



Wet canopy evaporation from a Puerto Rican lower montane rain forest: The importance of realistically estimated aerodynamic conductance

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SUMMARY

Rainfall interception (I) was measured in 20 m tall Puerto Rican tropical forest with complex topography for a 1-year period using totalizing throughfall (TF) and stemflow (SF) gauges that were measured every 2–3 days. Measured values were then compared to evaporation under saturated canopy conditions (E) determined with the Penman–Monteith (P–M) equation, using (i) measured (eddy covariance) and (ii) calculated (as a function of forest height and wind speed) values for the aerodynamic conductance to momentum flux ($g_{a,M}$). E was also derived using the energy balance equation and the sensible heat flux measured by a sonic anemometer (H_s). I per sampling occasion was strongly correlated with rainfall (P): $I = 0.21P + 0.60$ (mm), $r^2 = 0.82$, $n = 121$. Values for canopy storage capacity ($S = 0.37$ mm) and the average relative evaporation rate ($E/R = 0.20$) were derived from data for single events ($n = 51$). Application of the Gash analytical interception model to 70 multiple-storm sampling events using the above values for S and E/R gave excellent agreement with measured I . For $E/R = 0.20$ and an average rainfall intensity (R) of 3.16 mm h^{-1} , the TF -based E was 0.63 mm h^{-1} , about four times the value derived with the P–M equation using a conventionally calculated $g_{a,M}$ (0.16 mm h^{-1}). Estimating $g_{a,M}$ using wind data from a nearby but more exposed site yielded a value of E (0.40 mm h^{-1}) that was much closer to the observed rate, whereas E derived using the energy balance equation and H_s was very low (0.13 mm h^{-1}), presumably because H_s was underestimated due to the use of too short a flux-averaging period (5-min). The best agreement with the observed E was obtained when using the measured $g_{a,M}$ in the P–M equation (0.58 mm h^{-1}). The present results show that in areas with complex topography, $g_{a,M}$, and consequently E , can be strongly underestimated when calculated using equations that were derived originally for use in flat terrain; hence, direct measurement of $g_{a,M}$ using eddy covariance is recommended. The currently measured $g_{a,M}$ (0.31 m s^{-1}) was at least several times, and up to one order of magnitude higher than values reported for forests in areas with flat or gentle topography (0.03 – 0.08 m s^{-1} , at wind speeds of about 1 m s^{-1}). The importance of $g_{a,M}$ at the study site suggests a negative, downward, sensible heat flux sustains the observed high evaporation rates during rainfall. More work is needed to better quantify H_s during rainfall in tropical forests with complex topography.

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1. Introduction

Based on a preliminary comparison of the high rainfall interception losses reported for a temperate forest (Wales, UK) subjected to oceanic influences, with the much lower losses from a tropical rain forest situated in the middle of a continent (Manaus, Brazil), Shuttleworth (1989) hypothesized that tropical deforestation at continental-edge and island locations was likely to have a greater effect on streamflow than would deforestation at mid-continental

sites. Since then, a substantial number of interception studies conducted in tropical forests subject to 'maritime' climatic conditions have yielded interception values that were generally much higher (at 18–50% of incident rainfall; summarized by Schellekens et al. (2000), Roberts et al. (2005), and Wallace and McJannet (2008)) than those typically observed at mid-continental locations (9–17%; see discussion in Roberts et al., 2005; cf. Cuartas et al., 2007). Similarly, average wet-canopy evaporation rates (E) inferred from throughfall (TF) measurements under maritime tropical conditions (0.4 – 1.0 mm h^{-1}) are typically 4–5 times those calculated with the Penman–Monteith (P–M) equation (Schellekens et al., 1999; Roberts et al., 2005; Wallace and McJannet, 2008). No such discrepancy appears to have been encountered under continental tropical conditions (Lloyd et al., 1988; Cuartas et al., 2007).

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Several possible reasons for the high interception losses and for the discrepancy between *TF*-based and *P*-*M*-based rates of *E* under maritime tropical conditions have been suggested in the literature, including:

- (i) underestimation of *TF* due to inadequate sampling design (Dykes, 1997; Schellekens et al., 1999, 2000; cf. Lloyd and Marques, 1988; Ziegler et al., 2009; Zimmermann et al., 2010);
- (ii) advection of sensible heat from the nearby ocean (Dykes, 1997; Waterloo et al., 1999; Schellekens et al., 1999, 2000; Wallace and McJannet, 2008; cf. Shuttleworth and Calder, 1979; Shuttleworth, 1989; Roberts et al., 2005); and
- (iii) underestimation of the aerodynamic conductance (Schellekens et al., 1999, 2000; cf. Gash et al., 1980).

Holwerda et al. (2006) investigated the influence of underestimated *TF* by quantifying the variability of *TF* and the standard errors of the means associated with the deployment of 60 fixed, 30 fixed, and 30 roving gauges in the same Puerto Rican forest for which Scatena (1990) and Schellekens et al. (1999, 2000) had previously obtained very high interception losses. Holwerda et al. (2006) reported total *TF* values as measured with the fixed and roving sampling designs to be nearly equal, although the variability of the estimate was about half as high in the case of roving gauges. Furthermore, the sampling methods used by several *TF* studies reporting high interception losses under maritime tropical conditions, e.g. Dykes (1997) (17 roving gauges with 314 cm² orifice each), Waterloo et al. (1999) (20 roving gauges, 100 cm² orifice each), and Wallace and McJannet (2008) (4–6 large troughs with dynamically calibrated tipping buckets) all seem to have been adequate. As such, there is no reason to believe that an underestimation of *TF* caused the high interception losses observed at these maritime sites.

Roberts et al. (2005) discussed the role of advected energy and estimated that for an advection-driven evaporation rate of 100 W m⁻² (equivalent to ~0.15 mm h⁻¹), a horizontal temperature gradient of 1 K per 100 m would be required. Hence, to explain a typically observed difference of 0.5 mm h⁻¹ between *TF*-based and calculated (*P*-*M*-based) rates of *E* for maritime tropical sites (see Table 4 below), a (persistent) horizontal temperature gradient of 3–4 K per 100 m would be necessary. This is considered unrealistic given the average distance from the coast of these sites is approximately 17 km.

The alternative explanation for greatly enhanced *E*, i.e. underestimation of the aerodynamic conductance for momentum transfer ($g_{a,M}$), has received comparatively little attention under tropical and temperate conditions alike (e.g. Gash et al., 1980; Schellekens et al., 1999, 2000). With very few exceptions (e.g. Gash et al., 1999; Van der Tol et al., 2003; Cuartas et al., 2007; Herbst et al., 2008), rainfall interception modeling studies have typically estimated *E* with the *P*-*M* equation using the equation for momentum transfer of Thom (1975) to calculate $g_{a,M}$ (see Muzylo et al. (2009) for an overview of interception modeling studies world-wide). However, strictly speaking, the Thom equation is valid over uniform, extensive, and flat surfaces only (Monteith and Unsworth, 2008) and its application to forests located in complex terrain, where patterns of wind flow and turbulence are affected by topography, is problematic (cf. Chen and Schwerdtfeger, 1989; Raupach and Finnigan, 1997; see also Section 2).

This paper investigates whether underestimation of $g_{a,M}$ through the conventional use of the Thom formula can explain the major discrepancy between observed (*TF*-based) and calculated rates of wet-canopy evaporation for the Puerto Rican lower montane rain forest studied previously by Scatena (1990) and Schellekens et al. (1999). Micrometeorological and net precipitation data

were collected between November 2000 and October 2001. *E* rates were calculated using the *P*-*M* equation with $g_{a,M}$ either estimated from the Thom equation or measured directly by the eddy covariance method (see Section 2). Wet canopy evaporation was also calculated using the energy balance equation and the sensible heat flux as measured by a sonic anemometer (see Section 2). The evaporation rates obtained by the respective methods are then compared with values inferred from the rainfall, throughfall, and stemflow measurements.

2. Theory

The evaporation from a wet canopy is conventionally calculated using the Penman–Monteith equation (Monteith, 1965):

$$\lambda E = \frac{\Delta A + \rho c_p VPD g_a}{A + \gamma} \quad (1)$$

in which Δ is the slope of the saturated vapor pressure–temperature relationship at temperature T (kPa K⁻¹), A the available energy (W m⁻²) (in this study assumed to be equal to the net radiation R_n), ρ the density of air (kg m⁻³), c_p the specific heat of air at constant pressure (J kg⁻¹ K⁻¹), VPD the vapor pressure deficit (kPa), γ the psychrometric constant (kPa K⁻¹), g_a the aerodynamic conductance (m s⁻¹), and λ the latent heat of vaporization of water (J kg⁻¹).

The aerodynamic conductance (g_a) is usually estimated with the equation for momentum transfer of Thom (1975), with an added term to correct for non-neutral atmospheric stability conditions (Paulson, 1970; Webb, 1970; Monteith and Unsworth, 2008):

$$g_{a,M} = \left\{ \frac{k}{\ln[(z-d)/z_{0,M}] - \psi_M} \right\}^2 u \quad (2)$$

in which k is von Karman's constant (0.41), d the zero-plane displacement (m), $z_{0,M}$ the roughness length for momentum (m), u (m s⁻¹) the wind speed measured at height z (m), and ψ_M is an integrated stability correction function. Atmospheric stability is evaluated as $(z-d)/L$, where L is the Monin–Obukhov length (see Section 3; Monteith and Unsworth, 2008).

The aerodynamic conductance may also be derived in a more direct manner using the friction velocity (u_* , m s⁻¹) as obtained from sonic anemometer measurements (Gash et al., 1999; Van der Tol et al., 2003):

$$g_{a,M} = \left(\frac{u_*}{u} \right)^2 u \quad (3)$$

The friction velocity (u_*) is given by (Stull, 1988; Weber, 1999):

$$u_* = (\overline{u'w'^2} + \overline{v'w'^2})^{0.25} \quad (4)$$

where $\overline{u'w'}$ and $\overline{v'w'}$ are the co-variances of the turbulent fluctuations of the horizontal (u' , v') and vertical (w') wind components as measured with a sonic anemometer.

It is pertinent to note that Eq. (2) is based on the Monin–Obukhov similarity theory, which only holds under conditions of an extended, uniform, and flat surface (e.g. Monteith and Unsworth, 2008). In addition, similarity theory is only valid for the inertial sub-layer of the atmospheric boundary layer, whilst closer to the surface (i.e. in the roughness and transitional sub-layers) ‘similarity breakdown’ occurs (e.g. Thom et al., 1975; Simpson et al., 1998). As discussed by Chen and Schwerdtfeger (1989), the sub-layer with similarity breakdown tends to become more extended as the topographic complexity of the terrain is more pronounced. In such environments, u_* is enhanced (compared to ‘uniform’ environments experiencing the same large-scale flow), whereas at the same time u at the measurement level tends to decrease. The latter is related to the increase of the average du/dz above the measurement level, where du/dz scales with the increased u_* . The decrease in u is reflected in the decreased

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