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Short communication

Difference in the Priestley–Taylor coefficients at two different heights of a tall micrometeorological tower

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ABSTRACT

The eddy covariance method is highly recommended for measuring in situ water vapor flux. In this study, eddy covariance systems were installed at two heights (30 m and 100 m) on a micrometeorological tower located in a tropical mixed forest. We have identified a difference in the Priestley–Taylor coefficients (α), calculated at 30 m and 100 m, with vertical variation of water vapor flux, and this height-dependent difference is smaller when the vertical difference in wind speeds is higher. It will be useful to consider this result in the analysis of measurement data for model validation.

1. Introduction

Evapotranspiration (*ET*) is a critical phase in the hydrological cycle of terrestrial ecosystems. The energy required for the water vapor flux, i.e., latent heat exchange (*LE*), is also considered a significant heat source that affects the interactions between the atmosphere and the biosphere and changes which can dampen or amplify atmospheric circulation (Henderson-Sellers et al., 1993; Bonan, 2008). Flux measurement sites (e.g., FLUXNET, AMERIFLUX, EUROFLUX, and ASIAFLUX) have been established accordingly, to coordinate the global analysis of exchanges of water vapor and energy between diverse ecosystems, in order to understand the environmental, biological, and climatological controls of net surface exchange between the biosphere and the atmosphere (Baldocchi et al., 2001). In addition, global flux data could be compared with the simulation results of climate and terrestrial hydrological models to develop biophysical schemes and to enable parameterization which is related to the processes of energy exchange, CO_2 uptake, and evapotranspiration in lower boundary layer on land surface (Liu et al., 1999; Wang et al., 2006).

The eddy covariance system is highly recommended for measurement of water vapor flux because it is the most direct micrometeorological technique for measuring in situ heat energy exchange (Baldocchi et al., 2001; Meyers and Baldocchi, 2005). However, it is evident that the energy fluxes measured by the eddy covariance method do not usually indicate a closure of the surface energy balance (Foken and Oncley, 1995; Wohlfahrt and Widmoser, 2013). For example, two types of available energy (A), which are represented as the sum of turbulent fluxes ($LE + H$; where, H is sensible heat flux) at atmospheric height level and the rest of R_n minus G ($R_n - G$; where, R_n is net radiation, and G is the ground heat flux) at ground level, might be not commonly exactly same (Foken, 2008). In addition, A of turbulent fluxes is possibly different depending on measurement heights (Lindroth et al., 2010). This is true even for measurements performed over flat, homogeneous surfaces and over short vegetation, which are both considered ideal conditions for eddy covariance measurements (Foken and Oncley, 1995; Kanda et al., 2004).

One of the main reasons for the above observations could be that the eddy covariance system measures turbulent fluxes contributed by small eddies in the lower boundary layer and does not measure fluxes contributed by larger eddies (Finnigan et al., 2003; Kanda et al., 2004). Such measurement characteristics of aerodynamic turbulence can cause height dependency in the energy

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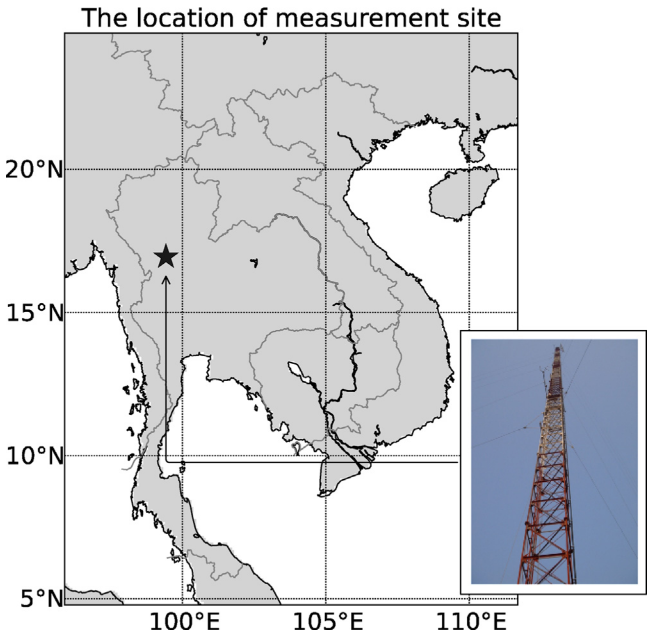

Fig. 1. The location of eddy covariance measurement flux tower in northwestern Thailand.

balance (e.g., Lee and Black, 1993; Lee, 1998; Lindroth et al., 2010). For example, there are a sensible heat flux divergence of 0.5–0.7 (W m^{-2}) between data of eddy covariance systems installed at the 35 m and 70 m heights (Lindroth et al., 2010). Indeed, aerodynamic motion characteristics near the ground, such as wind speed (u) and friction velocity (u^*), are commonly used to identify systematic patterns of the energy balance closure problem in flux measurement studies (Twine et al., 2000; Wilson et al., 2002; Kosugi et al., 2007).

In this study, we have identified the aerodynamic characteristics (e.g., the vertical differences in u and u^*) at two different heights, where height-dependent difference of water vapor flux is observed. The Priestley–Taylor coefficient (α) has been used to characterize water vapor flux measured by eddy covariance system. The measurement data used in this study have been obtained from a micrometeorological tower, which has eddy covariance systems installed at two heights (30 m and 100 m), and is located over mixed vegetation cover in northwestern Thailand.

2. Methodology

2.1. Eddy covariance measurement site

The experimental site is located at 16°56′23.64″ N, 99°25′47.64″ E in northwestern Thailand (Kim et al., 2003; see Fig. 1). The fairly flat fetch area is more than 10 km long and located at an elevation of 130 m above sea level. The climate of the experimental area is divided into rainy (June–November) and dry (September and January–May) seasons, because the area is a part of the Southeast Asia Monsoon region. The vegetation consists of tall deciduous trees (70%), which are 10–15 m in height and have a leaf area index between 3 and 5, and agricultural areas (30%) of rice paddy, corn, and tobacco.

Measurements were taken by eddy covariance systems installed at two heights (30 m and 100 m) of the tower using LI-7500, an open-path infrared gas analyzer (LI-COR, Lincoln, NE, USA), and CSAT-3, a three-dimensional sonic anemometer (Campbell Scientific Inc., Logan, UT, USA). These fast-response instruments were operated at a sampling rate of 10 Hz. The air temperature and relative humidity were observed using a temperature and humidity probe (HMP45C; Vaisala, Helsinki, Finland) installed at the same heights (30 m and 100 m) in the eddy covariance systems. The raw data from the fast-response instruments were processed in 1-h time steps. Flux data processes were determined using a moving average with a time constant of 200 s. The wind field (u, v, w) (m s^{-1}) coordinates were rotated so that the mean v and w values were zero over 10-min periods (McMillen, 1988); the CSAT-3 (a three-dimensional sonic anemometer) output already incorporates this transform. To correct the frequency-response losses associated with fast-response sensors, we used Moore's (1986) method. We also the density fluctuation correction method (Webb et al., 1980) using the sonic temperature corrected for the air pressure and water vapor concentration. We used only daytime flux measurements collected between January 2004 and December 2009, in order to avoid the condition of weaker turbulent mixing, which frequently occurs at night.

2.2. Priestley–Taylor coefficient

The Priestley–Taylor coefficient (α) is indicated as the ratio of evapotranspiration (ET) to equilibrium evapotranspiration (ET_{equ}) (Priestley and Taylor, 1972). ET_{equ} indicates the lower limit of evaporation from a wet surface under the condition of minimal advection (Monteith and Unsworth, 2008). ET_{equ} is commonly lower than the actual ET in a vegetated land due to the resistances between the land surface and the atmosphere (Baldocchi, 1994; Cho et al., 2012). The parameter α is widely used to evaluate the water vapor flux rate to enable a comparison of ET or LE data obtained under different meteorological conditions (Komatsu, 2005; Cho et al., 2012). Further, α can be represented in terms of the ratio of H to LE (called Bowen ratio; β) (Holtslag and Van Ulden, 1983):

$$\alpha = \frac{1 + (1/\varepsilon)}{1 + \beta} \tag{1}$$

where ε (=Δ/γ) is a functional parameter that depends on the air temperature and water vapor. Δ is the slope of the saturation vapor pressure curve with respect to temperature at a specified temperature (Pa °C^{-1}). γ is the psychrometer constant (Pa °C^{-1}). In this study, we estimated hourly daytime α using air temperature and β which calculated by LE and H from eddy covariance measurement. Systematic underestimation has been reported in both LE and H measurements collected by the eddy covariance method during the daytime (e.g., Blanken et al., 1998; Mahrt, 1998), which is characterized by unstable conditions and well-developed large eddies. Nevertheless, the values of β would be less affected due to the systematical underestimation of both H and LE by similar relative fractions at daytime (Turnipseed et al., 2002).

3. Results

Fig. 2(a) shows the comparison of the Priestley–Taylor coefficient, α, obtained using daytime data at 30 m and 100 m. It was found that α was higher at 100 m (α_{100}) than at 30 m (α_{30}), regardless of rainy and dry seasons. The difference between the coefficients obtained at 100 m and 30 m ($\alpha_{100} - \alpha_{30}$) does not indicate any significant patterns dependent on the vertical differences of u^* and air temperature (not shown here). When the vertical difference in wind speed ($u_{100} - u_{30}$) is small, the plots of $\alpha_{100} - \alpha_{30}$ are extensively distributed, particularly as larger positive values (Fig. 3). However, it was observed that the difference between the Priestly–Taylor coefficients at different heights (α_{100} and α_{30}) decreases with an increase in the vertical variation of wind speed ($u_{100} - u_{30}$). This characteristic is clearer in rainy season than in dry