

Characteristics of fog and fogwater fluxes in a Puerto Rican elfin cloud forest

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Abstract

The Luquillo Mountains of northeastern Puerto Rico harbours important fractions of tropical montane cloud forests. Although it is well known that the frequent occurrence of dense fog is a common climatic characteristic of cloud forests around the world, it is poorly understood how fog processes shape and influence these ecosystems. Our study focuses on the physical characteristics of fog and quantifies the fogwater input to elfin cloud forest using direct eddy covariance net flux measurements during a 43-day period in 2002. We used an ultrasonic anemometer–thermometer in combination with a size-resolving cloud droplet spectrometer capable of providing number counts in 40 droplet size classes at a rate of 12.5 times per second. Fog occurred during 85% of the time, and dense fog with a visibility < 200 m persisted during 74% of the period. Fog droplet size depended linearly on liquid water content ($r^2 = 0.89$) with a volume-weighted mean diameter of 13.8 μm . Due to the high frequency of occurrence of fog the total fogwater deposition measured with the eddy covariance method and corrected for condensation and advection effects in the persistent up-slope air flow, averaged 4.36 mm day^{-1} , rainfall during the same period was 28 mm day^{-1} . Thus, our estimates of the contribution of fogwater to the hydrological budget of elfin cloud forests is considerable and higher than in any other location for which comparable data exist but still not a very large component in the hydrological budget. For estimating fogwater fluxes for locations without detailed information about fog droplet distributions we provide simple empirical relationships using visibility data.

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1. Introduction

The frequent occurrence of dense fog is a common climatic characteristics of so-called cloud forests

around the world (Hamilton et al., 1995). However, the way fog shapes and influences cloud forest ecosystems is poorly understood (Bruijnzeel and Veneklaas, 1998). Compared with montane rain forests below the cloud belt, cloud forests tend to be mossier (Frahm and Gradstein, 1991), shorter-statured and more xerophyllous (Grubb, 1977). All of these characteristics are generally thought to be influenced by the frequent occurrence of fog although it is still not fully understood

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how and to what extent fog achieves this (Bruijnzeel and Veneklaas, 1998). Several decades ago, the German geobotanist Heinrich Walter expressed his doubts as to whether fog really is the dominant environmental factor in structuring cloud forest ecosystem as a whole, or only is important for the survival of certain types of epiphytes and mosses on trees (Walter, 1973). In recent decades, the hydrological importance of fog deposition as a source of sustained dry season base flows from cloud forested headwater areas has been emphasized repeatedly (Zadroga, 1981; Brown et al., 1996; Bruijnzeel, 2001).

The process of fogwater deposition is driven by impaction and interception of cloud droplets by the vegetation. Cloud droplets in fog follow the turbulent motion of the air in which they are dispersed (see La Porta et al., 2001; Voth et al., 2002). As soon as they touch the surface of a plant leaf, a stem of a tree, or any other obstacle they are intercepted and form a thin film on that obstacle. If the contact is more violent, the more correct term is impaction, with the same result that the droplet becomes incorporated into the water film of the wet canopy. Using the eddy covariance method we quantified the resulting ecosystem-scale fogwater deposition. Our results suggest that measured fog characteristics and deposition rates in a wet tropical montane elfin cloud forest are important for the hydrological budget. However, as long as there is abundant rainfall the additional input by fogwater deposition cannot be considered a very large budget component in terms of its amount. We therefore suggest to focus rather on the timing and duration of fog (affecting tree physiological functioning, including the light climate and soil water uptake), and on its role as a source of dissolved nutrients in generally nutrient-poor tropical mountain ecosystems (cf. Hafkenschied, 2000).

Mountainous areas are among the most diverse ecosystems on Earth (Orme et al., 2005) and tropical mountain cloud forests are known for their high share of endemic plants and animals (Bruijnzeel, 2001; Leo, 1995; Kappelle and Brown, 2001). Cloud forests have been reported to be very susceptible to changes in environmental conditions (Lawton et al., 2001; Pounds et al., 1999, 2006). The conservation of the remaining tropical mountain cloud forests is therefore one of the primary goals of the UNEP Cloud Forest Agenda (Bubb et al., 2004).

Within the Long-Term Ecological Research (LTER) network of the U.S.A., the Puerto Rican Caribbean National Forest in the eastern Luquillo Mountains harbours important fractions of tropical montane cloud forests, both of the montane and elfin cloud forest type

(Bruijnzeel, 2001) on its upper most peaks. Scientific investigations in the elfin cloud forest began as early as the late 1960s (Baynton, 1968, 1969; Brown et al., 1983). The area became part of the LTER network in 1988 and observations are still ongoing (<http://luq.lternet.edu>). Here we report the first direct measurements of fog droplet distributions and their turbulent exchange with the forest canopy conducted in any tropical montane cloud forests of the world. Since our results compare well with measurements from other mountain areas outside of the tropical zone, we provide more generally valid recommendations on how to empirically estimate fogwater deposition based on easily measured variables such as horizontal visibility.

2. Materials and methods

Fog is not defined equally by all scientific disciplines or researchers. Here we adhere to the following definition: fog occurs if the horizontal visibility is less than 1000 m (this is the most common meteorological definition according to Glickman, 2000), and the cloud droplets in the air reducing the visibility are less than 200 μm in size (an extension advanced by Glickman, 2000). The first criterion was determined in the present study with a forward-scattering instrument (Section 2.2.3), the second criterion was based on measurements made with a high-speed cloud droplet spectrometer (Section 2.2.1). All statistical analyses were performed with R version 2.0.0 (R Development Core Team, 2005).

2.1. Site description

Our measurements took place near Pico del Este ($65^{\circ}45'39''\text{W}$, $18^{\circ}16'17''\text{N}$, 1015 m a.s.l.) in the Luquillo Experimental Forest (LEF), in the northeastern part of Puerto Rico. Measurements were made ≈ 100 m below the crestline of a 17° slope of east-northeasterly aspect directly facing the prevailing winds (Brown et al., 1983; see also map shown in Fig. 1). Accordingly, the site is strongly influenced by the northeasterly trade winds (Garcia-Martino et al., 1996). Furthermore, occurrence of fog is very frequent due to the forced lifting and cooling of humid air masses arriving from the Atlantic Ocean (Baynton, 1969; Holwerda et al., 2006). Over the roughly 20 km transect from the Atlantic coast to the site there is a dramatic change in forest types, ranging from subtropical dry forest near the coast through lower montane rain forest (Tabonuco) to montane palm brakes, upper montane rain forest (Colorado), and elfin cloud forest (Brown et al., 1983). The vegetation at Pico del

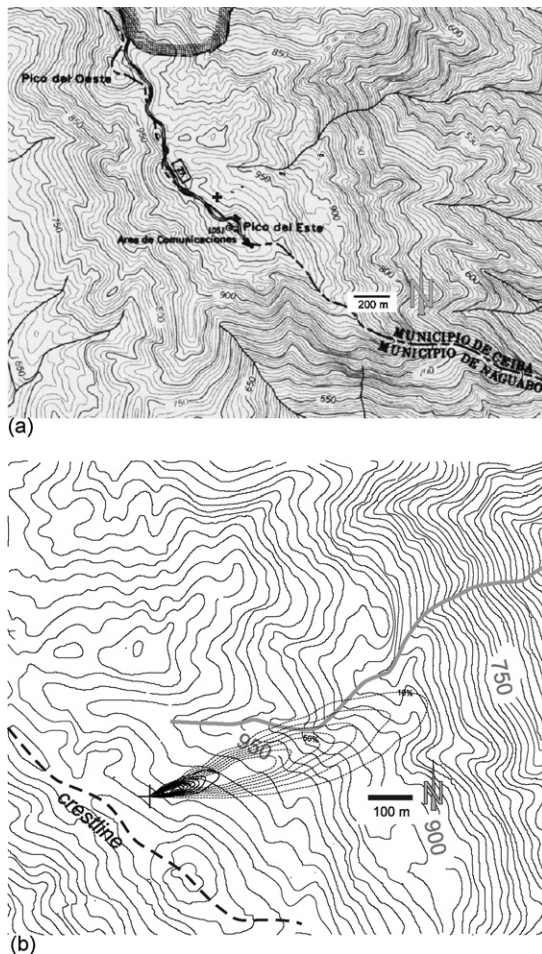


Fig. 1. Topographic map of the flux site (top) with the position of the eddy covariance tower shown by a black cross. Contour interval is 10 m. The flux footprint area computed with the Kljun et al. (2004) model for all cases with fog is shown on the magnified contour map at bottom. Solid isolines show the footprint area of liquid water flux measurements, whereas broken isolines show the corresponding area for momentum flux measurements, computed for a virtual height that is four times the true local measurement height. Isolines are drawn at 10% intervals from 10% to 70% of the maximum contribution found in the footprint area.

Este consists of elfin cloud forest. The average height of the canopy is 2.5–3 m, with a maximum height of individual trees of about 5.25 m. Leaf area index (LAI) as derived from light extinction measurements was 2.1 ± 0.7 (Holwerda et al., 2006) and close to the value of 2.0 derived for the same forest using destructive harvesting methods (Weaver et al., 1986). The dominant tree species include *Tabebuia rigida*, *Ocotea spathulata*, and *Calyptanthes krugii*. The forest is rich in epiphytes, with mosses and bromeliads covering branches and stems and parts of the soil surface (Weaver, 1995). The horizontal distances to the nearest public road and small

town are 2 and 5 km, respectively. The site can be considered an undisturbed site (cf. McDowell et al., 1990) without human management and without canopy openings. The homogeneous slope of 17° extends over roughly 800 m with a very smooth transition from higher-statured vegetation at lower elevations towards the tower. No abrupt changes in vegetation roughness upwind of the tower were present, but at roughly 1 km distance the slope angle changes to a steep but also rather uniform 35° towards the coastal plain. A topographic map of the surroundings can be found in Fig. 1.

2.2. Instrumentation

During an intensive measurement campaign of 43 days between 26 June and 7 August 2002 an eddy covariance system (Eugster et al., 2001; Burkard et al., 2002, 2003) and standard meteorological sensors were installed on top of a 6 m high meteorological tower above the canopy of the elfin cloud forest (total height 7.25 m from ground level, 4.58 m above aerodynamic displacement height, which corresponds to roughly 2–3 times mean canopy height). The tropical climate at this site is strongly dominated by the diurnal cycle of weather conditions and is only subject to a rather small seasonality in climate (e.g. Brown et al., 1983; Holwerda, 2005). Therefore, our 43-day period of measurements can be considered to be sufficiently representative of average conditions at this site.

2.2.1. Direct measurements of fog droplet distribution

A high-speed cloud droplet spectrometer (model FM-100, Droplet Measurement Technologies Inc., Boulder, CO, U.S.A.; see Fig. 2) was used to measure the size spectrum of fog droplets. The operating principle of the FM-100 is the following: a strong fan pulls air isokinetically at a constant rate of around 13 m s^{-1} perpendicularly through a laser beam. In order not to disturb the wind field, the fan was placed in a weatherproof box on the ground from which a vacuum cleaner hose roughly 10 m long was connected to the back of the FM-100 (in Fig. 2 this looks like a tail). In detail, the vacuum hose was attached to both ends of a metal tube that acted as the rotating axis of the turntable motor. The instrument detects the number and size of individual cloud droplets by the forward scattering principle (Bruijnzeel et al., 2005). The basic assumption is that the amount of light scattered by a particle is proportional to its size, composition and shape. The FM-100 allows the counting and the characterization of fog droplets with

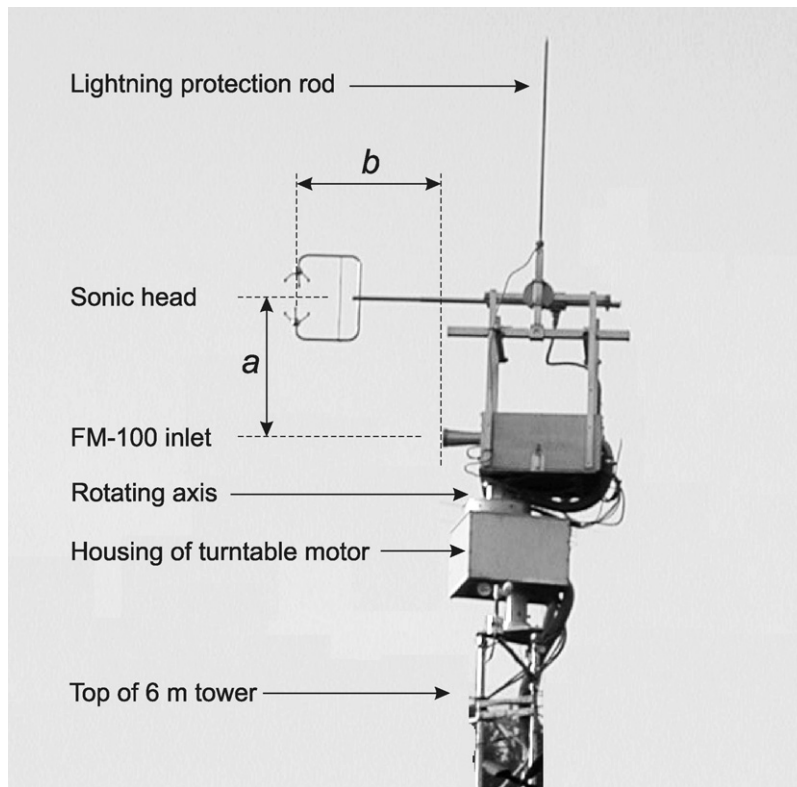


Fig. 2. Photograph showing the eddy covariance instrumentation from the side as it is aligned automatically with the mean wind. The vertical separation between the inlet of the fog droplet spectrometer (below) and the ultrasonic anemometer (above) is $a = 54$ cm and $b = 56$ cm. There is no lateral separation.

a diameter from roughly $1\text{--}2\text{ }\mu\text{m}$ up to $50\text{ }\mu\text{m}$ in up to 40 user-selectable size classes (Burkard et al., 2002). The calibration of the instrument was carried out by the manufacturer using glass beads of various sizes (7.8, 15.4, 19.9, 20.6, and $40.0\text{ }\mu\text{m}$). The difference in optical properties of glass beads (refractive index of 1.58) as compared to water was taken into account in the calibration process. In this way no size-dependent sampling bias is expected to confound the measurements.

2.2.2. Direct turbulent flux measurements of fog droplets

Fogwater fluxes were determined by means of the eddy covariance method, which combines a three-dimensional ultrasonic anemometer (Solent HS, Gill Ltd., Lymington, U.K.) with the FM-100 spectrometer. We used an upgraded anemometer with improved sensor stability and with firmware 3.00 2076_300 to enable the system to cope better with the problems of sonic anemometry under conditions with fog and heavy rain. The ultrasonic anemometer and FM-100 spectrometer were mounted on a turntable to align the

instruments with the mean wind speed every 30 min to minimize flow distortion (Burkard et al., 2002). Fig. 2 shows the two instruments on the turntable. A four-bit digital signal was used to adjust the directional position of the instruments such that the maximum angle between the FM-100's nozzle and the true mean wind direction was always less than 12° .

The raw data collection rate was set at 12.5 Hz which is the maximum continuous sampling rate sustained by the FM-100. Both instruments were attached to two serial ports on a laptop computer. The continuous data from the sonic anemometer were used as a time reference, and the FM-100 data records were synchronized with the anemometer data. Both data streams were merged during data acquisition before saving the merged data onto hard disk. Data processing included the following steps. First, the extensive house-keeping variables of the FM-100 were analyzed and spurious data rejected. This was occasionally the case when for example the apparent shape of a droplet was too far from the idealized spherical form (as is unavoidable if two droplets pass the laser beam within a distance that is shorter than their size). Single records of such rare

occurrences were replaced by the previously measured accepted record, thereby reducing the true sampling rate of the FM-100 to somewhat below 12.5 Hz. Next, the liquid water content $\rho_{L,d}$ for each of the 40 droplet size classes was computed based on an idealized mean volume of spherical droplets with aerodynamic diameter D . For each size class the geometric mean volume computed from the lower and upper D boundaries of the size class was multiplied by the number of droplets counted in the respective sampling interval to yield $\rho_{L,d}(D)$. The sum of all size classes then yielded the total liquid water content LWC or $\rho_L = \sum \rho_{L,d}(D)$. The turbulent flux of fogwater $F_{L,t}$ was then determined as

$$F_{L,t} = \overline{w' \rho_L'}, \quad (1)$$

where w is the wind speed component perpendicular to the mean streamlines of the atmospheric flow. Overbars denote the time average (30 min in this case), and primes denote the instantaneous or turbulent deviation from that average, thus $w' = w - \bar{w}$ and $\rho_L' = \rho_L - \bar{\rho}_L$ for every record within an averaging interval. Negative values of w and flux densities denote the direction towards the vegetation surface, whereas positive values indicate the reverse.

Before being able to use Eq. (1) a few additional data preparation steps were necessary. (1) The three-dimensional wind vector data were subjected to a coordinate rotation procedure that aligned the u -component with the local streamlines (which are along the prevailing wind direction) for each 30-min interval (the first two rotation steps of [McMillen, 1988](#)), and which forced the lateral v and surface-normal w components to be zero on average (i.e. $\bar{v} = 0$ and $\bar{w} = 0$). (2) To determine the lag time between ρ_L and w we performed a cross-correlation computation with w and ρ_L , then shifted the ρ_L time series relative to the wind speed series to eliminate possible time delays caused by the FM-100 data processing and delivery system and also to account for possible effects of sensor separation (see [Appendix A](#) for details). All these computations are analogous to standard eddy covariance flux measurements of trace gases (see e.g. [Baldocchi, 2003](#)). No detrending and no autoregressive filtering were used in the computation, and fluxes were computed as block averages for each 30-min interval. Work by [La Porta et al. \(2001\)](#) and [Voth et al. \(2002\)](#) has shown that under fully developed turbulence fluid particles of 46 μm follow the trajectories of a zero-size particle within a few percents, such that we do not expect any significant inertia effects of the droplet sizes measured by the FM-100 (up to 50 μm).

Since the FM-100 provides dynamic and static pressures (measured with a Pitot tube) together with temperature inside the air stream, the true density of air could be computed for each LWC sample. Therefore, no additional correction following [Webb et al. \(1980\)](#) was necessary. The net flux component of this turbulent LWC flux is mainly a result of removal of droplets by impaction to the vegetation surface below the sensors. There have been earlier attempts by others to estimate droplet fluxes via the quantification of impaction (e.g. [Collett et al., 1991](#)) or the differential measurements upwind and downwind of a group of trees (e.g. [Asbury et al., 1994](#)) whereas our approach should be able to quantify the same droplet removal mechanisms using the eddy covariance technique. However, for larger cloud droplets the settling velocity becomes additionally relevant, and thus we determined the additional settling flux of these larger cloud droplets from

$$F_{L,s}(D) = -v_d(D) \rho_{L,d}(D), \quad (2)$$

where D is the geometric mean droplet size per size class of the FM-100, $v_d(D)$ the settling velocity for spherical droplets with an aerodynamic diameter D according to Stoke's law, and $\rho_{L,d}(D)$ is the liquid water content of the respective size class. The total liquid water flux from fog droplets F_L is the sum of turbulent and settling fluxes

$$F_L = F_{L,t} + F_{L,s} = F_{L,t} + \int_{D=0 \mu\text{m}}^{50 \mu\text{m}} F_{L,s}(D) dD, \quad (3)$$

where the last term is approximated by the summation of the 40 discrete size classes up to 50 μm measured by the FM-100. Note that it is extremely difficult to determine the minimum resolvable size of droplets with the forward scattering technique, but as our results will show this is not of any significance for the present analysis.

2.2.3. Ancillary meteorological measurements

In addition to these high-frequency eddy covariance data we also recorded the outputs of various ancillary sensors with a data logger (Campbell Scientific, Inc., model CR10X) at a lower sampling rate of 0.1 Hz which were then stored as 10-min averages. To detect the presence or absence of fog and to determine its density (generally expressed as horizontal visibility) a Present Weather Detector (model PWD11, Vaisala OY, FI) was mounted at 5 m height. A detailed description of this device can be found in [Nylander et al. \(1997\)](#). A Kipp & Zonen CNR1 net radiometer was used to measure global

radiation ($R_{s,\downarrow}$), reflected short-wave radiation ($R_{s,\uparrow}$), plus incoming and outgoing long-wave radiation ($R_{L,\downarrow}$, $R_{L,\uparrow}$). These four sensors were heated to minimize condensation on the glass domes, but the most critical issue with radiation measurements under such wet tropical conditions was the slow but unavoidable growth of algae on all instruments, including the glass domes. The photosynthetically active photon flux density (PPFD) was measured using a Skye SKP215 PAR quantum sensor. Air temperature (T_a) and relative humidity (h) were measured by a Rotronic hygrometer–thermometer (MP100A, with radiation protection shield). Air pressure (p) was measured using a Vaisala PTB101B analog barometer, whereas wind speed (u) and wind direction were determined with a Vector Instruments A100R switching cup anemometer and W200P potentiometer windvane, respectively. All ancillary instruments were mounted at 5 m above ground level.

3. Results and discussion

During the 43-day measurement period (Table 1) the elfin cloud forest was immersed in fog for 85% of the time. For 74% of the time the fog was so dense that visibility was less than 200 m (Fig. 3). This persistent fog resulted from the low cloud condensation lifting level (typically between 600 and 800 m above mean sea level) in this humid environment. Only during the middle of the daytime and in the early afternoons did the fog tend to disappear (Figs. 3 and 4), although not on all days (cf. Holwerda, 2005). The disappearance of fog may have two reasons: due to the diurnal warming of the

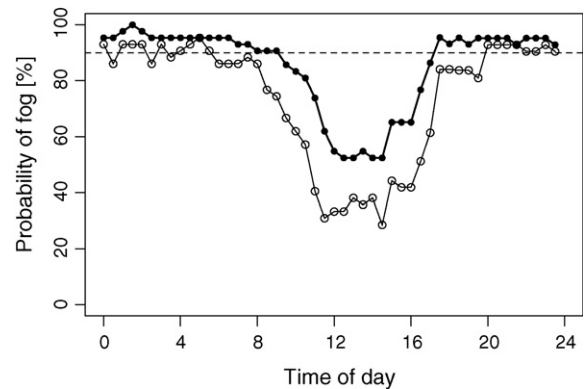


Fig. 3. Diurnal cycle of probability of fog (solid circles) and dense fog (visibility < 200 m, open circles) near Pico del Este, Puerto Rico, between 26 June and 7 August 2002 using 30-min averages. Broken horizontal line: 90% probability of fog occurrence.

air (Fig. 4b), fog droplets can evaporate (e.g. Leizerson et al., 2003), but sometimes synoptic weather conditions can bring in drier air with only partial cloudiness within the overall northeasterly flow (median wind direction 74° , with 80% of values in the range 66 – 86°).

The diurnal cycle of the probability of fog (Fig. 3) and its density (Fig. 4a) is also reflected in the (median) diurnal cycle of directly measured droplet flux above the cloud forest canopy (Fig. 5). It is remarkable that even around noon time, when fog is least frequent, there is still a significant liquid water input by fog. The reduction of the flux during that time is a result of the lower ρ_L (Fig. 4c) and smaller droplet sizes (Fig. 4d) when fog is less dense. Since ρ_L and volume-weighted mean droplet diameter D_v are closely related to one

Table 1

Mean values observed between 26 June and 7 August 2002 above an elfin cloud forest near Pico del Este, Puerto Rico

Variable (unit)	24 h	Daytime ^a	Nighttime ^b
Global radiation (W m^{-2})	116.0	212.1	3.0
Reflected short-wave radiation (W m^{-2})	16.3	27.7	3.2
Net radiation (W m^{-2})	92.9	176.1	−3.6
PPFD ($\mu\text{mol m}^{-2} \text{s}^{-1}$)	241.1	435.9	11.9
Air temperature ($^\circ\text{C}$)	20.5	20.8	20.1
Pressure (hPa)	909.2	909.4	909.0
Wind speed (m s^{-1})	5.86	5.31	6.52
Volume weighted mean diameter ^c (μm)	13.8	12.4	15.1
Visibility during fog ^c (m)	165.9	200.2	133.5
Cloud liquid water content in air ^c (g m^{-3})	0.079	0.061	0.095
Gaseous water content in air ^c (g m^{-3})	18.3	18.6	18.1
Deposition rate during fog events ^c (mm h^{-1})	0.21	0.19	0.24
Total fogwater deposition (mm day^{-1})	4.36	4.42	4.31
Total rainfall (mm day^{-1})	28.0	13.7	14.3

^a PPFD > $20 \mu\text{mol m}^{-2} \text{s}^{-1}$.

^b PPFD ≤ $20 \mu\text{mol m}^{-2} \text{s}^{-1}$.

^c Mean computed for conditions with fog (visibility < 1000 m).

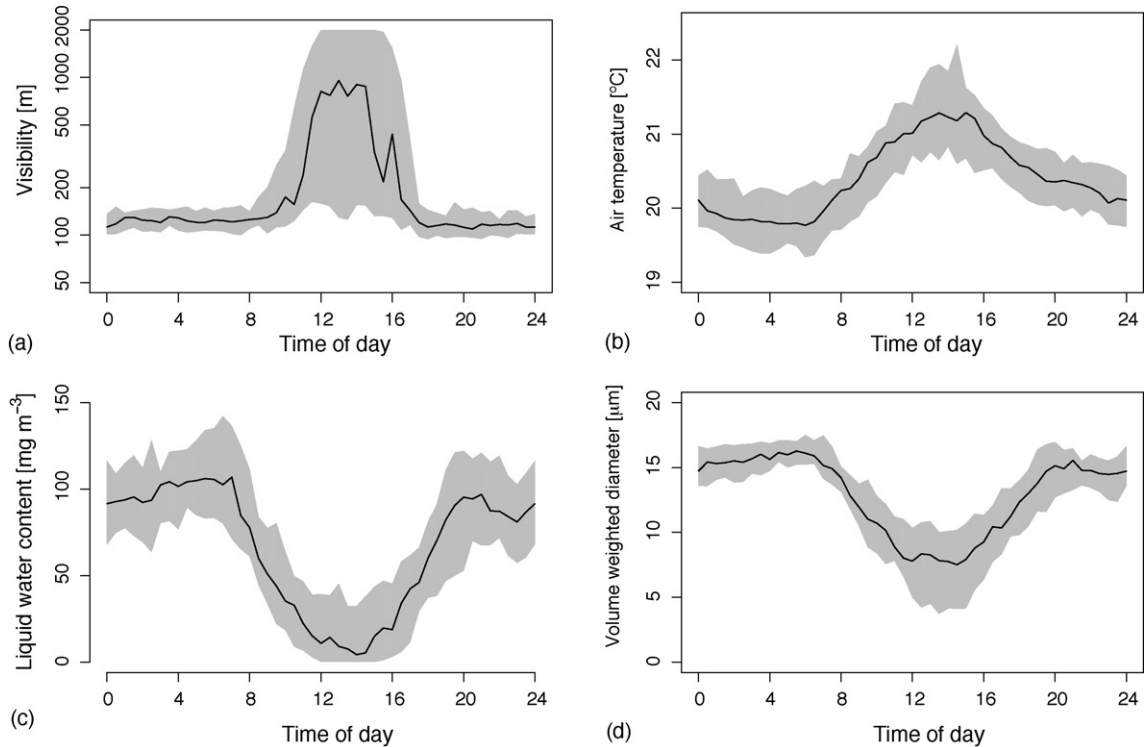


Fig. 4. Median diurnal cycle (solid lines) of: (a) visibility, (b) air temperature, (c) liquid water content, and (d) volume-weighted mean droplet diameter. Ranges given by the first and third quartiles (shaded area) are based on 30-min averages. In the absence of fog the Present Weather Detector reports a visibility of 2000 m.

another (Fig. 6), a simpler device than the FM-100 that is still capable of measuring total LWC would also allow to estimate D_v using the best fit relationship shown in Fig. 6a,

$$D_v = \exp[(1.23 \pm 0.01) + (0.327 \pm 0.003) \ln(\rho_L)], \quad (4)$$

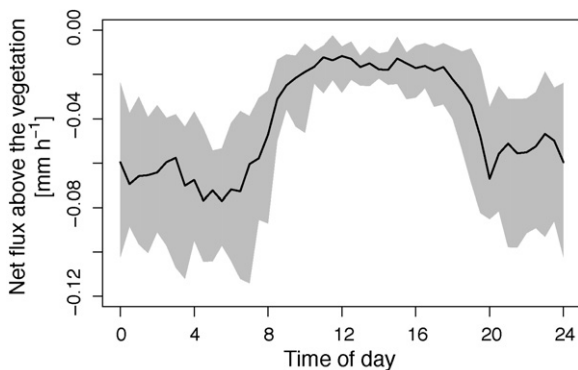


Fig. 5. Median diurnal cycle of fogwater flux measured above the elfin cloud forest canopy at Pico del Este (solid line). Range given by the first and third quartiles (shaded area) is based on 30-min averages. Negative fluxes denote a fogwater flux from the air to the canopy.

with ρ_L in mg m^{-3} and D_v in μm , adjusted $r^2 = 0.89$, $P < 0.001$, $n = 1611$. In contrast, although ρ_L and visibility X are also related to each other in a hyperbolic way (Kunkel, 1984), the scatter in the data is much larger, especially when fog is dense (Fig. 6b),

$$\rho_L = \frac{7120}{X - 2.515}, \quad (5)$$

adjusted $r^2 = 0.44$, $P < 0.001$, $n = 1611$. Thus, a conversion from visibility to ρ_L is only sufficiently meaningful for $X > 200$ m. However, because visibility estimates are much easier to perform than ρ_L measurements, we will present several simple approaches in Section 3.2 to use Eqs. (4) and (5) as efficiently as possible.

The fogwater flux per droplet size class (Fig. 7) showed clearly that droplet sizes below $7 \mu\text{m}$ and those at the upper end of the range resolved by the FM-100 do not contribute to the overall flux. It is therefore of no concern that the FM-100 is not capable of very accurately detecting the smallest droplet sizes. Moreover, Fig. 7 strongly suggests that a cloudwater collector capable of collecting droplets $> 7 \mu\text{m}$ is an ideal instrument for ion flux measurements since there is no

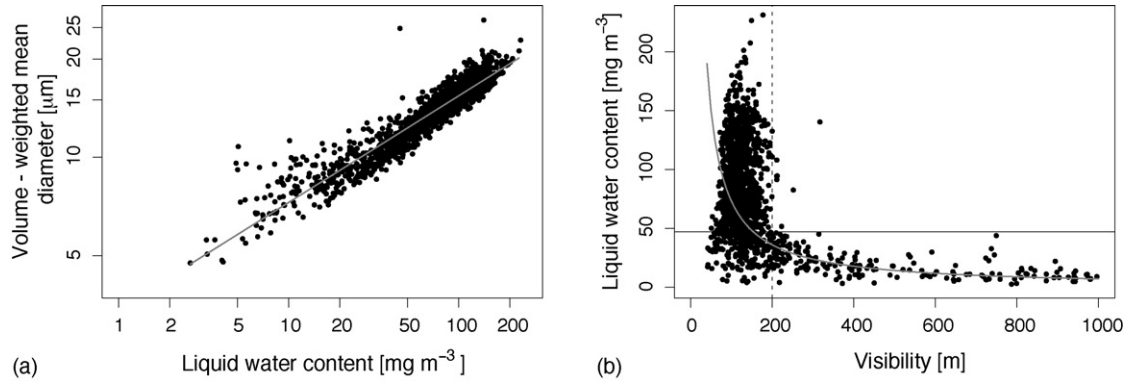


Fig. 6. (a) Volume-weighted mean diameter as a function of total liquid water content of fog and (b) relationship between liquid water content and visibility using 30-min averages. At visibilities below 200 m (broken vertical line), liquid water content does not exceed 47 mg m^{-3} in 95% of all observed cases.

detectable flux from smaller droplets in this environment.

3.1. Changes in radiation fluxes due to fog

The effect of fog is clearly seen in the diurnal course of net long-wave radiation (Fig. 8a) and the short-wave albedo (Fig. 8b) which is strongly dependent on light level, solar angle, and fraction of direct versus diffuse light (Monteith and Unsworth, 1990). The albedo is highest (> 0.13) when the fog is densest, and lowest in the absence of fog (≈ 0.09 , Fig. 8b). Thus, when light levels are already low during dense fog, there is a 44% greater loss of short-wave radiation via surface reflection. Not only is there more reflected radiation, but also incoming short-wave radiation is reduced by a similar proportion under foggy conditions. During dense fog (visibility $< 200 \text{ m}$) the photosynthetically active

photon flux density (PPFD, wavelengths 400–700 nm) is only $47 \pm 2\%$ of values observed at the same time of day under fogfree conditions (Fig. 9). The corresponding value for global radiation is $48 \pm 2\%$ (data not shown). The ratio of actual PPFD (Q_p) versus PPFD under fogfree conditions ($Q_{p,0}$) shows a log-linear relationship during peak daytime (10:30–15:00 h) as follows (Fig. 10)

$$Q_p/Q_{p,0} = (-0.18 \pm 0.19) + (0.14 \pm 0.03) \ln(X), \quad (6)$$

adjusted $r^2 = 0.62$, $P < 0.001$, $n = 20$. Similar values are found for global radiation $R_{s,\downarrow}$ (Fig. 10; wavelengths 300–3000 nm) with

$$R_{s,\downarrow}/R_{s,\downarrow,0} = (-0.21 \pm 0.19) + (0.14 \pm 0.03) \ln(X), \quad (7)$$

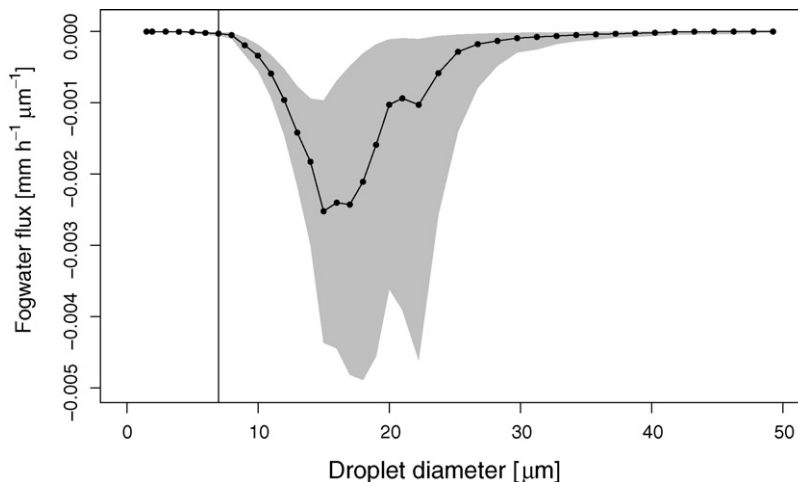


Fig. 7. Size-resolved median fogwater fluxes (solid line) in 40 size classes. Symbols represent the geometric centre of each size class. The shaded area shows the interquartile range of all 30-min flux values.

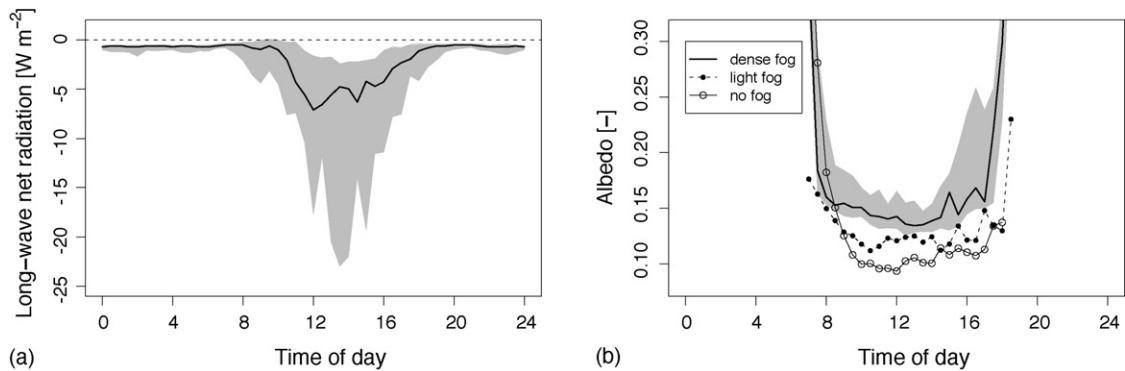


Fig. 8. Median diurnal cycles (solid lines) of (a) long-wave net radiation and (b) daytime short-wave albedo under conditions of dense fog (visibility < 200 m), and the interquartile range (shaded area) using 30-min averages. Additional lines in (b) indicate the albedo under fogfree (open circles) and light fog conditions (visibility 200–1000 m, filled circles). No albedo was computed when reflected short-wave radiation was < 10 W m^{-2} .

adjusted $r^2 = 0.64$, $P < 0.001$, $n = 20$. In combination, the reduction in incoming short-wave radiation due to dense fog (by 48%) and the increased reflectivity (+44%) under foggy conditions implies that amounts of net short-wave radiation are affected even more by fog than its two components (incoming and reflected radiation). However, because both the albedo and the effect of fog on albedo are strongly dependent on solar angle and thus on the time of day (Fig. 8), the overall observed reduction in net short-wave radiation $R_{s,n}$

differs only slightly from that of $R_{s,\downarrow}$,

$$R_{s,n}/R_{s,n,0} = (-0.26 \pm 0.20) + (0.15 \pm 0.03) \ln(X), \quad (8)$$

adjusted $r^2 = 0.65$, $P < 0.001$, $n = 20$. In all three cases the intercept is not significantly different from zero ($P > 0.2$) and could be omitted.

Fetcher et al. (2000) have found by experimental displacement of plant species from other locations that maximum photosynthetic rates do not show a significant adaptation to lower light levels that could be

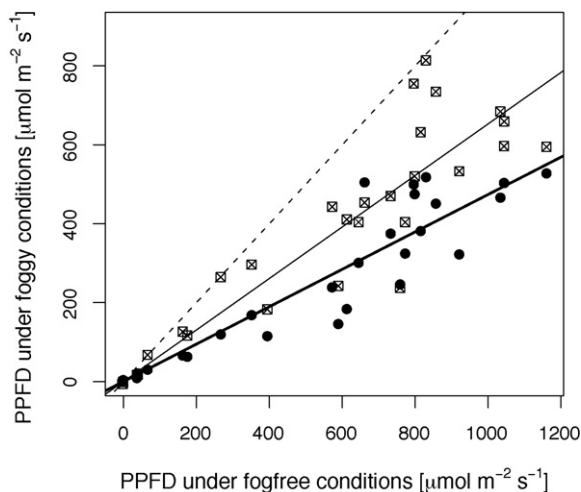


Fig. 9. Available photosynthetic photon-flux density (PPFD) under foggy conditions compared to corresponding values under fogfree conditions. Data were grouped by time of day for this comparison, only the median values for each group are displayed. Solid circles: dense fog (visibility < 200 m); squares: light fog (visibility 200–1000 m). The broken line indicates the 1:1 relationship, the solid line is the linear regression obtained for dense fog ($y = (0.474 \pm 0.020)x$, adj. $r^2 = 0.95$), and the thin line is the linear regression fit for light fog ($y = (0.653 \pm 0.030)x$, adj. $r^2 = 0.94$). Regressions were forced through the origin.

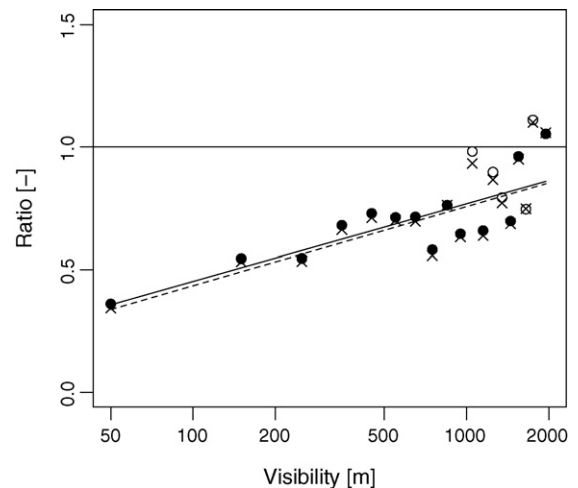


Fig. 10. Ratios between actual incoming radiation and corresponding values under fogfree conditions. Data were grouped in 50 m intervals of visibility. Circles: PPFD, crosses: global radiation. Solid circles indicate that at least 5 individual measurements were available, open circles have < 5 values. The solid line marks the linear regression fitted to the data points depicted by solid circles. The broken line represents the corresponding best fit to the global radiation data and is only marginally different from that for the PPFD data (see text).

considered a function of the fog frequency, but that the lower light levels reduce their growth rate.

3.2. Empirical estimation of liquid water content based on visibility

In most fog studies there may be a measurement or estimate of visibility available, but not this kind of detailed information presented here. Also, most tropical montane cloud forests are located at wet and remote sites where sophisticated measurements are difficult anyway. Therefore, there is a merit in addressing the possibility of obtaining realistic fog deposition estimates from visibility measurements in some detail.

Fig. 11 shows the relatively smooth transition in fog droplet size distributions from very dense fog with low visibility but small liquid water content (Fig. 11c) towards less dense fog (Fig. 11a). The highest liquid water contents are found in the visibility range of 95–175 m (Fig. 11b). Here, a shift occurs in the size distribution peak from the 15–18 μm typical of the lowest visibilities (Fig. 11c) towards the 22–23 μm typical of less dense fog (Fig. 11a). In this intermediate visibility range no clear dependency of fog deposition on visibility was found (Fig. 12a) and the best estimate of net

fog deposition is $-0.051 \pm 0.002 \text{ mm h}^{-1}$ (as determined from median values of 10-m bins of visibility). In the 175–200 m visibility range this net deposition flux reduces to $-0.018 \pm 0.003 \text{ mm h}^{-1}$. Above 200 m visibility, only $-0.008 \pm 0.001 \text{ mm h}^{-1}$ is intercepted by the canopy. The only visibility range showing a linear dependence of fog deposition rate on visibility is that below 95 m ($r^2 = 0.909$; Fig. 12). However, visibilities $< 95 \text{ m}$ only occurred in 14.6% of all cases, and thus our statistics for this range are not as robust as for the 95–175 m visibility range which covers 68.2% of all cases.

Provided the above-mentioned average relationships are valid elsewhere, it might thus be sufficient in future studies to determine fog density for the following density classes: below 95 m, 95–175 m, 175–200 m, and 200–1000 m, and then assign the mean fogwater flux determined for the respective ranges. When also the probability of occurrence of fog is included as a criterion, then it even appears that three classes are sufficient: (1) dense fog with $X < 200 \text{ m}$; (2) light fog with X in the range 200–1000 m; (3) fogfree conditions with $X > 1000 \text{ m}$. It should be noted, however, that measuring liquid water content instead of visibility yields much better correlations ($r^2 > 0.91$, Fig. 12b) and thus more accurate estimates of fogwater fluxes. Since

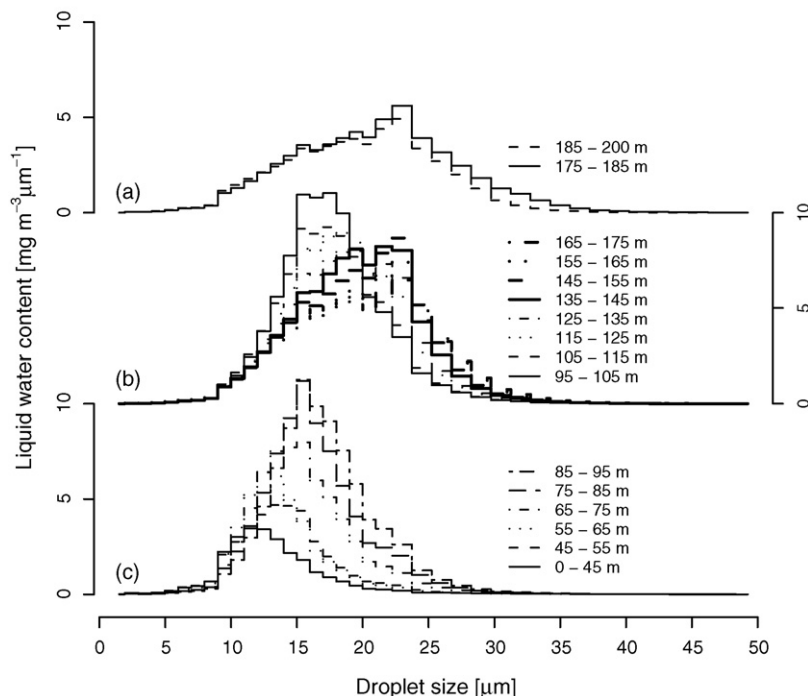


Fig. 11. Mean fog droplet size distribution by visibility range in dense fog (visibility $< 200 \text{ m}$). Highest liquid water contents are found in the visibility range 95–175 m (center panel, b) where a transition from the size distribution peak around 15–18 μm typical of the lowest visibilities (bottom panel, c) towards the peak around 22–23 μm typical of less dense fog (top panel, a) occurs.

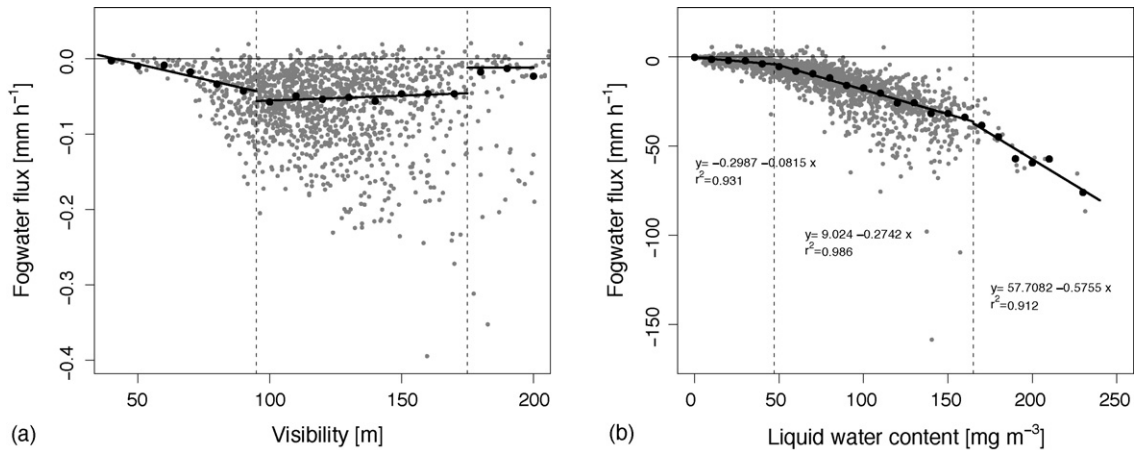


Fig. 12. Relationships between fogwater flux and (a) visibility and (b) fog liquid water content at Pico del Este between 26 June and 7 August 2002. Gray symbols show individual 30-min flux averages, and black symbols are the median value of each bin of (a) 10 m and (b) 10 mg m⁻³, respectively. Solid lines are fitted to the median values of three size classes (see text).

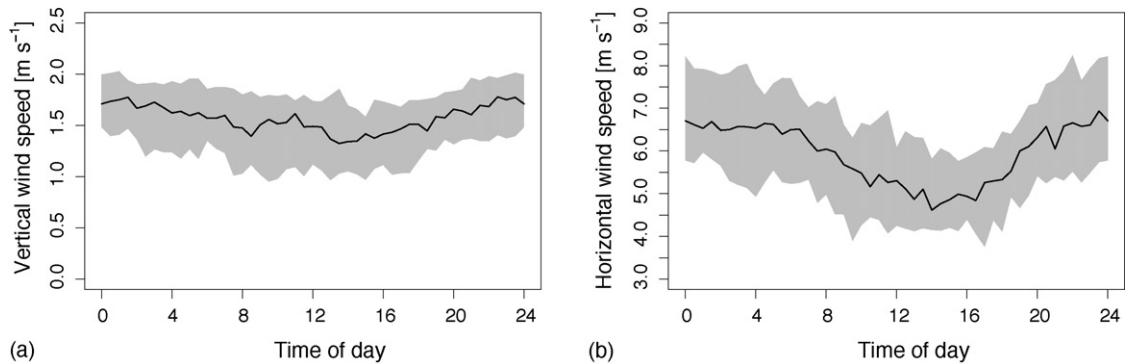


Fig. 13. Median diurnal cycle of (a) true vertical and (b) horizontal wind speeds (solid lines), and interquartile ranges (shaded area) using 30-min averages.

this is the first attempt to directly measure fogwater fluxes to a tropical cloud forest we were unable to assess the relationship between canopy structure parameters and deposition rates in more detail. In general, however, it should be recalled that heterogeneous canopies with highly variable tree heights expose much more surface area where cloud droplets can impact or be intercepted than a homogeneous canopy with comparable leaf area index but with a smooth canopy surface.

3.3. Net versus gross fogwater fluxes

Thus far we have only considered fogwater flux as measured above the canopy. By definition, a flux measured with the eddy covariance technique is a net flux and does not represent the true gross flux at the vegetation surface height (see [Appendix A](#)). We expect a substantial additional flux component from concurrent condensation of water vapour due to the upslope

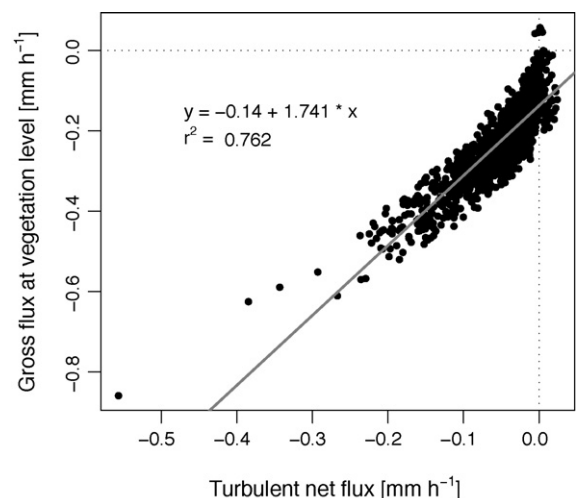


Fig. 14. Relationship between fogwater fluxes at canopy level ($F_L + F_c + F_A$) and the fluxes measured 3.0–3.5 m above the canopy (F_L) using 30-min averages.

movement of air by the mean wind. This vertical motion was measured with reference to a built-in inclinometer which allowed the computation of the true vertical wind speed with respect to the local gravity vector. Compared with the mean wind speed (Fig. 13 b) the vertical wind proved even more persistent (Fig. 13a).

A direct estimate of the condensation flux (see also Holwerda et al., 2006) due to vertical motion in this humid environment with the atmosphere being practically saturated with water vapour can be derived using the Clausius–Clapeyron relationship between saturation vapour pressure e_s and temperature T (Garratt, 1994):

$$\frac{\partial e_s}{\partial T} = \frac{L_v e_s}{R_v T^2}, \quad (9)$$

or

$$\frac{\partial \rho_{v,s}}{\partial T} \approx \frac{L_v e_s}{R_s^2 T^3} \quad (10)$$

if absolute humidity at saturation $\rho_{v,s}$ is used instead of e_s . L_v is the latent heat of vaporization ($L_v = [2.501 - 0.00237\{T - 273.15\}] \times 10^6 \text{ J kg}^{-1}$) and R_v the specific gas constant for water vapour ($461.53 \text{ J kg}^{-1} \text{ K}^{-1}$), and the conversion from Eq. (9) to Eq. (10) uses the approximation that $\rho_{v,s} \approx e_s / (R_v T)$.

With Eq. (10) and our measurements of true vertical wind speed $\bar{w}$ averaged over each 30-min interval it is now possible to estimate the condensation flux F_c between roughness height for liquid water deposition z_1

Table 2
Fogwater deposition at various mountain sites in comparison to concurrent rainfall

Location	Liquid water content (g m^{-3})	Deposition rate during event (mm h^{-1})	Total deposition			Reference
			Fog (mm d^{-1})	Rain (mm d^{-1})	Δ (%)	
Pico del Este, Puerto Rico	0.079	0.21	4.36	28.0	15.6	This study
Pico del Este, Puerto Rico	0.016	0.07	1.3 ^a	15	8.7	Asbury et al. (1994), published
Pico del Este, Puerto Rico	0.060	0.26	4.9 ^a	15	32.6	Asbury et al. (1994), scaled ^b
Humbolt Co., CA, U.S.A.	–	0.12 (maximum 1.7 ^c)	0.08	–	–	Azevedo and Morgan (1974)
Sequoia Park, CA, U.S.A.	0.15	0.12	0.08	–	–	Collett et al. (1990)
Coastal Sites, CA, U.S.A.	0.05–0.25	–	–	–	–	Jacob et al. (1985)
Clingman's Peak, NC, U.S.A.	0.16	0.03 (maximum 0.3 ^c)	0.5	6.8	7.4	Dasch (1988)
Whitetop Mt., VA, U.S.A.	–	0.25 (maximum 1.0)	1.6	3.8	42.1	Mueller and Weatherford (1988)
Mt. Moosilauke, NH, U.S.A.	0.4	0–0.32	2.3	4.9	46.9	Lovett (1984)
Great Dun Fell, Cumbria, U.K.	–	0.07	0.12	–	–	Dollard et al. (1983)
	0.144 ^d	0.02 ^d				
Laegeren, Switzerland			0.05	1.5	3.3	Burkard et al. (2003)
	0.063 ^e	0.002 ^e				
Fichtelgebirge, Germany	0.066	0.02	0.10	2.7 ^f	3.7	Thalmann et al. (2002)

Δ is the percentage by which fog and fogwater deposition increase measured rainfall.

^a Based on 75% cloud cover.

^b Assuming a factor 3.75 for their fog collector data, see text.

^c Rate for single tree in exposed position.

^d Radiation fog events.

^e Advection fog events.

^f Value from Thalmann (2001).

and the eddy covariance measurement height z_m , expressed as the finite distance $\Delta z = z_m - z_1$,

$$F_c = \cos(\theta) \int_{z_1}^{z_m} \bar{w} \frac{\partial \rho_{v,s}}{\partial T} \frac{\partial T}{\partial z} dz \approx \cos(\theta) \Delta z \bar{w} \frac{\partial \rho_{v,s}}{\partial T} \frac{\partial T}{\partial z}. \quad (11)$$

Here θ is the angle between the local streamline coordinates and the horizontal plane, and $\partial T / \partial z$ is the moist adiabatic temperature gradient of -0.005 K m^{-1} . F_c is the expected difference between the net fogwater flux measured by eddy covariance and the flux at the leaf surface if only condensation would be responsible for this difference in fluxes. Eq. (11) is only used during times with fog and $F_c = 0$ is set under the absence of fog.

The total flux at canopy level is thus the sum of the measured turbulent and settling fluxes F_L (Eq. (3)), the estimated condensation flux F_c (Eq. (11)), and a possible horizontal advective flux divergence F_A (see Appendix A). Fig. 14 shows how $F_L + F_c + F_A$ relates to measured F_L . The linear regression suggests that the true flux at vegetation level is 74% greater than our directly measured turbulent flux at $\approx 4.5 \text{ m}$ above the canopy, with an offset of -0.14 mm h^{-1} ($r^2 = 0.762$).

In an earlier paper by Asbury et al. (1994) it was reported that liquid water contents of clouds at Pico del Este were much smaller than those derived for comparable sites. It was argued that there were more “thin” clouds at Pico del Este than at other mountain sites. Our data do not support this contention. To examine this important discrepancy we were given access to the original data of Asbury et al. (1994) (William McDowell, personal communication). Since no errors could be found, the discrepancy must be attributed to an inadequate estimate of sampling efficiency (arbitrarily assumed to be 85%) and/or an overestimation of the sampled air volume (assumed to be $19.6 \text{ m}^3 \text{ min}^{-1}$; Asbury et al., 1994). A comparison of the quantile distribution of our LWC measurements with the corresponding original values from Asbury et al. (1994) suggests a linear scaling factor of 3.75 to be required to correct the older data. This implies that, in contrast to earlier reports, the mountain fog at Pico del Este is not very different from that at other comparable sites, but due to the high frequency of occurrence of fog Pico del Este collects more fogwater than any other location for which comparable data exist (Table 2).

4. Conclusions

Under conditions of dense fog at Pico del Este, Puerto Rico, we found very similar reductions in incoming solar radiation (both PPFD and $R_{s,\downarrow}$) and net short-wave energy ($R_{s,n}$). This finding suggests that the relative influence of fog must be the same for plant photosynthesis or plant transpiration (which both are a function of PPFD), and surface evaporation (which is a function of net radiation and thus also of $R_{s,n}$). Although there are important differences between the processes of photosynthesis and surface evaporation from a wet canopy (which also responds to wind speeds and vapour pressure deficits), our analysis suggests that the effect of fog must be very similar for plant photosynthesis, plant transpiration, and surface evaporation.

Fog characteristics vary widely for comparable visibilities, but it appears that good approximations can be achieved by considering the following three categories of (1) dense fog with visibility $< 200 \text{ m}$, (2) light fog with visibility ranging between 200 and 1000 m, and (3) fogfree conditions. The fog droplet size spectrum at Pico del Este has a characteristic volume-weighted mean diameter of 12–13 μm during very dense fog, which shifts towards larger droplets of ca. 22–23 μm at the transition between dense and light fog.

In the absence of liquid water content measurements the best estimate for net fogwater deposition to the elfin cloud forest canopy is 0.06 mm h^{-1} or $-17.7 \text{ mg m}^{-2} \text{ s}^{-1}$ (0.2 mm h^{-1} or $-63.4 \text{ mg m}^{-2} \text{ s}^{-1}$ gross flux at vegetation level) for the most frequently encountered condition of dense fog (visibility range 95–175 m). If there is a choice to either measure visibility or liquid water content, then the latter is preferred since it shows a much closer correlation with fogwater fluxes ($r^2 > 0.9$) than does visibility (typical $r^2 < 0.5$).

An important limitation of the eddy covariance method in comparison with other approaches to fog deposition quantification is the formation of fog due to ascending air masses along the mountain slope, which concurs with the deposition of fogwater to the vegetation canopy (cf. Kowalski and Vong, 1999; Holwerda et al., 2006). The eddy covariance method only measures net deposition whereas at the leaf level the gross deposition rate is considerably larger. We found a good correlation ($r^2 = 0.72$) between gross and net fluxes but with a substantial intercept of -0.14 mm h^{-1} (or $-38.9 \text{ mg m}^{-2} \text{ s}^{-1}$) due to steady upslope winds with a vertical component of 1–2 m s^{-1} . It should however be noted that this correction for condensation and advection effects does still not

eliminate the discrepancy between fog deposition estimates made by micrometeorological methods on the one hand and the hydrological budget approach on the other hand (cf. Holwerda et al., 2006).

Our results show strong evidence that the hydrologic input by fogwater deposition to the elfin mountain cloud forest at Pico des Este is substantial but still quite small as compared to precipitation and may thus not contribute significantly to dry season base flows. However, the reduction in solar radiation inputs and its indirect effect on base flow through reduced evaporation losses may be a much more relevant effect that fog – including dense and persistent fog – imposes on cloud forests in the tropics and most likely also elsewhere (cf. Mulligan and Burke, 2005).

Acknowledgements

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Appendix A

The site we chose for our flux measurements is located on a relatively steep slope of 17° at 100 m below the crestline of the mountain chain of El Dunque (Fig. 1). One important aspect to consider is the fact that the vast majority of eddy covariance studies focuses on turbulent exchange over flat and horizontally homogeneous terrain, which is not the customary setting one finds in complex terrain. Thus, a priori assumptions on how the eddy covariance point measurements relate to the land surface and vegetation in the footprint of that measurement may not be the same on mountain slopes as compared to idealized terrain (see also Wyngaard, 1990).

Others have successfully carried out fogwater deposition measurements, first with indirect methods (see, e.g. Vong et al., 1990; Collett et al., 1991; Vong et al., 1991), then also with the direct eddy covariance approach (Beswick et al., 1991; Vong and Kowalski, 1995; Kowalski et al., 1997; Vermeulen et al., 1997; Kowalski and Vong, 1999; Burkard et al., 2001, 2002, 2001a; Wrzesinsky et al., 2001; Eugster et al., 2001; Thalmann et al., 2002; Burkard et al., 2003; Holwerda et al., 2006). Still, a generally accepted protocol on how to measure fluxes with the eddy covariance technique in sloping terrain is lacking, and therefore we provide additional information on turbulence characteristics at our site, and explain how we estimated the along-streamline flux divergence term F_A which might be an additional important flux term in such terrain.

A.1. Turbulence characteristics at the site

The site is exposed to the steady east-northeasterly trade winds (Fig. 13). During fog events the fog reduces both short-wave and long-wave radiation components (Fig. 8). The combination of these two factors, high mechanical turbulence and small thermal convection, leads to near-neutral atmospheric stratification throughout fog events. The range of the Monin–Obukhov stability z/L during our measurement campaign showed 95% of all values between -0.163 and 0.042 ($n = 1675$ half-hour average values).

Examples of turbulence spectra and cospectra are shown in Fig. A.1. We do not show any spectra of temperature fluctuations nor cospectra of sensible heat flux since they are not relevant for near-neutral conditions. Moreover, the sensible heat flux cospectra do not show any organized structures since the heat flux is almost zero. The spectra of the along-streamline wind component (Fig. A.1a) shows a slight shift of the energy containing range and its maximum spectral density towards lower frequencies than would be expected over a flat and horizontally homogeneous surface (solid curve). The streamline-normal wind component (Fig. A.1b) agrees well with an idealized spectrum that we would expect at a height z that is four times the nominal height measured at the tower (i.e. 18.3 m above canopy), and still there is consistently more variance found in the lowest frequencies that differ from flat and homogeneous surface conditions. For reference, the expected spectrum for measured z is drawn with a thin broken line. For liquid water spectra no theoretical curves exist, and thus we compare the measured spectral densities in Fig. A.1c with an idealized spectrum for temperature over a flat and

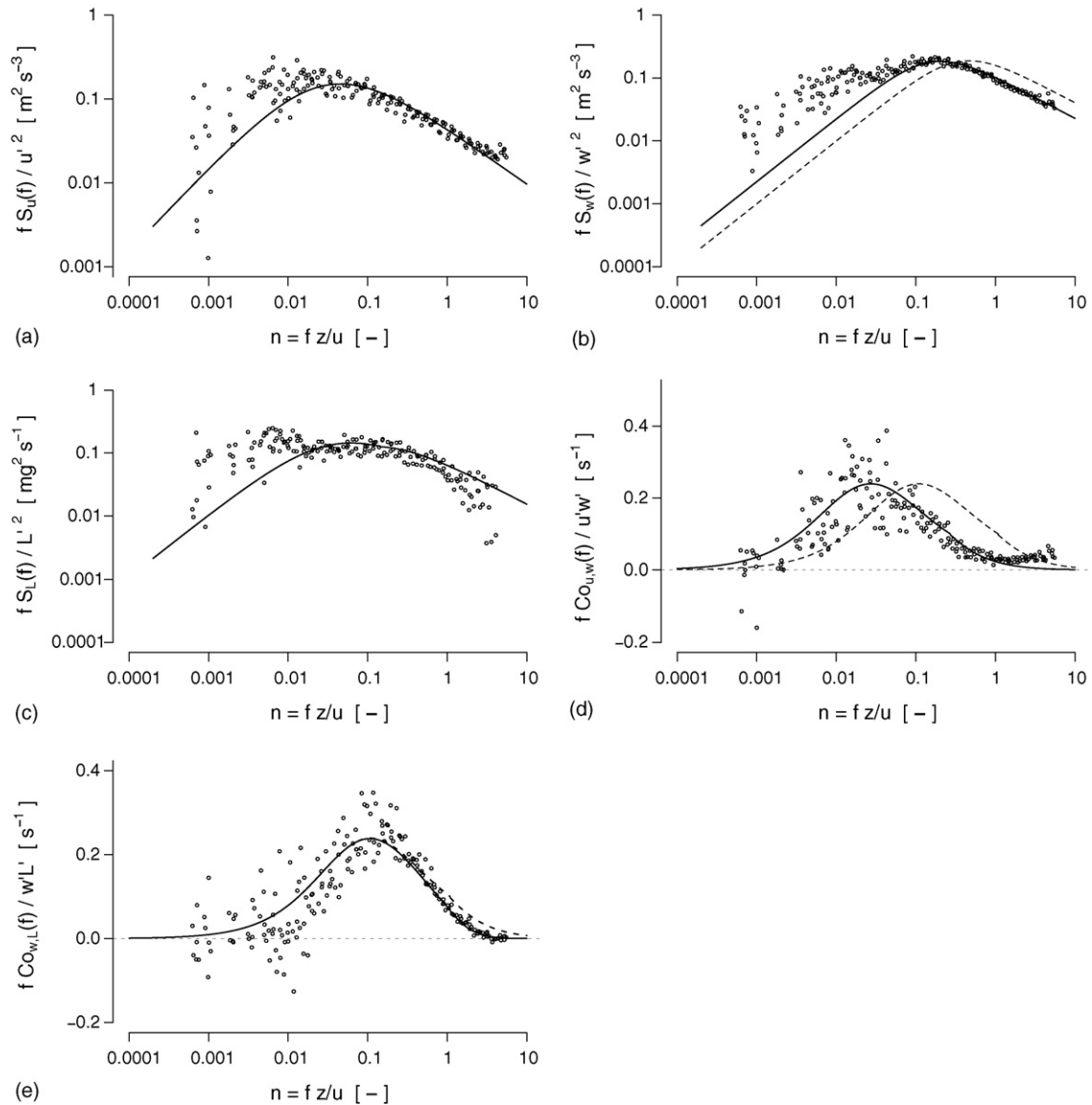


Fig. A.1. Example spectra of (a) along-streamline wind component, (b) streamline-normal wind component, (c) liquid water (white noise removed), and cospectra of (d) momentum flux and (e) liquid water flux. Each panel is a composite of six spectra or cospectra of one-hour data segments from 25.06.2002, 01–07 h during persistent and relatively stationary fog conditions. Idealized spectra in panels a and b are based on Højstrup (1981) for neutral conditions, the shape of the one in panel c is taken from Kaimal et al. (1972) for temperature variations. The idealized cospectra in panel d and e are based on the modified Kaimal et al. (1972) cospectra presented in Eugster and Senn (1995). See appendix for details.

horizontally homogeneous surface. Similar to what we found in the wind velocity spectra we see an increased importance of the lower frequencies. The increasing scatter at the highest frequencies is explained by the fact that the sampling rate of 12.5 Hz the air analyzed by the FM-100 fog droplet spectrometer very frequently does not contain a single droplet even during fog. This has been explained in more detail in Eugster et al. (2001). Since this oversampling leads to a high white-noise level

we already eliminated the white-noise effect from the measured spectrum in Fig. A.1c. We used the same method as was described in Eugster et al. (2003), that is, by assuming that the band-width averaged spectral density of the highest frequency band corresponds to the white noise level. This marginally overestimates the effect of random noise but should not bias our interpretation of the overall performance of the FM-100 in sloping terrain.

Thus, for the spectra we find acceptable differences from conditions over flat and uniform terrain and do not expect any serious problem with eddy covariance flux measurements at our site. The frequency shift seen in the streamline-normal wind component (Fig. A.1b) translates to a corresponding shift in the frequency of the maximum covariance towards lower values in the cospectra of momentum flux (Fig. A.1d, solide curve), but not so in fogwater fluxes (Fig. A.1e, solid curve). The broken curve in Fig. A.1d shows the expectation for flat terrain. This indicates that for momentum flux – but not for liquid water flux – the true footprint area and fetch must be much larger than what would be expected from the local height of the tower. One reason could be that the change in steepness of the slope roughly 800 m upwind (see Fig. 1) leaves a marked trace in the momentum flux footprint. Another interpretation could be that the upslope movement generates additional mechanical turbulence, and since turbulence is generated at the low frequency end of the cospectrum this may be a typical feature that might be found on any other sloping terrain as well.

Sensor separation is an issue (see Fig. 2), which leads to a time lag between LWC and wind velocity measurements, and a damping loss in the highest frequencies. The typical time lag is 2 or 3 records (accounting for 73.7% and 18.3% of all cases, respectively), which corresponds to 0.16–0.24 s. The horizontal separation between instruments is 0.56 m. At a mean horizontal wind speed of 5 m s^{-1} this translates to a lag time of roughly 0.11 s. The lag introduced by the time it takes the FM-100 to process the LWC sample is 0.05 s, giving the typical time lag of 0.16 s. When synchronizing data online it should be recalled that if the FM-100 lags the sonic anemometer, then an additional lag of one record or – at 12.5 Hz – 0.08 s is introduced, which gives the less frequently observed lag time of 0.24 s (or 3 records).

This correction for the time lag between the two data streams does not perfectly account for sensor separation, as is clearly seen in the high frequency range in Fig. A.1e: the theoretical cospectrum for a scalar flux of an ideal undamped instrument at true z is shown as a dashed curve, whereas the best fit is the typical damped cospectrum (solid curve) with a damping constant of 0.1 s for this specific set-up (see Eugster and Senn, 1995). It should be noted that for the computation of the cospectra of fogwater fluxes we did not remove white noise (true white noise should disappear by definition in the cospectrum). While LWC cospectra nicely near the expected zero-value at the highest frequencies, this is not the case for momentum flux, where some systematic

noise in the high frequencies is responsible for the fact that the cospectra in Fig. A.1d do not approach zero in the highest frequencies. It should be recalled that measuring turbulent fluxes with sonic anemometry is rather challenging under foggy conditions. There are instruments that completely fail under such circumstances, whereas the one we used obviously performed acceptably well for momentum flux, and excitingly well for liquid water fluxes.

From these spectral and cospectral analyses we do not find any indication that there should be a real concern about the turbulent flux measurements of liquid water in such a sloping terrain. Thus, in the following section we will use a general footprint model to estimate the extent of surface area that influenced our measurements. This however does not necessarily mean that there is no influence of advection, the last factor to be assessed in the final section of this appendix.

A.2. Flux footprint areas

We used the Kljun et al. (2004) flux footprint model. This model similar to others actually assumes horizontally homogeneous terrain, and thus should be rather considered an order-of-magnitude estimate (see Schmid, 2002 for a more thorough discussion of usefulness and shortcomings of the footprint modeling approaches used in micrometeorology). We proceeded as follows. For each half-hourly data record with fog (visibility < 1000 m) we computed the along-wind integrated footprint for the 80% footprint area with the parameterized version of the Kljun et al. (2004) model. We repeated the computations for an estimated planetary boundary layer (PBL) height of 1000 and 100 m. The results did not differ significantly and thus we continued with the 1000 m PBL height data in our analysis presented here. All output was then mapped cumulatively to a spatial map according to mean wind direction, and isolines were drawn relative to the local maximum of cumulative footprint information found over the spatial extent of all footprints. These lines are shown in Fig. 1 as solid lines and nicely show a small footprint area that has its maximum roughly 40–70 m upwind from the tower location. The cospectra of the liquid water flux were very similar to what would be expected over flat terrain, such that this footprint area should actually be a good estimate for fogwater fluxes. In the case of momentum fluxes we found cospectra that best fitted the theoretical curve for four times true measurement height. Therefore, we repeated all footprint calculations for $z = 18.3 \text{ m}$, assuming that this will give the best order-of-magnitude estimate for

the momentum flux footprint area (broken lines in Fig. 1). The 10% isoline nicely extends to the edge of the 17° slope where the steeper part of the lower slope begins. This is consistent with our argumentation about the lower than expected frequencies found in the momentum flux cospectra, but one should recall that the assumptions made by a flux footprint model are not very different from the assumptions made with idealized cospectral curves. To summarize, we conclude that the turbulent fogwater fluxes have a much smaller footprint than the momentum fluxes and nicely represent the conditions of the surrounding vegetation, whereas momentum fluxes are much more severely affected by the steep topography of the site.

A.3. Estimating horizontal advection

In this final section we start with the continuity equation describing the high-frequency point measurements collected by eddy covariance equipment in general,

$$\frac{\partial \bar{\rho}_L}{\partial t} = F_c(z) - \left(\frac{\partial \bar{u} \bar{\rho}_L}{\partial x} + \frac{\partial \bar{v} \bar{\rho}_L}{\partial y} + \frac{\partial \bar{w} \bar{\rho}_L}{\partial z} \right) - \left(\frac{\partial \bar{u}' \bar{\rho}_L'}{\partial x} + \frac{\partial \bar{v}' \bar{\rho}_L'}{\partial y} + \frac{\partial \bar{w}' \bar{\rho}_L'}{\partial z} \right) - D_L(z), \quad (12)$$

where the x , y , and z directions are aligned with the mean streamlines. In such a rotated coordinate system the $\partial \bar{v} \bar{\rho}_L / \partial y$ and $\partial \bar{w} \bar{\rho}_L / \partial z$ terms are zero by definition. For the simplified case of a two-dimensional slope – which is a good approximation of the conditions found at the location of our measurements – we also can neglect $\partial \bar{v}' \bar{\rho}_L' / \partial y$.

In order to relate the eddy covariance point flux measurement to the local surface, Eq. (12) needs to be integrated in the vertical direction from $z = 0$ to instrument height z_m , which allows to further simplify the continuity equation as follows for our mountain slope site. Measured along-streamline turbulent fluxes $\bar{u}' \bar{\rho}_L'$ were only $52 \pm 1\%$ of the surface-normal component ($r^2 = 0.649$). Moreover, since there is no strong discontinuity for turbulent fluxes in the along-streamline direction, we can neglect the horizontal turbulent flux divergence term $\partial \bar{u}' \bar{\rho}_L' / \partial x$ while the surface-normal divergence term would directly correspond with the net flux $F_c - D_L$ under flat and horizontally homogeneous conditions. We consider these simplifications adequate for the conditions experienced near Pico del Este, which allows us to reduce Eq. (12), integrate in vertical direction, and rearrange for the gross deposition term D_L (which

equals the uptake at roughness height for liquid water deposition z_1) as

$$D_L = F_c - \int_{z_1}^{z_m} \left(\frac{\partial \bar{u} \bar{\rho}_L}{\partial x} + \frac{\partial \bar{w}' \bar{\rho}_L'}{\partial z} + \frac{\partial \bar{\rho}_L}{\partial t} \right) dz. \quad (13)$$

F_c represents $\int_{z_1}^{z_m} F_m(z) dz$ and is approximated by Eq. (11). $\int_{z=0}^{z_m} (\partial \bar{w}' \bar{\rho}_L' / \partial z) dz$ is replaced by F_L according to Eq. (3), which includes the settlement of larger droplets in addition to the pure turbulent motion, and $\int_{z_1}^{z_m} (\partial \bar{\rho}_L / \partial t) dz$ is zero under stationary conditions, or could be approximated as $\Delta \rho_L / \Delta t \cdot z_m$ otherwise, where Δ denotes the finite difference between two subsequent 30-min averages. The only term that cannot be directly measured with instruments at one single point is $\int_{z_1}^{z_m} (\partial \bar{u} \bar{\rho}_L / \partial x) dz$, which we denoted with F_A in the main text. To estimate the order of magnitude of that term we made the following assumptions: (1) The relatively steady tradewinds approaching the mountain slope where we carried out our measurements accelerate due to the topographical constriction by the crestline (Fig. 1) such that the divergence term can be computed via the divergence of wind speed at constant ρ_L ; (2) the speed-up of horizontal flow is restricted to 100 m above the crestline; (3) the influence of the slope and crestline can be approximated as a two-dimensional homogeneous slope of infinite lateral extent (that is, no effect of curvature of the slope is considered); (4) the acceleration effect increases linearly between z_1 and measurement height z_m .

As expected the additional downward flux F_A is much smaller than the one due to condensation effects (F_c ; Eq. (11)), ranging from -3.9 to 21.2% of the condensation effect with a median and mean of 7.5 and 7.6% , respectively. In the final flux estimate as presented in Fig. 14 this horizontal advective component F_A accounts for 0.2 – 12.3% of the gross flux, with a median and mean of 5.5 and 5.2% , respectively, thus confirming our initial assumption that the condensational effect on sloping terrain (see also Holwerda et al., 2006) is the most critical additional factor that needs to be considered in addition to the turbulent flux measured by eddy covariance.

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