CHARACTERIZATION OF THROUGHFALL HETEROGENEITY IN A

TROPICAL PRE-MONTANE CLOUD FOREST IN COSTA RICA

An Undergraduate Research Scholars Thesis

by

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ABSTRACT

Characterization of Throughfall Heterogeneity in a Tropical Pre-
Montane Cloud Forest in Costa Rica. (May 2014)

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Understanding the water budget in tropical forests is essential because of its role in ecosystem health, drinking water supply, land, and resource management. Throughfall, the amount of precipitation reaching the forest floor, plays an important role in the balance between precipitation, runoff, and other components of the water budget. Previous research has indicated that vegetation and precipitation variables are the main drivers of throughfall variability. During the data collection stage of this study, rain gauge networks were deployed in a 2.2-hectare watershed within a tropical pre-montane transitional cloud forest in Costa Rica. Throughfall data were collected daily for a total of 39 events from 28 June–17 July 2012 and 12 June–16 July 2013. To quantify vegetation cover, leaf area index was estimated above each gauge using hemispheric photography. Precipitation and its intensity were also recorded for each event based on portable weather stations. The purpose of this thesis is to use these observational datasets to provide a comprehensive quantitative assessment of the spatial and temporal throughfall variability, and determine its main drivers. This study demonstrates that rainfall intensity and canopy density significantly affect throughfall patterns. However, throughfall variability is also

driven by complex interactions between coupled factors.

ACKNOWLEDGEMENTS

CHAPTER I

INTRODUCTION

Tropical regions

Characteristics of tropical regions in Central America

Tropical regions are typically located between 23.27°N and 23.27°S, where the high atmospheric moisture contents and narrow temperature ranges provide a favorable environment for cloud formation (Balek, 1983a). Temperatures in tropical regions typically range from 15 to 30°C (Balek, 1983a). The humid tropics of Central America, as defined by Chang and Lau (1983), typically have a tropical − equatorial climate, which is characterized by high temperatures and high amounts of rainfall year-round. Annual precipitation in these regions ranges from 2000 to 3000 mm, and is characterized by high spatial and temporal variability (Griesinger and Gladwell, 1993). The main sources of precipitation are tropical hurricanes, convectional showers and thunderstorms due to diurnal heating, converging trade winds, and orographic uplift (Balek, 1983a). Equatorial regions located outside of the hurricane belt such as Panama and Costa Rica still receive high rainfall amounts due to their narrow land extent and temperature range (Griesinger and Gladwell, 1993). Precipitation seasonality occurs because of the shifting Intertropical Convergence Zone (ITCZ), which migrates over Central America during the wet seasons around June and October, and is located farther south during the relatively dry season between December and April (Balek, 1983a; Griesinger and Gladwell, 1993). Because of the generally high rainfall amounts year-round, dry seasons are not always observable in some regions of Central America. The spatial variability of rainfall is in part due to the mountain range

dividing Central America into eastern and western slopes. The Pacific side receives about half as much rain as the Caribbean side (Balek, 1983a). The humid Central American tropics are valued for their rich biodiversity, pristine forests, and hydroelectric potential (Chang and Lau, 1983). However, growing populations and ecotourism put stress on their water resources. The increase in water demand and land-use changes in these regions put the current hydrologic system at risk, and has put the fate and management of tropical rain forests at the center of attention over the past decades (Balek, 1983b).

Water use in tropical regions

Water supply in terms of quantity and quality has become a concern for tropical areas with fast growing populations. The proper management of local and regional water resources is essential to meet the growing water demands for domestic use, agriculture and forestry, and hydroelectric power supply (Griesinger and Gladwell, 1993; Balek, 1983c). Water resource management is also essential to controlling land-use changes that may put local populations and farmers at risk from flooding (Tejwani, 1981). In some areas, farmers represent 90% of the population; the countries' economies in Central America therefore greatly depend on the productivity of their lands (Tejwani, 1981). In the tropical regions of Central America, 80% of the water is used for agriculture. Proper land and water resource management for irrigation, crop selection, and operation timing is therefore essential to control the water supply needed on a local and regional scale (Balek, 1983c). A change in these management practices and/or a change in climate could greatly affect the area's water yield. To increase the water yield and take advantage of more fertile soils, some farmers burn forested areas and start exhaustive cropping (Balek, 1983c; Pereira, 1981). However, a watershed's outflow relies on the relationship between rainfall and

runoff. Modifying the vegetation can cause drier conditions and a loss of soil fertility, thus resulting in a loss of crops (Pereira, 1981).

Water sources in tropical regions

Sources of water in the tropics are rivers and watersheds. Most of the water used in agriculture comes from rivers, as they flow from upper to lower catchment areas. Land uses in the upper catchment areas affect the water quantity and quality received in the lower catchment areas (Russell, 1981). If the vegetation and soil are disturbed upstream, the balance between rainfall and evapotranspiration will be disturbed, as it depends on vegetation, soil water holding capacity, and soil depth (Russell, 1981). The seasonal river flow will therefore be modified, with higher sediment load at peak flow due to soil erosion of upstream lands. Arable lands downstream will therefore be vulnerable to flooding during the wet season, and low river flow during the dry season (Russell, 1981). In areas of subsistence agriculture, it is essential that the river flow be controlled by farmer cooperation for proper watershed management in the upper catchment areas (Pereira, 1981; Russell, 1981).

Another source of water is represented by groundwater. Water for domestic use and to sustain rural areas typically comes from this source. Groundwater is an essential water supply, particularly when surface waters are polluted or insufficient during dry periods (Balek, 1983d). Tropical forests themselves play an essential role in protecting and controlling the natural flow of the river leaving the watershed (Russell, 1981). Depending on the forest type, tropical forests may even increase the water yield. Farmers responsible for deforestation of certain areas could therefore be negatively impacting their water supply. In some cases, deforestation increases the water supply, as newly exposed soil is compacted and loses its porosity. This increases water

runoff and stream flow during high intensity rain events (Balek, 1983b; Pereira, 1981). Farmers therefore become vulnerable to flooding as well as soil erosion, which depletes the land of nutrients, thus rendering once arable land unproductive and impacting food production (Pereira, 1962).

Tropical forest hydrology

Water budget terms for a forested watershed

To best study the hydrology of a forested watershed, it is essential to quantify each individual term coming into play in the water budget. Ward and Robinson (1990) created a water-balance equation for tropical forests:

$$P_i = Q + ET + \Delta S + \Delta G + \Delta L \tag{1}$$

where P_i is the incident precipitation, Q is the streamflow leaving the watershed, ET is the evapotranspiration, ΔS and ΔG are the changes in soil moisture and groundwater storage, respectively, and ΔL represents any leakage in or out of the watershed. Incident precipitation can also be separated into multiple terms according to the following equation by Fleischbein (2010):

$$P_i = I + TF + SF \tag{2}$$

where I is the interception loss, TF is throughfall, and SF is stemflow. In the case of cloud forests, initial water input P is the sum of P_i and cloud water interception (CWI; Giambelluca et al., 2010).

Incident gross precipitation designates the amount of rainfall and/or cloud water that enters the canopy, whereas the precipitation that falls through a canopy is designated as throughfall. Throughfall can be divided into free TF, which reaches the forest floor without coming in contact

with the vegetation, and release TF, which is intercepted and redistributed before it reaches the forest floor (Levia and Frost, 2006). TF redistribution by the canopy results in small-scale spatial and temporal variability (e.g., Loustau et al., 1992; Staelens et al., 2006). While TF represents about 70 to 90% of gross precipitation in temperate forests, it typically ranges from 60 to 95% of gross precipitation in tropical forests (e.g., Brauman et al., 2010; Berger et al., 2008; Bruijnzeel, 2004; Levia and Frost, 2006; Lloyd and Marques, 1988; Zimmermann et al., 2007). This wider range indicates there is a higher variability of TF in tropical forests.

Stemflow represents another fraction of gross precipitation reaching the forest floor, as it is defined as water that flows down along the stems and tree trunks (Levia and Frost, 2006). Together, TF and stemflow make up the net precipitation reaching the forest floor, and also contribute to surface runoff (Balek, 1983b).

A fraction of gross precipitation does not reach the forest floor due to interception.

Interception is defined as the process by which rainfall is intercepted by the canopy before ever reaching the forest floor, and is subsequently evaporated (Holwerda et al., 2010; Cavelier et al., 2007). Evapotranspiration refers to the water that is evaporated from the canopy. This water can simply be from interception, accumulated on leaf surfaces, or water that is stored in the foliage and mosses. Quantifying evapotranspiration is challenging, and although various methods exist, it is difficult to obtain an accurate estimate (Balek, 1983b). Evapotranspiration is typically calculated using meteorological variables such as air temperature, humidity, wind speed and direction, and net radiation, as it depends on evaporative energy and demand of the atmosphere, which is typically driven by wind speed and vapor pressure deficits. During the wet season, interception is an essential component of evapotranspiration (Balek, 1983b). As shown above in equation (2), interception can be calculated as the difference between rainfall and the sum of TF

and stemflow. Interception is essential in controlling the rate at which net rainfall reaches the forest floor and penetrates into the soil, particularly during high rainfall intensity events with a potential for high runoff (Balek, 1983b). The water stored in forest soils represents soil moisture and depends partly on soil water storage capacity, exposure to solar radiation, and vegetation (Balek, 1983d). The infiltration of condensed cloud water, rainfall, or water from streams and lakes into the soil forms groundwater.

Soil moisture and groundwater depletion typically occur through transpiration, root uptake, and anthropogenic use (Balek, 1983d).

Water ultimately leaves the watershed through streamflow. Water runoff from TF and stemflow, along with surfacing groundwater all contribute to streamflow.

Characteristics of tropical rainforests

Tropical rainforests are typically defined as forests for which the rainfall exceeds the water needs of the plants (Balek, 1983b). Tropical rainforests can be divided into two subtypes based on their elevation: pre-montane rainforests typically occur between 1000 and 2000 m a.s.l., while tropical wet forests occur between 0 and 1000 m a.s.l. (La Bastille and Pool, 1978). Rainforests are typically characterized by their dense vegetation, which intercepts rainfall and directs it into the soil rather than flowing away (Balek, 1983b). The high leaf area associated with dense vegetation increases evapotranspiration rates. Scientists have therefore concluded that more water is evaporated from forests than from deforested areas (Balek, 1983b; Lal, 1981). High evapotranspiration rates, porous soils, and rainfall interception therefore allow rainforests to retain water and lower the outflow for the watershed (Balek, 1983b; Lal, 1981). In the case of rainforests, it may seem beneficial for farmers to use deforestation to increase their water yield.

However, the water evaporated from the vegetation can cause an accumulation of cool air above the canopy, thus creating favorable conditions for further vapor condensation and contributing to higher amounts of rainfall (Balek, 1983b; Lal, 1981). Land use changes must therefore be tightly managed, as reducing the extent of forested areas could cause drier conditions.

Characteristics of tropical cloud forests

Cloud forests are a unique subtype of rainforests, defined as forests that are frequently or persistently immersed in low clouds or mist (Stadtmüller, 1987). They generally occur on steep mountain slopes, where cloud belts form due to ascending air masses (Zadroga, 1981). Cloud forests therefore occur at relatively high altitudes, typically between 1500 and 3000 m a.s.l. in Central America (Zadroga, 1981). However, cloud forests worldwide have been found anywhere between 220 and 5005 m a.s.l. (Jarvis and Mulligan, 2010). Cloud forests are wetter, cooler, and less seasonally variable than rainforests, especially at altitudes higher than 1000 m a.s.l., where they are exposed to high amounts orographic rain and cloud moisture (Zadroga, 1981; Jarvis and Mulligan, 2010). Because of the long periods of cloudiness, there is negligible evapotranspiration from the vegetation (Zadroga, 1981; Häger and Dohrenbusch, 2011). These high amounts of precipitation and low rates of evapotranspiration favor high volumes of streamflow, which would make these soils highly susceptible to erosion without protective vegetation (Mulligan, 2010). The climate of cloud forests differs depending on location: annual rainfall amounts for cloud forests range from 600 to 2000 mm. The average temperature of cloud forests is about 17°C, but can range from 12 to 21°C depending on altitude, latitude, and exposure to solar radiation (Jarvis and Mulligan, 2011; Zadroga, 1981). Cloud forest vegetation is characterized by dense growths of trees and shrubs, with a high epiphytic load such as mosses,

ferns, and bromeliads, which allow the canopy to efficiently intercept precipitation and store large amounts of water (Zadroga, 1981).

Cloud water interception

Whereas rainforests retain the majority of incident precipitation, cloud forests can be a source of freshwater due to their ability to intercept cloud water. Plant surfaces intercept water droplets from fog, clouds, and mist, therefore eventually adding to the net precipitation in the form of drip or stemflow (Zadroga, 1981). In areas with relatively dry seasons such as Peru and Chile, cloud forests are of great hydrological importance, as cloud water interception can exceed rainfall and represent an important source of freshwater to local populations (Zadroga, 1981). Cloud water interception depends on many factors, including vegetation type, canopy saturation, and wind and solar radiation exposure (Tobón et al., 2010). While it is important to quantify the water input to a watershed, the interactions between these factors and the limitations in measurement strategies make it difficult to determine the contribution cloud forest interception to net precipitation (Häger and Dohrenbusch, 2011; Tobón et al., 2010; Zadroga, 1981; Rhodes et al., 2010; Schmid et al., 2010). Cloud water interception is usually quantified by comparing rainfall and crowndrip during periods with or without fog, by using fog collectors, or through modeling (Harr, 1982; Sigmon et al., 1989; Goodman, 1985; Bruijnzeel et al., 2005; Ritter et al 2008). Schmid et al. (2010) used the isotopic composition of the net precipitation to determine the fraction of cloud water. They reported that cloud water interception during the dry season in a Costa Rican cloud forest accounted for 4–7 % of incident rainfall. Other studies have estimated a larger contribution of cloud water: Holwerda et al. (2010) found that mean daily cloud water interception was 10–12% of the mean daily rainfall, and McJannet et al. (2010) found a

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contribution of 19% and 29% of annual precipitation at two forest sites in Northern Australia, which increased to about 65% of the monthly water input for both sites during the dry season.

Threats to cloud forests

While the deforestation of tropical rainforests, though not without negative consequences, is sometimes seen as a solution to increase runoff, the deforestation of a montane cloud forest could actually cause a decrease in water yield (Zadroga, 1981). As it was shown previously, cloud forests intercept cloud water, which drips down to the forest floor and contributes to the water yield of the watershed. However, the lack of knowledge regarding cloud forest hydrometeorology and the hydrological impacts of forest conversion make it difficult to implement cloud forest conservation and management plans (Scatena et al., 2010). Additionally, some studies in areas where input from cloud water is low report an increase in flow after forest conversion to pasture (Gomez-Cardenas, 2009). Schellekens (2006) even found that cloud forest conversion had a neutral effect on the hydrologic cycle in areas such as northern Costa Rica. However, changes in water flow are not the only potential impacts of deforestation. For example, stripping steep mountainous slopes of vegetation could cause landslides (Scatena et al., 2010). The effects of lowland deforestation on precipitation have also been investigated, and Rhodes et al. (2010) argue that lowland deforestation would decrease orographic precipitation because of lower evaporation rates over the pastures, or that it might raise the cloud base, therefore increasing the numbers of mist-free days (Ray et al., 2006). Others have observed the opposite trends at their study sites (Roy and Avissar, 2002; Van der Molen et al., 2006; Pounds et al., 1999). Some of these trends could also be due to global climate change, and not so much atmospheric drying after land use changes (Foster, 2010). Cloud forest hydrology is complex,

and it is difficult to determine the impacts of deforestation on precipitation inputs. The response of the hydrologic system depends on the location, pattern, and extent of land use changes (Mulligan et al., 2010).

Because tropical montane cloud forests occur within specific ranges of altitude, temperature, precipitation, and fog conditions, they are extremely sensitive to climate change and habitat loss (Bubb et al., 2004; Häger and Dohrenbusch., 2011; Mulligan, 2010). Forest conversion to pastures and agricultural lands has been increasing rapidly since the 1970s, and is still currently occurring (Scatena et al., 2010; Aide et al., 2010). More knowledge about forest hydrology is essential to decision making about land use changes in tropical regions.

Study site: watershed near San Isidro, Costa Rica

Characteristics of Costa Rica

Costa Rica is divided into the Caribbean and Pacific slopes by the Cordillera de Tilarán, a central mountain range running from northwest to southeast Costa Rica. This divide is responsible for regional differences in rainfall amounts and their seasonality (Häger and Dohrenbusch., 2011). In general, the Caribbean slope receives about twice as much total annual precipitation than the Pacific slope, due to its windward position within the trade wind flow (Häger and Dohrenbusch., 2011). On the Caribbean slope, where this study is located, most of the precipitation is due to convective showers from orographic lifting and diurnal heating. The Caribbean slope owes its higher precipitation amounts to the north-easterly trade winds, which force orographic uplift of moist Atlantic air (Waylen et al., 1996). This source of precipitation is particularly important during the dry season, which corresponds to an ITCZ position south of Costa Rica (Rhodes et al.,

2010). The main wet season in Costa Rica occurs from May to October, when the ITCZ is positioned over Costa Rica and generates intense convective events (Clark et al., 2000). Clark et al. (2000) also report a dry transitional season from November to January, and a dry season from February to May. Some studies report a short, weaker dry season called *veranillos* within the main wet season in July–August (e.g., Waylen et al., 1996; Chazdon and Fetcher, 1984; Newstrom et al., 1993). Meteorological data from our study site indicates a longer wet season from May to December. Differences in observed seasonality are due to differences in locations, year-to-year variability, and degree of seasonality (Chazdon and Fetcher, 1984). Over the past decade, Costa Rica has seen an increase in ecotourism, particularly during the dry season. This adds stress to water resources when they are most in need by locals and farmers. Therefore, changes in climate or land–cover in Costa Rica could have significant impacts on local and regional water supply (Rhodes et al., 2010). Studying the interaction of surface characteristics, local water budgets, and microscale climate is therefore essential to addressing and implementing local resource management plans.

Characteristics of our study site

The 2.2-ha watershed in this study is located at the Texas A&M University Soltis Center for Research and Education in San Isidro de Peñas Blancas, Costa Rica (Figure 1). The Soltis Center is adjacent to the Monteverde Cloud Forest Reserve, in north-central Costa Rica. The watershed is located at an elevation of 455 m a.s.l., with an elevation range of 120 m and slopes from 12° to 55°. The mean annual rainfall at this site is approximately 4500 mm (Buckwalter et al. 2012), and monthly averages range from 136 mm during the dry season, to 512 mm during the wet season. Temperatures range from 21 to 24°C, however there is no significant difference between

temperature means among seasons. Although this forested watershed has long been classified as a tropical pre-montane cloud forest (TMCF), it appears to be transitioning into a tropical rainforest. According to Scatena et al. (2010), lower montane cloud forests usually transition into lowland evergreen rainforests at elevations where the temperature increases above 18°C. Whether this transition is due to changing climate or land-cover changes in the area, there is a need to study and quantify the present water budget to determine how the water supply from this watershed will impact and be impacted by local communities.
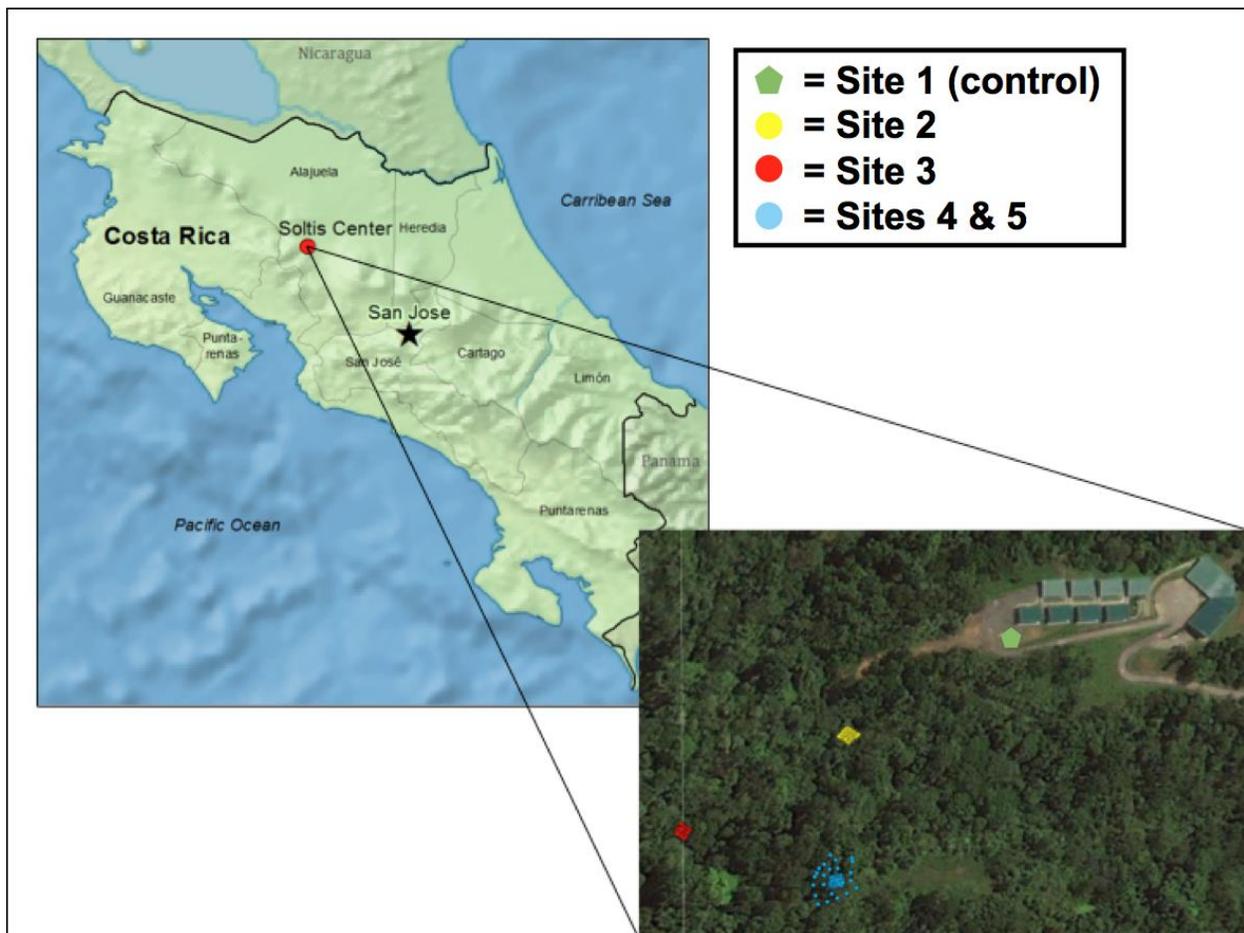

Figure 1. Watershed location on the Caribbean slope of the Cordillera de Tilarán, in San Isidro de Peñas Blancas, Costa Rica, and site locations within the watershed (based on Teale et al., 2014).

Quantification of throughfall amounts and variability

Local communities near this study site get their water from streams leaving the watershed. Streamflow and runoff leaving the watershed are tightly related to TF. This relationship, however, can be quite complicated due to the spatial and temporal variability caused by the redistribution of precipitation as it falls through the canopy. This study therefore focuses on providing a comprehensive quantitative assessment of the spatial variability of TF and determining its main drivers.

Influence of the canopy on throughfall variability

Previous research indicates that vegetation density has a strong influence on the spatial variability of throughfall (Lloyd and Marques, 1988; Whelan and Anderson, 1996; Zimmermann et al., 2009). As rain falls through the canopy, it is intercepted and redistributed by the canopy. This results in small-scale spatial variability of TF (Loustau et al., 1992; Staelens et al., 2006). According to Zimmerman et al. (2009), the spatial variability of TF increases with understory density, regardless of the canopy structure above.

Several studies have also found a negative relationship between canopy cover and the amount of TF reaching the forest floor (Burghouts et al., 1998; Llorens and Gallart, 2000; Loescher et al., 2002; Whelan and Anderson, 1996). More coverage would therefore lead to lesser TF amounts reaching the forest floor, as more leaf area implies more interception (Scatena, 1990; Gomez-Peralta, 2008).

Vegetation absorption during the process of interception is larger when a canopy is under unsaturated conditions (Brauman et al., 2010). An unsaturated canopy will therefore hold more precipitation than a saturated one, thus lowering the fraction of rainfall reaching the forest floor.

Canopy saturation depends on duration between events, vapor pressure deficits, event duration, and rainfall intensity (Cuartas et al., 2007; Levia and Frost, 2006; Staelens et al., 2008). Brauman et al. (2010) also found that the spatial variability of TF decreases when the canopy is saturated, as the differences in vegetation no longer influence canopy storage capacity uniformity.

Epiphytes play an important role in the interception process, as their high water storage capacity allows them to increase interception loss and canopy saturation (Pócs, 1980; Veneklaas and van Ek, 1990). However, Kohler et al. (2007) found that the impact of epiphytes on rainfall interception in Costa Rica is negligible, particularly during the wet season, when they remain close to saturation.

Vegetation morphology plays an important role in TF heterogeneity. TF percentages at a given point can range from 0 to 1000% of incident rainfall (Cavalier et al., 1997). Such values are due to variations in the canopy structure above the rain gauges, such as drip points from leaves and branches (Zimmermann et al., 2009). Drip points therefore contribute to the spatial heterogeneity of TF by concentrating TF. This heterogeneity is amplified by the fact that drip points are not active during all events, depending on the duration and intensity of the events (Loustau et al., 1992; Zimmerman et al., 2009). The impact of vegetation morphology on TF distribution is difficult to quantify, and may mask other relationships between canopy cover and TF.

Influence of meteorological parameters on throughfall variability

TF amounts and variability are largely influenced by the interaction between rainfall and canopy structure. Rainfall intensity is known to have a negative relationship with TF variability (Carlyle-Moses et al., 2004; Holwerda et al., 2006; Staelens et al., 2006; Sato et al 2011; Zimmermann et al., 2009). Events of higher intensity therefore cause a less heterogeneous redistribution of

rainfall as it falls through the canopy, as vegetation interception and absorption is weakened due to the higher momentum and larger size of raindrops (e.g., Brandt, 1990; Calder et al., 1996; Loustau et al., 1992; Sato et al 2011; Zimmermann et al., 2009). Longer events with larger rainfall amounts tend to have the same impact on TF variability. Low rainfall intensity is also attributed to higher evaporation losses. Scatena et al. (1990) therefore suggests that evaporation should be taken into account during such events.

During low intensity events, wind influence could lead to an underestimation of precipitation (Crockford and Richardson, 2000). However, some studies have reported that wind influence on precipitation over tropical forests is minimal (e.g. Bruijnzeel and Veneklaas, 1998; Proctor et al., 1988; Hafkenscheid, 1994).

Study objectives

The main goals of this study are to determine the main drivers of heterogeneous redistribution of throughfall, and quantify their effects on throughfall variability. This study will focus on identifying spatial throughfall patterns, and examining how vegetation density, terrain, rainfall intensity, and wind speed potentially drive them.

CHAPTER II

METHODS

Data collection

Throughfall measurements

The data for this study were collected at the above-mentioned 2.2-ha watershed at the Texas A&M University Soltis Center for Research and Education in San Isidro de Peñas Blancas, Costa Rica (Figure 1). Precipitation and TF measurements were made daily at five gauge networks throughout the watershed. Four of these networks were located under the canopy, and one control site was installed in a clearing at the edge of the forest (Figure 1). Among the four sites in the forest, three were classified as "hyper-dense" networks with 36 gauges each at 2-meter spacing (sites 2, 3, and 4), and one was an "extensive" (coarser) network of 21 gauges at 10-meter spacing (site 5), encompassing one of the hyper-dense networks. The gauges used were wedge-shaped Tru-Chek® Direct-Reading Rain Gauges that could measure from 0.1 mm to 150 mm of precipitation, and were installed at a height of 1 m from the ground. The total number of gauges sampled was 164, 129 of which were in the forest. Daily measurements were made at each gauge from 28 June–17 July 2012 and 12 June–16 July 2013. Precipitation and TF were typically measured during morning hours, or at the end of a rain event. Events for which the total amount of rainfall was less than 1 mm were not included in the data set for analysis (Staelens *et al.,* 2006).

Throughfall was characterized by calculating the percent TF received at each gauge compared to total precipitation. Coefficients of variation (CVs) across events were also calculated for each gauge under the canopy, based on mean TF received under the canopy.

Canopy density

Canopy density was determined above each gauge using hemispheric photographs. These photographs were taken at gauge height with a fish-eye lens on a Nikon D3200 digital camera. They were then analyzed using HemiView V. 2.4 to obtain leaf area index (LAI) and visible sky values for each gauge. These values were calculated at a range of angles from the normal (zenith angles) for a more accurate and complete estimate of canopy density (Zimmermann *et al.,* 2009). Subsequent analyses will be presented based on the 2.5° angle and for total (180°) LAI only.

Meteorological data

Tipping bucket rain gauges installed on portable Onset HOBO weather stations at each site measured precipitation amounts, therefore providing data such as event timing, duration, and intensity. site 1 also had a 10 m weather station, used as a control for the HOBO data. Average 10-m wind speeds and gusts were available from the control site weather station for the 2012 events only.

Site topography

Relative elevation data for each gauge were obtained through site surveys during the summer 2012. These data were produced for all sites relative to site 2.

Precipitation intensity and wind gusts

The effects of precipitation intensity and wind gusts on TF variability were evaluated using basic statistical analyses such as analysis of variance (ANOVA). All statistical analyses were evaluated for significance using a 95%-confidence level. Similar methods to those of Sato *et al.* (2011) and Staelens *et al.* (2006) were used to determine a 7.5 mm/hr threshold (Teale et al., 2014) for rainfall intensity, thus differentiating high intensity (convective rain) from low intensity (stratiform rain) events. Gusty and non-gusty events were separated based on the distribution of maximum winds speeds over 5 minute intervals. Events for which wind gusts exceeded the mean wind gust based on all 13 events (2.3 m/s) were classified as gusty. 9 events were therefore classified as gusty, with the remaining 4 being non-gusty.

ANOVA was performed for each site on the CVs and mean TF amounts of each gauge. The first ANOVA was performed with event classifications based on rainfall intensity, while the second ANOVA used wind gust distinctions.

Principal component analyses (PCAs) were then performed on the standardized TF values for each site. TF values were standardized with respect to the mean TF and standard deviation for all events under consideration. PCAs were therefore performed on the correlation matrix of TF values. PCAs identify the main patterns of variability (modes) and extract them from the data. These modes can then be related to physical parameters such as LAI or elevation. The relationship between the dominant modes and these parameters was then examined under varying intensity and gustiness conditions.

Canopy variability

The effects of vegetation density and canopy cover were determined using LAI values above each gauge. The loadings from the first and second patterns of variability (PCA modes 1 and 2) were then correlated with total LAI above each gauge, as well as LAI at a 2.5° zenith angle (Teale *et al.*, 2014; Zimmermann *et al.,* 2009). These analyses were conducted 1) on all 39 events, 2) the summer 2013 events separately, 3) all events separated by intensity, and 4) the 2012 events separated by gustiness.

Site topography

Correlation coefficients between relative elevation and the loading associated with modes 1 and 2 in the above-mentioned PCAs were calculated for each site.

CHAPTER III

RESULTS

Spatial variability of throughfall

The PCA indicated that each site had one very dominant PC, which accounted from 82.3% to 92.1% of the total variance over both years of the study (Table 1). The loadings from the first PC were plotted as a grid, representing the layout of each site. The first mode of sites 2 through 5 is shown in Figure 2.

Table 1. Percent of variance (%) explained by the 1st, 2nd, and 3rd principal components at each site under the canopy.

Site #	1st PC	2nd PC	3rd PC
2	92.1	2.1	1.7
3	88.9	4.4	1.8
4	88.1	6.6	1.2
5	82.3	6.5	3.4

Influence of topography

The topography at extensive network 5 seems to match the first and second patterns of variability at this site (Figure 3). The gauge network at site 5 was installed throughout a valley for which the difference in elevation is 25 m.
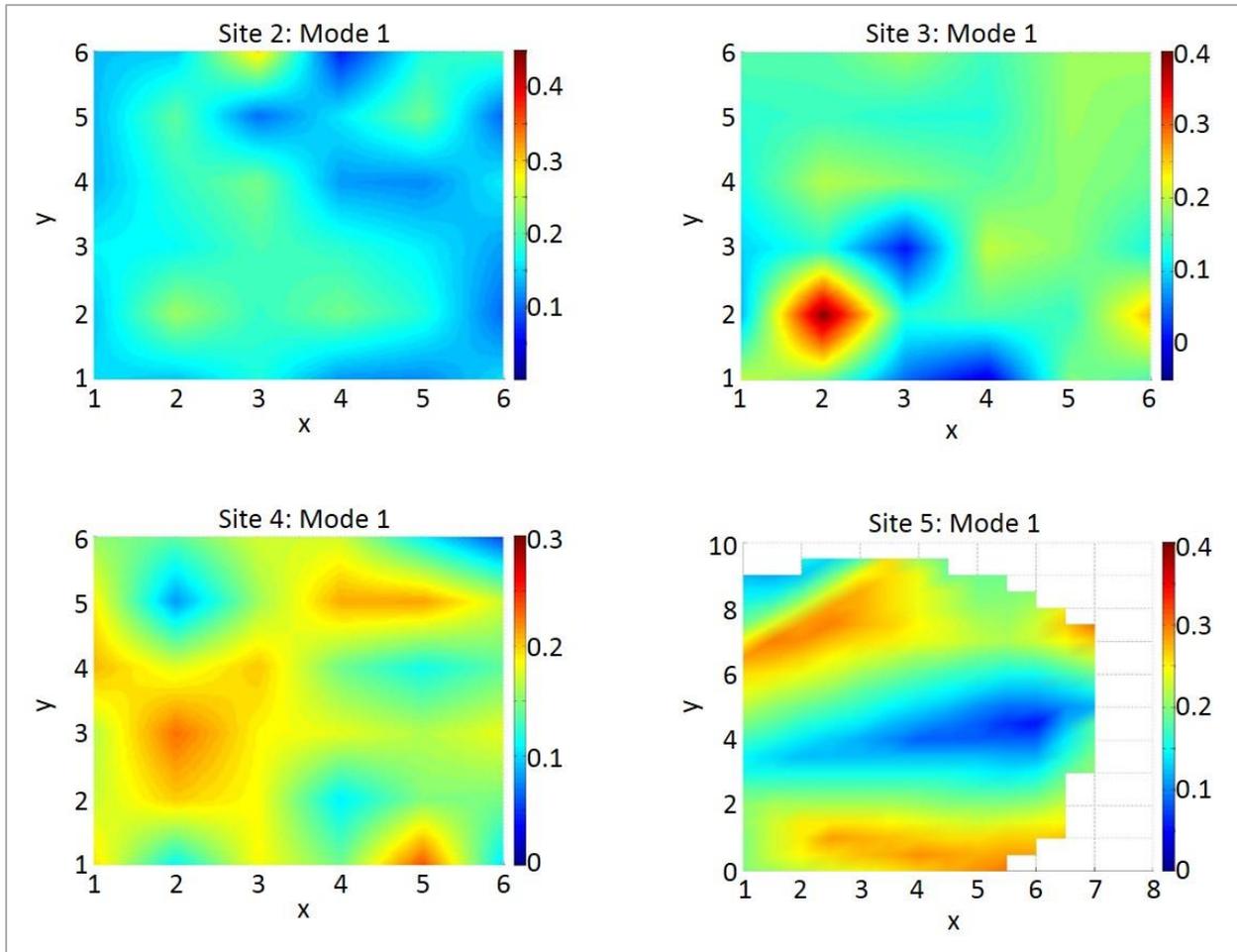
Figure 2. First pattern of throughfall variability (mode 1) for sites 2–5.
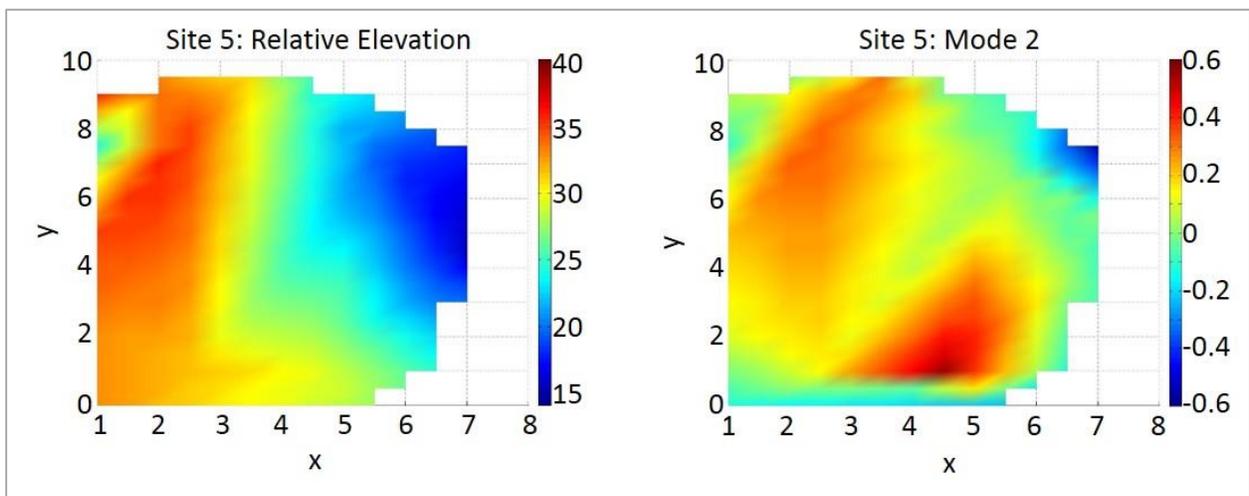

Figure 3. Site 5 topography (left) and second pattern of throughfall variability (right).

Results from the correlations between the PC loadings and elevation show a statistically significant correlation between the second pattern of variability (mode 2) and relative elevation at site 5 ($r^2 = 0.22$). Therefore, for the 2012–2013 period, site topography accounts for 22% of the variance of mode 2. However, this correlation becomes insignificant at site 5 during low intensity rainfall events.

While site 5 has the largest difference in elevation, relative elevation at site 4 is significantly correlated with mode 1 during low intensity events ($r^2 = 0.12$). This means that for low intensity rainfall events, site topography accounts for 12% of the variance in the first pattern of variability at site 4. No significant correlations were found between relative elevation at site 4 and mode 1 over the 2012–2013 period in general.

Although sites 2 and 3 have larger elevations differences (8.2 m and 10.9 m, respectively) than the elevation difference at site 4 (5.2 m), no correlations were found between relative elevation and any of the first 3 modes.

Influence of canopy density

Correlations between LAI and the loadings from the first PC were determined for the 2012–2013 period at each site. The coefficients of determination are reported in Table 2. There is no correlation between LAI and mode 1 at site 4. Mode 1 and LAI at a 2.5° zenith angle correlate at site 2 ($r^2 = 0.18$), so 18% of the variance of the first mode is explained by LAI at a small zenith angle. The first modes of sites 3 and 5 both correlate with the total LAI at each gauge ($r^2 = 0.12$ and $r^2 = 0.23$, respectively). Total LAI therefore accounts for 12% and 23% of the first pattern of variability, respectively, at sites 3 and 5.

Table 2. Coefficients of determination (r^2) at each site between loadings from the first PC (mode 1) and total LAI or LAI at 2.5° above each gauge for 2012–2013; ns indicates that correlations were not significant.

Site #	LAI 2.5°	total LAI
2	0.18	ns
3	ns	0.12
4	ns	ns
5	ns	0.23

Influence of meteorological parameters

Influence of rainfall intensity on TF variability

The ANOVA performed on gauge CVs under the canopy for high and low intensity rainfall events indicates that there is a statistically significant effect of rainfall intensity on CV. The mean gauge CV across low intensity rainfall events was statistically different from the mean gauge CV during high intensity events. This suggests that rainfall intensity affects TF variability. The percent of variance explained by the first PC decreases, and that of the second PC increases, for high to low intensity events (Table 3). However, a t-test reveals that these differences are not statistically significant.

Table 3. Percent of variance (%) explained by the 1st, 2nd, and 3rd principal components at site 4 for high and low intensity rainfall events.

	High intensity	Low intensity
1st PC	88.1	81.9
2nd PC	6.1	11.3
3rd PC	1.6	1.9

Effect of rainfall intensity on LAI contribution to throughfall variability

Correlations at each site between LAI and mode 1 were determined after separating rainfall events into "low" and "high" intensity events. The coefficients of determination for each analysis are listed in Table 4. At site 2, the correlation between LAI at 2.5° and mode 1 is larger during low intensity rainfall events ($r^2 = 0.31$). While LAI at 2.5° accounts for 18% of the variability of mode 1 at site 2 during high intensity events, it accounts for 31% of the variability during low intensity events. The same pattern is observed at site 4, for which there is no correlation between total LAI and mode 1 during high intensity events. During low intensity events, however, total LAI accounts for 12% of the variability of mode 1 at this site.

Table 4. Coefficients of determination (r^2) at each site between loadings from the first PC (mode 1) and total LAI or LAI at 2.5° above each gauge, for high and low intensity events; ns indicates that correlations were not significant.

Site #	High intensity		Low intensity	
	LAI 2.5°	total LAI	LAI 2.5°	total LAI
2	0.18	ns	0.31	ns
3	ns	0.15	ns	ns
4	ns	ns	ns	0.12
5	ns	0.22	ns	ns

The relationships at sites 3 and 5, however, are different. LAIs at sites 3 and 5 account for a greater percent of the total site variability during high intensity events. At site 3, total LAI accounts for 15% of the mode 1 variability during high intensity events, but does not relate to TF variability during low intensity events. This is also true at site 5, where total LAI during high intensity events initially accounts for 22% of the site's first mode of variability.

Influence of wind gusts

The ANOVA performed on TF CVs under the canopy for gusty and non-gusty events in 2012 indicate that wind gusts have a statistically significant effect on CVs, and therefore TF variability. The mean CV for all gauges under the canopy for gusty events (0.58) is statistically different from the mean CV for calm events (1.21).

The ANOVA was repeated for the percent TF received by all gauges during gusty and calm events and reveals that gustiness significantly influences the percent TF reaching the forest floor. The mean percent TF under the canopy for gusty events is 43%, whereas the mean percent TF for calm events is significantly lower at 30%.

The PCAs performed on TF patterns for gusty versus calm events did not lead to any significant correlations between the physical factors listed above and the main modes of variability. This may be due to a low event sampling (9 gusty events and 4 calm events).

Temporal persistence of spatial patterns

The scores associated with the first PCs for each site were plotted as a time-series in Figure 4. These graphs represent the amplitudes of mode 1 for each site across all 39 events. The amplitudes for the events in 2012 (events 1 through 13) are always negative. The oscillation between positive and negative scores in 2013 also demonstrates the variable nature of the system.
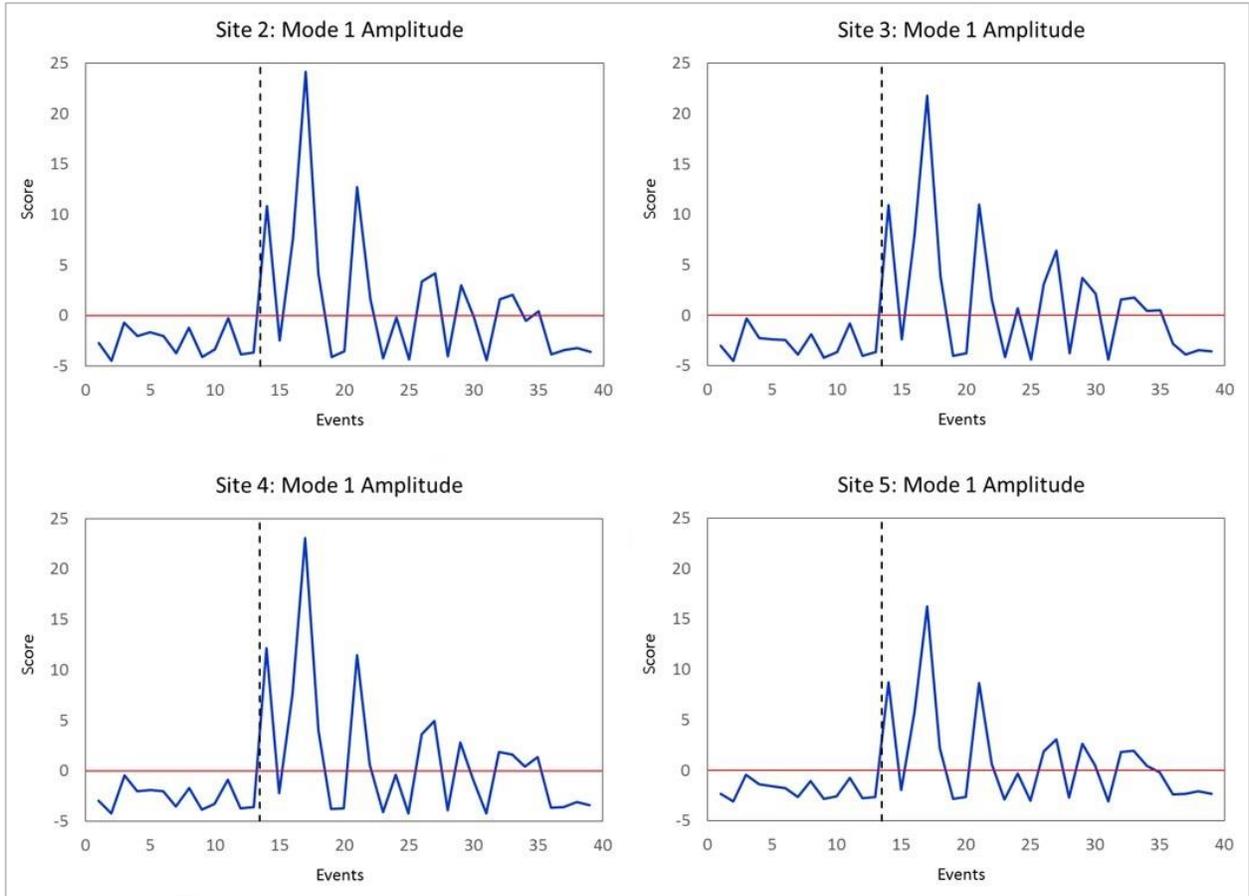
Figure 4. Amplitude of mode 1 for each site. The dashed line represents the separation between the 2012 and 2013 events.

CHAPTER IV

DISCUSSION

The dominant PCs in the spatial variability patterns at each site suggest that TF variability is indeed driven by certain physical factors. This study focused on topography, vegetation density, and meteorological parameters to explain the heterogeneous spatial patterns of TF.

Influence of topography

This study has shown that the effect of topography on TF variability varies from site to site. Because the spatial patterns associated with the first and second PCs at site 5 resemble its valley-shaped elevation variability, topography was expected to play a significant role in the spatial redistribution of TF. However, topography only explained about 22% of the variance of mode 2 at that site.

Even more surprisingly, topography at site 3 accounted for 12% of the TF variability in mode 1 during low intensity events. Because site 3 is one of the hyper-dense gauge networks and has a relatively low elevation range, the observed correlation between topography and mode 1 probably results from factors related to elevation, such as plant type and leaf morphology.

Vegetation types varied for each site. The ridges at site 5 typically had larger trees and denser mid or upper-story, while the lower parts of the valley had a denser under-story. The gauges at site 3 were also placed under different vegetation conditions, which could be the cause of the apparent relationship between site topography and TF variability.

Influence of vegetation density

While LAI has been shown to account from 0% to 31% of the variance in the first patterns of variability, this study demonstrated that the relationship between LAI and TF variability is inconsistent. Previous studies have also found this (Lloyd and Marques, 1988; Whelan and Anderson, 1996; Zimmermann *et al.,* 2009; Teale et al., 2014). The use of either total LAI or LAI at a 2.5° zenith angle, as suggested by Zimmermann *et al.* (2009), to characterize TF variability suggests that LAI may not be the best way to quantify vegetation density at such a small scale. This study also suggests that vegetation density is not a dominating driver of TF heterogeneity. Denser vegetation is expected to intercept more TF, and therefore redistribute water more heterogeneously. However, other vegetation parameters such as plant species, leaf morphology, and branch inclination play an important part in micro-scale TF variability by generating drip points (Zimmermann et al., 2009). Much of the TF spatial variability can therefore be attributed to interactions between TF and canopy structure, rather than vegetation density alone.

Influence of meteorological parameters

Rainfall intensity

This study has shown that rainfall intensity has a significant effect on TF variability, as well as on the relationship between spatial patterns of variability and vegetation parameters. However, this relationship seems unpredictable. Changes in rainfall intensity cause changes in rain drop kinetics, therefore changing the interaction between rain drops and the canopy (Brandt, 1990). During intense rainfall events, rain drops fall through the canopy with a higher kinetic energy.

This leads to a decrease in interception, and a less heterogeneous redistribution of TF. This study shows that for sites 2 and 4, the contribution of LAI to TF variability patterns becomes more significant when rainfall intensity decreases. However, this relationship was not observed for sites 3 and 5. The relationship between total LAI and TF variability patterns became more significant for events of higher rainfall intensity. This could be due to the fact that high intensity events tend to be accompanied by strong wind gusts. Stronger winds reduce the heterogeneous redistribution of TF and thus increase the significance of the relationship between open sky above each gauge and the amount of TF received (Levia, *personal communication*). Low intensity events and more still winds allow for a more heterogeneous redistribution of TF, which could mask the relationship between LAI and TF variability.

Wind gusts

Despite the expected effect of wind gusts on TF variability, this study suggests that differentiating gusty from non-gusty events does not affect the relationship between TF variability patterns and LAI. However, the event sampling for this part of the study was probably insufficient for characterizing TF variability.

2012 versus 2013 differences

The shift in mode amplitudes from exclusively negative in 2012 to more highly variable in 2013 suggest a change in TF variability amplitude from one year to the next. This pattern shift could be due to changes in vegetation parameters and/or varying meteorological conditions between 2012 and 2013. Timing between events and event characteristics affect canopy saturation conditions, which in turn affect TF redistribution. A saturated canopy will not intercept as much

rainwater, which will result in a less heterogeneous redistribution of TF (Brauman et al., 2010). The drying of the canopy between events can increase the canopy storage capacity, which leads to more water interception during the following event, thus contributing to a more heterogeneous spatial redistribution of TF. The negative scores in 2013 could therefore be due to drier canopy conditions before the measured event. A more in-depth study of the process of interception would therefore be useful for characterizing TF variability.

Conclusion

The findings of this study suggest that throughfall heterogeneity in this tropical pre-montane cloud forest in Costa Rica is driven by the complex interactions between multiple coupled factors. While rainfall intensity and vegetation density have been shown to significantly influence throughfall patterns, a more comprehensive and spatially extensive study also focusing on tree and plant species, canopy structure, and canopy saturation conditions would allow for a better characterization of throughfall heterogeneity in this watershed.

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