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Spatial modelling of evapotranspiration in the Luquillo experimental forest of Puerto Rico using remotely-sensed data

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Summary Actual evapotranspiration (aET) and related processes in tropical forests can explain 70% of the lateral global energy transport through latent heat, and therefore are very important in the redistribution of water on the Earth's surface [Mausser, M., Schädlich, S., 1998. Modelling the spatial distribution of evapotranspiration on different scales using remote sensing data. *J. Hydrol.* 212–213, 250–267]. Unfortunately, there are few spatial studies of these processes in tropical forests. This research integrates one Landsat Thematic Mapper (TM) image and three Moderate Resolution Imaging Spectroradiometer (MODIS) images with a hydrological model [Granger, R.J., Gray, D.M., 1989. Evaporation from natural nonsaturated surfaces. *J. Hydrol.* 111, 21–29] to estimate the spatial pattern of aET over the Luquillo Experimental Forest (LEF) – a tropical forest in northeastern Puerto Rico – for the month of January, the only month that these remotely sensed images were acquired. The derived aETs ranged from 0 to 7.22 mm/day with a mean of 3.08 ± 1.35 mm/day which were comparable to other estimates.

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Simulated aET was highest in the low elevation forest and decreased progressively toward higher elevations. Because of differences in solar radiation at different elevations, aspects and topographic positions, aET tended to be higher on south slopes and along ridges than on north slopes and in valleys. In addition, the Bowen ratio (the ratio of sensible heat to latent heat) varied across different vegetation types and increased with elevation, thus reflecting differences in the distribution of net solar radiation incident on the earth's surface. Over a day, the highest simulated aET occurred at around noon. We also applied this model to simulate the average monthly aET over an entire year based on the cloud patterns derived from at least two MODIS images for each month. The highest simulated aET occurred in February and March and the lowest in May. These observations are consistent with long term data. The simulated values were compared with field measurements of the sap flow velocity of trees at different elevations and in different forest types. These comparisons had good agreement in the low elevation forest but only moderate agreement in the elfin forest at high elevations.

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Introduction

Tropical forests are responsible for a large part of the world's actual evapotranspiration (aET), and are very important in the redistribution of water on the Earth's surface (Mauser and Schädlich, 1998). Research on the water budget, the role of evapotranspiration in mineral pumping, and the energy components of water flow requires data on rates of evapotranspiration from various topographies, during different diel regimes, and from different conditions of rainfall and season (Odum et al., 1970a). Evapotranspiration is also closely related to the net primary production of different ecosystems (Rosenzweig, 1968), tree height and tree growth (Holdridge, 1947, 1962; Holdridge and Tosi, 1967; Ryan and Yoder, 1997).

Although aET is one of the most important components of the water balance, it is one of the most difficult to measure. It is especially difficult to measure aET of forests under natural conditions due to the large size of trees and the heterogeneity of forest stands. This heterogeneity includes species and canopy diversity, topographical exposure and the horizontal extension of the forest (Granier, 1987).

A number of techniques have been developed to measure the water use of large trees, e.g., water balance, lysimeters, large tree potometers, tent enclosures, chemical tracers, radioisotopes such as tritium, stable isotopes such as deuterium, energy balance, heat dissipation and heat techniques. Before 1990, evapotranspiration generally was estimated indirectly from water balances by subtracting rainfall interception, changes in soil water storage and surface runoff from rainfall. This catchment water balance approach resulted in considerable uncertainty because the estimates are affected by errors in all of the measurements (Bosch and Hewlett, 1982; Moran and O'Shaughnessy, 1984). Deep drainage losses were also often ignored and contributed to uncertainty unless the catchment was watertight (Ward and Robinson, 1990). Several studies have measured transpiration using micrometeorological techniques such as the Bowen ratio and eddy covariance methods. However, these are difficult to apply in rough terrain and can be used only in areas where the surface is homogeneous (Granier, 1987). More recently, energy-balance, heat-dissipation and heat-pulse systems are gaining widespread acceptance. Wullschlegel et al. (1998) reviewed 52 publications between

1970 and 1998 in which different techniques were used to provide quantitative estimates of whole-plant water use for trees growing in stands or plantations. Thirty-two of these were published during the past decade. Of these, 30 used energy-balance, heat dissipation or heat-pulse methods.

Formalizing a conceptual model using mathematical relations and simulation models allows us to make explicit quantitative projections based on the relations, assumptions, and initial conditions we assume are reasonable for the system being considered. The most widely applied model to simulate the relation between evaporation and other environmental factors for a wet environment with an unlimited water supply was developed by Penman (1948) and modified by Monteith (1965). It combines the concepts of the energy-balance of the land surface with a species-dependent surface resistance (i.e. the inverse of stomatal conductivity) for water-vapor release (Mauser and Schädlich, 1998). Using an approach similar to that used by Penman (1948), Granger and Gray (1989) derived a general combination equation to describe evaporation from nonsaturated surfaces.

One can generate formal spatial models to derive a quantitative assessment of the spatial distribution of aET. Spatial simulation models specifically include the geographical arrangements of key elements of the system being studied and can address questions about the spatial implications of interacting processes (Gergel and Turner, 2001). Most importantly, such models allow us to test our assumptions and models through measurements on the ground. Net radiation and vapor pressure deficit are the principal driving parameters for most models used to estimate aET. Progress in both remote sensing and GIS-based modeling techniques make it easier to derive these parameters optically. Visible light channels can be used to estimate the surface albedo, from which net radiation and hence the absorption of solar energy can be estimated. Thermal channels provide an estimate of the land surface temperature from which the vapor pressure deficit can be estimated (Granger, 2000).

The objective of this study was to model the spatial distribution of aET over the Luquillo Experimental Forest in Puerto Rico by integrating remotely-sensed data and the Granger and Gray evapotranspiration model. We compared the results against the field measurements of transpiration using the thermal dissipation method.

Study area

The Luquillo Experimental Forest (LEF) is located in the Luquillo Mountains in the northeastern part of Puerto Rico, between 18°14'45.8" and 18°20'58.2" N latitude and between 65°42'46.6" and 65°53'53.3" W longitude (Wang, 2001). Elevations range from about 100 to 1075 m above mean sea level (Weaver and Murphy, 1990). Mean annual rainfall and temperature increases with elevation from approximately 2450 mm/year and 23 °C at lower elevations to over 4000 mm/year and 19° at higher elevations (Brown et al., 1983; Weaver and Murphy, 1990; Scatena and Lugo, 1995; Silver et al., 1999). There are four main forest ecosystem types in the LEF: tabonuco, Colorado, palm and elfin (also called cloud or dwarf) forests. These four ecosystems are stratified roughly by altitude and soil moisture. Tabonuco forest is found below 600 m, is best developed on low, protected, well-drained ridges and occupies nearly 70% of the LEF's area. Colorado forest, which covers about 17% of the LEF, is above the average cloud condensation level (600 m) and is expected to exhibit lower evapotranspiration. Elfin forest, with its short, gnarled vegetation, is found on peaks and ridges above 750 m in elevation and represents only 2% of the LEF by area. Palm forests are limited to areas of steeper slopes, poor drainage and saturated soils at all elevations and cover 11% of the LEF's area (Brown et al., 1983).

Model description

Actual evapotranspiration (aET) was estimated using Granger and Gray's model (1989) which makes use of the concept of relative evaporation to account for the departure of evaporation from what it would be at saturated conditions, and it does not require prior calculation of potential evaporation. Since potential evaporation has not been clearly defined in a universally-accepted manner, this model successfully avoids calculating this yet-to-be-defined parameter (Granger and Gray, 1989). The parameters of this model are derived from the Landsat TM image, MODIS images and a digital elevation model (DEM).

According to Granger and Gray (1989), the energy balance in the vertical direction above a horizontal surface, assuming advection is negligible, can be written as:

$$\lambda ET + H = Q \quad (1)$$

where λ is known as the latent heat of vaporization, and is a function of the water temperature; ET is the evapotranspiration rate; λET is the latent heat flux, the energy used to evaporate and transpire; H is the sensible heat flux; Q is the total energy available from net radiation after subtracting soil heating.

Granger and Gray derived a general combination equation following a development similar to that used by Penman (1948). This equation is similar in form to the

Penman equation, but differs through the inclusion of relative evaporation, G , which accounts for departure from saturated conditions:

$$ET = \frac{\Delta GQ}{\Delta G + \gamma} + \frac{\gamma GEa}{\Delta G + \gamma} \quad (\text{Granger and Gray, 1989}) \quad (2)$$

where Δ is the slope of the saturation vapor pressure curve (kPa/°C); γ is the psychrometric constant (kPa/°C) (a function of atmospheric pressure); Δ and γ are calculated using the method of FAO56 (Allen et al., 1998); G represents relative evaporation ($\frac{ET}{\text{Potential ET}}$) and is calculated using the method of Granger and Gray (1989):

$$G = \frac{1}{1 + 0.028e^{8.405D}} \quad (2.1)$$

where

$$D = \frac{Ea}{Ea + Q} \quad (2.2)$$

where Ea is referred as the drying power of the air, the product of a wind function $f(u)$ and the vapor pressure deficit, where u is the wind speed,

$$f(u) = \frac{0.622\rho}{P \times r_a} \quad (2.3)$$

where ρ is the air density (g/cm³); P is the atmospheric pressure (mb); r_a is aerodynamic resistance (s/cm).

The empirical relation between G and D (Eq. (2.1)) should be used with caution since the lack of measurements when $G > 0.7$ does not allow the development of a functional relation between G and D that can be treated with confidence over the entire range in G (Granger and Gray, 1989). However, we used this empirical relation in the ET simulation. In order to account for the uncertainty, we performed sensitivity analysis on the empirical coefficients in this relation.

Flow diagram of the model and data source

The Landsat Thematic Mapper (TM) imagery used in this study was obtained at 10:14 a.m. local time on January 21, 1985 (Fig. 1, Wang, 2001). The properties of this image are shown in Tables 1 and 2. The important inputs for the hydrological model (Granger and Gray, 1989), which include surface albedo, net short wavelength radiation, surface temperature, net long wavelength radiation and soil heat, were derived from the Landsat image.

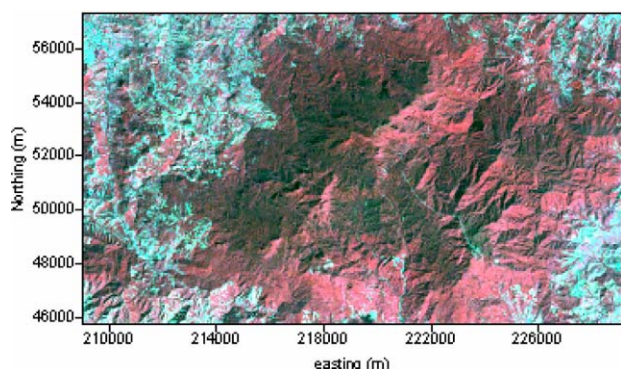
MODIS level 1B (radiometrically and geometrically corrected) images acquired on 8 January 2003, 4 January 2004, and 18 January 2004 were used to derive the probability of cloud cover (defined as "percentage of cloud cover per unit area") for January using the method of Wu et al. (2006). In addition, at least two images for every other month were used to derive the monthly average probability of cloud cover in those months. They were obtained at approximately

Table 1 Spectral and spatial characteristics of Landsat-5 TM images

Band	Band 1	Band 2	Band 3	Band 4	Band 5	Band 6	Band 7
Wavelength (μm)	0.45–0.52	0.52–0.60	0.63–0.69	0.76–0.90	1.55–1.75	10.4–12.5	2.08–2.35
Resolution (m)	30					120	30

Table 2 Important properties of the Landsat-5 TM image used in this study (from metadata of the image)

Sun elevation (°)	Sun azimuth (°)	Cloud cover	Datum	Coordinate system
39	137	0–9%	NAD27	State Plane (Zone 5201)

**Figure 1** The Landsat-5 TM image of the LEF obtained on January 21 of 1985 in the coordinate system of state plane (Zone 5201) (displayed as false color: band 4 as red, band 3 as green, band 2 as blue).

11:00 a.m. local time. Two MODIS bands were used: one visible band, and the other a near-infrared band. The spatial resolution of the images is 250 m. They were georeferenced using state plane coordinates, NAD 27 datum.

From a Digital Elevation Model (DEM) with 30-m resolution, the hill/terrain slopes and aspects of the LEF were derived using the "SURFACE" function in Idrisi32. Other key inputs of the hydrological model, including the slope of saturation vapor pressure, the psychrometric constant and the

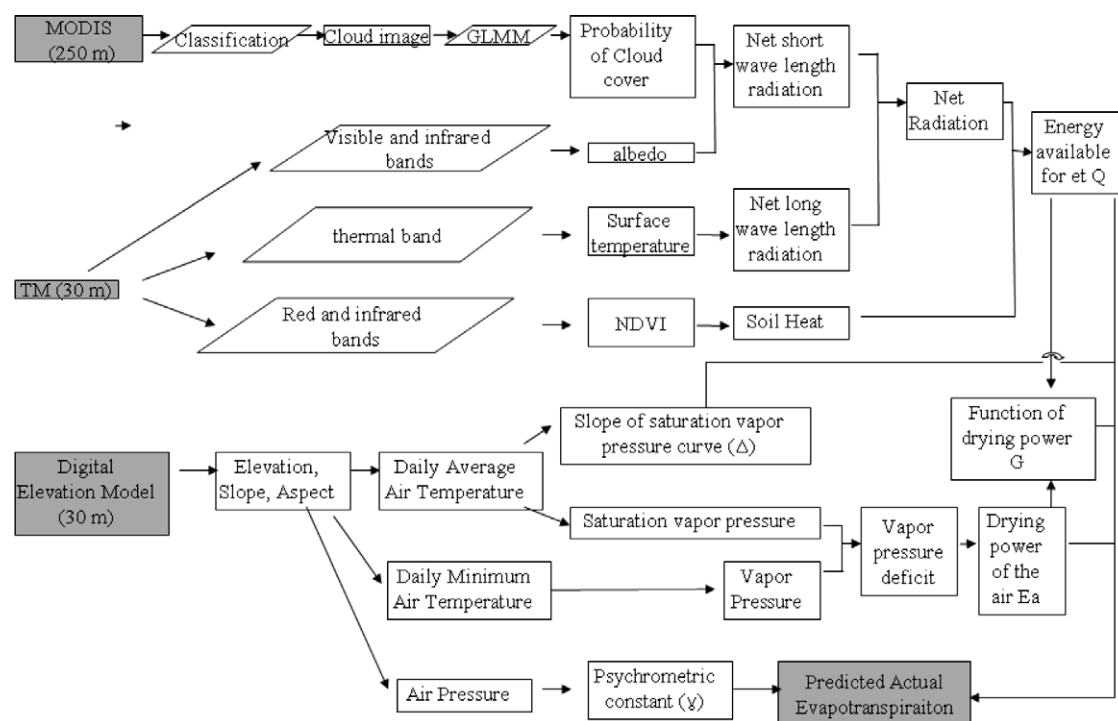
drying power of the air, were calculated from the DEM and the meteorological data collected in the LEF (refer to "Derivation of the model parameters" section).

These remote sensing images and maps are very important inputs to the model (Fig. 2).

Derivation of the model parameters

The probability of cloud cover calibrated by MODIS images

A logarithmic relation of the probability of cloud cover ($\log(\frac{\text{probability}}{1-\text{probability}})$) and three independent topographic variables was derived using a general linear mixed model (GLMM) that incorporated an exponential spatial covariance structure to account for the spatial autocorrelation in the data of cloud cover (Wu et al., 2006). The three variables are the difference between Elevation and Lifting Condensation Level (Elev-LCL), Aspect (Asp) and Slope (Slp). The MODIS images were used to calibrate the model. Because multiple MODIS images were being used, accurate registration was necessary to mitigate the distortions in remote sensing images that occur as a result of variations in factors such as platform position, rotation of the earth, and relief displacement (Chen et al., 2003). We used manual procedures in ERDAS Imagine 8.6 to register the images. The reference image used was the Landsat TM scene, which was

**Figure 2** Flow diagram of the inputs, data processing, and outputs of the evapotranspiration model. (MODIS: Moderate Resolution Imaging Spectroradiometer; TM: Landsat Thermal Mapper; GLMM: Generalized Linear Mixed Model).

orthorectified by the vendor that supplied this imagery. These orthorectified scenes are the most accurate commercially available base maps of the world created from Landsat imagery, and have a better positional accuracy (50 m RMSE) than the vast majority of the world's 1:200,000 maps (http://www.geocover.com/gc_ortho). Before the MODIS images were registered (250-m pixel size) to the reference image (30-m pixel size), both images were resampled to have 240-m pixels. The relative mean square errors (RMSEs) of registration were all within one pixel. It is difficult to obtain a registration of RMSE below half of one pixel due to the existence of clouds in the images.

The "unsupervised classification" module in ERDAS Imagine was used to distinguish clouds from clear sky and cloud shadows. The cloudy areas were coded as 1, and the other areas were coded as 0s. Because the clouds were formed mainly due to orographic uplift in the mountainous area for most times of the year, the cloud cover was closely associated with the lifting condensation level and the topographical variables such as elevation, aspect and slope. We used the dew point temperature as well as temperature data from the El Verde meteorological station (18°19'22" N, 65°49'13" W, elevation = 350 m AMSL) in the LEF to calculate the temperature at the condensation pressure level according to the (AWIPS) (Advanced Weather Interactive Processing System) method (http://meted.ucar.edu/awips/validate/li_as.htm). We then derived the LCL, using the coefficient associated with the variable "elevation" in the regression models that quantified the relation between temperature and elevation. The relation was based on January temperature data from 10 locations along a windward elevation gradient in the (LEF) (available at <http://luq.lter.net.edu/data/temp/bistempdata/Bis-temp.htm>).

The derivation of lifting condensation level (LCL) is consistent with that of Briscoe (1966), who based his derivation on the independent meteorological data from Cape San Juan located at the northeastern tip of Puerto Rico, 12 miles northeast of El Yunque peak and situated on a low hill (Odum et al., 1970b): LCL is low in elevation in the morning, moves up slope until early afternoon, then moves down slope in the afternoon and again at night. Since LCL changes with time of day, we estimated hourly probabilities of cloud cover (P) over a day. The relation in the GLMM is formed as (Agresti, 1990):

$$\ln \frac{P}{1-P} = \beta_0 + \beta_1(\text{Elev} - \text{LCL}) + \beta_2(\text{Asp}) + \beta_3(\text{Slp}) \quad (3)$$

$$P = \frac{e^{\beta_0 + \beta_1(\text{Elev} - \text{LCL}) + \beta_2(\text{Asp}) + \beta_3(\text{Slp})}}{1 + e^{\beta_0 + \beta_1(\text{Elev} - \text{LCL}) + \beta_2(\text{Asp}) + \beta_3(\text{Slp})}} \quad (4)$$

(P is the probability of cloud cover as a function of the difference between elevation and lifting condensation level (Elev-LCL), aspect (Asp) and slope (Slp); β_0 , β_1 , β_2 and β_3 are the regression coefficients in the GLMM.)

Incidence angle of solar radiation

The incidence angle of solar radiation was determined by the solar position related to the slope and aspect of the land surface (Tasumi et al., 2000) for each pixel in the Landsat image at each hour during daytime (18.25° N, 65.83° W). The incidence angle was later used to calculate the solar radiation incident on the land surface.

$$\begin{aligned} \text{CosIncAg} = & \text{SIN}(\text{SDc}) \times (\text{SIN}(\text{Lat}) \times \text{COS}(\text{Slp}) \\ & - \text{COS}(\text{Lat}) \times \text{SIN}(\text{Slp}) \times \text{COS}(\text{Asp})) \\ & + \text{COS}(\text{SDc}) \times (\text{COS}(\text{Lat}) \times \text{COS}(\text{Slp}) \times \text{COS}(\text{RHA}) \\ & + \text{SIN}(\text{Lat}) \times \text{SIN}(\text{Slp}) \times \text{COS}(\text{Asp}) \times \text{COS}(\text{RHA}) \\ & + \text{SIN}(\text{Asp}) \times \text{SIN}(\text{Slp}) \times \text{SIN}(\text{RHA})) \end{aligned} \quad (5)$$

where CosIncAg, Cosine of the incident angle of solar radiation; SDc, Solar declination (negative in the winter of the northern hemisphere); Lat, Latitude; Slp, Slope; Asp, Aspect from due south; RHA, hour angle. 0 representing solar noon, negative values representing morning, positive values representing afternoon.

Surface albedo (α)

Albedo is the ratio of reflected to incident solar radiation at a surface and is computed as the ratio of outgoing short wave radiation to incoming short wave radiation. The reflectivity of a surface is wavelength-dependent, with few natural surfaces being uniform reflectors across the portion of the electromagnetic spectrum of interest (Dirmhirn, 1968; Kondratyev, 1969; Pease and Pease, 1972; Goward, 1985; Brest and Goward, 1987). We calculated the reflectance of TM bands 1–5 and 7 (band 6 is the thermal band) as the ratio of outgoing radiation of the band measured by the satellite to the incoming radiation of the band at the top of the atmosphere (Tasumi et al., 2000). The outgoing radiation for each band was calculated using Eq. (6) and the calibration constants are shown in Table 3 (Tasumi et al., 2000).

$$\text{Rad}_{\text{out}} = \left(a + (b - a) \times \frac{\text{DN}}{255} \right) \times \pi \quad (6)$$

where DN, original digital number recorded for the pixels of each band in the satellite image; a , b , calibration constants (Table 3).

The incoming radiation of the band at the top of atmosphere was computed as:

$$\text{Rad}_{\text{in}} = \text{Gsc} \times \text{CosIncAg} \times \text{dr} \quad (7)$$

where Gsc, solar constant for the band (Table 3); dr, Inverse relative distance from the earth to the sun, which corrects for the earth's elliptical orbit and was calculated as

Table 3 Calibration constants for Landsat-5 TM images Tasumi et al. (2000) (mW – microwatts; ster – solid angle)

	Band 1	Band 2	Band 3	Band 4	Band 5	Band 7
Gsc (mW/cm ² /μm)	195.7	182.9	155.7	104.7	21.93	7.452
a (mW/cm ² /ster/μm)	−0.15	−0.28	−0.12	−0.15	−0.037	−0.015
b (mW/cm ² /ster/μm)	15.21	29.68	20.43	20.62	2.72	1.44
Weighting coefficient	0.293	0.274	0.233	0.157	0.033	0.011

$$dr = 1 + 0.033 \times \cos(\text{JulDay} \times 2 \times \pi/365) \quad (8)$$

where JulDay = Sequential day of a year.

After we calculated the reflectance of each band ($\frac{\text{Rad}_{\text{out}}}{\text{Rad}_{\text{in}}}$), we used the following equation to compute the albedo at the top of the atmosphere (Tasumi et al., 2000):

$$\text{toa} = \sum_{i=1,2,3,4,5,7} \text{weight} \times \text{reflectance}(\text{band}_i) \quad (9)$$

where toa, albedo at the top of the atmosphere; weight, weighting coefficient from Table 3.

Then according to Tasumi et al. (2000) surface albedo (α) was calculated as:

$$\alpha = \frac{\text{toa} - \text{pathradiance}}{\tau^2} \quad (10)$$

where τ is one way transmittance; and path radiance refers to that component of radiation received by a sensor at the top of the atmosphere but did not reach the earth because it was lost through scattering in the earth's atmosphere. A general value for one-way transmittance for clear sky is (Tasumi et al., 2000):

$$\tau = 0.75 + 2 \times 10^{-5} \times \text{Elevation} \quad (11)$$

In our model, transmittance was calculated by multiplying the transmittance for a clear sky by (1 – probability of cloud cover). Path radiance values are typically between 0.025 and 0.04 (Tasumi et al., 2000). We used a value of 0.0325, the average of 0.025 and 0.04, for path radiance.

Surface temperature (T_{surf})

Surface emissivity and radiance temperature in the thermal band are required to estimate surface temperatures. Surface emissivity is a factor that describes how efficiently the surface radiates energy compared to a blackbody (Lillesand and Kiefer, 2000). Normalized difference vegetation index (NDVI), one of the vegetation indices that can be calculated from TM images (Eq. (12)), was used by Van de Griend and Owe (1993) to estimate surface emissivity (e) according to Eq. (13).

$$\text{NDVI} = \frac{\text{reflec}(\text{band } 4) - \text{reflec}(\text{band } 3)}{\text{reflec}(\text{band } 4) + \text{reflec}(\text{band } 3)} \quad (12)$$

where reflec = reflectance

$$e = \begin{cases} 1.009 + 0.047 \ln(\text{NDVI}) & \text{when NDVI} > 0 \\ 0, & \text{otherwise} \end{cases} \quad (13)$$

The integrated thermal radiance of band 6 (L) was calculated according to Scott and Volchok (1985):

$$L = 0.0056322 \times \text{DN}_{\text{band } 6} + 0.1238 \quad (14)$$

Scott and Volchok (1985), Wukelic et al. (1989), Goetz et al. (1995) and Sospedra et al. (1998) proposed relations similar to Planck's function with two parameters K_1 and K_2 that were adjusted to give the radiant temperature (T) with the maximum accuracy.

$$T = \frac{K_2}{\ln(\frac{K_1}{T} + 1)} \quad (15)$$

where $K_1 = 67.162 \text{ mW cm}^{-2} \text{ sr}^{-1} \mu\text{m}^{-1}$ and $K_2 = 1284.3 \text{ K}$ (Kelvin degrees).

Then with e (Eq. (13)), surface temperature (T_{surf}) is calculated as (Tasumi et al., 2000):

$$T_{\text{surf}} = \frac{T}{e^{0.25}} \quad (16)$$

Long wave radiation

The interaction of electromagnetic radiation with the earth's atmosphere is complex. The atmosphere modifies the radiance that reaches the ground and also that part reflected from the ground and contributes an additive path radiance term. Therefore, it is necessary to correct for the atmospheric effect to estimate net radiation. Many studies (Holbo and Luvall, 1989; Richter, 1990; Duguay and LeDrew, 1992; Gratton et al., 1993; Goetz et al., 1995; Gillies et al., 1997; Wang et al., 2000) used the atmospheric radiative transfer models LOWTRAN-7 (Kneizys et al., 1988) and MODTRAN (Berk et al., 1989) to perform the correction. However, these models require many physical parameters that are not available for the Luquillo Experimental Forest. Thus we used a simpler model (Idso, 1981; Tuzet, 1990) to compute downward longwave radiation.

$$\text{RL} \downarrow (\text{Wm}^{-2}) = 0.175 \times \text{VP}^{1/7} \times \exp(350/T_{\text{surf}}) \times \sigma \times T_{\text{surf}}^4 \quad (17)$$

where VP, vapor pressure (mb) (Marley, 1998); T_{surf} , surface temperature (K) as calculated from Eq. (16); σ , Stefan–Boltzman constant $5.67 \times 10^{-8} \text{ W/m}^2/\text{K}^4$.

We then calculated net long wavelength radiation (Rlnet) as (Tasumi et al., 2000):

$$\text{Rlnet}(\text{Wm}^{-2}) = (\text{RL} \downarrow - e \times \sigma \times T_{\text{surf}}^4) - (1 - e) \times \text{RL} \downarrow \quad (18)$$

Net radiation (Rn)

Net incoming radiation is the sum of incoming short wave net radiation and incoming long wave net radiation. It was computed as (Brest and Goward, 1987; Boegh et al., 2002):

$$\text{Rn}[\text{Wm}^{-2}] = \text{Rs} \downarrow (1 - \alpha) + \text{Rlnet} \quad (19)$$

where $\text{Rs} \downarrow$, incoming short wave solar radiation; α , surface albedo.

Incoming short wave solar radiation $\text{Rs} \downarrow$ was calculated as the product of the solar constant (1367 W m^{-2}), the cosine of the solar zenith angle, the inverse relative distance from the earth (dr) to the sun and one-way transmittance (Tasumi et al., 2000).

Soil heat

An empirical equation was applied to estimate soil heating (Tasumi et al., 2000):

$$\text{Soil Heat} = 0.30(1 - 0.98\text{NDVI}^4)\text{Rn} \quad (20)$$

Aerodynamic resistance (r_a) in wind function

Aerodynamic resistance (r_a), required to calculate drying power, is normally obtained using Eq. (21) (Thom, 1975; Schellekens et al., 2000):

Table 4 The comparison between the values of r_a (aerodynamic resistance) we used with van der Molen's

Unit: m/s	Tabonuco forest	Colorado forest	Palm forest	Elfin forest
r_a from Van der Molen (2002) based on multiple methods	0.1–58 from the observations of the forests in Northeast Puerto Rico, 8–10 from parameterized method, 1–50 using modeled temperature and mixing ratios and their parameterization			
r_a in this paper	30.0	13.73	13.73	3.96

$$r_a = \frac{(\ln \frac{z-d}{z_0})^2}{k^2 u} \quad (21)$$

where, z , measurement height above the ground surface; d , zero plane displacement height, estimated as $0.83 \times$ vegetation height (Shuttleworth, 1989); z_0 , roughness length, estimated as $0.06 \times$ vegetation height (Shuttleworth, 1989); k , von Karman's constant, 0.41; u , wind speed at height z .

We used the measured data (Van der Molen, 2002) of heights of wind speed, wind speed and vegetation heights in the palm and elfin forests to calculate r_a , which is 13.73 s/m and 3.96 s/m respectively in those two forest types, and apply the same r_a of the palm forest to the Colorado forest. We used 30.0 s/m as the r_a in the tabonuco forest (Schellekens et al., 2000) (Table 4).

Other inputs

We derived linear relations between elevation and both monthly daily mean air temperature and monthly daily minimum air temperature using temperature data from 10 locations along a windward elevational gradient in the LEF (<http://luq.lternet.edu/data/lterdb90/data/bistempdata/Bis-temp.htm>) for each month (Table 5).

Saturation vapor pressure at a given temperature was estimated using the method developed by Murray (1967). We calculated vapor pressure and vapor pressure deficit using the equations from TOPOCLIM (Marley, 1998) for each 30×30 m area in the LEF. Other parameters we calculated were: atmospheric pressure (Ham, 2000), psychrometric constant (FAO, 1998) and the slope of the

saturation vapor pressure curve by the method of Allen et al. (1998).

Field methods and statistical analysis

Field work

We measured sap flow of different tree species at different elevations in the LEF using a thermal dissipation sap velocity probe-80 (TDP-80). The transpiration rate of whole plants (E_t) is closely approximated by the sap flow velocity (g/h) in the main stem or trunk of the tree (Dynamax, 1997). The TDP probe measures the temperature of a heat source implanted in the sapwood of a tree, referenced to the sapwood temperature at a location well below the heated needle. The principle behind this is the Granier method, which is based on the liquid velocity heat dissipation principle, rather than on a specific model of heat transport in plant stems or tree trunks. Therefore it is reasonable to use this technique on palm and other non-woody plants in our studies.

Granier defined a dimensional parameter K as:

$$K = \frac{dT - dT_m}{dT} \quad (22)$$

dT is the measured difference in temperature between that of the heated needle, referenced to the lower non-heated needle. dT_m is the value of dT when there is no sap flow, for example, at night.

Granier found empirically that the average sap flow velocity V (cm/s) could be related to K by an exponential expression:

$$V = 0.0119 \times K^{1.231} \text{ cm/s} \quad (23)$$

Table 5 Coefficient estimates in the regression model between temperature ($^{\circ}\text{C}$) and elevation (m.a.s.l.) in each month

	Daily mean temperature			Daily minimum temperature		
	Intercept in the regression model	Slope in the regression model	Adj- R^2	Intercept in the regression model	Slope in the regression model	Adj- R^2
January	24.822	-0.0054	0.89	21.718	-0.0051	0.96
February	23.809	-0.0059	0.92	20.613	-0.0048	0.98
March	24.330	-0.0053	0.87	20.005	-0.0050	0.97
April	25.935	-0.0060	0.90	21.183	-0.0060	0.92
May	26.851	-0.0060	0.90	21.601	-0.0040	0.57
June	28.374	-0.0062	0.89	24.094	-0.0053	0.98
July	28.280	-0.0064	0.90	23.851	-0.0048	0.97
August	27.991	-0.0055	0.85	23.862	-0.0047	0.98
September	27.455	-0.0051	0.83	23.665	-0.0047	0.98
October	26.505	-0.0047	0.85	23.121	-0.0048	0.97
November	25.786	-0.0050	0.91	22.946	-0.0046	0.97
December	25.581	-0.0052	0.87	22.237	-0.0050	0.95

A data logger recorded the temperature difference every minute, from which we calculated sap flow velocity according to Eqs. (22) and (23). Since we did not see significant fluctuations over each 30-min time period, we averaged the data every 30 min to reduce the data volume.

Statistical analysis and validation methods

ANOVAs were used to test for the effects of species, diameter at breast height (DBH) and solar radiation on measured sap flow.

The estimates of evapotranspiration (mm/day) derived from Granger and Gray's model were compared to the sap flow velocity (cm/h) measurements of the representative trees we sampled. We scaled the estimates of evapotranspiration from the model to the sap flow velocity of an "average" tree, and then compared them to the sap flow velocity of an "average" tree derived from our field data.

Deriving the sap flow velocity of an "average" tree from the modeled results

Sapwood area is often suggested as the appropriate scalar to scale up sap flow of trees to estimate field evapotranspiration (Cadler, 2001). In this study, "sap wood area" was used to scale down rather than up. We applied the equation of the equilibrium of volume of water transpired (Eq. (24)), and scaled the derived aET down to sap flow velocity of an "average" tree by using sapwood area.

$$SF(\text{cm/h}) \times \text{Area}_{\text{sw}} (\text{ha}) = \text{aET}(\text{mm/day}) \times \text{Area}_1 \text{ ha} (\text{ha})/24/10 \quad (24)$$

where SF, sap flow velocity of an average tree; Area_{sw} , sap wood area per hectare; Area, one hectare.

From the data of species types and DBHs along elevational transects collected by John Barone et al. (personal communication), we did not find a significant difference in basal area per hectare along the elevation gradient (p -value = 0.9959) at $\alpha = 0.05$. Therefore, we used the average basal area 42.72 m²/ha for all elevations, a value similar to Weaver and Murphy's (1990) measurements before Hurricane Hugo in 1989: 40 m²/ha in the tabonuco forest, 42 m²/ha in the Colorado forest, 40 m²/ha in the palm forest and 45–65 m²/ha in the elfin forest.

According to Teskey and Sheriff (1996):

Sap wood area (m²/ha)

$$= -0.00747 + 0.83671 \times \text{basal area (m}^2/\text{ha)} \quad (25)$$

Thus the average area of sap wood in the LEF is 35.74 m²/ha.

Then we applied Eq. (24) to derive the sap flow velocity of an "average" tree.

Deriving the sap flow velocity of an "average" tree from the field data

The degree to which an individual tree contributes to the daily evapotranspiration at a given elevation is a function of the sapwood area (derived from the basal area) compared to the total sapwood area of that elevation and whether the tree is in shade or in sunlight. Thus we needed to assign different weights to the sampled trees of different sizes and growing in different light environments in order to calculate the sap flow velocity of an "average" tree using sap flow data we collected in the field.

We classified trees according to their DBHs (class 1: DBH 1 cm–10 cm; class 2: DBH 10 cm–20 cm; class 3: DBH 20 cm–30 cm; up through class 18: 170 cm–180 cm) using the extensive field survey of Barone, J., (personal communication). We calculated the proportion of basal area of the trees for each size category at each elevation. Since sapwood area was derived here as a linear function of basal area (Teskey and Sheriff, 1996), the proportion of basal area can be seen as the proportion of total sapwood area, and approximately as the contributing factor of each-size class to the total evapotranspiration of the forest at that elevation. If we assigned this proportion of each class to the sap flow velocity of the trees in that size class we sampled, and calculated the weighted average, we could get the sap flow velocity of an "average" tree. However, we needed to correct for the shading factor because a tree in the shade would transpire less than a tree of the same size (DBH) in the sunlight. We assigned different shading probabilities to different size classes (Table 6): the smaller the tree, the higher the probability that the tree was in the shade. We assumed that all the trees larger than size class 4 (>40 cm DBH) were going to be in the sun all the time. If the trees were in the shade, we decreased the evapotranspiration per unit of sapwood area by a factor that we varied arbitrarily from 30% to 70%. We considered this to account adequately for the uncertainty of the shading correction (last column of Table 6).

We assigned the contributing proportions based on the above discussion (summarized in Table 7) as the weights to the sap flow velocity of the trees in each size class we sampled to calculate a weighted average. This approximated the sap flow velocity of an "average" tree.

Table 6 Definitions of three situations to account for uncertainties of shading factors

Cases	Percentage of trees in the shade in each size class				Sapwood area shading scalar(%)
	Class 1 DBH 1–10 cm	Class 2 DBH 10–20 cm	Class 3 DBH 20–30 cm	Class 4 DBH 30–40 cm	
Situation 1	0.50	0.25	0.10	0	50
Situation 2	0.80	0.40	0.20	0.10	70
Situation 3	0.30	0.15	0.07	0	30

Note: If we assign 0.50 to the percentage of trees in shade in each size class, it means 50% of the trees in that class would be in the shade, and for the trees in shade, the sap wood area will be decreased by "sapwood area shading factor", e.g., in situation 1, it will be reduced by 50%.

Table 7 Contributing proportion of aET as a function of the tree size and shading factor (see Table 6 for the definitions of situation 1, 2 and 3)

Site	Size class	Proportion of number of trees (%)	Contributing proportion of each size class in situation 1 (%)	Contributing proportion of each size class in situation 2 (%)	Contributing proportion of each size class in situation 3 (%)
Bisley upper tower (385 m)	1	81.84	7.28	4.74	8.46
	2	15	27.72	25.13	28.98
	3	1.75	10.16	10.13	10.03
	4	0.44	5.00	5.12	4.79
	5	0.53	10.53	11.61	10.09
	6	0.26	8.34	9.20	7.99
	7	0.088	0.04	4.69	4.08
	>7	0.088	0.27	29.42	25.58
El Verde station (451 m)	1	74.3	9.35	9.35	8.14
	2	22.18	41.00	41.00	39.68
	3	1.64	8.49	8.49	8.70
	4	0.94	9.19	9.19	9.71
	5	0.47	8.40	8.40	8.88
	6		0.00	0.00	0.00
	7	0.23	10.04	10.04	10.61
	8	0.12	6.53	6.53	6.90
	9	0.12	7.00	7.00	7.39
Cloud forest (991 m)	1	90.16	21.01	21.01	23.74
	2	9.29	30.00	30.00	30.50
	3	0.46	4.90	4.90	4.70
	>3	0.092	44.09	44.09	41.06

Results

Our results indicate that simulated aET was highest at lowest elevations and decreased progressively up the mountain (Fig. 5). They also indicate that aET progressively decreased from south-facing slopes to north-facing slopes and from ridges to valleys and suggest that the highest aET occurred around noon and during the months of February and March. These observations are in agreement with other direct observations from the Luquillo Forest and indicate that our methods have generated a useful and fairly accurate model that can be applied to other tropical environments.

Results of the probability of cloud cover

Modeled probabilities of cloud cover increased with elevation. Highest values occurred in the elfin forest at high elevations, and decreased progressively in the Colorado forest, palm forest at middle elevations and the tabonuco forest at low elevations (Fig. 3). In addition, modeled probabilities of cloud cover were highest at night and early morning, then decreased in the morning after the sun rose until early afternoon and increased again from the afternoon through night (Fig. 4). This trend was apparently in response to the movement of the lifting condensation level over a day (Odum et al., 1970b). Since cumulus clouds appear to be the dominant cloud type in the LEF (Wooster, 1989), we assumed that the existence of clouds attenuated solar radiation by 75% (Page, 1986) when we modeled the net solar radiation incident on the earth. For example, if the probability of cloud cover for one pixel on the map was 0.80 for

a given hour, this translated into 80% of the area of that pixel was covered by clouds. However this 80% clouded area could be located anywhere within the pixel during that hour. Therefore solar radiation in 80% of the pixel area received full sun solar radiation attenuated by 75% (Page, 1986) while the other 20% of the area in that pixel received full sun solar radiation with no attenuation.

Results from the evapotranspiration model

The simulated aET from our application of Granger and Gray's model ranged from 0 to 7.22 mm/day over the entire LEF with a mean of 3.08 mm/day and a standard deviation of 1.35 mm/day in January (Fig. 5). Most of the simulated data fell around 3.0 mm/day. Modeled aET was the highest in the tabonuco forest at low elevations (close to the boundary of the forest) and decreased with elevation (Fig. 5), thus becoming progressively less from the Tabonuco forest to the Colorado forest, to the palm forest to the elfin forest (Table 8). At the same elevation, south facing slopes tended to have higher aET than north facing slopes (Figs. 6, and 7), apparently in response to different solar intensities (Fig. 8). Trees in the elfin forest appeared to be better adapted to reduced solar radiation since they had higher rates of aET than trees exposed to the same low level of solar radiation but located at lower elevations. Conversely, as solar radiation increases, trees at low elevations increased their rates of transpiration more rapidly than trees in the elfin forest. This resulted in higher aET rates at low elevation trees at high levels of solar radiation (Fig. 8). Over a day, the highest aET occurred at around noon (Fig. 9), consistent

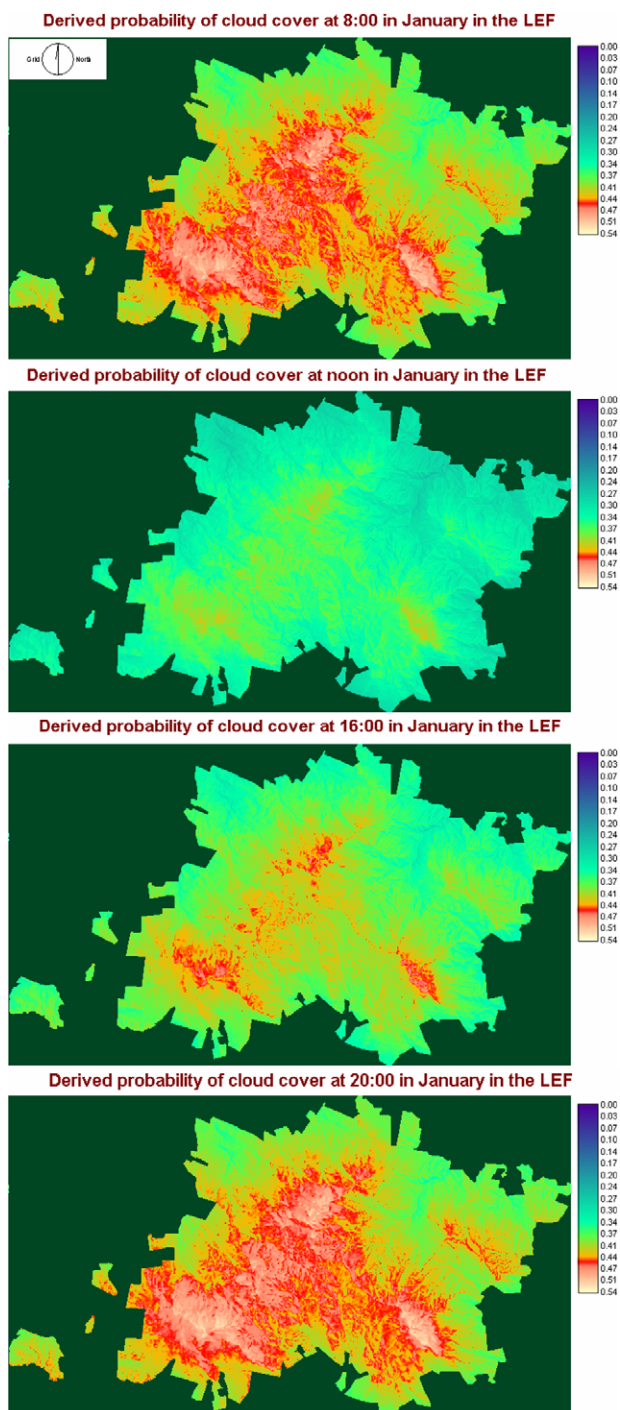


Figure 3 The probability of cloud cover at 8:00, noon, 16:00 and 20:00 in January in the LEF in the coordinate system of state plane.

with Van der Molen's results (2002, pp. 81–82: Fig. 4.5 and 4.6, pp. 92: Fig. 4.13, pp. 100–101: Fig. 4.18 and 4.19).

The "TOPOSHAPE" function in the "Feature Extraction" module of IDRISI was used to extract three topographic features: ridge, valley, and hillside (Fig. 10). Rates of aET at the ridges were higher than those at the valleys (Table 9), which was most likely due to the higher solar radiation received at the ridge locations.

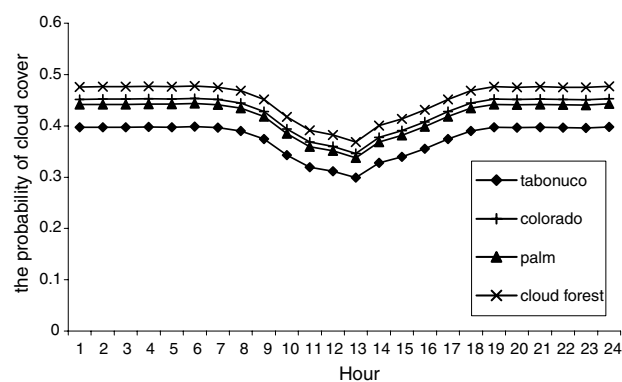


Figure 4 The probability of cloud over a day in January.

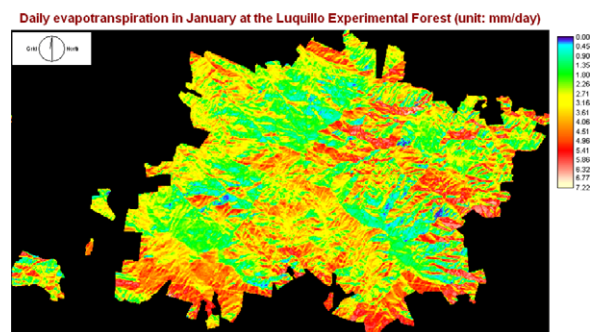


Figure 5a Derived daily average aET in January (unit: mm/day).

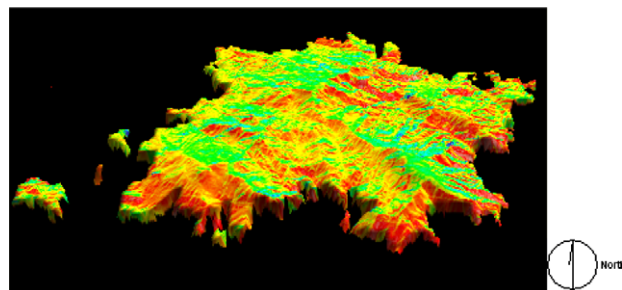


Figure 5b Derived daily average aET in January displayed in 3-D (same color scheme as in Fig. 5a unit: mm/day).

Table 8 Simulated average aET with standard deviation at different forest types in the LEF

	Simulated mean (mm/day)	Standard deviation (mm/day)
Whole forest	3.08	1.35
Tabonuco	3.28	1.42
Colorado	2.84	1.15
Palm	2.96	1.43
Elfin	1.72	0.58

Since the Luquillo forest is a tropical forest, and the albedo of the vegetation does not change much over a year (Van der Molen, 2002), we simulated aET over a year using

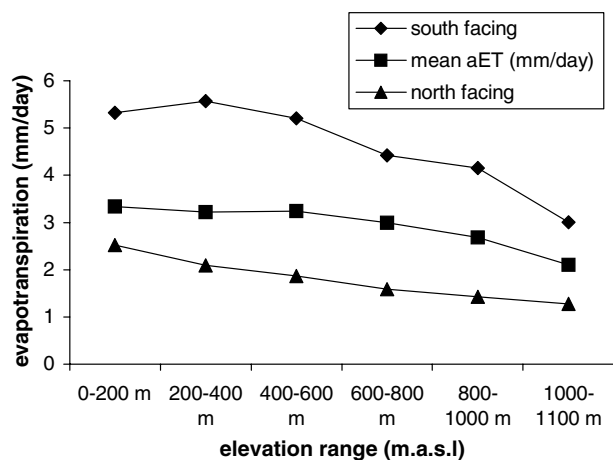


Figure 6 The simulated average aET and aET on south slopes and on north slopes along the elevation gradients in the LEF.

the same albedo we derived from the Landsat image for the month of January. Surface temperatures were estimated from air temperatures and the regression relation between air temperatures and surface temperatures derived in Janu-

ary (surface temperature = $6.188 + 0.647 \times \text{air temperature}$, $R^2 = 0.61$). The probability of cloud cover was derived from at least two MODIS images every month. The highest simulated aET occurred in February or March at different forest types when there was abundant (but not overabundant) precipitation and high net solar radiation incident on the earth because of low probability of cloud cover. The lowest aET occurred in May, when the probability of cloud cover was highest (Fig. 11). The aET at the elfin forest was more constant than at the other forest types, probably due to the low variation of cloud cover over a year at the high elevations.

At van der Molen's MICRO_MET sampling site in the palm forest ($18^\circ 16' 59''$ N, $65^\circ 46' 05''$ W), our simulated monthly aET (Fig. 12) was in the range of measurements but underestimated summer values. The inconsistency may be due to differences in cloud cover during the year of van der Molen's measurements.

Results from the sap flow velocity in the field

We measured sap flow velocity of different-sized individuals of major tree species at different elevations in the LEF in August of 2001 and in January of 2002 using the thermal dissipation method (Table 10). Sap flow velocity generally

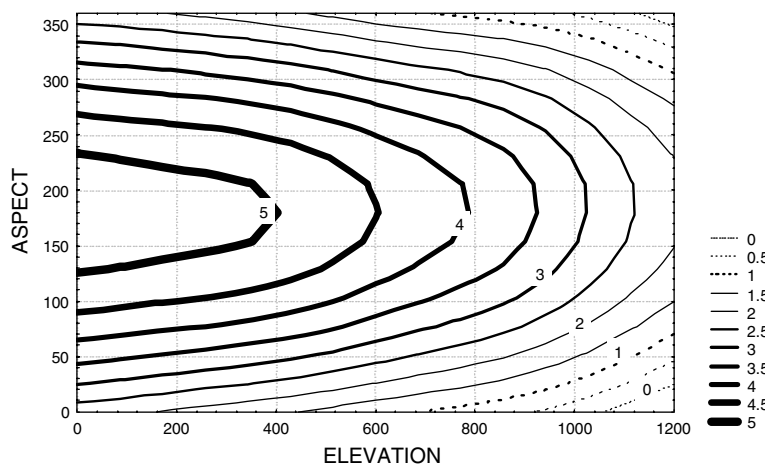


Figure 7 Contour plot between derived aET and elevation and aspect (aspect: $^\circ$ from the north, elevation: m; aET: mm/day).

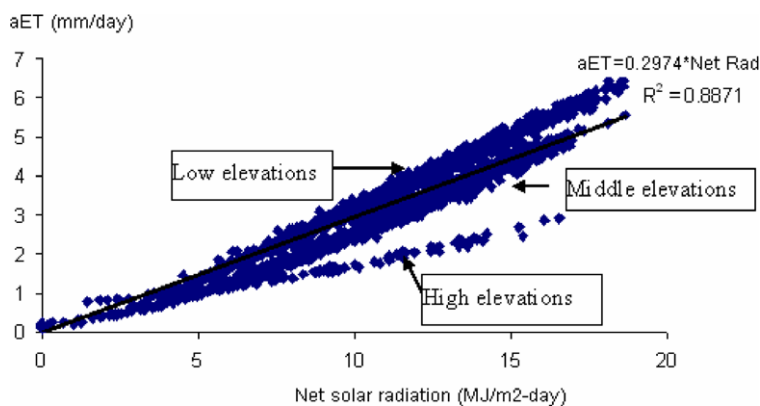


Figure 8 Modeled aET with modeled solar radiation.

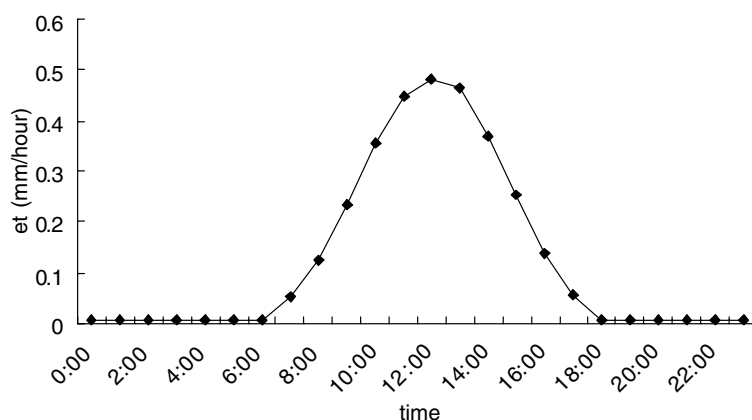


Figure 9 Derived hourly average aET in January over the whole LEF (unit: mm/hour).

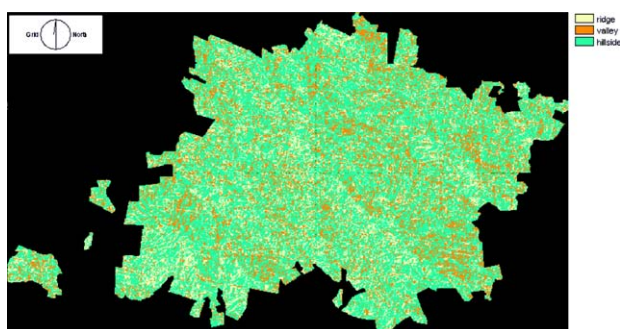


Figure 10 Topographic shapes in the LEF.

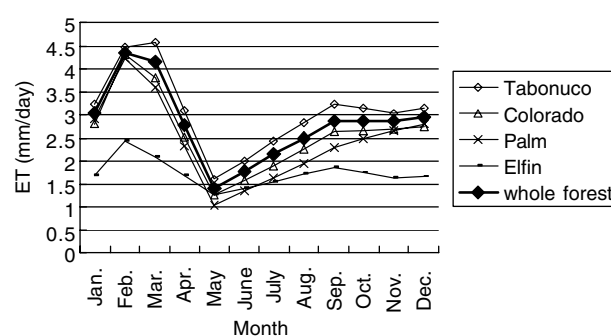


Figure 11 Monthly predicted aET in the LEF.

Table 9 Average aET at different topographic features in the LEF

Forest types	Topographic positions	Mean AET (mm/day)
Tabonuco forest	Ridge	3.71
	Valley	3.09
	Hillside	3.31
Colorado forest	Ridge	3.17
	Valley	2.61
	Hillside	2.81
Palm forest	Ridge	3.35
	Valley	2.49
	Hillside	2.99
Elfin forest	Ridge	1.89
	Valley	1.58
	Hillside	1.72

increased after sunrise, and reached a maximum between 10:00 a.m. and 3:00 p.m. (Figs. 13–16). The larger the tree size, the larger the sap flow velocity. The exception is the Colorado trees that were measured. In general the sap flow velocity in these trees was small and stable over a day. In addition, the few Colorado trees we measured were similarly sized, and there was no consistently clear relation be-

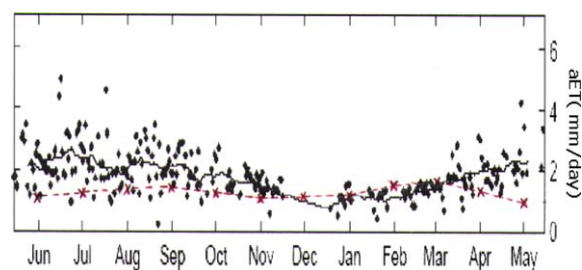


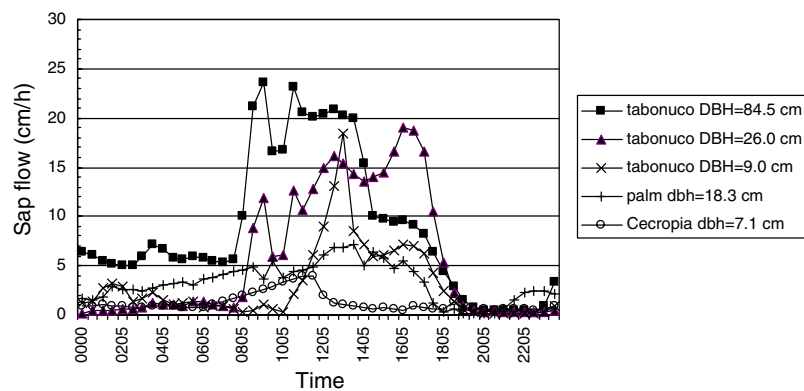
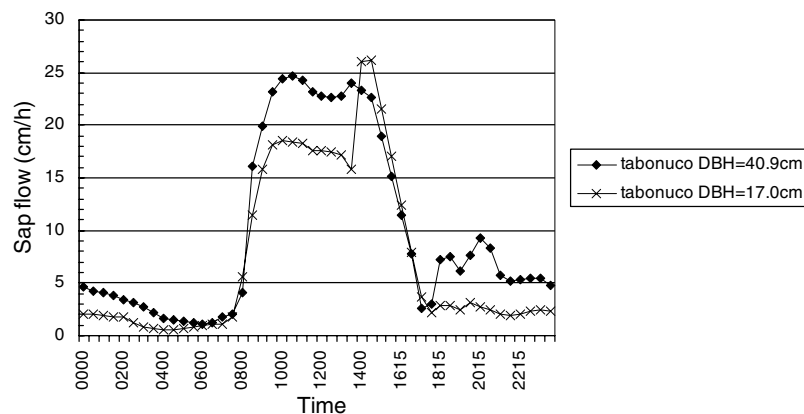
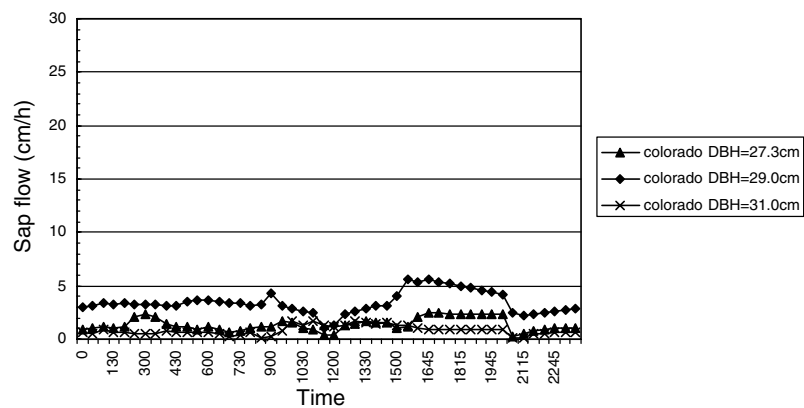
Figure 12 Monthly simulated and observed aET at van der Molen's sampling site in the palm forest (18°16'59" N, 65°46'05" W). (y axis: aET in mm/day; point represents observation data; solid line represents a 20-day moving average of the observations; x-dotted line represents simulated aET from my model).

tween tree size and sap flow velocity. One possible explanation for the small decrease of sap flow velocity of the Colorado tree at noon is that the sampled trees were in a relatively open area where they received strong sunlight at noon. This may have caused the stomata to close in order to keep water in the plants. In any case the field data from the Colorado trees was harder to reconcile with the model output.

We treated the data collected at the Bisley upper tower (about 385 m in elevation) and in the elfin forest (about 1000 m in elevation) in the LEF separately. The data

Table 10 Sampling sites

Time	Location
August of 2000 (including rainy days and sunny days)	Tabonuco forest near the El Verde meteorological station (18°19' 22" N, 65°49' 13" W, elevation = 350 m.a.s.l.); Colorado forest along 191 road; Cloud forest at Peak El Este
January of 2001 (including rainy days and sunny days)	Tabonuco forest at the Bisley upper tower (18°18'53" N, 65°44'39" W, elevation = 385 m.a.s.l.) Cloud forest at Peak El Este

**Figure 13** Measured sap flow velocity of different trees over a day in the tabonuco forest at Bisley upper tower in January of 2002.**Figure 14** Measured sap flow velocity of different trees over a day in the tabonuco forest near El Verde station in August of 2001.**Figure 15** Measured sap flow velocity over a day in the Colorado forest along road 191 in August of 2001.

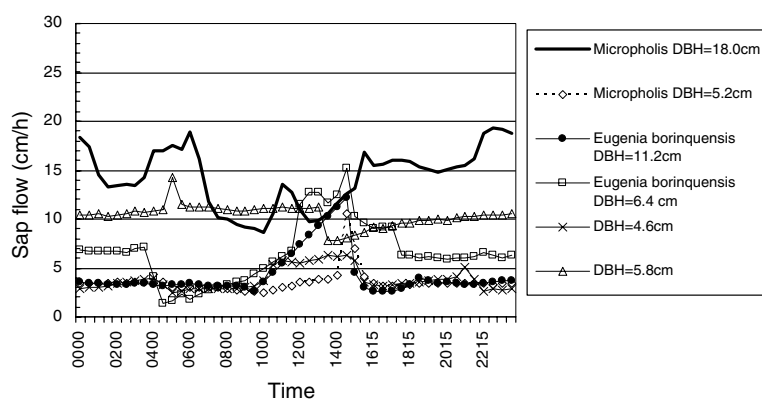


Figure 16 Measured sap flow velocity over the same day in the cloud forest in January of 2002.

collected in the winter of 2002 did not show a significant effect of different species at $\alpha = 0.1$ at the Bisley upper tower (p -value = 0.1065) or in the elfin forest (p -value = 0.5575). Diameter at breast height (DBH) had a significant effect on sap flow at $\alpha = 0.1$ at Bisley (p -value = 0.0363), but not in the elfin forest (p -value = 0.1864). This could be due to the small range of DBHs in the elfin forest, making it difficult to determine the effect of DBH. We derived a linear relation between sap flow velocity and DBH for the tabonuco forest and in the elfin forest respectively (Figs. 17, 18). So within each site, different species did not transpire

at significantly different rates, but trees with different DBHs had a significant difference in transpiration at the Bisley upper tower.

When all the species at different elevations were combined there was a significant effect of species (p -value = 0.0817) but not DBH (p -value = 0.7856) at $\alpha = 0.1$. However, if we added in the solar radiation factor in the ANOVA test, species became not significant (p -value = 0.2137), while the solar radiation was found to have a significant effect on sap flow velocity (p -value = 0.0471 at $\alpha = 0.1$) for all the trees we measured. Granger and Gray's model (1989),

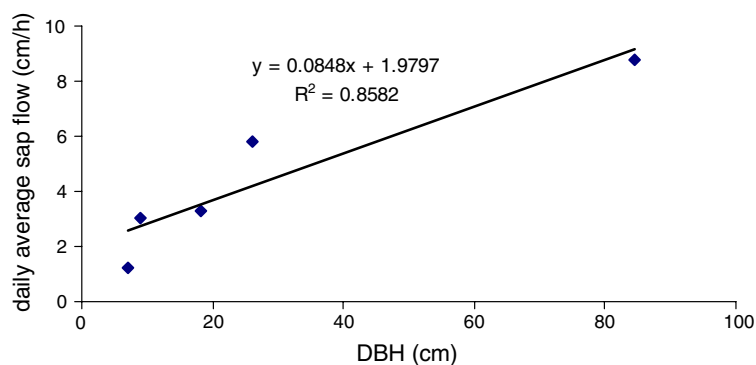


Figure 17 Measured sap flow velocity vs. DBH in the tabonuco forest.

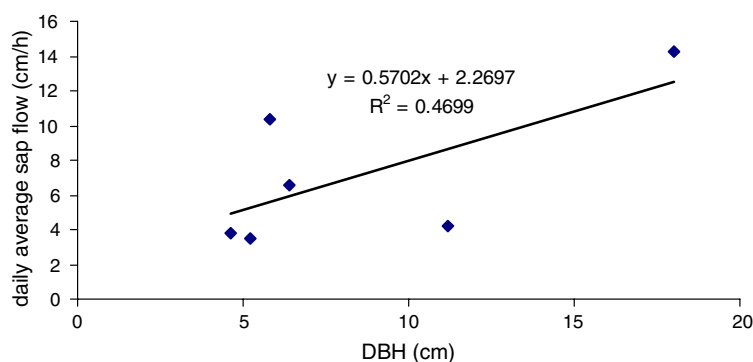


Figure 18 Measured sap flow velocity vs. DBH in the cloud forest.

which we used in this study, appears to quantify the factor of radiation correctly (Fig. 8).

Validation

The three situations given in Table 6 yielded very similar results, and since we could not differentiate them visually from the graph we plotted only the results from the high

shading situation (situation 2 defined in Table 6) (Figs. 19–21).

We also considered the uncertainty involved in the sapwood area used to scale aET from the model down to sap flow velocity of “average” trees. When we increased the values of the sapwood area by 20%, the daily average velocity of the sap flow of “average” trees derived from the model would decrease by 16.7%. When we decreased

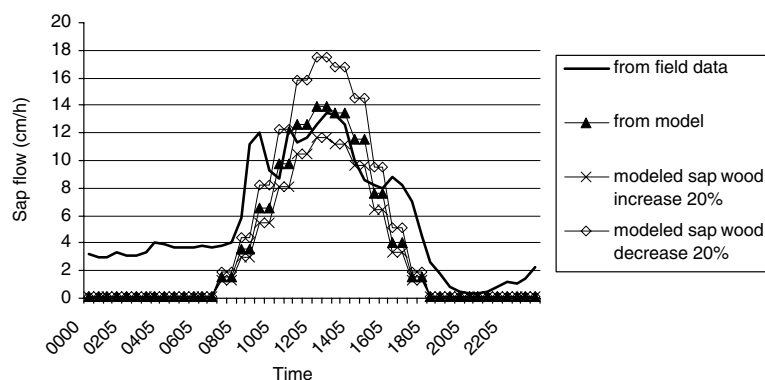


Figure 19 Comparing sap flow velocity of an “average” tree (DBH = 5.47 cm) derived from the field data and from the model at the bisley upper at tower (385 a.s.l.m) in the winter of 2002.

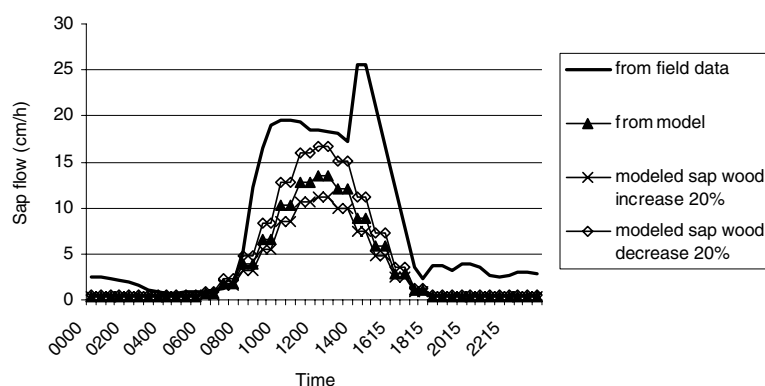


Figure 20 Comparing sap flow velocity of an “average” tree (DBH = 6.69 cm) derived from the field data and from the model near El Verde station (450 a.s.l.m.) in the summer of 2001.

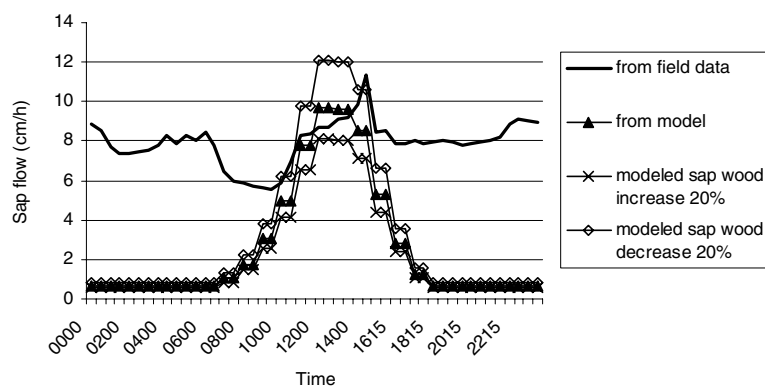


Figure 21 Comparing sap flow velocity of an “average” tree (DBH = 3.91 cm) derived from the field data and from the model near Pico del Este (990 a.s.l.m.) in the winter of 2002.

the values by 20%, sap flow velocity would increase by 25%.

The modeled ET agreed reasonable well with the measured and scaled up ET for the tabonuco forest, and moderately well with the data for the elfin forest during the day (Figs. 19–21). The model also captured the daily change of ET both in the tabonuco forest and in the elfin forest but was much less than the measured values at night in the elfin forest. Why the measured sap flow velocity during night in the elfin forest was large compared with the tabonuco and the Colorado forest is uncertain but may be related to the windy conditions of the elfin forest.

Overall, the model validation was very general at best. Since the relation of DBH and sap flow is not necessarily linear, the use of this relation perhaps causes an incorrect sap flow estimate from the modeled aET. Increasing the number of sampled trees with different sizes may help evaluate the relation better between the DBH and the velocity of sap flow, which can make scaling up and down more reliable.

Sensitivity analysis

Increasing or decreasing the probability of cloud cover by 10% in the model would reduce or increase simulated aET by 7.1% (Table 11). Usually, path radiance has a value from 0.025 to 0.04. We used 0.0325 in our model simulation. When we changed it to 0.025, the simulated aET decreased 5.5% (Table 11). When we changed it to 0.04, the simulated aET increased 2.6% (Table 11). When we increased (decreased) aerodynamic resistance in the model, the simulated aET at the tabonuco, Colorado and palm forests did not decrease (increase) as we expected, only the simulated aET in the elfin forest decreased (increased) (Table 11), which implied that aerodynamic resistance played a more important role in driving evapotranspiration in the elfin forest than at other forest types. Since the relation between G and D is an empirical one, we changed the empirical coefficients in the exponential term to 7.045 and 9.045 from 8.045. Decreasing the coefficient made the simulated aET increase in all forest types, and to a greater extent at the elfin (108%), Colorado (32.7%), and palm forests (30.7%), than at the tabonuco forest (13.4%) (Table 11). Increasing the coefficient made the simulated aET decrease, and also at a larger extent at the elfin (54.1%), Colorado (26.8%), and palm forests (25.3%) than at the tabonuco forests

(12.5%) (Table 11). This shows that the simulated aET in the wetter environment at the higher elevations in the LEF is more sensitive to the empirical coefficients, indicating that a relation between G and D in the wet environment ($G > 0.7$) is necessary.

Discussion

Our derived mean aET in January over the LEF (3.15 mm/day) is between Wang's (2001) derivation of annual aET (2.06 mm/day) using the Penman-Monteith equation and the water balance estimates of Brown et al. (1983) and Garcia-Martino et al. (1996) which were 4.8 mm/day and 4.7 mm/day, respectively. Our simulated aET at the Bisley watershed, was 2.87 mm/day, which was roughly comparable to the average transpiration rate of 2.0 mm/day in 1996, and 2.4 mm in 1997 at the Bisley II catchment, which were calculated by using a combination of the temperature fluctuation method and the Penman-Monteith equation from Schellekens et al. (2000). The simulated aET at El Verde, was 2.85 mm/day, which was similar to Odum and Jordan's (1970) conservative estimate of 3.0 mm/day. Our simulated aET also compared reasonably with ET ranging from 3.1 mm/day (Lance and Lesack, 1993; Schellekens et al., 2000) to 4.6 (Leopoldo et al., 1982; Schellekens et al., 2000) in the small forested catchments in Central Amazon Basin, another tropical forest site but not in maritime climate as the LEF (Table 12). The advantage of our method, which seems to give numbers similar to other methods, is that it is spatial.

Characterizing vegetation coverage such as leaf area index (LAI) in a tropical forest in relation to evapotranspiration is important for understanding how tropical ecosystems work and assessing how land use and land cover interact on the water yield of catchments. We calculated LAI using the empirical equation derived by Wang (2001). Evapotranspiration tends to be lower at higher elevations unless LAI is very high, although there is a curiously low LAI and hence aET rate at about 300 m in elevation (Fig. 22).

We calculated the ratio of ET to rainfall over the whole elevation range in order to quantify the relative importance of ET. Average annual rainfall was derived using a rainfall-elevation equation that was developed for the LEF by Garcia-Martino et al. (1996). The ratio of ET to rainfall varied from 0% to 89.3%, with a mean of 30.5% (Fig. 23 and Table

Table 11 Results from sensitivity analysis (G is relative evaporation)

aET (mm/day)	The whole forest	Tabonuco forest	Colorado forest	Palm forest	Elfin forest
Model (Path radiance = 0.035 $G = 1/(1 + 0.028 \times \exp(8.045 \times D))$)	3.08	3.28	2.84	2.96	1.72
Cloud cover (+10%)	2.86	3.07	2.61	2.71	1.61
Cloud cover (−10%)	3.30	3.49	3.08	3.21	1.84
Path radiance (low: 0.025)	2.91	3.12	2.67	2.78	1.64
Path radiance (high: 0.040)	3.16	3.36	2.93	3.05	1.76
Aerodynamic resistance (−20%)	2.89	3.14	2.59	2.71	1.76
Aerodynamic resistance (+20%)	3.21	3.36	3.04	3.14	1.74
$G = 1/(1 + 0.028 \times \exp(9.045 \times D))$	2.49	2.87	2.08	2.21	0.79
$G = 1/(1 + 0.028 \times \exp(7.045 \times D))$	3.79	3.72	3.77	3.87	3.58

Table 12 Comparisons between the modeled aET and evapotranspiration data from the literatures at different vegetation types in the LEF

	Mean (mm/day)	Estimate from the literature (mm/day)
Whole forest	3.15	2.06 Wang (2001) 4.8 Brown et al. (1983) 4.7 Garcia-Martino et al. (1996)
Tabonuco	3.43	2.92 Van der Molen (2002) 3.5–5.0 (El verde station in the tabonuco forest, Odum (1970), pp. H-5) 2.0 in 1996, 2.4 in 1997 (Bisley watershed in the tabonuco forest, Schellekens et al. (2000)
Colorado	2.93	2.86 Van der Molen (2002)
Palm	2.88	1.99 Van der Molen (2002)
Elfin	1.62	1.93 Van der Molen (2002)

13), which is similar to the 35% derived from Garcia-Martino et al. (1996). The higher the elevation, the lower the ratio, which corresponds to the wet soil in the elfin forest at high elevation. The ratio on the south slopes was higher than north slopes due to the higher solar radiation they received (Table 14).

Our estimates of the Bowen ratio increased from 0.397 in the tabonuco forest to 0.695 in the Colorado forest, to 0.727 in the palm forest (similar to Van der Molen's (2002) result in the Colorado and palm forests) and to 1.317 in the elfin forest. The Bowen ratio we estimated in the tabonuco forest was smaller than Van der Molen's 0.73 (2002). An average of 58.82% of the net solar radiation was used for evapotranspiration in the tabonuco forest, 46.08% in the Colorado forest and palm forest, and 36.67% in the elfin forest. The Bowen ratio varied across different vegetation types. Any change of vegetation, due to global warming or deforestation, will alter the Bowen ratio, thus will change the pattern of hydrological cycling.

Rainfall is distributed fairly evenly throughout the year in the LEF, although May and November are relatively wet and

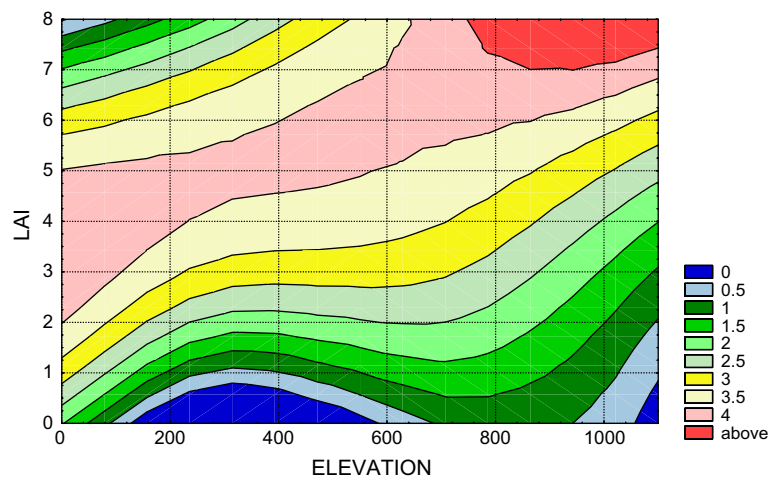
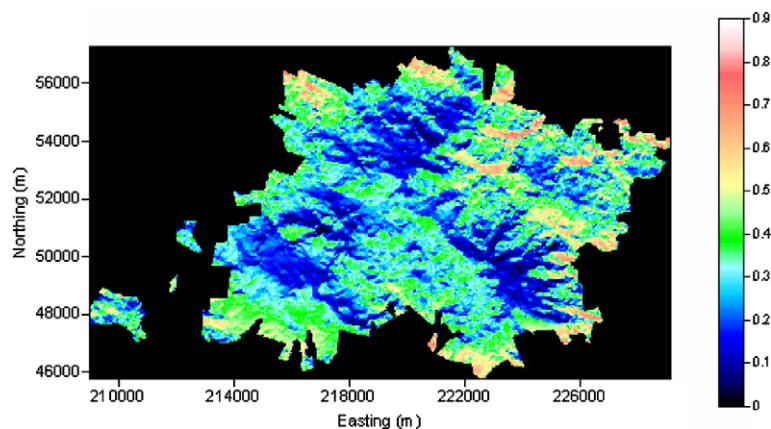
**Figure 22** Predicted aET as a function of elevation and LAI.**Figure 23** The ratio of aET to rainfall over the whole LEF.

Table 13 Average ratio of aET to rainfall at each forest type in the LEF

Forest types	Average ratio of aET to rainfall
Tabonuco	0.36
Colorado	0.26
Palm	0.25
Cloud forest	0.13

Table 14 Ratio of aET to rainfall at different aspects in the LEF

Aspect	Average of aET/rainfall
North	0.203
North-East	0.273
East	0.349
South-East	0.439
South	0.405
South-West	0.361
West	0.277
North-West	0.216

January–March are relatively dry (Schellekens et al., 2000). The higher net solar radiation available for evapotranspiration, the stronger drying power of the air and the absence of limiting soil moisture were the probable reasons that explained why the evapotranspiration in February and March was higher than in the other months.

Scaling is an important issue in spatial modeling and needs more field data to complete. The conversion from evapotranspiration per unit area to sap flow velocity of an “average” tree was dealt with in only a very general way here. More data points are needed especially for the sap flow velocity of large trees that contribute a significant part to evapotranspiration even though there are not many of them per unit area. Further research on the relations between sap flow velocity and DBH, between sapwood area and basal area, and the spatial autocorrelation of sap flow velocity is also necessary to find a more reliable scaling relation.

It is common to believe field data more than model results. However, both field work and modeling involve uncertainties. Fortunately we can quantify uncertainties in models, for example, by using sensitivity analysis described above. However, the uncertainties associated with field work tend not to be discussed and can be harder to evaluate. For example, in the thermal dissipation method, the data logger in the field recorded temperature differences between the heated probe and a reference probe. An empirical equation was then used to calculate sap flow velocities. However uncertainties in the temperature differences and the empirical equation are unknown and hard to evaluate. Ideally, it would be helpful to try different field methods to measure ET and validate and calibrate the spatial models. In this study in particular, additional measurements in the forests at high elevations including Colorado, palm and elfin forests are recommended.

The evapotranspiration model in this paper is based on the energy balance in the vertical direction above a horizontal surface, assuming advection is negligible. The universal characteristics of energy balance, the model’s spatial characteristics and the fact that simulation results agree well with other approaches applied to the LEF make the model ready to apply to other areas. However it is necessary to have cloud free Landsat images for that region (they can be mosaics) so the parameters associated with short wave radiation and long wave radiation can be derived. Since ET is a very important factor of hydrological cycling, the modeling practices of trying to obtain more accurate estimates of it will be very useful for watershed modeling, especially in the context of climate and land use changes.

Conclusion

Remote sensing can be a very useful tool in ecological modeling since the imagery covers large areas and can provide estimates at high spatial resolution, although each image is still a snap shot and the temporal resolution of different satellites varies. Field data are usually helpful in interpreting imagery, and parameterizing and validating models for large areas, although extensive field work is not necessarily required (Kite and Droogers, 2000), and the issues of scaling up and down can be complicated.

By comparing the modeled aET with the measured transpiration, we found the model behaved well in the tabonuco forest at low elevations but only moderately well in the elfin forest at high elevations during the day. Since the tabonuco forest occupies 70% of the LEF, this model captured the main characteristics of aET in the LEF. Because the empirical data during night in the elfin forest made little sense we trust the model more than the data. The variation of the Bowen ratio across different vegetation types implies that a change in vegetation will induce a change in the distribution of solar radiation, thus a change of aET, and therefore affect water yield in the stream and the water resource available to humans. Since high aET occurs in the tabonuco forest at low elevations, and land use change is more likely to occur at lower elevations where people have easier access than at higher elevations, land use change can bring a big change of hydrological cycles.

No single method of measuring ET is without problems. In addition, some other methods, such as the water balance equation, cannot be applied spatially. We have introduced fairly new spatial techniques to modeling and experimentation for the tropics that seem to give good spatial results and that are similar to other approaches in the same area. We believe that it is ready to apply to other areas.

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