 108 days) DC days. The soil moisture likely became insufficient beyond the rooting depth limitations on soil water use because of the duration of drought conditions along with the stronger atmospheric evaporative demand.

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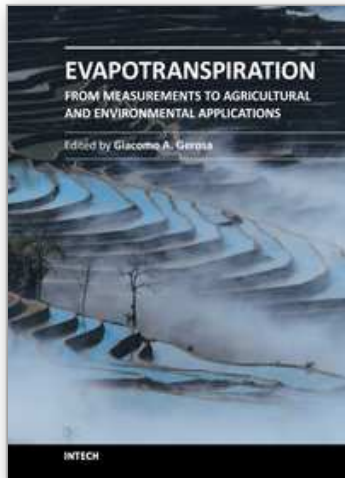
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