HYDROLOGY, HYDROCHEMISTRY AND IMPLICATIONS FOR WATER SUPPLY OF A CLOUD FOREST IN CENTRAL AMERICA

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HYDROLOGY, HYDROCHEMISTRY AND IMPLICATIONS FOR WATER SUPPLY OF A CLOUD FOREST IN CENTRAL AMERICA

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Cloud-forest ecosystems are important sources of water supplies for the expanding cities of Central America. Surprisingly, few stream flow records are available for watersheds in Central America which has a climate with distinct wet and dry periods. Consequently, the tropical hydrology of cloud-forest watersheds is not well studied. To contribute to the understanding of the hydrology of this important ecosystem and to narrow the knowledge gap of tropical hydrology with that of temperate zones, we instrumented four neighboring catchments, located within La Tigra National Park in Central Honduras. This experimental watershed site is part of the headwater catchment of the Choluteca River Basin which drains into the Pacific Ocean. Although rainfall increased with the elevation gradient, it could not explain the greater water yield from the cloud forest watershed compared with the neighboring three watersheds at lower elevations. Additionally, analysis of the stream flow records suggests that subsurface flow paths are the primary mechanisms in all watersheds. Baseflow and interflow were greater and lasted longer for the cloud forest watershed. Any direct runoff was originated from saturated areas or from rock outcrops. Statistical analysis (MANOVA) of the of the stream flow chemistry confirms that the cloud forest watershed has a longer residence time because all the elements linked to parental material (Ca, Mg, SO_4-S, Na, and SiO_2-Si) had significantly lower concentrations in the cloud-forest watershed than in the non-cloud-forests watersheds.

On the other hand, most elements associated with the immediate effect of rainfall events such as Dissolved Organic Carbon (DOC), nitrates, potassium, phosphorous, and iron were not significantly different between the four watersheds. Finally, using a simple water balance model we were able to simulate the observed daily discharges with the Nash Sutcliffe model efficiency index ranging from 0.67 to 0.91. The cloud-forest watershed had a distinctly smaller amount of available plant water and greater groundwater storage than the three non-cloud forest watersheds. This result is similar to the results obtained for an undisturbed paramo system in the Andes of South America. Consequently, protecting cloud-forests to maintain hydrologic processes overtime is critical for the sustained provision of clean water for the growing population of Central America.

Key Words: Central America, cloudforest, watershed hydrology, solutes concentration, hydrologic modeling.

BIOGRAPHICAL SKETCH

Luis Alonso Caballero Bonilla was born to Juan and Marta Caballero in La Libertad, Comayagua, Honduras. After graduation from high school and a year of social service as a school teacher, he won a scholarship to attend the National School of Agriculture in Catacamas, Olancho. He earned an agronomist degree in 1983 and subsequently worked as an agricultural technician for the Honduran Coffee Producer Association where he observed the relationships between people, forest, agriculture, clean water and the management of watersheds first hand.

Luis attended Kansas State University with sponsorship of the United States Agency for International Development (USAID) earning a Bachelor of Science degree in Agriculture in 1990. He then worked as a watershed technician for The Honduran Ministry of Natural Resources and later joined the Agronomy Department at the Pan-American School of Agriculture where he taught soil and water conservation. In 1995 he received a Master of Science in Earth Resources from Colorado State University.

Upon his return to the Pan-American School of Agriculture in 1996, he was appointed first as an Assistant Professor and then next promoted to Associate Professor. While teaching, he was also actively involved in undergraduate research and community outreach projects. As Luis's scientific interest in tropical watershed hydrology and water resources continue to evolve, in 2006 he was given the opportunity to pursue his Ph.D. at the Biological and Environmental Engineering Department at Cornell University.

In 2007 he was awarded a scholarship from the Organization of American States (OAS). He was also named and awarded a grant as one of the 2007 Canon National Park Science Scholar Program administered by the American Association for the Advancement of Science (AAAS) which allowed him to establish in Honduras the first experimental watershed in Central America to study the hydrological processes.

This dissertation is dedicated to those who passionately served as my teachers.

Dedicated also from my heart to our Honduran youth.

No matter what your economic or social conditions you can realize your dreams.

To my parents who knew that education is a door for success.

To my brothers and sisters.

To my beloved wife and children.

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I owe an immense gratitude and appreciation to a large number of people, without whom, my education and research would not have been possible. First, I would like to express my deep gratitude to my Special Committee Chair, Advisor and mentor Dr. Tammo Steenhuis for his invaluable teaching, support and guidance throughout my academic and dissertation research. His countless hours of work, sharing ideas, reviewing and editing manuscripts and, yes, pushing me to do what was possible are and will always be deeply appreciated. I know one thing; I could never have gotten this far without his encouragement and support.

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I must reserve the greatest measure of gratitude for my family and friends as they made the most sacrifices of all. Traveling for more than three years half of the American Continent collecting data and maintaining my close family relationships hasn't been easy, but your patience and understanding has been magnificent and is deeply appreciated. Ending this journey successfully would not have been possible without the help, support and encouragement of my parents Juan Francisco Caballero and my mom Marta Beatriz Bonilla. You are, dad and mom, after my mighty God, the roots of who I turned out to be. I am also in debt to my brother and sisters for pushing and helping me through my long schooling life.

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My academic program and research at Cornell University would not have been possible without the funding support from various sources, among them: A research assistantship by Cornell University through a USDA project lead by Professor Tammo

Steenhuis, an academic teaching assistantship granted by Dr. Steve DeGloria of The Crop and Soils Science Department of Cornell, The Association of American States (OAS) which partially supported tuition, fees and personnel expenses and the Canon National Parks Scholar Program which paid the research construction, equipment, travel and personal expenses, and finally, my brothers in Colorado Juan, Mario and Nerio who were there always to assist me when financial and other support was needed.

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INTRODUCTION

GENERAL INTRODUCTION

In many parts of the world water is in short supply, often in regions where the population growth is the most rapid (Kok, 2000). In Central America, the quantity and the quality of water is deteriorating in part due to increasing population and subsistence agriculture-linked forest clearing (UNEP, 2003; FAO, 1986; Bonell and Bruijnzeel, 2004; Bruijnzeel, 1992, 2002, 2004). Deforestation and the expected impacts of climate change on the hydrology have raised concerns for national governments (WMO, 1990; Middelkoop et al. 2001).

Water shortage in Honduras has already led to a crisis. According to the Pan-American Health Organization and World Health Organization., (2000) 98% of Honduran water systems provided water on an intermittent basis in 2000, averaging six hours per day. The drought of 2009-2010 resulted in an even more severe urban water supply crisis in the capitol city of Tegucigalpa. Water was provided only every three days due to low reservoir levels (La Prensa, 12/11/2009). In Honduras only 51% of urban drinking water systems disinfect the water, and only 3% of wastewater receives treatment before disposal (OPS-OMS, 2000). To make things worse, a high percentage of water in rural water systems becomes polluted during handling (Trevett, et al., 2005).

As the water crisis escalates, governments in Central America are desperately seeking dependable water supplies for their growing cities.

The major option is to access water flowing through springs and creeks from the remaining cloudforest. There is surprisingly little information available for water resource design and management purposes.

IMPORTANCE OF THE RESEARCH

This dissertation project is important for several reasons. First, few studies have been carried out in tropical watersheds. Even fewer in the upper headwater cloudforest influenced catchments, despite dramatic land use changes during the last three decades. Secondly, the hydrology of these ecosystems is poorly understood due to the minimal stream flow gauging data which along with the potential effects of climate change could further exacerbate the water supply crisis in Central America (UNDP, 2010; Fisher, 2010). Finally, as stated by Fisher (2010) “many nations are struggling to provide adequate supplies of good quality fresh water to their citizens, and in many countries most reservoir sites already have been utilized. To resolve these issues and others, water managers will require solid hydrologic data and descriptions of the hydrologic systems, particularly aquifers that we as water scientists must provide”.

To contribute to the understanding of the hydrology of this important ecosystem and narrow the knowledge gap with that of temperate zones, we instrumented four neighboring catchments, locating them within La Tigra National Park, Central Honduras. This experimental watershed site is part of the headwater catchment of the Choluteca River Basin, which drains into the Pacific Ocean (Figure 1.1).

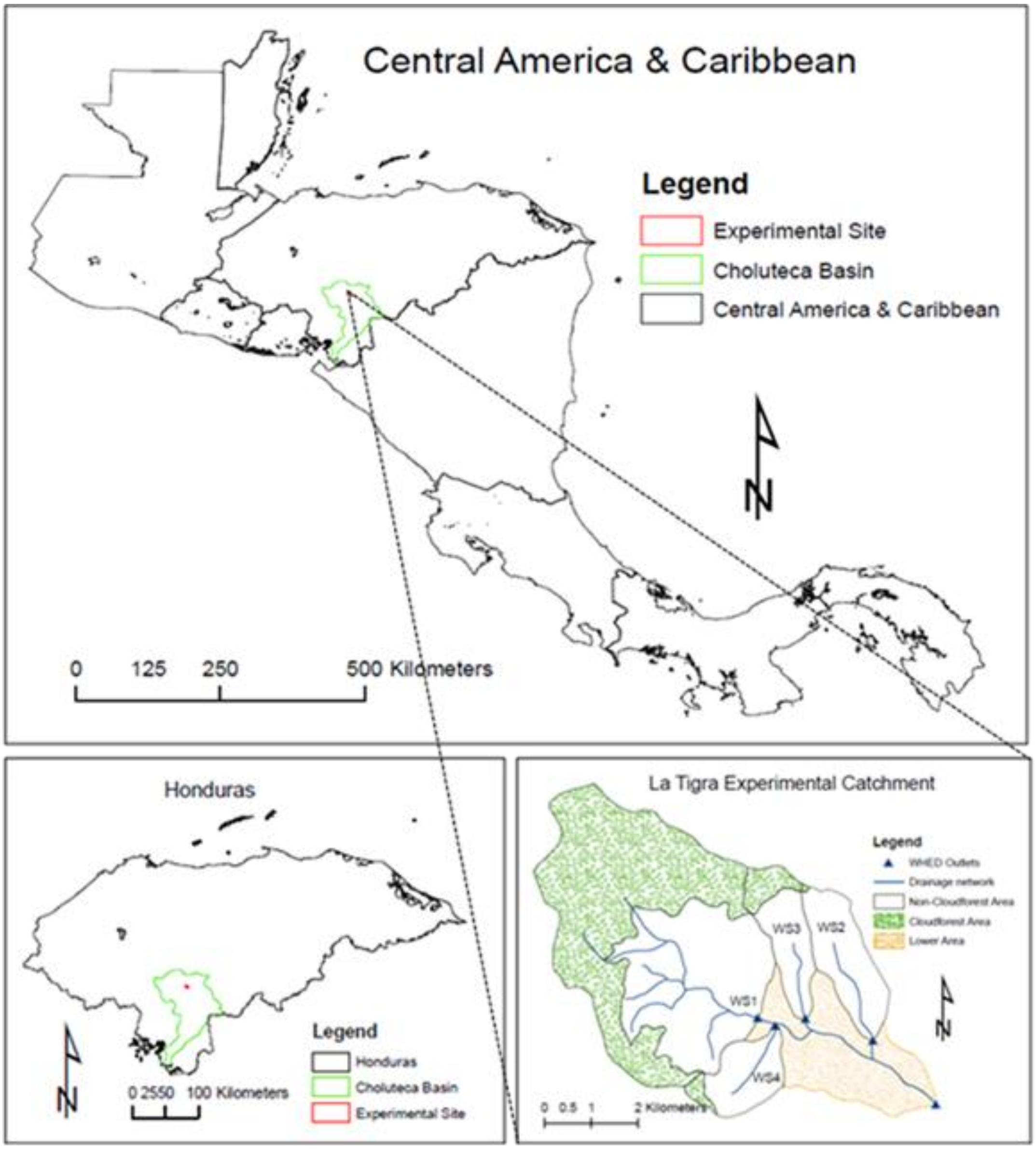

Figure 1.1. Location of La Tigra Experimental Watershed, Municipality of Valle de Angeles, Honduras, Central America.

DESCRIPTION OF THE PAPERS

Paper One (Chapter 2) is a contribution towards greater understanding of the hydrology of tropical catchments with and without cloudforest. It incorporates and expands on plot studies such as of Hanson et al. (2004) and Mendoza and Steenhuis (2002) in Honduras. The watershed instrumented for this purpose was the first in Central America. Although the period of observation is relatively short, according to Montanari et al. (2006), any insights that can be gained from the few well-gauged catchments that do exist can be valuable for engineering design practice and water resource assessments in other poorly gauged or ungauged catchments in the region. The paper focuses on the characterization of precipitation as being the major driving force in hydrologic processes, streamflow analysis and water balance comparison among watersheds. Final results suggest watersheds precipitation varies greatly along the elevation gradient, therefore a suitable number of rain gauges must be placed to reduce modeling error. Likewise, streamflow records analysis suggest that subsurface flowpaths are the primary mechanisms by which water flows out of these catchments. So consequently protecting these processes overtime is critical for the sustained provision of clean water.

The purpose of Paper Two (Chapter 3) is to evaluate the spatial and temporal variability in water chemistry between the cloud-forest and the non-cloudforest sub-watersheds, and elucidate the differences that can be attributed to variations in runoff mechanisms and flowpaths. Based on the results of streamflow hydrograph separation (Paper One), we hypothesized that despite large fluctuations in streamflow amounts between the cloud forest and the pine forest, the effects on stream chemistry would be minimal. This was dependent on both geology and water flowpaths which would be similar in the different watershed study sites. Finally, statistical analysis (MANOVA) suggests that the cloudforest does not exert significant influence on the chemistry of water flowing out of the catchments. Most elements such as Total Organic Carbon (TOC), nitrates, potassium, phosphorous, and iron were not significantly different in concentration between the cloudforest and the non-cloudforest watersheds. However, the cloudforest watershed had considerably different concentration for almost all elements (Ca, Mg, SO_4-S, Na, and SiO_2-Si) which are linked to parental material chemistry and water resident time. Thus, confirming that the cloudforest watershed being 7-fold bigger had a longer residence time for water and thus gave a different signature to its water chemistry.

Finally, in Paper Three (Chapter 4) we compared the hydrology of a cloud forest micro-watershed with the other three predominantly pine tree forested micro-watersheds, using a simple water balance model suitable for these environments. Our objective here was to test if the model was able to simulate the observed runoff hydrograph from a cloud-forest watershed and other micro-watersheds in the study site, and then use the model to infer differences in hydrologic behavior between cloud forests and other non-cloudforest watersheds. Final model results indicate that the model simulated well total streamflow discharges in all watersheds with Nash Sutcliffe model efficiency index (Nash and Sutcliffe, 1970) ranging from 0.67 to 0.91. This result was similar to what was found in paper two. The cloudforest watershed had a distinctly smaller amount of available plant water and greater groundwater storage which means longer residence times, resulting in watershed discharges that were four times greater than those of the other watersheds.

REFERENCES

Bonell, M. and L. A. Bruijnzeel. (2004). Forest, water and people in the humid tropics: past, present and future hydrological research for integrated land and water management. Cambridge University Press.

Bruijnzeel, L.A. (1992). Managing tropical forest watersheds for production: where contradictory theory and practice co-exist. In: Miller, F.R., Adam, K.L. (Eds.), Wise Management of Tropical Forests 1992. Oxford Forestry Institute, Oxford, pp. 37–75.

Bruijnzeel, L.A. (2002). Hydrological impacts of converting tropical montane cloud forest to pasture, with initial reference to northern Costa Rica. Project Memorandum Form, Project No. R7991 within the Forestry Research Programme ofthe Department for International Development of the UK, Aylesford, UK, 60 pp.

Bruijnzeel, L.A. (2004), Hydrological functions of tropical forests: not seeing the soil for the trees? Agriculture, Ecosystems and Environment 104:185–228

FAO. (1986). Tropical Forestry Action Plan. Committee on Forest Development in the Tropics, UN Food and Agriculture Organization, Rome.

Fisher, J. (2010). Hydrology and Water Management in the Humid Tropics. Proceedings of the Second International Colloquium 22-26 March 1999, Panama Republic of Panama. Technical Documents in Hydrology No. 52 UNESCO, Paris 2002.

Hanson, D.L., T. S. Steenhuis, M. F. Walter and J. Boll. (2004). Effects of Soil Degradation and Management Practices on the Surface Water Dynamics in the Talgua River Watershed in Honduras. Land Degradation & Development. 15:367-381

Kok, W. (2000). Prime Minister of the Netherlands, Welcome words in the second announcement for the Second World Water Forum and Ministerial Conference. http://www.worldwaterday.org/wwday/2000/

La Prensa News Paper. (2009). Tegucigalpa con la peor crisis de agua potable. Edición 12/11/2009. www.laprensa.hn.

Mendoza, G. and T. S. Steenhuis. (2002). Determination of hydraulic conductivities of hillsides with a hillside infiltrometer. J. Soil Science Society of America Journal: 66: 1501-1504

Middelkoop, H., K. Daamen, D. W. Gellens, J. C. J. Kwadijk, H. Lang, W. A. H. Parmet, B. Schädler, J. Schulla and K. Wilke. (2001). Impact of climate change on hydrological regimes and water resources management in the Rhine Basin. *Climatic Change* 49: 105–128, 2001. © 2001

Montanari, L., M. Sivapalan and A. Montanari. (2006). Investigating of dominant hydrological processes in a tropical catchment in a monsoonal climate via the downward approach. Hydrol. Earth Syst. Sci., 10: 769-782.

Nash, J. E. and J. Sutcliffe. (1970). River flow forecasting through conceptual models, Part I. A discussion of principles. J. Hydrol. 10: 282–290.

Pan-American Health Organization and World Health Organization. (2000). Evaluación de los Servicios de Agua Potable y Saneamiento en las Américas (www.bvsde.ops-oms.org/eswww/eva2000/Honduras/caliserv/calidad.html

Trevett, A.F., R. C. Carter and S. F. Tyrrel. (2005). Mechanisms leading to post - supply water quality deterioration in rural Honduran communities. Int. J. Hyg. Environ.-Health 208: 153–161

UNDP. (2010). Adaptation Fund Board Secretariat. Addressing Climate Change Risks on Water Resources in Honduras: Increased Systemic Resilience and Reduced Vulnerability of the Urban Poor. UNDP PIMS 4399; Atlas IDs-Proposal 00060323, Project 00075904, HND10).

World Meteorological Organization (WMO). (1990). Second World Climate Conference. Geneva, Switzerland.

CHAPTER 2

RAINFALL RUNOFF RELATIONSHIPS FOR A CLOUDFOREST WATERSHED IN CENTRAL AMERICA: IMPLICATIONS FOR WATER RESOURCE ENGINEERING

Luis A. Caballero,[1] Alon Rimmer[2] and Tammo S. Steenhuis[3,4]

ABSTRACT

Understanding the basic relationships between rainfall and runoff is vital for effective management and utilization of scarce water resources in Central America with widespread potable water shortage during the dry months of the monsoon. Potential good water sources for the dry season are the forested lands and especially cloud forests, but little information concerning its potential is available to water supply engineers. Our objective is to define rainfall-runoff-baseflow relationships for forested watershed. Flumes were installed for measuring river discharge in four sub-watersheds in La Tigra National Park, Central, Honduras. This included a 636 ha sub-watershed with more than 60% cloud forest coverage. Precipitation averaged 1130 mm/year. About half of the total rainfall became runoff for the cloud forest watershed while the discharge was less that 20% of the precipitation in the adjacent undisturbed forested watershed. Infiltration rates were generally greater than rainfall rates. Direct

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runoff was generated over a maximum 20% of the watershed that consisted of the saturated areas near the river and exposed covering This research provides compelling evidence that baseflow from the cloud forests was the primary contributor to stream flow during dry season. Protecting these catchments is critical for the sustained provision of potable water.

Key Terms: Central America, tropical hydrology, cloudforest headwater catchments, runoff generation, water balance, rainfall-runoff relationships.

INTRODUCTION

In many parts of the world, water is in short supply, often in regions where the population growth is the most rapid (Kok, 2000). In Central America, the quantity and the quality of water is deteriorating, in part because of increasing population and subsistence agriculture-linked forest clearing (UNEP, 2003; FAO, 1986; Bonell, 1993; Bonell and Bruijnzeel, 2004; Bruijnzeel, 1992, 2002, 2004). Deforestation and climate change have national government’s concerned (WMO, 1990; Middelkoop et al. 2001). In Honduras, water shortage already has led to a crisis. According to the Pan-American Health Organization (PHO-WHO, 2000), 98% of Honduran water systems provided water on an intermittent basis in 2000, for an average duration of 6 hours per day. The drought of 2009-2010 resulted in an even deeper urban water supply crisis in the capitol city of Tegucigalpa, where water was provided only twice a week or even once every week due to reservoir low levels. In Honduras, only 51% of urban drinking water systems disinfect water, and only 3% of wastewater receives treatment before disposal (PHO-WHO, 2000). A high percentage of water in rural water systems becomes polluted during handling (Trevett et al., 2005).

One of the main problems for the governments in Central America to solve the water supply crises is that there is surprisingly little information available for water resource design purposes (Montanari et al., (2006). Since most hydrologic engineering research has been carried out in the temperate regions, which are not directly applicable to tropical monsoon climates (Bonell, 1993; Bruijnzeel 1990; Bonell and Bruijnzeel, 2004), it can be argued that the only similarity between temperate and monsoonal climates is that both have dormant periods and a growing period. In temperate climates, growth is temperature-limited in the dormant season, during which there is typically plentiful precipitation and little evaporation, with the result that soils wet up and watershed outflow increases. In contrast, in monsoonal climates, the limiting dormant season factor is insufficient rainfall, with the consequence that the soils dry out and discharge from the watershed decreases. The effects of climate on the hydrology during the growing season are more complicated, but a simplified comparison is that while the landscape dries out in temperate climate growing seasons, the opposite is true for a monsoonal climate. As a consequence, the amount of runoff resulting from a given storm increases as the watershed becomes wetter throughout the rainy season (Liu et al, 2008; Kohl, and Markart, 2002; Merz and Blöschl, 2009). Thus for a monsoonal climate, the runoff response increases throughout the wet season until some steady state is reached. For example, in Ethiopia, this plateau is reached after 500 mm of effective rainfall (Liu et al., 2008). These differences in how climate interacts with the hydrology indicates that only engineering (mechanistically-based) models can be realistically applied in both climate regimes, whereas statistical techniques, such as the SCS curve number approach (Steenhuis, 2009), will be less successfully transferable.

One of the main runoff mechanisms in a monsoon climate is the saturation excess runoff. Saturation excess best explained runoff patterns in Australia (Steenhuis et al., 1995), the highlands of Ethiopia (Steenhuis et al., 2009; Bayabil et al., 2010), the monsoonal climate in China (Hu et al., 2005), Spain with a long dry season (Merz et al., 2006), and Nepal (Lange et al., 2003). For Honduras, Mendoza and Steenhuis (2002) found that the infiltration rates were generally greater than the rainfall intensity, implying saturation excess. Hanson et al. (2004) found that infiltration rates drastically decreased when land use changed in the Talgua River watershed of Honduras from primary forest (>840 mm hr^{-1}) to coffee plantation (89-109 mm hr^{-1}) to degraded grassland (8-11 mm hr^{-1}) where all macropores were filled with sediment. Thus, the switch from primary (cloud) forest to heavily grazed areas usually leads to a strong increase in storm flow volumes and peak flows (Gupta et al., 1974; Gupta et al., 1975; Bruijnzeel and Bremmer, 1989) due to shifts toward infiltration excess runoff.

This paper is a contribution towards greater understanding of the overall hydrology of tropical catchments with and without cloud forests. This study focuses on the characterization of precipitation and long- and short- term runoff relationships in Central America and specifically Honduras. The watershed that was instrumented for this purpose was the first in Central America. Although the period of observation is relatively short, according to Montanari et al. (2006), any insights that can be gained from the few well-gauged catchments that do exist can be valuable for engineering design practice and water resource assessments in other poorly gauged or ungauged catchments in the region. The study incorporates and expands on plot studies such as those of Hanson et al. (2004) and Mendoza and Steenhuis (2002) in Honduras.

MATERIALS AND METHODS

Experimental catchments: The study catchments are located within the La Tigra National Park of central Honduras and are part of the headwater catchments of the Choluteca River Basin which drains into the Pacific Ocean (Figure 2.1). Watershed relief is dominated by moderate to steep slopes, with mean slopes ranging from 20 to 30 percent. Elevation ranges from 1374 to 2270 m. The experimental site consists of four contiguous small catchments together comprising 880 ha. The largest catchment, WS1 (635 ha), located at the upper part of the experimental basin is the most important water source. This catchment is drained by a second order perennial stream, the Carrizal River, which serves as the main water source from the park for Tegucigalpa, the capital city of Honduras. Catchments WS2 and WS3 have relatively similar contributing areas of 93 and 82 ha, respectively (Table 2.1). They also share similar geomorphic characteristics related to catchment form, stream length, stream slope, elevation range and drainage density. The fourth catchment WS4 (70 ha) has an oval shape with a higher relief (30%) and shorter stream length. Table 2.1 provides a description of the primary geomorphologic features of the four catchments.

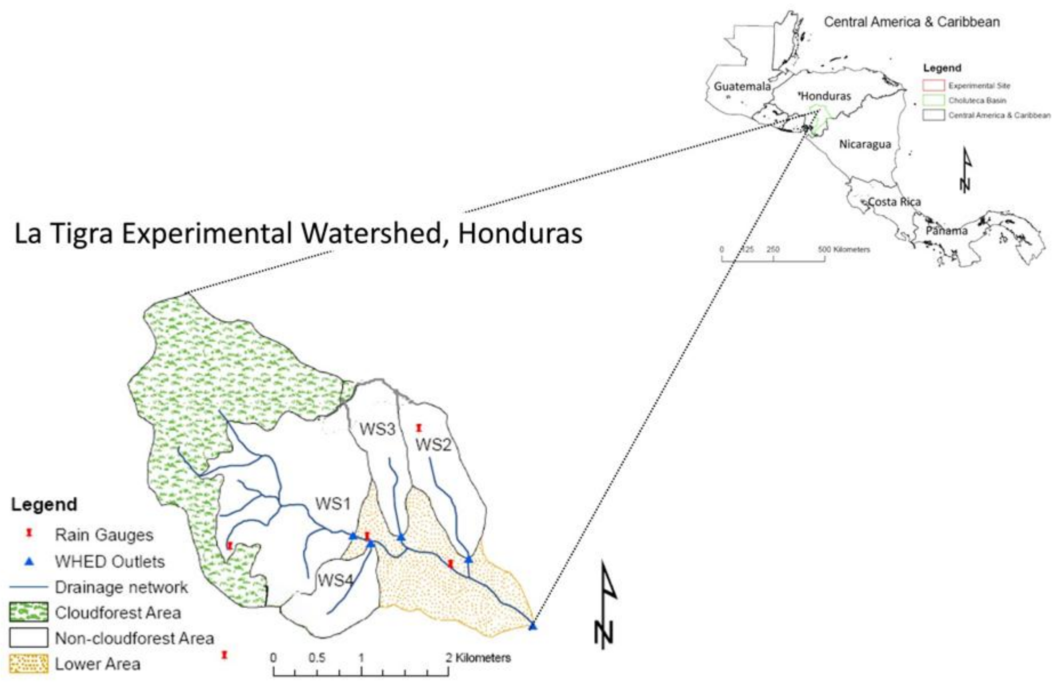

Figure 1:

Figure 2.1. Study site, La Tigra Experimental Catchment, Honduras C.A.

Table 2.1 Characteristics of the four study catchments and their rivers in La Tigra National Park, Honduras, Central America.

	CATCHMENT			
	WS1	**WS2**	**WS3**	**WS4**
Catchment area (ha)	635	93	82	70
Cloud forest area %	58	0	4	0
Other forested %	41	100	96	84
Deforested %	1	0	0	16
Weir elevation at outlet (m)	1505	1374	1431	1486
Elevation range (m)	1505-2270	1374-1850	1431-2000	1486-1960
Mean elevation (m)	1905	1625	1730	1715
Mean slope (%)	22	20	27	30
Main stream channel length (m)	6600	1508	1105	994
Main stream channel slope (%)	18	14	18	21
Drainage density (km/km^2)	1.00	1.62	1.35	1.42
Mean annual temperature (ºC)	16-20	16-20	16-20	16-20
Mean annual precipitation (mm)	1085	1085	1085	1085
Mean annual discharge (mm)	520	-	-	-
Geology (bedrock formation)	Volcanic	Volcanic	Volcanic	Volcanic
Period of measurements	Apr 2008-Dec 2009	Apr 2008-Dec 2009	Wet season 2008-2009	Wet season 2008-2009

Climate: The climate is characteristic of monsoonal regions with distinctive wet and dry phases. The wet phase begins in the end of May and continues through October and contributes to 90% of the annual precipitation (Hastenrath, 2002). Convective storms dominate from May through mid-July, and frontal systems from mid-August to the end of the wet season. Precipitation amounts are reduced from mid-July through mid-August during a period called *canicula* (lack of precipitation) throughout Central America (Guswa et al., 2007). The dry season begins in November and lasts through May. Long-term climate data recorded in Zamorano University (35 km from the watersheds and at 800 m elevation) indicates a long term (>50 yrs) average precipitation of 1100 mm. Monthly available precipitation data for Zamorano and La Tigra is presented in the supplementary material (Figure A2).

Land cover: La Tigra National Park is the major water source for Tegucigalpa and has been under protected status since 1950 when it was declared a Forest Reserve. Pine forest is predominant in the lower elevations (1300-1700m), with mixed forest at higher elevations (1800-2400 m). WS1, the largest and most important water source, is 99% forested and has the greatest area (58%) under cloudforest (Figure 2.1). Two other watersheds (WS2 and WS3) are in pine forests. WS4 is the smallest watershed. Sixteen percent of the watershed is deforested, with one small farming site (2.6 ha) dedicated to horticulture, and another (8.5 ha) to grain production (Table 2.1). An underground tunnel crosses the entire catchment for conducting the water from WS1 to a treatment plant for the City of Tegucigalpa.

Soils and geology: The study site is underlain by soils formed on volcanic parent material mapped as primarily basaltic and andesitic magma (Rogers and

O'Conner, 1993). Igneous rocks, such as basaltic and andesitic ash, and tuff and carbonate rocks and clastic sediments are predominant in the area (Carpenter, 1954). According to the geologic classification by the Honduran Geographic Institute, the dominant geological groups are the Padre Miguel (Tpm) and Valle de Angeles group (Kva). Tpm formation belongs to the Cenozoic era, composed of andesitic and ryolithic pyroclastic rocks and volcanic tuffs, dominating more than half of the northeast of the study site. Kva formation belongs to the Mezosoic era and it is primarily composed of heterogeneous redbeds and Jaitique limestone from continental marine environments (Simmons and Castellanos, 1968; IGN, 1956).

Field installation for data collection: Field equipment installed for climate and hydrologic monitoring included four digital rain gauges, four controlled weir streamflow measurement stations, and eight shallow water table wells (Figure 2.1).

Gross precipitation was measured from May 17, 2008 to January 1, 2010 using three digital tipping bucket rain gauges (RGM-3 HOBO® Data Logging Rain Gauge; 0.2 mm resolution per tip) at elevations of 1350, 1450, and 1800 m. Data from July 17, 2008 through July 30, 2008 was lost due to electronic problems.

To assess whether infiltration excess could occur, we selected 45 storm events greater than 15 mm day^{-1} for intensity analysis (Table A2 in Supplementary Material), for which duration, maximum intensity, hourly average intensity and overall mean and standard deviation were calculated. Soil saturated hydraulic conductivity ranged from 1.6 to 7 cm hr^{-1} (Lavaire. and Fiallos, 2010).

Stream discharge was measured at each catchment outlet using a controlled concrete weir structure (Figure A1 in Supplementary Material). At WS1 catchment, a suppressed rectangular weir was used while at the other three catchments (WS2, WS3 and WS4), a concrete V-shaped weir with metal edges was constructed. We followed the recommended U.S. Bureau of Reclamation (USBR) construction standards for these weirs (USBR, 1997). A Global Water WL-15 or WL-16 data logger measured water height in all gauges every 10 minutes during the wet season and every hour during the dry season. Manual readings were taken at each gauge twice a week as a means of quality control. Logger calibration was carried out when needed. For all weir sites, water height was transformed into streamflow rates using standardized equations developed by USBR (1997) and proposed in ISO (1980). (Table A3 in the supplementary material) provides a description of the main characteristics and stage discharge relations for all weirs. Reliable streamflow records were available for WS1 (October 2008 – October 2009), and thus rainfall-runoff relationships and water balance analysis focus on that micro-catchment. Continuous streamflow records for WS2, WS3 and WS4 were available for the 2009 rainfall season (June to October, 2009), and thus we used them for preliminary analysis of water balances.

Hydrograph separation: Hydrograph separation was performed using an automated Recursive Digital filter originally proposed by Lyne and Hollick (1979) and Nathan and McMahon (1990) for signal processing and modified by Arnold and Allen (1999) and Arnold et al. (1995) for baseflow separation.

$$\text{for } q_t > 0 \tag{1}$$

where q is the filtered rate of quickflow (cm/day), Q is the observed total discharge (cm/day), and α is the filter parameter (in this paper set to 0.925). The subscripts t and t-1 are the time indices. When Eq. 1 calculates a negative quickflow, then the quickflow q_t is set to zero. The baseflow is obtained by subtracting the quickflow from the total flow for each time step.

RESULTS AND DISCUSSION

Precipitation: A total of 2,314 mm of spatially averaged rain fell during the measurement period (May 17, 2008 through December 31, 2009, Table A1: supplementary material). According to Zamorano's weather station data, 2008 had average rainfall while 2009 was 29% below the long-term average due to the occurrence of the so-called El Nino Southern Oscillation (ENSO) phenomenon. For hydrology, we evaluated the amount of rain per storm, the rainfall gradient with elevation and the intensity of the rainfall. Ninety percent of the 368 registered events had less than 20 mm of precipitation per day. The average amount of rainfall (using the three rain gauges) for each event was 5.9 mm day^{-1} and the maximum was 69 mm day^{-1} (Figure 2.2).

The three rain gauges were located at 1350, 1450, and 1800 m elevation and offer an opportunity to evaluate the rainfall increase with elevation. The data were plotted as cumulative amounts for the period with mainly convective storms from the period June 10 through August 27, 2009 and with mainly frontal systems from September 10 through October 16, 2008 (Figure 2.3a). In June, the rain gauge at 1350 m had about 100 mm less than the other two gauges. The rain gauge at 1800 m recorded a storm in the beginning of July comprising 75 mm of rainfall while the other two locations had only 10 mm of rain. In the remaining part of the July and August the precipitation at all three locations were similar with some orographic effect for the upper gauge (Figure 2.3b). There was a small but consistent orographic effect in frontal rainfall (Figure 2.3a). The lower elevation rain gauge registered 321 mm while the middle and the upper registered 344 and 343 mm, respectively. One study, 25 kilometers south of our site in Uyuca Mountains, found similar increase in rainfall with elevation,

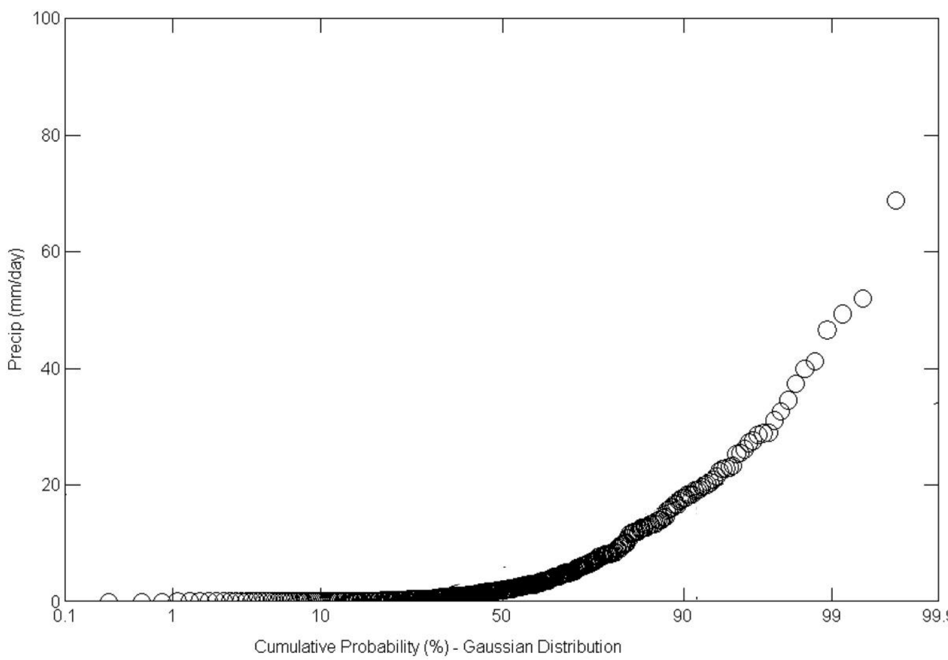

Figure 2.2. Cumulative probability of 368 rainfall events ranging from 0.5 to 68.9 mm. during measurement period 2008 and 2009.

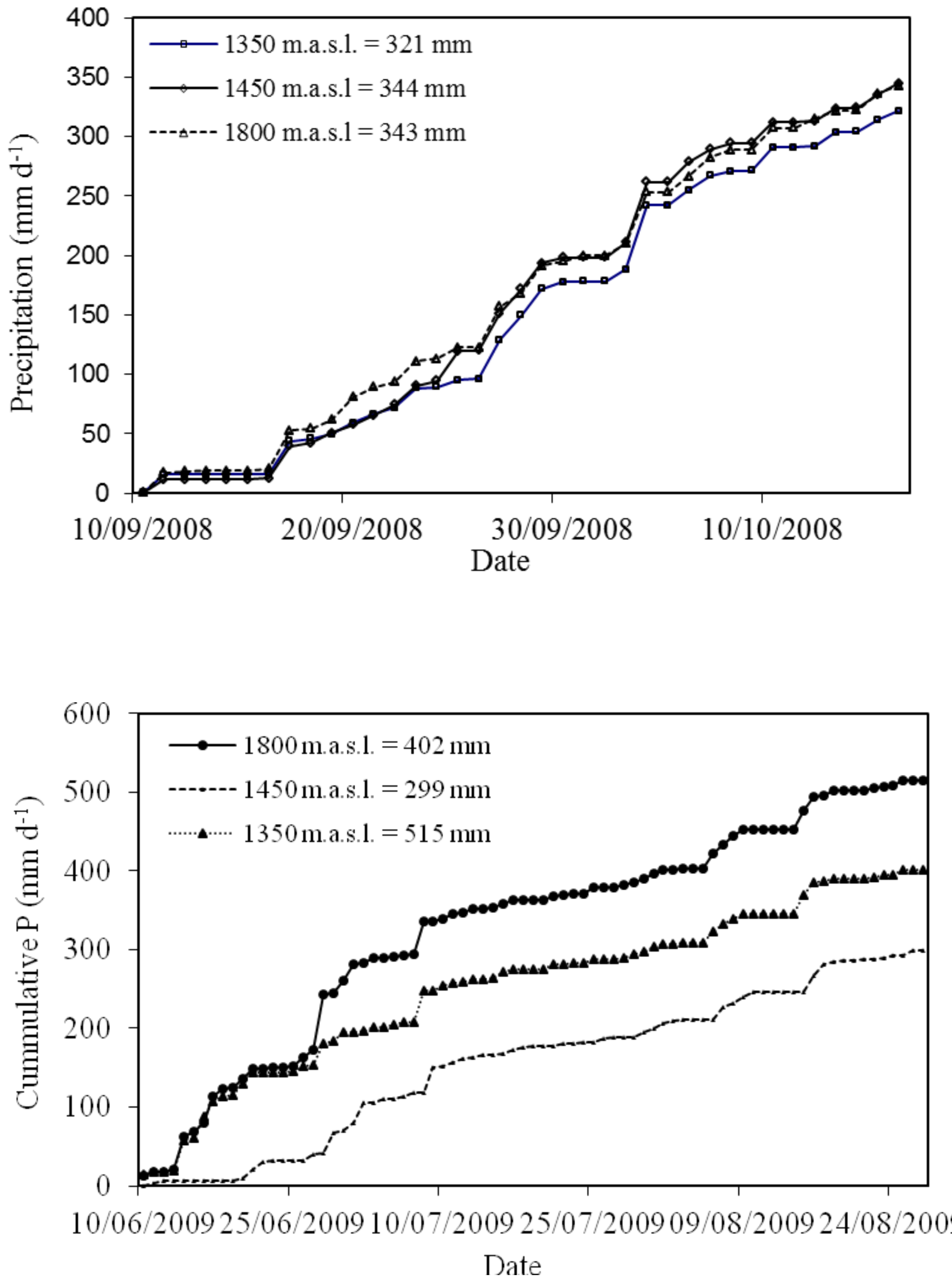

Figure 2.3. Cumulative rainfall amount for three rain gauges located along the elevation gradient reflecting effects of frontal systems and elevation: a) September-October 2008 rainfall season; b) May-August 2009.

including fog contribution in the order of 0.25-1.3 mm day^{-1} (Stadtmüller and Agudelo, 1990). In Arizona's Walnut Gulch Experimental Watershed, Goodrich et al. (1995) and Chaubey et al, (1999) found a first-order rainfall drift between 4-15% over 100 m elevation difference.

Rainfall intensity and infiltration measurements: The instantaneous rainfall intensity was calculated for the 45 events with more than 15 mm of rain per day (Table A2; Supplementary material). The maximum event precipitation intensity ranged from 3.2 to 36.6 mm hr^{-1} over periods ranging from 1 to 2.3 h**.** The mean maximum intensity was 15.3 mm hr^{-1}. The mean storm intensity for all events was 7.1 mm hr^{-1} ± 5 mm hr^{-1} (Table A2). Two of the most intense storm events during the period of measurements (47 mm hr^{-1} for 10/4/2008 and 35.4 mm hr^{-1} for 6/6/2009, Table A2) were analyzed in 5-minute time steps and resulted in a maximum recorded 115 mm hr^{-1} as the greatest intensity observed. These storms intensities were nevertheless well below the soil's infiltration capacity (809 mm hr^{-1}) reported by Hanson et al. (2004) and Mendoza and Steenhuis (2002). Likewise, recent laboratory measurement of disturbed soil samples resulted in saturated hydraulic conductivity values ranging from 1.6 to 7 cm hr^{-1} for both the cloud and non-cloud-forests (Lavaire and Fiallos, 2010). Thus, infiltration excess runoff would be a rare occurrence on these forested catchments. This confirms that surface runoff from the forested land is generated by saturation excess flow or exposed bedrock and roads.

In Honduras, precipitation measurements are taken on a daily basis and there is no published local data available to compare our results. The closest station for which hourly data have been analyzed is Monteverde, Costa Rica. During a 66 day study period in 1996, the mean precipitation intensity value was 3.0 mm h^{-1} and median 1.9 mm h^{-1} (Schellekens et al., 1999). In another study performed in the same area

(November 2004 through May 2005), only one storm had a peak precipitation of 11 mm h^{-1} (Guswa et al., 2007). Studies in two tropical mountain rain forests in Colombia found averages of 3.0-5.0 mm h^{-1} (Veneklaas and van Ek, 1990).

Streamflow discharge: Daily baseflow was separated from quickflow (surface runoff and interflow) for the available data of the four watersheds with the automated recursive digital filter (Eq 1) with one pass and the filter value α set to 0.925. The resulting baseflow separated hydrographs for watershed are shown as Figure 2.4**.**

Two analyses were made: all four watersheds were analyzed for the period from June to October 2009 when the discharge for all watersheds was available. To see if this period was representative over a longer time period, we also analyzed WS1 for a whole year starting in October 2008.

The amount of runoff between the watersheds during the 2009 rainy season differed greatly (Table 2.2). Of the 390 mm rainfall, 194 mm became runoff at the outlet of WS1 reflecting a water yield (i.e., quotient of total discharge and precipitation) of 50%. For the other watersheds at a lower elevation and with significantly less cloud forest (WS2, WS3 and WS4), the yield was much smaller (Table 2.2) ranging from 7 to 18%. Most discharge generated was in the form of baseflow. The ratio of direct runoff to precipitation can be used to estimate the average area that contributes to saturation excess overland flow. This was 8% or less for all watersheds. Thus most in-coming rainfall infiltrates the soil and either leaves the watershed via baseflow processes or as evaporative losses.

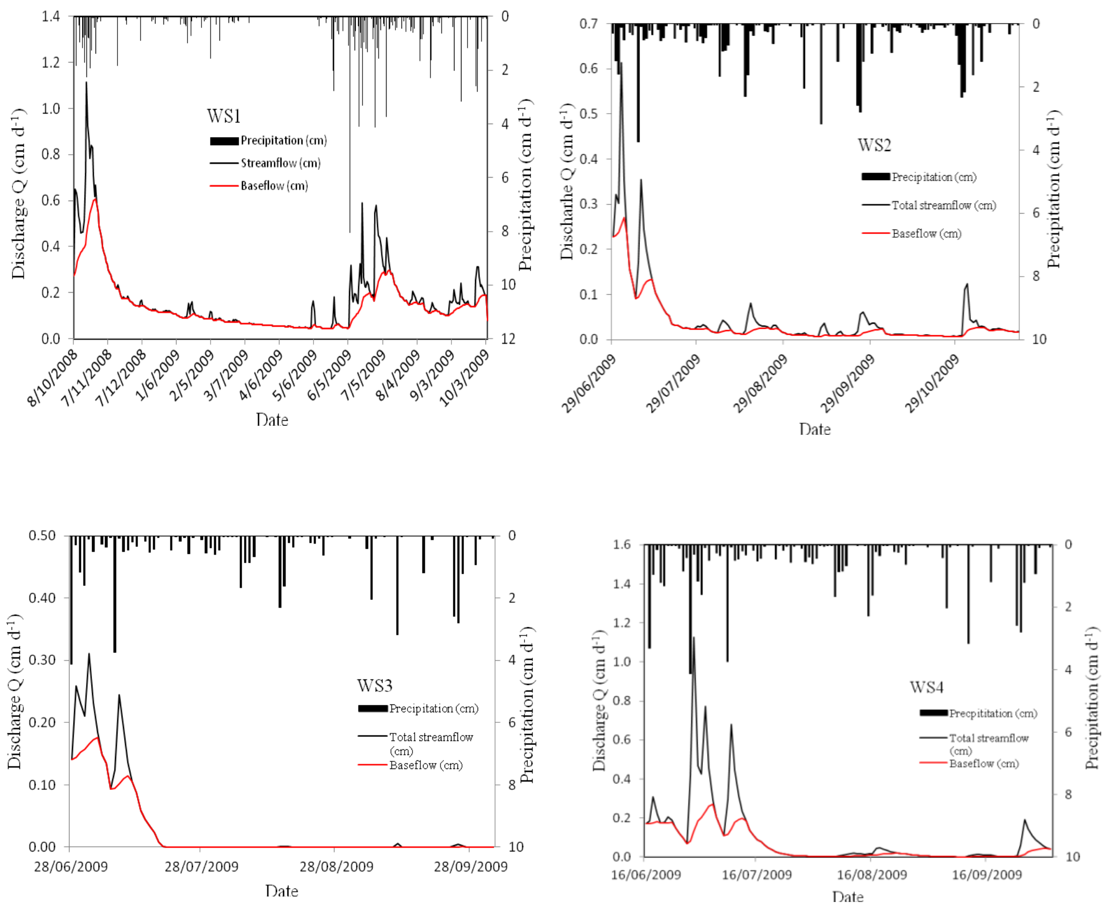

Figure 2.4. Baseflow separated hydrographs for the study catchments.

Table 2.2. Summary of water balances for watershed W1 from October 1 2008 to September 30 2009 and for watersheds W2, W3 and W4 from June 1 –September 30 2009

Parameter	WS1	WS2	WS3	WS4	WS1*
Precipitation, P (mm)	390	390	390	390	1100
Total discharge, Q (mm)	194	60	29	71	630
Baseflow, BF (mm)	161	43	21	38	520
Direct runoff, RF (mm)	32	17	7	33	110
Yield Q/P (%)	0.50	0.15	0.07	0.18	0.57
RF/P (%)	0.08	0.04	0.02	0.08	0.10
Baseflow index BF/Q	0.83	0.72	0.74	0.54	0.83

The wet season data for WS1 (June to October, 2009) compared well with the whole year data of 1110 mm annual precipitation; 630 mm became streamflow of which 520 mm was baseflow and 110 mm quickflow (i.e., interflow and overland runoff). Thus only 10% of the rainfall became runoff and 90% of the rainfall infiltrated, split almost equally between evaporation and baseflow. The baseflow index of 0.83 for the whole year was the same as for the wet season. The overall results obtained here are comparable to those found in small forested catchments throughout the Central and South America with a monsoonal climate where a significant portion of the discharge is the result of base and interflow (Norclift and Thornes, 1984; Motohisa et al, 1997; McGlynn and McDonnel 2003; Fujieda et al. 1998).

Rainfall runoff relationships: To further explore the relationship between rainfall and runoff, we performed a separate analysis of rainfall-runoff data for several storm events. In this subtropical region, precipitation usually occurs in late afternoon between 2:00 and 8:00 p.m. due to heat accumulation of convective systems. Therefore, rising and falling hygrograph limbs usually spread across two days as peak flow occurs overnight and streamflow reach pre-storm levels usually before 10:00 a.m. of the next morning. Hence, for this analysis, we altered day accounting to begin at noon (12:00 p.m.) and end at 11:00 a.m. the next day. By doing so, we found that the rainfall and runoff were linearly related, especially for precipitation events greater than 20 mm hr^{-1} (Figure 2.5). The slope of regression line was 0.19 in Figure 2.5 indicating that 19% of the area is hydraulic active and contributes interflow and direct runoff to the outlet for large storms. On the other hand, precipitation events ranging from 5 to 20 mm, which are prevalent in this area (92%), are more scattered, meaning that runoff response is affected by antecedent moisture content and thus the amount of rainfall needed before the area becomes hydrologic active.

These results, although coming from a limited number of rainfall events, indicate that there is a strong linear correlation (R^2 = 0.85) between precipitation and runoff generation, especially after the catchment's soils have gone through an initial period of wetting. This behavior is similar to that found by Lui et al. (2008) and Collick et al. (2009) in a study of four small catchments in the upper Blue Nile Basin.

Baseflow recession analysis: Baseflow characteristics during the dry phase of the monsoon for cloud forests are especially important for water supply systems -- especially during the months without any rain from the end of February through early May. This recession period is characterized by a zero order reservoir in which the outflow declines by a constant amount each day and thus discharge decreases linearly with time. A zero order reservoir indicates that the gravity dominates the flow process (Stagnitti et al., 1986; Steenhuis et al., 1999; 1998) and thus water flows from the hill downwards. In Figure 2.6, the discharge is plotted for watershed W1 as a function of time for the period March 3 to May 3. The flow on March 3 was 3.95 10^3 m^3/day (or 0.62 mm/day over the entire watershed) and on May 3, the discharge was 2.7 10^3

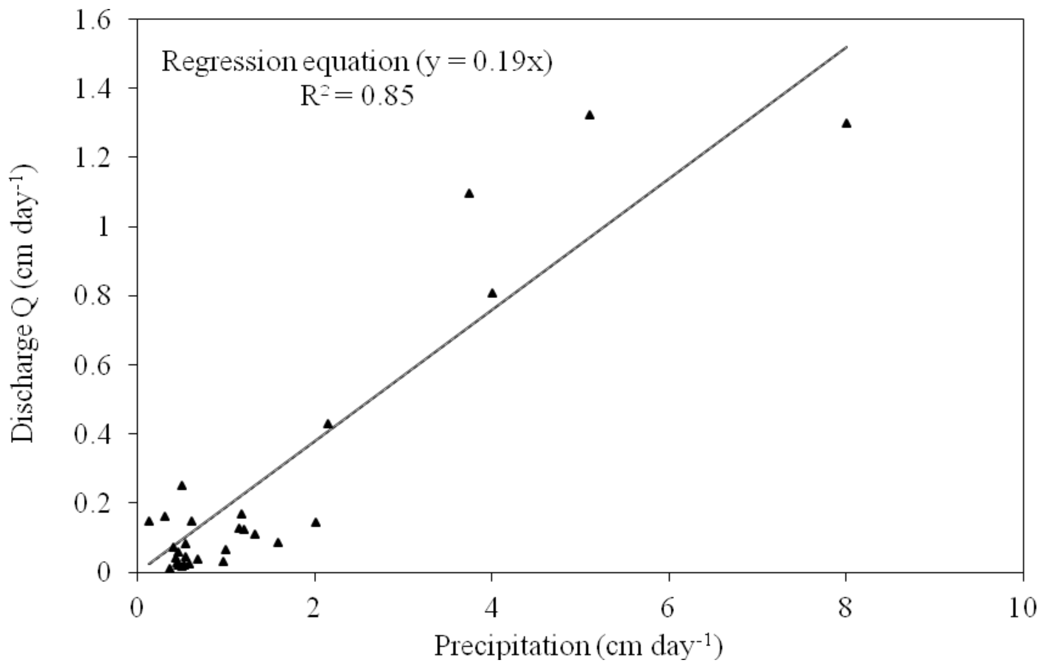

Figure 2.5. Rainfall-runoff relationship 29 storm events measured at WS1 from October 8 2008 through October 4 2009 water year.

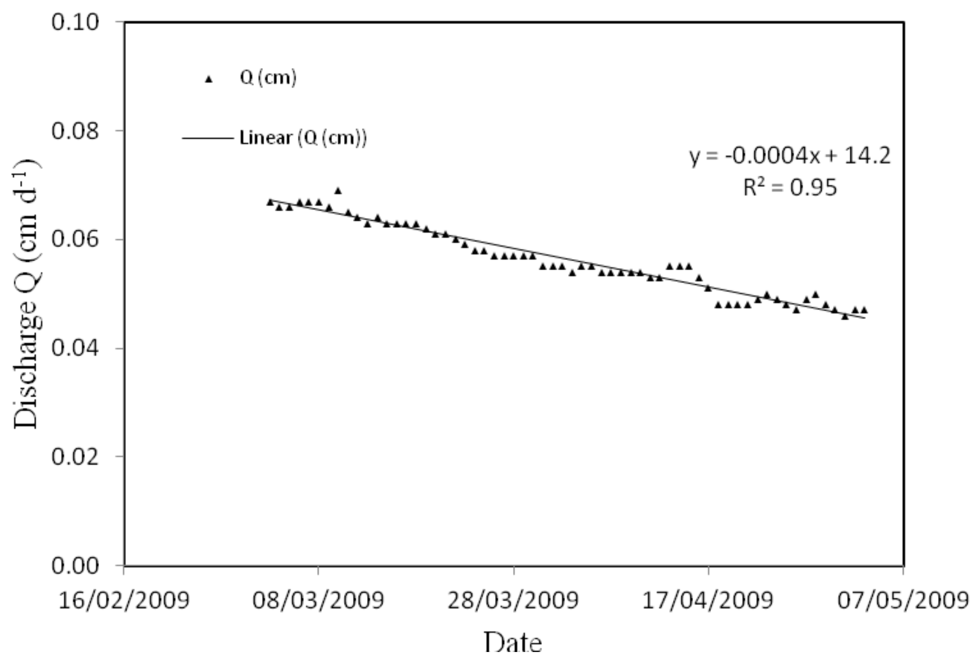

Figure 2.6. WS1 flow recession for the late part of the dry season (March 3-May 3 2009). The solid line is the linear regression line

m^3/day (or 0.44 mm/day). The discharge decreased by 22 m^3/day (0.035 mm/day); calculated by a linear regression of the discharge, the R squared is 0.95. Since the flow decreases linearly, we can calculate the time that the flow will stop in case there would be no recharge from rainfall. This is approximately 120 days after March 3 or around July 1. This is theoretically the average travel time down the hill of a drying front (Stagnitti et al., 1986). The travel time is constant since it depends on physical factors while the flow on February 1 depends on the amount of rainfall during the wet phase of the monsoon.

CONCLUSIONS

The amount of precipitation that became stream flow for the cloud forest was approximately half while for the adjacent forested watersheds less than 15% of the precipitation became stream flow. In addition, these forested watersheds had 10% less rainfall. Despite these big differences in streamflow response to rainfall, only a small portion was direct runoff from saturated areas or exposed bed rock. On average, more than 90 % of the rainfall infiltrated.

This research provides compelling evidence that baseflow is the primary contributor to streamflow during both wet and dry season in cloud forest catchments. Preserving these flow processes over time is critical for the sustained provision of water, especially when demand is high and supply is short in the dry season. Following on these research results, an economic evaluation of the impact of land use changes on water quality and quantity should lead to improved economic and environmental policies to protect these critical water source areas in Central America. In the long run, it would be more economically feasible and environmentally sustainable to continue to allow forested watersheds to serve as both water producing areas as well as underground water storage systems.

ACKNOWLEDGMENTS

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SUPPORTING INFORMATION

Supplementary materials containing additional information of the watersheds and mentioned in the text are available as part of the online paper.

APPENDIX 1:

Baseflow analysis

Literature review

To separate the storm runoff hydrograph from the base flow signal, several methods been proposed by Nathan and McMahon (1990), Wilson (1990) and Lin (2007), among others, including graphical approaches, chemical composition, an analytical approach, and digital filters. The simplest is the graphical approach, which assumes that baseflow prior to a runoff event is the same after the runoff (Linsley et al., 1958). Analytical methods consist of solving the water balance equation through hydrologic modeling. Other methods include the use of recursive digital filters which are routine tools in signal analysis and processing, used to remove the high-frequency quickflow signal in order to derive the low-frequency baseflow signal (Nathan and McMahon, 1990; Lynn and Hollick, 1979; Arnold and Allen, 1999; Eckhart, 2002, 2004). These are the most sophisticated and are used in this study.

The rising limb of a hydrograph is usually associated with surface runoff while recession flow is related to subsurface flow. Based on hydrograph analysis in Arizona, Horton (1933) reasoned that runoff was being generated by infiltration excess, which is when rainfall intensity exceeded the infiltration capacity of the soil. Hewlett (1967) later argued that in well-vegetated and undisturbed soil conditions, saturation excess runoff is common, with runoff generated when a soil profile is saturated. In watersheds where saturation excess runoff predominates, typically only a small (and relatively wetter) part of the catchment area contributes directly to storm flow (Hewlett and Hibbert, 1967; Hewlett and Nuter 1970; Dune 1978; Ambroise, 1999; Steenhuis, *et al*, 1995). The extent of these wet source areas is dynamic in time and space in response to changes in rainfall inputs and antecedent soil moisture conditions, and they are thus termed Variable Source Areas (VSA) (Hewlett, 1967; Steenhuis 1985).

Analysis of flow recession curves allows the determination of characteristics for the groundwater reservoir (Wittenberg, 1999). Maillet (1905) was the first to introduce an exponential function for characterization of baseflow recession, subsequently used by others (Nattan and McMahon 1990; Arnold et al., 1995, Tallaksen1995, and Rivera et al., 2002):

$$Q_t = Q_0 \cdot \exp(-t/k) \tag{2}$$

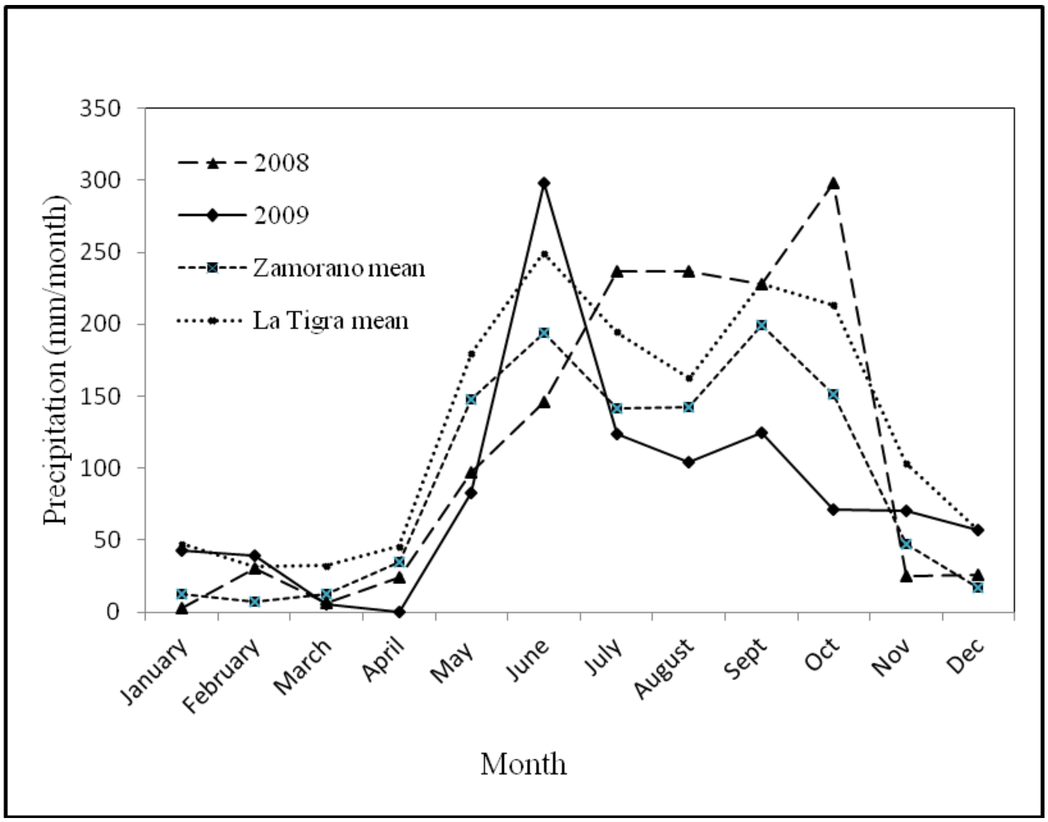
where Q_t is the discharge at time t, Q_0 the initial discharge, and k the recession constant which can be considered to represent the average response time in storage. This function has been widely used to describe baseflow recession, and implies that the aquifer reacts like a single linear reservoir where outflow is dependent on storage, thus $S=kQ$. However, the outflow from aquifers is not always linearly proportional to storag, and to allow for nonlinearity the storage discharge relationship has been modified by using a power function:

$$S = aQ^b \tag{3}$$

where a and b are constants. The recession part of the hydrograph provides information related to the release of water from a catchment after a daily or seasonal precipitation input. In temperate climates, different coefficients are typically used for winter and summer conditions.

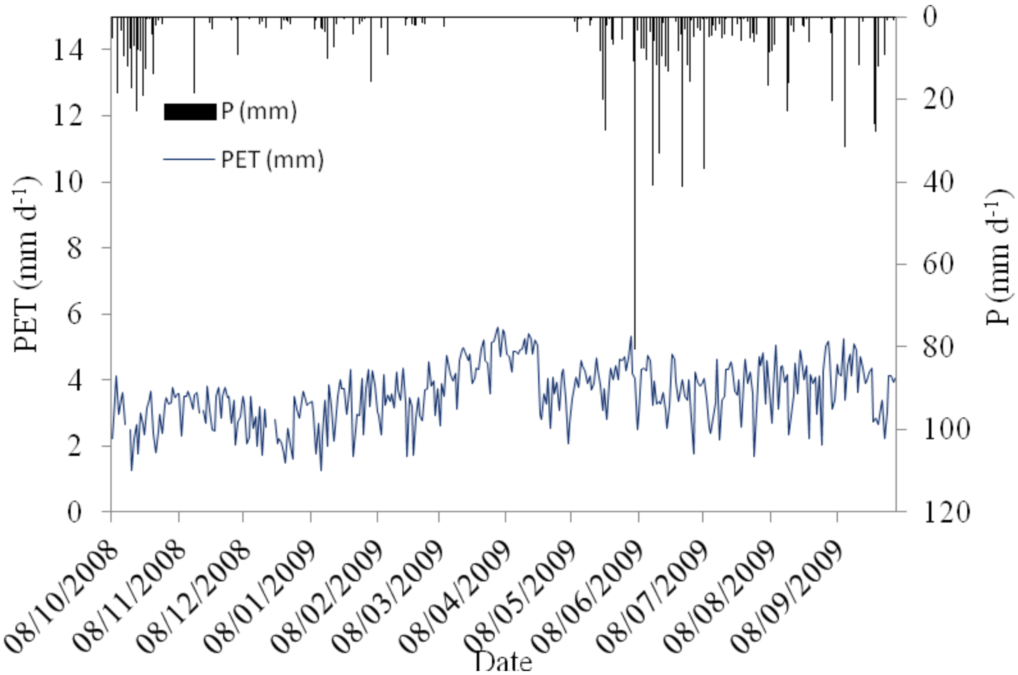
APPENDIX 2:

Evaporation and Precipitation in Study Area



Source: Zamorano University (1942-2009) and SANAA (1963-2008)

Figure A2.1. Monthly precipitation at experimental site compared to long-term average Zamorano weather station and La Tigra SANAA.



(PET = Potential Evapotranspiration, P = Precipitation).

Figure A2.2. Evaporation and precipitation in the study area

Table A2.1. Monthly precipitation at the study site for 2008 and 2009 compared to long-term averages at Zamorano and La Tigra

Months	Our site P (mm)		Weather Station P (mm)		Long-Term Average	
	2008	2009	2008	2009	Zamorano*Mean	La Tigra (SANAA) *
January	2.4*	42.3	2.4	12.0	12.3	47.4
February	30.7*	39.4	30.7	6.0	7.6	31.3
March	6.1*	5.1	6.1	1.0	12.1	32.4
April	24.1*	0.0	24.1	4.0	34.8	45.7
May	96.6	83.1	72.9	67.0	147.5	180.1
June	146.0	291.4	151.7	363.0	194.2	249.4
July	236.9	123.7	236.9	85.0	141.7	194.5
August	236.9	104.2	143.3	78.0	142.6	162.7
Sept	227.9	124.6	178.9	62.0	199.4	228.1
Oct	298.5	71.4	246.8	32.0	151.3	213.2
Nov	24.7	70.6	4.6	37.0	46.8	103.5
Dec	26.0	56.9	8.6	36.0	17.3	56.8
Total	1293.5	1013	1107	783.0	1108	1545

*Values taken form Zamorano's weather station now added for the year 2008 in order to match the field data (Auxiliary material A2)

Table A2.2. Rainfall intensities for 45 rainfall events greater than 15 mm day^{-1} occurring in 2008 and 2009.

No. Event	Date	Duration	Maximum P. intensity mm hr^{-1}	Average P intensity mm hr^{-1}	Total P storm event mm day^{-1}
1	05/29/08	7:00-23:00	4.8	1.7	29
2	05/30/08	0:00-16:00	15.2	2.91	46.6
3	06/5/08	19:00-23:00	4.4	3.68	18.4
4	06/10/08	21:00-23:00	14.8	11.53	34.6
5	06/29/08	12:0017:00	16.6	4.0	20
6	07/2/08	14:00-21:00	7.4	3.57	25.6
7	07/4/08	18:00-21:00	17.4	6.15	24.6
8	8/12/08	16:00-23:00	24	7.31	51.2
9	8/19/08	15:00-20:00	7.4	4	28.6
10	8/25/08	19:00-21:00	14.4	5.53	16.6
11	8/29/08	1:00-6:00	4.2	3.4	17
12	8/31/08	0:00-2:00	11.4	9.27	27.8
13	9/3/08	2:00-3:00	19.6	16.9	33.8
14	9/11/08	17:00-19:00	14	5.73	17.2
15	9/17/08	14:00-23:00	15.4	8.0	31.2
16	9/27/08	19:00-22:00	16.8	7.7	30.8
17	9/29/08	19:00-21:00	12.8	7.67	23
18	10/4/08	16:00-19:00	36.6	10.65	42.6
19	10/18/08	01:00-23:00	15.4	3.21	41.8
20	10/19/08	16:00-23:00	12.6	2.8	16.8
21	10/22/08	17:00-23:00	9	4.35	17.4
22	3/6/09	13:00-14:00	29.8	20.9	41.8
23	3/7/09	12:00-21:00	15.2	9.2	46
24	5/22/09	13:00-23:00	6.6	3.03	18.2
25	5/23/09	01:00-16:00	10	2.77	25
26	6/5/09	19:00-0:00	10.4	3.53	21.2
27	6/6/09	18:0021:00	35.4	20.15	80.6
28	06/14/09	21:00-0:00	24	10.25	41
29	06/17/09	0:00-6:00	12.2	4.15	33.2
30	6/28/09	14:00-21:00	33	14	70
31	7/1/09	19:00-22:00	17.4	7.33	22
32	7/8/09	17:00-19:00	30.6	13.67	41
33	8/6/09	18:00-19:00	16.6	9.8	19.6
34	8/15/09	18:00-20:00	16.4	10.8	21.6
35	8/16/09	20:00-22:00	11.2	5.1	15.4
36	9/5/09	15:00-16:00	14.8	8.8	17.6
37	9/11/09	18:00-21:00	19.6	8.35	33.4
38	9/24/09	17:00-20:00	20.4	8.5	34.0
39	9/25/09	16:00-19:00	16.6	7.75	31.0

Table A2.2 (Continued)

No. Event	Date	Duration	Maximum P. intensity mm hr^{-1}	Average P intensity mm hr^{-1}	Total P storm event mm day^{-1}
40	10/30/09	15:00-20:00	13.6	5.63	33.8
41	10/31/09	17:00-23:00	15.2	5.33	32
42	11/1/09	19:00-03:00	9.6	2.69	35
43	11/4/09	16:00-20:00	5.8	3.68	18.4
44	11/7/09	10:00-13:00	8.2	3.46	15.8
45	12/21/09	00:00-11:00	3.2	2.54	27
Mean			15	7	30
StdDev			8	4	14

Table A2.3 Total number of rainfall events during the measurement period 2008-2009.

N_0. events	Precipitation (mm d^{-1})	N_0. events	Precipitation (mm d^{-1})	N_0. events	Precipitation (mm d^{-1})
1	68.80	50	14.00	99	6.73
2	52.00	51	13.80	100	6.60
3	49.33	52	13.40	101	6.50
4	46.60	53	13.40	102	6.40
5	41.27	54	13.30	103	6.27
6	40.00	55	13.20	104	6.15
7	37.40	56	12.95	105	6.15
8	34.60	57	12.80	106	6.13
9	32.67	58	12.80	107	6.07
10	31.13	59	12.80	108	6.00
11	29.00	60	12.67	109	5.80
12	28.93	61	12.40	110	5.80
13	28.60	62	12.07	111	5.73
14	27.67	63	12.00	112	5.40
15	27.30	64	11.93	113	5.40
16	26.20	65	11.93	114	5.27
17	25.60	66	11.70	115	5.13
18	25.40	67	11.67	116	5.10
19	23.40	68	11.00	117	5.07
20	23.10	69	11.00	118	5.00
21	22.93	70	10.13	119	5.00
22	22.80	71	10.00	120	5.00
23	22.33	72	9.87	121	5.00
24	21.55	73	9.80	122	4.93
25	21.00	74	9.60	123	4.80
26	20.50	75	9.20	124	4.80
27	20.20	76	9.13	125	4.60
28	20.00	77	9.10	126	4.53
29	19.70	78	8.60	127	4.47
30	19.30	79	8.47	128	4.40
31	19.20	80	8.47	129	4.40
32	19.10	81	8.40	130	4.40
33	18.47	82	8.20	131	4.20
34	18.47	83	8.20	132	4.20
35	18.40	84	8.20	133	4.10
36	17.87	85	8.20	134	4.00

Table A2.3 (Continued)

N_0. events	Precipitation (mm d^{-1})	N_0. events	Precipitation (mm d^{-1})	N_0. events	Precipitation (mm d^{-1})
37	17.87	86	8.15	135	3.93
38	17.60	87	8.00	136	3.80
39	17.30	88	8.00	137	3.70
40	16.67	89	7.80	138	3.60
41	16.67	90	7.73	139	3.60
42	16.20	91	7.73	140	3.60
43	16.13	92	7.30	141	3.60
44	15.87	93	7.27	142	3.55
45	15.73	94	7.20	143	3.53
46	14.60	95	7.00	144	3.40
47	14.40	96	7.00	145	3.40
48	14.07	97	6.85	146	3.40
49	14.05	98	6.80	147	3.27
148	3.27	198	1.73	248	0.73
149	3.13	199	1.73	249	0.73
150	3.07	200	1.70	250	0.70
151	3.00	201	1.70	251	0.70
152	3.00	202	1.67	252	0.65
153	3.00	203	1.67	253	0.60
154	3.00	204	1.65	254	0.60
155	3.00	205	1.60	255	0.60
156	3.00	206	1.60	256	0.60
157	2.93	207	1.60	257	0.60
158	2.90	208	1.50	258	0.60
159	2.87	209	1.47	259	0.60
160	2.80	210	1.45	260	0.60
161	2.80	211	1.40	261	0.60
162	2.80	212	1.40	262	0.60
163	2.73	213	1.20	263	0.60
164	2.73	214	1.20	264	0.60
165	2.67	215	1.20	265	0.60
166	2.65	216	1.20	266	0.60
167	2.60	217	1.20	267	0.53
168	2.60	218	1.20	268	0.53
169	2.53	219	1.20	269	0.50
170	2.50	220	1.15	270	0.47
171	2.45	221	1.15	271	0.47
172	2.40	222	1.13	272	0.47
173	2.33	223	1.13	273	0.47

Table A2.3 (Continued)

N_0. events	Precipitation (mm d^{-1})	N_0. events	Precipitation (mm d^{-1})	N_0. events	Precipitation (mm d^{-1})
174	2.30	224	1.13	274	0.47
175	2.30	225	1.13	275	0.45
176	2.30	226	1.10	276	0.40
177	2.27	227	1.10	277	0.40
178	2.20	228	1.07	278	0.40
179	2.20	229	1.05	279	0.40
180	2.15	230	1.00	280	0.40
181	2.13	231	1.00	281	0.40
182	2.13	232	1.00	282	0.40
183	2.10	233	1.00	283	0.40
184	2.10	234	1.00	284	0.40
185	2.00	235	1.00	285	0.33
186	2.00	236	0.95	286	0.33
187	2.00	237	0.93	287	0.33
188	2.00	238	0.90	288	0.33
189	2.00	239	0.90	289	0.33
190	1.93	240	0.87	290	0.33
191	1.90	241	0.80	291	0.33
192	1.87	242	0.80	292	0.30
193	1.80	243	0.80	293	0.27
194	1.80	244	0.80	294	0.27
195	1.80	245	0.80	295	0.27
196	1.80	246	0.80	296	0.27
197	1.75	247	0.75	297	0.20
298	0.20	348	0.10		
299	0.20	349	0.10		
300	0.20	350	0.07		
301	0.20	351	0.07		
302	0.20	352	0.07		
303	0.20	353	0.07		
304	0.20	354	0.07		
305	0.20	355	0.07		
306	0.20	356	0.07		
307	0.20	357	0.07		
308	0.20	358	0.07		
309	0.20	359	0.07		
310	0.20	360	0.07		
311	0.20	361	0.07		
312	0.20	362	0.07		

Table A2.3 (Continued)

N_0. events	Precipitation (mm d^{-1})	N_0. events	Precipitation (mm d^{-1})	N_0. events	Precipitation (mm d^{-1})
313	0.20	363	0.07		
314	0.20	364	0.07		
315	0.20	365	0.07		
316	0.20	366	0.05		
317	0.20	367	0.05		
318	0.15	368	0.05		
319	0.13	Total	2169		
320	0.13				
321	0.13				
322	0.13				
323	0.13				
324	0.13				
325	0.13				
326	0.13				
327	0.13				
328	0.13				
329	0.13				
330	0.13				
331	0.13				
332	0.10				
333	0.10				
334	0.10				
335	0.10				
336	0.10				
337	0.10				
338	0.10				
339	0.10				
340	0.10				
341	0.10				
342	0.10				
343	0.10				
344	0.10				
345	0.10				
346	0.10				
347	0.10				

This data reports only May 2008-December 2009 and without missing data. This is the reason it does not match Table 2. Monthly data reported for 2008 and 2009.

APPENDIX 3:

Experimental watershed layout

APPENDIX 4:

Supporting precipitation data and weir design and location

Table A4.1. Main characteristics of the weirs sites and formulas utilized to calculate discharge rates.

Site	Creek type	Location (UTM)	Weir type/size	Q (cfs) formula
WS1	Perennial	East: 492452 North: 1565177	Suppressed rectangular 2.275 m wide	Q = 3.33*(LH)^1.5
WS2	Perennial	East: 493746 North:1564875	V-notch 90º Hmax = 29 cm	Q = 2.49*(H)^2.48
WS3	Perennial	East: 493,130 North:1565032	V-notch 90º Hmax = 17 cm	Q = 2.49*(H)^2.48
WS4	Intermittent	East: 492676 North: 1565070	V-notch 90º Hmax = 19 cm	Q = 2.49*(H)^2.48

Datum: North American _1927
Projected Coordinate System NAD-1927_UTM-Zone_16

Table A4.2. Monthly precipitation at the study site for 2008 and 2009 compared to long-term averages at Zamorano and La Tigra.

Months	Our site P (mm)		Weather Station P (mm)		Long-Term Average	
	2008	2009	2008	2009	Zamorano*Mean	La Tigra (SANAA) *
January	2.4*	42.3	2.4	12.0	12.3	47.4
February	30.7*	39.4	30.7	6.0	7.6	31.3
March	6.1*	5.1	6.1	1.0	12.1	32.4
April	24.1*	0.0	24.1	4.0	34.8	45.7
May	96.6	83.1	72.9	67.0	147.5	180.1
June	146.0	291.4	151.7	363.0	194.2	249.4
July	236.9	123.7	236.9	85.0	141.7	194.5
August	236.9	104.2	143.3	78.0	142.6	162.7
Sept	227.9	124.6	178.9	62.0	199.4	228.1
Oct	298.5	71.4	246.8	32.0	151.3	213.2
Nov	24.7	70.6	4.6	37.0	46.8	103.5
Dec	26.0	56.9	8.6	36.0	17.3	56.8
Total	1293.5	1013	1107	783.0	1108	1545

*Values taken form Zamorano's weather station no added for the year 2008 in order to match the field data (Auxiliary material A2)

Table A4.3. Rainfall intensities for 45 rainfall events greater than 15 mm day^{-1} occurring in 2008 and 2009.

No. Event	Date	Duration	Maximum P. intensity mm hr^{-1}	Average P intensity mm hr^{-1}	Total P storm event mm day^{-1}
1	05/29/08	7:00-23:00	4.8	1.7	29
2	05/30/08	0:00-16:00	15.2	2.91	46.6
3	06/5/08	19:00-23:00	4.4	3.68	18.4
4	06/10/08	21:00-23:00	14.8	11.53	34.6
5	06/29/08	12:0017:00	16.6	4.0	20
6	07/2/08	14:00-21:00	7.4	3.57	25.6
7	07/4/08	18:00-21:00	17.4	6.15	24.6
8	8/12/08	16:00-23:00	24	7.31	51.2
9	8/19/08	15:00-20:00	7.4	4	28.6
10	8/25/08	19:00-21:00	14.4	5.53	16.6
11	8/29/08	1:00-6:00	4.2	3.4	17
12	8/31/08	0:00-2:00	11.4	9.27	27.8
13	9/3/08	2:00-3:00	19.6	16.9	33.8
14	9/11/08	17:00-19:00	14	5.73	17.2
15	9/17/08	14:00-23:00	15.4	8.0	31.2
16	9/27/08	19:00-22:00	16.8	7.7	30.8
17	9/29/08	19:00-21:00	12.8	7.67	23
18	10/4/08	16:00-19:00	36.6	10.65	42.6
19	10/18/08	01:00-23:00	15.4	3.21	41.8
20	10/19/08	16:00-23:00	12.6	2.8	16.8
21	10/22/08	17:00-23:00	9	4.35	17.4
22	3/6/09	13:00-14:00	29.8	20.9	41.8
23	3/7/09	12:00-21:00	15.2	9.2	46
24	5/22/09	13:00-23:00	6.6	3.03	18.2
25	5/23/09	01:00-16:00	10	2.77	25
26	6/5/09	19:00-0:00	10.4	3.53	21.2
27	6/6/09	18:0021:00	35.4	20.15	80.6
28	06/14/09	21:00-0:00	24	10.25	41
29	06/17/09	0:00-6:00	12.2	4.15	33.2
30	6/28/09	14:00-21:00	33	14	70
31	7/1/09	19:00-22:00	17.4	7.33	22
32	7/8/09	17:00-19:00	30.6	13.67	41
33	8/6/09	18:00-19:00	16.6	9.8	19.6
34	8/15/09	18:00-20:00	16.4	10.8	21.6
35	8/16/09	20:00-22:00	11.2	5.1	15.4
36	9/5/09	15:00-16:00	14.8	8.8	17.6
37	9/11/09	18:00-21:00	19.6	8.35	33.4
38	9/24/09	17:00-20:00	20.4	8.5	34.0
39	9/25/09	16:00-19:00	16.6	7.75	31.0

Table A4.3. (Continued)

No. Event	Date	Duration	Maximum P. intensity mm hr^{-1}	Average P intensity mm hr^{-1}	Total P storm event mm day^{-1}
40	10/30/09	15:00-20:00	13.6	5.63	33.8
41	10/31/09	17:00-23:00	15.2	5.33	32
42	11/1/09	19:00-03:00	9.6	2.69	35
43	11/4/09	16:00-20:00	5.8	3.68	18.4
44	11/7/09	10:00-13:00	8.2	3.46	15.8
45	12/21/09	00:00-11:00	3.2	2.54	27
Mean			15	7	30
StdDev			8	4	14

Table A4.4. Total number of rainfall events during the measurement period 2008-2009.

N_0. events	Precipitation (mm d^{-1})	N_0. events	Precipitation (mm d^{-1})	N_0. events	Precipitation (mm d^{-1})
1	68.80	50	14.00	99	6.73
2	52.00	51	13.80	100	6.60
3	49.33	52	13.40	101	6.50
4	46.60	53	13.40	102	6.40
5	41.27	54	13.30	103	6.27
6	40.00	55	13.20	104	6.15
7	37.40	56	12.95	105	6.15
8	34.60	57	12.80	106	6.13
9	32.67	58	12.80	107	6.07
10	31.13	59	12.80	108	6.00
11	29.00	60	12.67	109	5.80
12	28.93	61	12.40	110	5.80
13	28.60	62	12.07	111	5.73
14	27.67	63	12.00	112	5.40
15	27.30	64	11.93	113	5.40
16	26.20	65	11.93	114	5.27
17	25.60	66	11.70	115	5.13
18	25.40	67	11.67	116	5.10
19	23.40	68	11.00	117	5.07
20	23.10	69	11.00	118	5.00
21	22.93	70	10.13	119	5.00
22	22.80	71	10.00	120	5.00
23	22.33	72	9.87	121	5.00
24	21.55	73	9.80	122	4.93
25	21.00	74	9.60	123	4.80
26	20.50	75	9.20	124	4.80
27	20.20	76	9.13	125	4.60
28	20.00	77	9.10	126	4.53
29	19.70	78	8.60	127	4.47
30	19.30	79	8.47	128	4.40
31	19.20	80	8.47	129	4.40
32	19.10	81	8.40	130	4.40
33	18.47	82	8.20	131	4.20
34	18.47	83	8.20	132	4.20
35	18.40	84	8.20	133	4.10
36	17.87	85	8.20	134	4.00
37	17.87	86	8.15	135	3.93

Table A4.4. (Continued)

N_0. events	Precipitation (mm d^{-1})	N_0. events	Precipitation (mm d^{-1})	N_0. events	Precipitation (mm d^{-1})
38	17.60	87	8.00	136	3.80
39	17.30	88	8.00	137	3.70
40	16.67	89	7.80	138	3.60
41	16.67	90	7.73	139	3.60
42	16.20	91	7.73	140	3.60
43	16.13	92	7.30	141	3.60
44	15.87	93	7.27	142	3.55
45	15.73	94	7.20	143	3.53
46	14.60	95	7.00	144	3.40
47	14.40	96	7.00	145	3.40
48	14.07	97	6.85	146	3.40
49	14.05	98	6.80	147	3.27
148	3.27	198	1.73	248	0.73
149	3.13	199	1.73	249	0.73
150	3.07	200	1.70	250	0.70
151	3.00	201	1.70	251	0.70
152	3.00	202	1.67	252	0.65
153	3.00	203	1.67	253	0.60
154	3.00	204	1.65	254	0.60
155	3.00	205	1.60	255	0.60
156	3.00	206	1.60	256	0.60
157	2.93	207	1.60	257	0.60
158	2.90	208	1.50	258	0.60
159	2.87	209	1.47	259	0.60
160	2.80	210	1.45	260	0.60
161	2.80	211	1.40	261	0.60
162	2.80	212	1.40	262	0.60
163	2.73	213	1.20	263	0.60
164	2.73	214	1.20	264	0.60
165	2.67	215	1.20	265	0.60
166	2.65	216	1.20	266	0.60
167	2.60	217	1.20	267	0.53
168	2.60	218	1.20	268	0.53
169	2.53	219	1.20	269	0.50
170	2.50	220	1.15	270	0.47
171	2.45	221	1.15	271	0.47
172	2.40	222	1.13	272	0.47
173	2.33	223	1.13	273	0.47
174	2.30	224	1.13	274	0.47

Table A4.4. (Continued)

N_0. events	Precipitation (mm d^{-1})	N_0. events	Precipitation (mm d^{-1})	N_0. events	Precipitation (mm d^{-1})
175	2.30	225	1.13	275	0.45
176	2.30	226	1.10	276	0.40
177	2.27	227	1.10	277	0.40
178	2.20	228	1.07	278	0.40
179	2.20	229	1.05	279	0.40
180	2.15	230	1.00	280	0.40
181	2.13	231	1.00	281	0.40
182	2.13	232	1.00	282	0.40
183	2.10	233	1.00	283	0.40
184	2.10	234	1.00	284	0.40
185	2.00	235	1.00	285	0.33
186	2.00	236	0.95	286	0.33
187	2.00	237	0.93	287	0.33
188	2.00	238	0.90	288	0.33
189	2.00	239	0.90	289	0.33
190	1.93	240	0.87	290	0.33
191	1.90	241	0.80	291	0.33
192	1.87	242	0.80	292	0.30
193	1.80	243	0.80	293	0.27
194	1.80	244	0.80	294	0.27
195	1.80	245	0.80	295	0.27
196	1.80	246	0.80	296	0.27
197	1.75	247	0.75	297	0.20
298	0.20	348	0.10		
299	0.20	349	0.10		
300	0.20	350	0.07		
301	0.20	351	0.07		
302	0.20	352	0.07		
303	0.20	353	0.07		
304	0.20	354	0.07		
305	0.20	355	0.07		
306	0.20	356	0.07		
307	0.20	357	0.07		
308	0.20	358	0.07		
309	0.20	359	0.07		
310	0.20	360	0.07		
311	0.20	361	0.07		
312	0.20	362	0.07		
313	0.20	363	0.07		

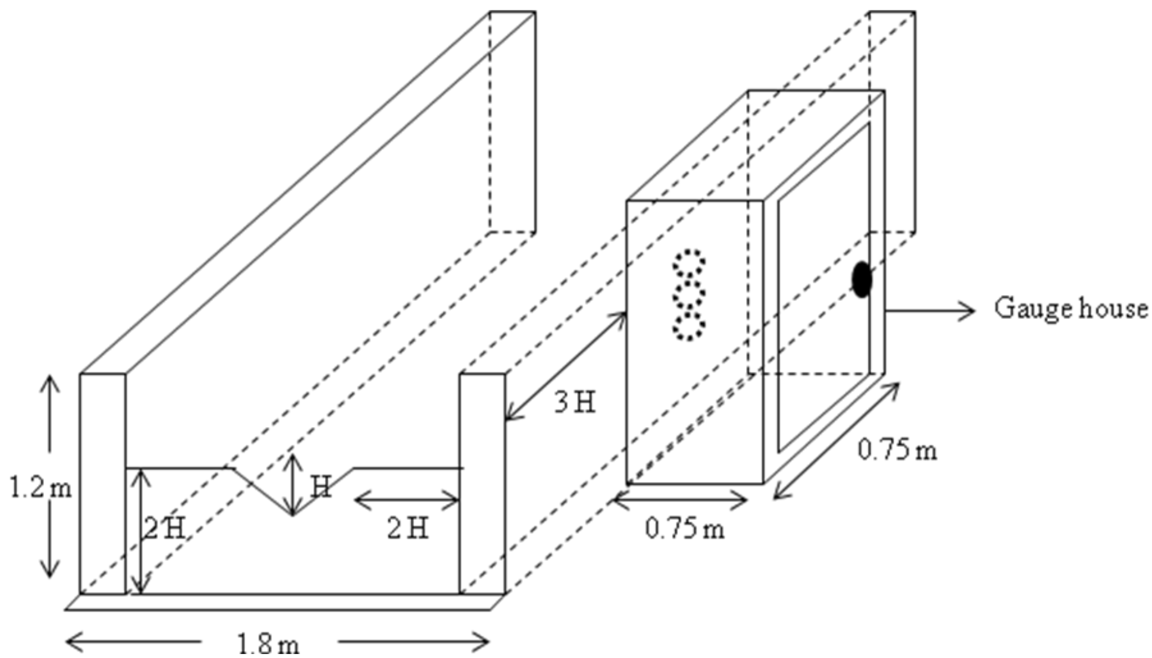
Table A4.4. (Continued)

N_0. events	Precipitation (mm d^{-1})	N_0. events	Precipitation (mm d^{-1})	N_0. events	Precipitation (mm d^{-1})
314	0.20	364	0.07		
315	0.20	365	0.07		
316	0.20	366	0.05		
317	0.20	367	0.05		
318	0.15	368	0.05		
319	0.13	Total	2169		
320	0.13				
321	0.13				
322	0.13				
323	0.13				
324	0.13				
325	0.13				
326	0.13				
327	0.13				
328	0.13				
329	0.13				
330	0.13				
331	0.13				
332	0.10				
333	0.10				
334	0.10				
335	0.10				
336	0.10				
337	0.10				
338	0.10				
339	0.10				
340	0.10				
341	0.10				
342	0.10				
343	0.10				
344	0.10				
345	0.10				
346	0.10				
347	0.10				

This data reports only May 2008-December 2009 and without missing data. This is the reason it does not match Table 4.2. Monthly data reported for 2008 and 2009.

APPENDIX 5:

V-notch 90º design constructed in WS2, WS3 and WS4 catchments (actual distances might vary from weir to weir)



REFERENCES

Arnold, J. G., P. M. Allen, R. Muttiah and G. Bernhardt. (1995). Automated base flow separation and recession analysis techniques, *Ground Water*, *33*(6), 1010–1018, 1995.

Arnold, J.G. and P.M. Allen. (1999). Automated methods for estimating baseflow and groundwater recharge from streamflow records. Journal of the American Water Resources Association VOL. 35, NO.2 AMERICAN WATER RESOURCES ASSOCIATION APRIL 1999.

Bayabil, H. K., S. A. Tilahun A. S. Collick, B. Yitaferu and T. S. Steenhuis. (2010). Are runoff processes ecologically or topographically driven in the (sub)humid Ethiopian highlands? The case of the Maybar watershed. Ecohydrology, 3:457-466.

Bonell, M. (1993). Progress in understanding of runoff generation dynamics in forest. Journal of Hydrology, 150 (1993) 217-275.

Bonell, M. and L.A Bruijnzeel. (2004). Forest, water and people in the humid tropics: past, present and future hydrological research for integrated land and water management. Cambridge University Press.

Bruijnzeel, L.A. (1990). Hydrology of most tropical forests and effects of conversion: a state of knowledge review. Amsterdam: UNESCO Humid Tropics Programme.

Bruijnzeel, L.A. (1992). Managing tropical forest watersheds for production: where contradictory theory and practice co-exist. In: Miller, F. R. and Adam, K. L. (Eds.), Wise Management of Tropical Forests 1992. Oxford Forestry Institute, Oxford, pp. 37–75.

Bruijnzeel, L.A. (2002). Hydrological impacts of converting tropical montane cloud forest to pasture, with initial reference to northern Costa Rica. Project Memorandum Form, Project No. R7991 within the Forestry Research Programme of the Department for International Development of the UK, Aylesford, UK, 60 pp.

Bruijnzeel, L.A. (2004), Hydrological functions of tropical forests: not seeing the soil for the trees? Agriculture, Ecosystems and Environment 104: 185–228.

Bruijnzeel, L.A. and C. N. Bremmer. (1989). Highland–Lowland Interactions in the Ganges Brahmaputra River Basin: A Review of Published Literature. ICIMOD Occasional Paper No. 11. International Centre for Integrated Mountain Development, Kathmandu.

Carpenter, R.H. (1954). Geology and ore deposits of the Rosario Mining District and the San Juancito Mountains, Honduras, Central America: Bull. Geo. Soc. Am. 65: 23-38.

Chaubey, I., C.T Hann, S. Grunwald and J.M. Salisbury. (1999). Uncertainty in the model parameters due to spatial variability of rainfall. Journal of Hydrology Volume 220: 4861.

Collick, A. S., Z. M. Eaton, T. Ashagrie, B. Biruk, S. A. Tilahun, E. Adgo, S. B. Awulachew, G. Zeleke and T. S. Steenhuis. (2009). A simple semi-distributed water balance model for the Ethiopian highlands. Hydrological Processes, 23:3718-3737. doi: 10.1002/hyd.7517.

FAO. (1986). Tropical Forestry Action Plan. Committee on Forest Development in the Tropics, UN Food and Agriculture Organization, Rome.

Fujieda, M., T. Kudoh, V. De Cicco and J. L. De Calvarcho. (1997). Hydrological processes at two subtropical forest catchments: The Serra do Mar, Sao Paulo, Brazil. Journal of Hydrology 196: 26-46.

Goodrich, D., C. Faures, J. M. Woolhiser, D. A. Lane and L. J. Sorooshian. (1995). Measurement and analysis of small-scale convective storm rainfall variability. Journal of hydrology. 173: 283-308.

Gupta, R. K., P. R. Mishra, P. Shankar, R. C. Kaushal and S. S. Sajwan. (1974). Studies on the effect of different land treatments on water yield in the Siwalik Chandigarh. In Annual Report 1974 pp 18–21. Central Soil and Water Conservation Research and Training Institute, Dehradun, India.

Gupta, R. K., P. R. Mishra, S. P. Mittal and K. Singh. (1975). Effect of different land treatments on water yield in the Siwalik-Chandigarh pp. 15–17. In Annual Report 1975 Central Soil and Water Conservation Research and Training Institute, Dehradun.

Guswa, A. J., A. L. Rhodes and S. E. Newell. (2007). Importance of Orographic precipitation to the water resources of Monteverde, Costa Rica. Advances in Water Resources 30: 2098-2112.

Hanson, D. L., T. S. Steenhuis, M. F. Walter and J. Boll. (2004). Effects of Soil Degradation and Management Practices on the Surface Water Dynamics in the Talgua River Watershed in Honduras. Land Degradation & Development 15:367-381.

Hastenrath, S. (2002). The Intertropical Convergence Zone of the Eastern Pacific Revisited. International Journal of Climatology. 22: 347-356.

Hu, C. H., S. L. Guo, L. H. Xiong and D. Z. Peng. (2005). A modified Xinanjiang model and its application in northern China. Nordic Hydrology, 36: 175-192.

Instituto Geográfico Nacional. (1956). Geología de Honduras 1:50,000 San Juan de Flores Quadrangle.

ISO (1980). International Organization of Standards. ISO 1438/1-1980(E). Water flow measurement in open channels using weirs and venturi flumes - Part 1: Thin plate weirs. Available from Global Engineering Documents at http://global.ihs.com.

Kohl, B. and G. Markart. (2002). Dependence of surface runoff on rain intensity-results of rain simulation experiments, paper presented at International Conference on Flood Estimation, Swiss Fed. Off. for Water and Geol., Berne, 6–8 March.

Kok, W. (2000). Prime Minister of the Netherlands, Welcome words in the second announcement for the Second World Water Forum and Ministerial Conference. http://www.worldwaterday.org/wwday/2000/.

Lange J., N. Greenbaum, S. Husary, M. Ghanem, C. Leibundgut and A. P. Schick. (2003). Runoff generation from successive simulated rainfalls on a rocky, semi-arid, Mediterranean hillslope. Hydrological Processes, 17: 279-296.

Lavaire, T. and E. Fiallos. (2010). Potencial de captura de agua de los suelos de la microcuenca El Carrizal, Parque Nacional La Tigra, Francisco Morazán, Honduras. Tesis de Pregrado, Escuela Agrícola Panamericana, Honduras.

Liu, B. M., A. S., Collick, G. Zeleke, E. Adgo, Z. M. Easton and T. S. Steenhuis. (2008). Rainfall-discharge relationships for a monsoonal climate in the Ethiopian highlands. Hydrol. Process. 22: 1059–1067.

Lyne, V. and M. Hollick, (1979). Stochastic time-variable rainfall-runoff modeling, *I.E. Aust. Natl. Conf. Publ. 79/10*, pp. 89–93, Inst. of Eng. Aust., Canberra, ACT.

McGlynn, B. L. and J. J. McDonnell. (2003). Quantifying the relative contributions of riparian and hillslope zones to catchment runoff. Water Res Res 39: 1310, doi:10.1029/2003WR002091.

Mendoza, G. and T. S. Steenhuis. (2002). Determination of hydraulic conductivities of hillsides with a hillside infiltrometer. Soil Science Society of America Journal; 66: 1501-1504.

Merz, R. and G. Blöschl. (2009). A regional analysis of event runoff coefficients with respect to climate and catchment characteristics in Austria, Water Resour. Res., 45, W01405, doi:10.1029/2008WR007163.

Merz, R., G. Bloschl and J. Parajka. (2006). Spatio-temporal variability of event runoff coefficients. J. Hydrol, 331: 591–604.

Middelkoop, H., K. Daamen, D. Gellens, W. Grabs, J. C. J. Kwadijk, H. Lang, B. W. A. H. Parmet, B. Schädler, J. Schulla and K. Wilke. (2001). Impact of climate change on hydrological regimes and water resources management in the Rhine Basin. *Climatic Change* 49: 105–128.

Montanari, L., M. Sivapalan and A. Montanari. (2006). Investigating of dominant hydrological processes in a tropical catchment in a monsoonal climate via the downward approach. Hydrol. Earth Syst. Sci. 10: 769-782.

Motohisa, F., T. Kudoh, V. De Cicco and J. L. Calvarcho. (1997). Hydrological processes at two subtropical catchments: The Sierra do Mar, Sao Pablo Brazil, Elsevier Science.

Nathan, R. J. and T. A. McMahon. (1990). Evaluation of automated techniques for base flow and recession analyses, Water Resour. Res., 26: 1465–1473.

Norcliff, S. and J. B.Thornes. (1984). Floodplain response of a small tropical stream. In catchments experiments in fluvial hydrology, 73-85 In: Burt TP, Walling DE (Eds.) Geo Books: Norwhich, UK.

Panamerican Health Organization and World Health Organizacion. (2000). Evaluacion de los Servicion de Agua Potable y Saneamiento en las Americas (www.bvsde.ops-oms.org/eswww/eva2000/Honduras/caliserv/calidad.html).

Rogers, R. D. and E. A. O'Conner. (1993). Mapa Geológico de Honduras: Hoja de Tegucigalpa (segunda edicion), Instituto Geográfico Nacional, Tegucigalpa, Honduras, escala 1:50,000.

Schellekens, J., F. N. Scatena, L. A. Bruijnzeel and A. J. Wickel. (1999). Modeling rainfall interception by a lowland tropical rain forest in northeastern Puerto Rico. J. Hydrol. 225: 168–184.

Simmons, F. W. and J. Z. Castellanos. (1968). Soil classification for Honduras. 1:500,000 Digitalization por IGN, Honduras.

Stadtmüller, T. and N. Agudelo. (1990). Amounts and variability of cloud moisture input in a tropical cloud forest. Int. Assoc. Hydrol. Sci. Publ. 193: 25–32.

Stagnitti, F., M. B. Parlange, T. S. Steenhuis and J.-Y. Parlange. (1986). Drainage from a uniform soil layer on a hillslope, Water Resour. Res., 22, 631– 634.

Steenhuis, T. S., J.-Y. Parlange, M. B. Parlange and F. Stagnitti. (1998). A Simple model for flow on hillslopes, Agric. Water Manage., 14, 153–168.

Steenhuis, T. S., J.-Y. Parlange, W. E. Sanford, A. Heilig, F. Stagnitti and M. F. Walter. (1999). Can we distinguish Richards' and Boussinesq's equations for hillslopes?: The Coweeta experiment revisited. Water Resources Research 35: 589-593.

Steenhuis T. S., M. Winchell, J. Rossing, J. A. Zollweg and M. F. Walter. (1995). SCS Runoff Equation Revisited for Variable-Source Runoff Areas. ASCE. Journal of Irrigation and Drainage 121: 234–238.

Steenhuis, T. S., A. S. Collick, Z. M. Easton, E. S. Leggesse, H. K. Bayabil, E. D. White, S. B. Awulachew, E. Adgo and A. A. Ahmed. (2009). Predicting Discharge and Erosion for the Abay (Blue Nile) with a simple model. In: Abtew & Melesse. Hydrology and Ecology of the Nile River Basin Under Extreme Conditions, Florida International University, Miami. ISBN 978-1-4276-3150-3. pp. 200-212

Trevett, A. F., R. C. Carter and S. F. Tyrrel. (2005). Mechanisms leading to post - supply water quality deterioration in rural Honduran communities. Int. J. Hyg. Environ.-Health 208: 153–161.

United Nations Environment Programme (2003). Water Resources Management in Latin America and the Caribbean. UNEP/LAC-GWG.XIV/Inf.5.

USBR. (1997). U.S. Department of the Interior, Bureau of Reclamation. Water Measurement Manual. 3rd ed. Available from http://www.usbr.gov/pmts/hydraulics_lab/pubs/wmm/ index.htm.

Veneklaas, E.J. and R. van EK. (1990). Rainfall interception in two tropical montane rain forests, Colombia. Hydrological Processes 4: 311-326.

WMO (World Meteorological Organization). (1990). Second World Climate Conference. Geneva, Switzerland.

CHAPTER 3

DISCHARGE-SOLUTE CONCENTRATION RELATIONSHIPS AND IMPLICATION FOR RUNOFF GENERATION ANALYSIS IN CLOUDFOREST WATERSHEDS IN CENTRAL HONDURAS

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ABSTRACT

Although it is generally known that cloudforest ecosystems are important sources of environmental services including water, biodiversity, and carbon sequestration, relatively few quantitative hydrologic and hydrochemical assessments have been made. Four adjacent watersheds (WS1-WS4) were studied in Honduras, one of which (WS1) included an undisturbed cloudforest. From April 2008 through December 2009, weekly or biweekly streamflow water samples were taken in four instrumented watersheds, and a single rainfall event was intensively sampled throughout all sites. Streamwater physical and chemical analysis included electrical

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conductivity, pH, turbidity and major ions (Ca, SiO_2-Si, Mg, Na, K, Al, PO_4-P, Fe, SO_4-S, NO_3^- N and Cl). Multivariate analysis of variance (MANOVA) was applied to elucidate major differences between the cloudforest watershed and non-cloudforest watersheds as well as seasonal trends. The streamflow ionic composition was dominated by Ca and SiO_2-Si, both having mean concentrations around 4.0 mg/L. Ca was highly variable among catchments, with WS1 having the lowest mean concentrations and WS2 the greatest. Water pH was similar among all catchments (7.0 +/- 0.5). Macronutrient (K, NO_3-N and PO_4-P) concentrations were low, with slight increases in P and K concentrations during rainfall events. Multivariate cross-correlation indicated significant differences ($P<0.0001$) in ionic solute concentrations among cloudforest and non-cloudforest stream flows for those elements linked to parent material composition of the catchments (Ca, Mg, SO_4-S, Na and to some extent SiO_2-Si), but not for those elements linked to plant-soil-water interfaces (DOC, NO_3-N, P, K).

Key words: Central America, tropical cloudforest catchments, biogeochemistry, ion concentrations, streamflow, water quality.

INTRODUCTION

Cloudforests are generally defined as “forests that are frequently covered in cloud or mist” (Stadmuller, 1987; Hamilton 1995), and are usually located on high mountains between 1,500 and 3,000 m a.s.l. (Bruijnzeel 2004) where moist ascending air masses form clouds (Zadroga, 1981). Bruijnzeel (2001) stressed the importance of temperature and humidity on montane forest zonation. Tropical cloudforest ecosystems are important sources of water, biodiversity, carbon sequestration and other environmental services. Despite their importance, these ecosystems are under a great threat in Central America due to deforestation and subsequent land conversion to agriculture (Bruijnzeel, 1990, 1996, 2004). Poor understanding of the hydrologic and biogeochemical functioning of these ecosystems hinders conservation efforts because the harm done by deforestation cannot yet be fully quantified (Feddema et al., 2005; Bucker et al., 2010).

Most studies in cloudforest hydrology deal with fog contribution to precipitation (Lowett, 1984; Cavelier and Golstein, 1989; Stadmuller et al., 1990; Bruijnzeel and Proctor, 1995; Caveleir, et al., 1996; Bruijnzeel, 2001; Holder, 2004 and Schmid, 2004) concluding that fog, with few exceptions, contributes an average of 1.0 mm d^{-1} during the dry season and 0.5 mm d^{-1} during the rainy season. In addition, fog minimizes evaporation with the result that cloudforests contribute more base flow than the surrounding non-cloudforest areas. Few researchers have performed water balances of cloudforests in Latin America (Cavelier, et al., (1997) in Panama); Bruijnzeel, et al., (2006) and Schellekens (2006) in Costa Rica; Charlier, et al., (2008) in Guadalupe). Very few studies have looked at the differences in chemical composition in runoff water from cloudforests vs. surrounding non-cloudforests.

The relationship between streamwater chemistry and stream discharge has been used for characterizing the origin and flow regime of water reaching the watershed outlet (Cirmo and McDonnell, 1997; Holko and Lepisto, 1997; Perakis, 2002). This characterization can help explain the interrelationships among physical, chemical and biological processes occurring within a catchment (Newbold, et al., 1995). Solute concentrations are dependent on the pathways and water residence time in the watershed system (Bishop et al., 1990; Mulholland et al., 1990, McDowell and Asbury 1994; Holloway et al., 2001). During the dry season when streamflow consists of baseflow, water chemistry is mainly controlled by bedrock type, climate and water residence time (e.g. Reynolds et al., 1987; Bishop et al., 1990; Durand et al., 1991). In contrast, during wet season high flows, streamwater consists of three source components: direct surface runoff, interflow and baseflow, each of which has distinctive chemical constituents because they have interacted to a different extent with plant canopy surfaces, forest litter, soil minerals, soil organisms and bedrock (Parker, 1983). Streamwater chemistry thus depends on the relative volumetric contributions of the three flow components as well as rainfall falling directly on the stream, weighted by their current contributions to streamflow. Using streamwater chemistry Genereux and Pingle (1997) found that, in La Selva Biological Station in Costa Rica, most runoff was baseflow. In Puerto Rico's Luquillo experimental forest, a chemical mass balance model provided valuable (albeit incomplete) information about water flow paths (Sckellekens, et al., 2004). Nevertheless very few studies have been carried out for regions with tropical monsoon climates, where hydrologic regimes differ substantially from temperate climates and where temperate-climate based models may fail to represent essential mechanisms, including cloudforest impacts.

The main objective of this paper was to evaluate the spatial and temporal variability in water chemistry between the cloud-forest and the non-cloudforest subwatersheds, and elucidate if any difference can be attributed to differences in runoff mechanisms and flowpaths. The study site was the La Tigra National Park, 20 km northeast of Tegucigalpa, which we monitored for eighteen months (May, 2008-December, 2009). The 880 ha experimental site consisted of 4 neighboring watersheds, the largest of which is covered with mainly broad leaved cloudforest, while the other three watersheds are primarily pine forests (Caballero, et al., 2011). The cloudforest had distinctly less water storage in the root zone than the other three watersheds resulting in four times as much discharge per area (Caballero, et al., 2011).

Since the geology was the same for the whole watershed, differences in chemistry can be attributed directly to differences in flow paths.

MATERIALS AND METHODS

Study sites

The study area was located (87º 5’ W Long., 14º 10’ N Lat., WGS84 Datum) within la Tigra National Park in the headwaters of the Choluteca River Basin which drains into the Pacific Ocean (Figure 3.1). Since the 1940’s the study area has been protected, first as a forest reserve and more recently as a national park. Although all catchments are forested, the amount of cloudforest area differs substantially among the catchments (Table 3.1). In La Tigra National Park, the cloudforest begins at 1500 m and reaches to mountain tops at 2270 m.a.s.l, consisting of a multistoried mixture of broadleaf trees are abundant and heavily covered by epiphytes, including bromeliads, orchids, mosses and orborecent ferms. Plant species are very similar to Uyuca Mountain which includes species *Fagaceae, Lauraceae, Aquifoliacea* and *Podocarpaceae* (Agudelo, 2010). Below 1500 m, evergreen pine tree species dominate the landscape.

Climate

The climate is characteristic of monsoonal regions with distinct dry and wet seasons. The wet season (in which almost 90% of the annual precipitation falls) begins in late May or early June when the Intertropical Convergence Zone (ITCZ) becomes active, bringing warm, moist clouds from the eastern Pacific to Central America and the Caribbean (Hastenrath, 2002) and continues through October. The dry season

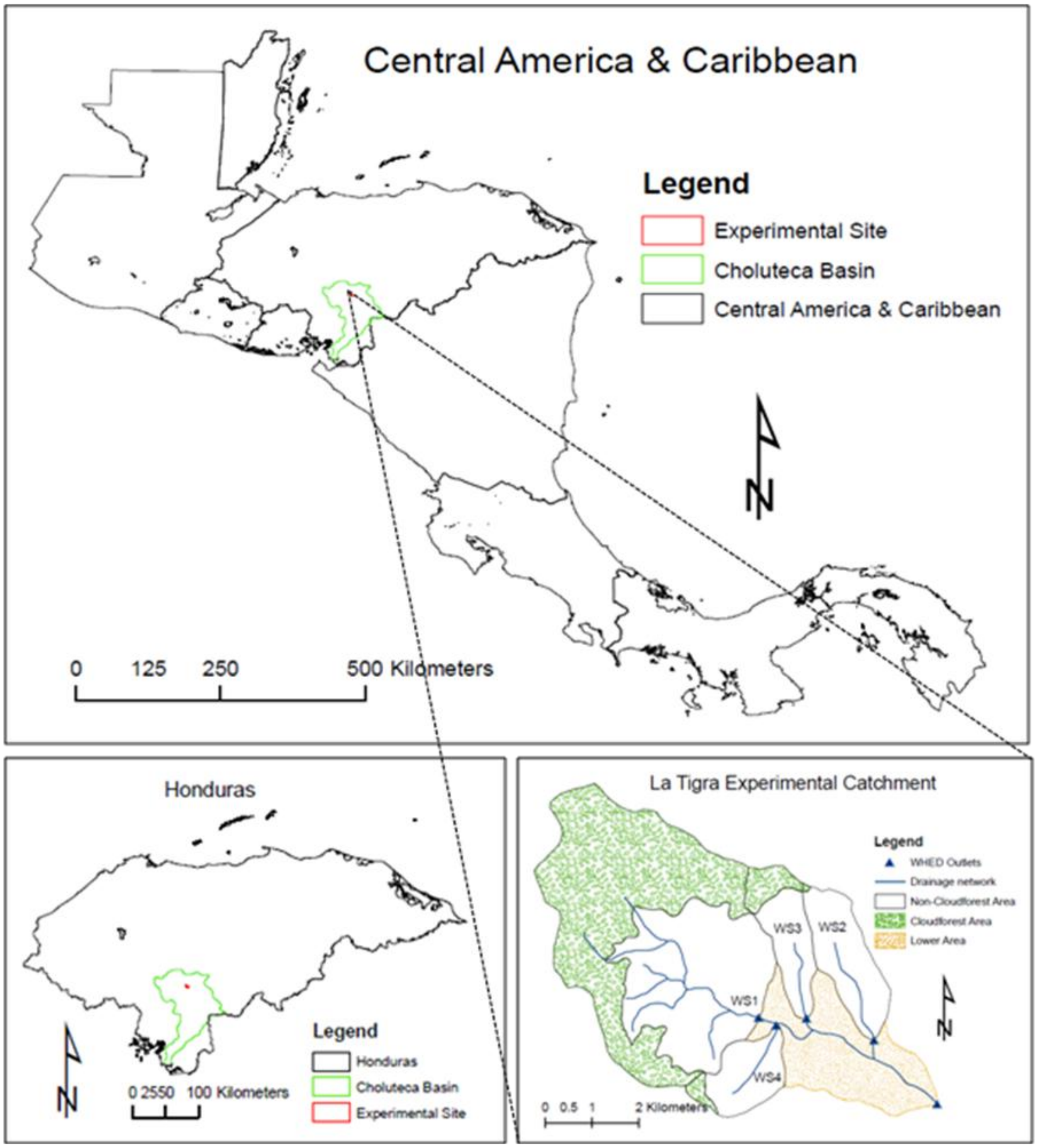

Figure 3.1. Study site, La Tigra Experimental Catchment, Honduras C.A.

Table 3.1. Characteristics of the four study catchments and their streams in La Tigra National Park, Honduras, Central America.

	CATCHMENT			
	WS1	**WS2**	**WS3**	**WS4**
Catchment area (ha)	635	93	82	70
Cloudforest area %	58	0	4	0
Other forested %	41	100	96	84
Deforested %	1	0	0	16
Weir elevation at outlet (m)	1505	1374	1431	1486
Elevation range (m)	1505-2270	1374-1850	1431-2000	1486-1960
Mean elevation (m)	1905	1625	1730	1715
Mean slope (%)	22	20	27	30
Main stream channel length (m)	6600	1508	1105	994
Main stream channel slope (%)	18	14	18	21
Drainage density (km/km^2)	1.00	1.62	1.35	1.42
Mean annual temperature (ºC)	16-20	16-20	16-20	16-20
Mean annual precipitation (mm)	1085	1085	1085	1085
Mean annual discharge (mm)	520	-	-	-
Geology (bedrock formation)	Volcanic	Volcanic	Volcanic	Volcanic
Period of measurements	Apr 2008-Dec 2009	Apr 2008-Dec 2009	Wet season 2008-2009	Wet season 2008-2009
Total number water of samples	132	130	129	115
Type of stream	Perennial	Perennial	Intermittent	Intermittent

begins with a very sharp reduction of daily precipitation from November through January, with no significant precipitation occurring from February through April. The total precipitation for the observation period (May 2008 through December, 2009) was 2170 mm, with annual averages (from four digital rain gauges) of 1150 for 2008 and 1020 mm for 2009. These gauge averages do not take into account cloudforest fog contributions which might be in the order of an additional 0.3 to 1.3 mm day^{-1} (Stadmuller and Agudelo, 1990).

Soil and Geology

Detailed soil maps were not available for the study catchments. The study site is underlain by Andisols formed on volcanic parent material mapped as primarily basaltic and andesitic magma (IGN, 1990). Igneous rocks, such as basaltic and andesitic (volcanic) ash and tuff and carbonate rocks and clastic sediments are predominant in the area (Carpenter, 1954, Mann, 2007). According to the geologic map of the Honduras Geological Institute (IGN, 1990), chemical composition of igneous rocks from the study area is dominated by silicon (50-59%), aluminum (17%), iron (9%), calcium (8%), magnesium (5%), sodium (3%), potassium (1.5%) and phosphorus (4%). In similar conditions, Martinez (2007) found organic matter contents of 8% and high infiltration rates, which we suspect are typical for our study site.

Field instrumentation

The four adjacent headwater catchments (WS1, WS2, WS3 and WS4) together comprise an area of 880 ha. WS1 is the largest (635 ha) with approximately 60% cloudforest coverage, and serves as the main water source for Tegucigalpa. The other catchments are similar in size and only WS3 has a small area (3%) under cloudforest

(Table 3.1 and Figure 3.2). Most of the land cover is old-growth forest, with deforested areas (approximately 6.0 ha in WS1 and 8.5 ha in WS4) occupying less than 2% of the total study area. Geomorphic characteristics such as catchment form, stream length, stream slope, mean elevation and drainage density are similar for WS1, WS2 and WS3 (Table 3.1), whereas WS4 has a more oval shape, steeper relief (30% mean slope) and shorter stream length (Table 3.1). WS1 and WS2 have year-round streamflow. WS3 has a community water supply intake upstream of the weir and streamflow ceases shortly after each rainfall. The stream from WS4 dries up after rainfalls cease in October.

Water sampling and analysis

Water grab samples (a total of 440 as detailed in Table 3.1) were collected either weekly or biweekly from April 2008 through December 2009 at the catchment outlet weirs. For the WS1 watershed discharge and solute concentration were available for the entire period from May, 2008 through December 2009. The WS2 watershed had solute concentrations for the entire monitoring period, but discharge data was available only for the wet season of 2009. Samples for WS3 and WS4 could only be taken during the 2009 wet season as the streams dry up at other times. WS4 had the lowest amount of available discharge and solute data due to the fact that the streamflow ceases even during the wet monsoon phase (Table 3.1). An intensive sampling event was performed during the period from September 5 to 11, 2009, during which 75 water samples were taken, including 51 samples collected twice a day prior to a 33 mm rainfall event on September 11, 2009, and 24 samples collected every two hours through the rising and falling limbs of the resulting storm hydrograph. Due to logistics, shallow piezometer well samples (20) and precipitation (15) water samples were only collected during October and November of 2009 and were intended to

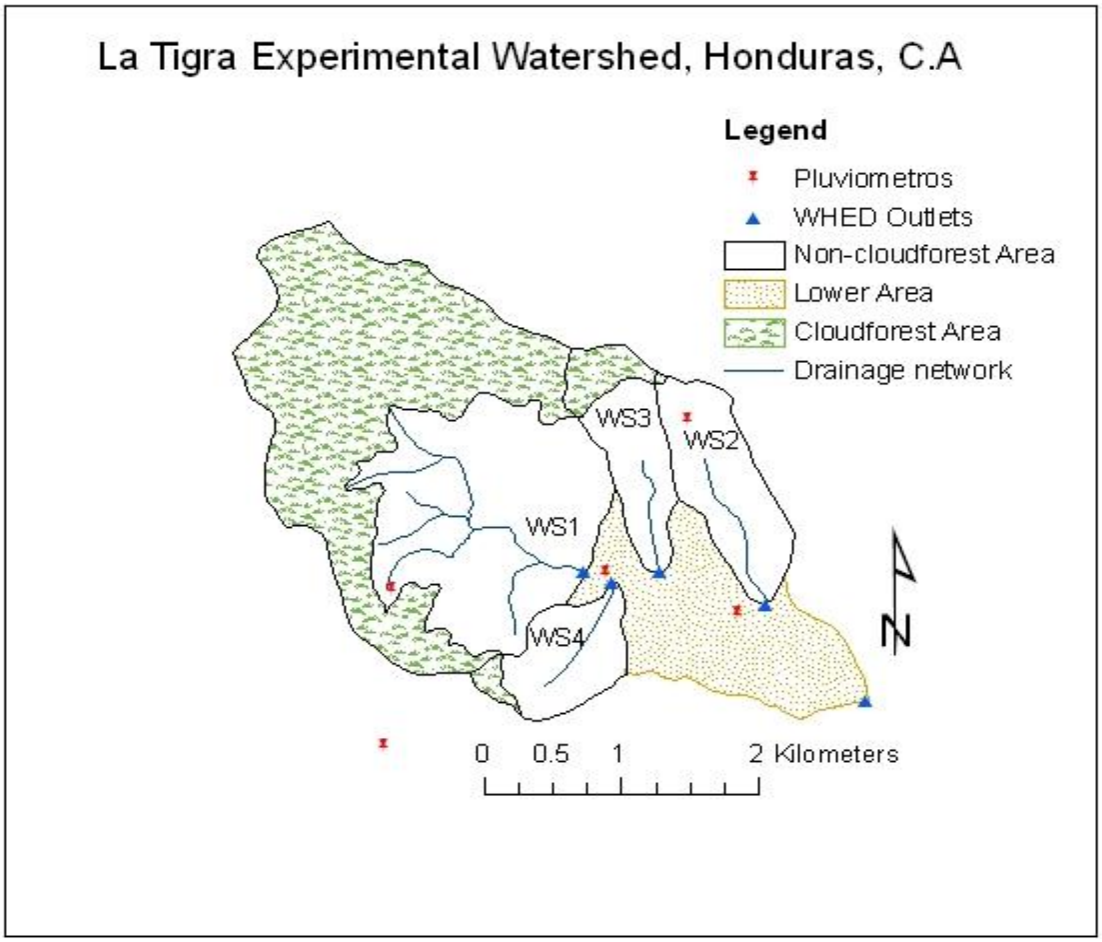

Figure 3.2. Land cover, watershed areas under cloud and non-cloudforest, and subwatersheds

provide only a broad picture of the soil water and precipitation chemistry. Long-term discharge–solute concentration relationships are presented only for WS1 due to insufficient long-term coupled hydrometric and hydrochemical data for the other sites. WS1 data encompassed both dry and wet seasons and therefore provided good grounds for solute-discharge evaluations. Information concerning the stream discharge measurements is given in Caballero (Caballero, et al., 2011). Here we use an intensive event sampling to explore preliminary discharge solute concentration relationships in the four sites.

Laboratory Analysis

Precleaned polystyrene 250-mL sample bottles were preconditioned by washing with stream water three times before sample filling, and were transported at 4ºC to Zamorano University's Water Quality Laboratory. Water samples were filtered (0.45 µm polypropylene membrane) and split into 65 mL bottles, with one subsample stored for ion screening while the other was acidified (0.15 ml 2% NHO_3) to pH <2. All samples were transported and stored at cool temperatures for later analysis at Cornell University's Soil and Water Laboratory.

Non-acidified samples were analyzed for Cl^-, NO_3-N and SO_4-S by ion chromatography (Dionex ICS-2000) with minimum detection limits of 0.9, 0.1 and 0.1 mg L^{-1} respectively. Acidified water samples were analyzed for total organic carbon (TOC) using an OIAnalytical Model 1010 TOC analyzer (persulfate oxidation), and for elemental composition via inductively coupled plasma spectrometry (Thermo Jarrell Ash TRACE ICP) including Ca, Mg, Fe, Na, P, K, Al, Mn, Mo, and Si as well as trace metals. Instrument detection limits (mg L^{-1}) for primary analytes were: Al (0.01), Ca (0.11), Fe (0.3), K (0.5), Mg (0.25), Na (0.65), P (0.02) and SiO_2 (0.025). We used these values as background levels that were therefore deducted from measured values for each chemical constituent. Other parameters such as pH, electric conductivity (µS/cm), turbidity, dissolved oxygen, and temperature were measured in the field during sampling by means of a Horiba U-10 portable water quality monitoring system. Quality control for pH, conductivity and turbidity was done in the laboratory and included the use of a pH/conductivity meter (AR50 Fisher Scientific) and a turbidimeter (Hach). Missing values on each catchment (typically 4 to 6) for both physical and chemical parameters were filled by linear interpolation using preceding and subsequent values.

Data analysis

Preliminary statistical analysis consisted of evaluating means and standard deviations of concentration for each element and flow condition (dry vs. wet season) after identification and removal of outliers. Outliers are atypical values which are often found in ecological studies frequently involving large numbers of variables and observations (Jackson and Chen, 2004). Outliers often come from various errors in the data sets and tend to bias the interpretation if they are not representative of the study population. These values (outliers) were identified visually or using statistical means such as identifying points exceeding 3 or more standard deviations from the trend line. In order to compare how solutes varied as a function of hydrology, we worked under the conceptual framework that streamwater solute concentration can exhibit one of the three general trends with respect to increasing stream discharge: dilution, enhanced hydrologic access (increasing with increased flow), or hydrologically constant (Salmon et al., 2001, Asano, et al., 2009). We also conducted a multivariate analysis of variance (MANOVA) to elucidate if any difference exist ($P<0.001$) in solute concentrations between the cloudforest watershed (WS1) and the non-cloudforest WS2, WS3, and WS4 watersheds.

RESULTS

Catchment hydrology and hydrochemistry

Two types of analysis are presented. First, we describe the chemistry of water entering the watershed system and the relative changes as it moves through the soil profile and subsequently leaves as streamflow. The data set (period of analysis) corresponds to water samples from October 2009 through November, 2009 when solute concentrations were available for precipitation, perched groundwater and the stream flow for the WS1 watershed (Table 3.2).

Secondly, two overlapping periods of analysis are presented: a) Solute chemistry results for 115 to over 130 samples for each watershed which correspond to the whole sampling period of May 2008 through December 2009 (Tables 3.3 and 3.4), and b) the data set corresponding to a single 33 mm storm event for which both concentration and discharge were measured during rising and falling limbs of the hydrograph in all watersheds. Table 3.3 presents the averages values of all watersheds for both, baseflow and peakflow conditions.

Solute concentrations in precipitation, soils, baseflow and peakflow conditions.

Precipitation chemistry

The short period (October-November 2009) of precipitation sampling in La Tigra National Park found that rainfall was slightly acidic (median 6.4, range 5.9-6.8; Table 3.2) and had a low ionic strength, being composed primarily of DOC and ions dominated by Cl^-, Ca, SO_4-S, NO3-N, and SiO_2-Si in that order (Table 3.2). As

Table 3.2. Rainfall composition at study site and other regional sites.

	Honduras*	Costa Rica*	Panama*	Puerto Rico*
TOC	1.10	0.17	-	0.96
Cl	0.70	2.27	1.95	3.44
NO_3-N	0.25	0.02	0.29	0.03
SO_4-S	0.34	0.12	0.85	-
Al^+	0.00	-	-	-
Ca^{2+}	0.38	0.14	0.74	0.37
Fe^{2+}	0.00	-	0.13	-
K^+	0.00	0.09	0.3	0.13
Mg^{2+}	0.00	0.06	0.14	0.24
Na^+	0.00	0.89	1.56	1.81
PO_4-P	0.00	-	-	-

La Selva Biological station, Costa Rica
La Fortuna, Panama (Cavelier et al., 1997)
El Verde, Puerto Rico ((MacDowell, 1998)
La Tigra National Park, Honduras (present study)

Table 3.3. Storm event mean concentrations (mg L^{-1}) of chemical constituents, pH, conductivity (µS cm^{-1}) and turbidity (NTU) sorted by water type: Precipitation, Shallow Groundwater, Baseflow, and Peak flow (n= number of samples).

Parameter	**Sources of Water**			
	Precipitation (n=15)	**Shallow groundwater (n=20)**	**Base flow (n=41)**	**Peak flow (n=4)**
pH	6.40	NA	6.90	6.70
Turbidity (NTU)	1.90	NA	4.20	51.20
Conductivity (µScm^{-1})	NA	NA	71.00	50.10
Q (mm/h)	-	-	0.04	0.05
DOC (mg L^{-1})	1.10	0.72	2.31	6.37
Cl	0.70	1.85	1.60	1.46
NO_3-N	0.25	0.02	0.20	0.31
SO_4-S	0.34	1.23	1.20	0.95
Al^+	0.00	0.00	0.01	0.01
Ca^{2+}	0.38	1.61	4.30	3.16
Fe^{2+}	0.00	0.00	0.00	0.03
K^+	0.00	0.23	0.14	0.44
Mg^{2+}	0.00	0.47	0.63	0.48
Na^+	0.00	0.29	0.09	0.04
PO_4-P	0.00	0.06	0.03	0.04
SiO_2-Si	0.12	1.81	3.67	3.14
Total chemical load	2.89	8.29	14.16	16.43

Precipitation (n=14) collected October-November 2009

Shallow groundwater (n=21) collected October and November 2009 (0.6-1.20 m depth)

Baseflow (n=41) based on average concentration during pre-event condition September 7-11 (4 sites)

Peakflow (n=4) based on average concentration at peak runoff from in 4 sites (September 11 at night)

Table 3.4. Means and standard deviation of solutes in all sample stations during dry and wet season. EC in µS cm^{-1}, turbidity (NTU), all other in mg L^{-1}

	Dry season					Wet Season					Total
Parameter	WS1	WS2	WS3	WS4	Mean	WS1	WS2	WS3	WS4	Mean	Mean
pH	6.30 (±0.8)	7.22 (±0.5)	6.9 (±0.5)	7.0 (±0.4)	6.9	6.2 (±0.6)	7.36 (±0.5)	7.0 (±0.6)	7.50 (±0.7)	7	6.90
EC	26.4 (±6.6)	112 (±47)	40.8 (±22)	53.0 (±9.0)	58	22 (±6)	88 (±44)	26 (±20)	42.8 (±7.7)	44.7	51.00
Turbidity	4.77 (±12.5)	1.6 (±2.0)	2.6 (±6.2)	19 (±17.0)	7	10.5 (±23)	9.3 (±12.0)	8 (±13)	24.8 (±23.3)	9.3	8.00
TOC	0.54 (±0.15)	0.88 (±0.45)	0.62 (±0.55)	1.24 (±0.41)	0.82	2.39 (±1.30)	2.92 (±0.75)	2.19 (±1.18)	2.52 (±1.25)	2.505	1.66
Cl-	1.05 (±0.80)	1.43 (±0.65)	1.11 (±0.79)	2.98 (±0.59)	1.62	0.78 (±1.21)	0.92 (±0.31)	0.82 (±0.61)	1.95 (±1.23)	1.118	1.37
NO_3-N	0.18 (±0.49)	0.22 (±0.47)	0.09 (±0.27)	0.19 (±0.52)	0.17	0.24 (±0.57)	0.00 (±0.00)	0.01 (±0.04	0.12 (±0.78)	0.093	0.13
SO_4-S	2.11 (±0.40)	2.80 (±1.38)	1.07 (±0.74)	0.52 (±0.19)	1.62	1.13 (±0.65)	1.93 (±0.75)	0.62 (±0.19)	0.36 (±0.31)	1.011	1.32
Al	0.00 (±0.00)	0.00 (±0.00)	0.01 (±0.01)	0.01 (±0.01)	0.01	0.00 (±0.00)	0.00 (±0.00)	0.03 (±0.01)	0.01 (±0.01)	0.01	0.01
Ca^{+2}	0.99 (±0.12)	11.40 (±5.0)	2.26 (±0.80)	3.27 (±0.77)	4.5	0.88 (±0.70)	8.57 (±3.3)	1.85 (±0.45)	3.02 (±0.66)	3.568	4.03
Fe	0.00 (±0.00)	0.01 (±0.02)	0.00 (±0.00)	0.01 (±0.01)	0.01	0.00 (±0.00)	0.01 (±0.01)	0.02 (±0.03)	0.03 (±0.15)	0.059	0.03
K+	0.06 (±0.09)	0.08 (±0.12)	0.03 (±0.05)	0.08 (±0.23)	0.06	0.15 (±0.15)	0.14 (±0.13)	0.16 (±0.14)	0.04 (±0.08)	0.123	0.09
Mg^{2+}	0.24 (±0.03)	1.34 (±0.56)	0.48 (±0.23)	0.72 (±0.13)	0.69	0.20 (±0.07)	0.95 (±0.34)	0.37 (±0.09)	0.65 (±0.13)	0.542	0.62
Na^{+}	0.02 (±0.04)	0.04 (±0.028)	0.05 (±0.08)	0.05 (±0.10)	0.04	0.02 (±0.03)	0.05 (±0.11)	0.16 (±0.07)	0.01 (±0.02)	0.059	0.05
P	0.01 (±0.01)	0.02 (±0.02)	0.02 (±0.01)	0.03 (±0.01)	0.02	0.02 (±0.01)	0.02 (±0.01)	0.03 (±0.01)	0.03 (±0.01)	0.024	0.02
SiO_4	3.16 (±1.63)	3.71 (±1.14)	3.80 (±0.63)	4.38 (±0.82)	3.76	2.71 (±1.61)	4.36 (±1.02)	4.33 (±0.56)	4.75 (±0.63)	4.038	3.90
Totol IS	8.0	21.0	9.0	12.0	12.5	6.0	17.0	9.0	11.0	9.5	11.6

expected, we did not detect any Al, Fe, K, Mg, Na, and P in precipitation. Comparative data from the nearest site (La Selva Biological Station in Costa Rica) and elsewhere in Central American and the Caribbean is provided in Table 3.2. DOC at our site was greater than that in Costa Rica but comparable to that in Puerto Rico (MacDowell, 1998; Cavelier, et al., 1997). Similar DOC concentrations occur in forested catchments in temperate environments (Bilby and Likens, 1979; Likens et al., 1983). Chloride, on the other hand, was much less concentrated than at other sites, indicating a weaker oceanic influence, also confirmed by the absence of sodium. Other elements such as calcium were somewhat similar to levels in Puerto Rico, but lower than in Panama (MacDowell, 1998; Cavelier, et al., 1997).

Soil water chemistry

The concentration of chemicals in precipitation and soil water (collected with piezometers 0.9-1.2 m deep) is compared with baseflow and peakflow concentrations during a period of three months (September-November 2009) which includes a rainfall event that was sampled over the hydrograph in Table 3.3. Except for DOC and NO_3-N, there was an average threefold increase in total ionic strength in the shallow wells as compared to rainfall water passing through the canopy. Rainfall DOC concentration measurements (data not shown) were higher than shallow well DOC levels possibly due to due to dry matter deposition (from the surrounding area) in the sampler funnel. NO_3-N levels decreased due to uptake and/or denitrification in the soil. There was a subsequent fivefold overall increase in ionic strength in water leaving the watershed as baseflow or peakflow. The greatest increases occurred for silica, calcium, sulfate, magnesium and sodium (in that order). This behavior closely correlated to soil and bedrock mineralogy of the site, as described by the Honduran Geological Institute

(IGN, 1990). Similar enrichment in soil water has been reported (Bilby and Likens, 1979; Bruijnzeel, 1983b; Waterloo, 1994; McDowell, Weneklaas 1990; MacDowell, 1998; Halloway and Dahlgren, 2001).

Biweekly streamflow concentrations for the period from May 2008-August 2009

Stream water characteristics

Streamwater physical and chemical characteristics for dry and wet seasons are summarized in Table 3.4. Streamwater across all stations and seasons was generally circumneutral, with WS1 notably lower pH from the others in both dry (6.3) and wet (6.2) seasons. Turbidity, as expected in forested catchments, varied little between seasons and sites. Mean values ranged from 2-5 NTU (dry) and 8-10 NTU (wet) in WS1, WS2 and WS3. WS4 had slightly higher turbidities of 19 NTU (dry) and 25 NTU (wet, Table 3.4). Field observations during water sampling suggest this relative high turbidity was associated with suspended colloidal material. Mean electric conductivity (EC) for all sites was 51µS/cm (ranging from 22 to 112). However, as Table 3.4 shows, station WS2 had notably greater EC values in both the dry (112 µS/cm) and wet (88 µS/cm) season, corresponding to total ionic strengths that were at least twofold greater than the other sites, except for WS4 (Table 3.4). Unsurprisingly, general water physical characteristics were typical of forested ecosystems having stable and well-defined runoff processes dominated by subsurface flowpaths.

Water quality of cloudforest and nearby forested watersheds

To find if significant differences existed in biogeochemical processes between the cloudforest and the other nearby watersheds, a multivariate analysis of variance (MANOVA) was conducted (Table 3.5). Streamwater concentrations of Ca, Mg, SO_4-S and SiO_2-Si in WS1 differed significantly from those in WS2, WS3 and WS4

Table 3.5. Relationship among the cloudforest and the non-cloudforest area for each chemical constituent (MANOVA)

Solute	Catchment	Estimate	Standard error	t value	P > [t]	Adj. P
Cl	WS2	-0.220	0.130	-1.740	0.080	0.300
	WS3	-0.001	0.130	-0.010	0.990	1.000
	WS4	-1.460	0.130	-11.210	<0.0001	<0.0001
DOC	WS2	-0.410	0.135	-3.010	0.003	0.015
	WS3	0.080	0.136	0.590	0.555	0.935
	WS4	-0.390	0.136	-2.870	0.004	0.023
NO3-N	WS2	0.110	0.070	1.600	0.110	0.380
	WS3	0.170	0.070	2.450	0.010	0.070
	WS4	0.070	0.070	0.990	0.320	0.760
K	WS2	-0.001	0.020	-0.050	0.960	1.000
	WS3	0.014	0.020	0.680	0.500	0.900
	WS4	0.045	0.020	2.130	0.030	0.010
PO4-P	WS2	-0.010	0.002	-1.830	0.068	0.262
	WS3	-0.010	0.001	-6.500	<0.0001	<0.0001
	WS4	-0.010	0.001	-6.950	<0.0001	<0.0001
Fe	WS2	-0.006	0.008	-0.770	0.440	0.870
	WS3	-0.009	0.008	-0.117	0.240	0.640
	WS4	-0.014	0.008	-1.830	0.070	0.260
Al	WS2	0.000	0.001	-0.340	0.736	0.987
	WS3	-0.015	0.001	-15.140	<0.0001	<0.0001
	WS4	-0.008	0.001	-7.390	<0.0001	<0.0001
Ca	WS2	-8.940	0.340	-26.140	<.0001	<.0001
	WS3	-0.850	0.340	-2.750	0.006	0.031
	WS4	-2.090	0.350	-6.030	<0.0001	<0.0001
Mg	WS2	-0.910	0.040	-22.540	<0.0001	<0.0001
	WS3	-0.180	0.040	-4.490	<0.0001	<0.0001
	WS4	-0.440	0.041	-10.760	<0.0001	<0.0001
SO4-S	WS2	-0.790	0.110	-7.320	<0.0001	<0.0001
	WS3	0.731	0.110	6.660	<0.0001	<0.0001
	WS4	1.120	0.110	10.220	<0.0001	<0.0001
Na	WS2	-0.030	0.020	-1.580	0.110	0.390
	WS3	-0.090	0.020	-4.630	<0.0001	<0.0001
	WS4	-0.010	0.020	-0.660	0.510	0.910
SiO2-Si	WS2	-0.420	0.100	-4.170	<0.0001	<0.0002
	WS3	-0.253	0.100	-2.490	0.013	0.063
	WS4	-0.765	0.100	-7.500	<0.0001	<0.0001

P <0.0001 adjusted by Tuckey-Kramer; DF = 342-345 (p<0.0002; Tables 3.4 and 3.5). These elements are related to the bedrock chemistry and we speculate that the greater amount of discharge from watershed WS1 in combination with a consistent mineralization rate and/or greater prior rates of leaching may be the cause for the lower concentrations in WS1 Concentrations of P and Al were significantly (p<0.0001) greater in the drier watersheds WS3 and WS4 than in WS1 (Table 3.5). The second group of constituents comprised of DOC, NO_3-N, K and Fe showed no significant (<0.0001) difference in concentrations between WS1 and the other sites (Table 3.5). As Table 3.5 shows, Cl was significantly (<0.0001) different only in WS4, while Na differed (<0.0001) only in WS3.

Discharge - concentration relationships for the WS1 cloudforest

In this section, the relationship between discharge and the concentration of the various chemicals is examined for the cloudforest (WS1) watershed using the data from October 2008 to December 2009. In general, the behavior of the various ions fitted one of the three discharge-solute relationships described by Salmon, et al., (2001) and Asano, et al., (2009): enhanced hydrological access (increasing concentration), dilution and hydrologically constant.

Increasing concentration control

Enhanced hydrological access (increasing concentration) refers to conditions in which chemical constituents increase in response to increasing stream discharge. This is thought to occur for those elements that leach from the plant canopy and soils only during rainfall/runoff events, as well as constituents (e.g. DOC, phosphorus, nitrate and iron) coming from hydrologically active areas of the catchments during periods of high flow, such as streamflow contribution from interflow and saturated areas. This type of relationship is quite complex due to dependence on plant-soil-water relationships which are highly dynamic; therefore (as our results show) relationships will vary depending on the season, antecedent watershed wetness, and frequency of sampling. DOC data plotted for the period May through September 2009 (earlier samples being unreliable due to inadequate preservation practices) had a strong streamflow correlation (r^2 = 0.68; Figure 3.3a). Likewise, data for the runoff event (September 5-13, 2009) sampling resulted in an even stronger correlation of (r^2 =0.91) for WS2 (Figure 3.3c) and (r^2 =0.75) for WS4 (Figure 3.3e), and a little less for the other two sites (Figure 3.3b and 3.3d).

Decrease in concentration with discharge (dilution) control

Decrease in concentration with increasing discharge (or dilution) occurs for those elements that are present in the system's parent material and which are only slowly mineralized and transported by water exiting the watershed. Dilution generally occurs when the net delivery of water to the stream is greater than the increase in chemical delivery (Salmon, et al., 2001). This type of relationship is expected in those chemical constituents having strong linkage to deep soil weathering, in which case water chemistry carries the signal associated with rock and parent material of the site. Dilution behavior has been widely observed in the tropics, including for sodium (MacDowell and Asburry 1994, MacDowell et al., 1998). In the WS1 cloudforest

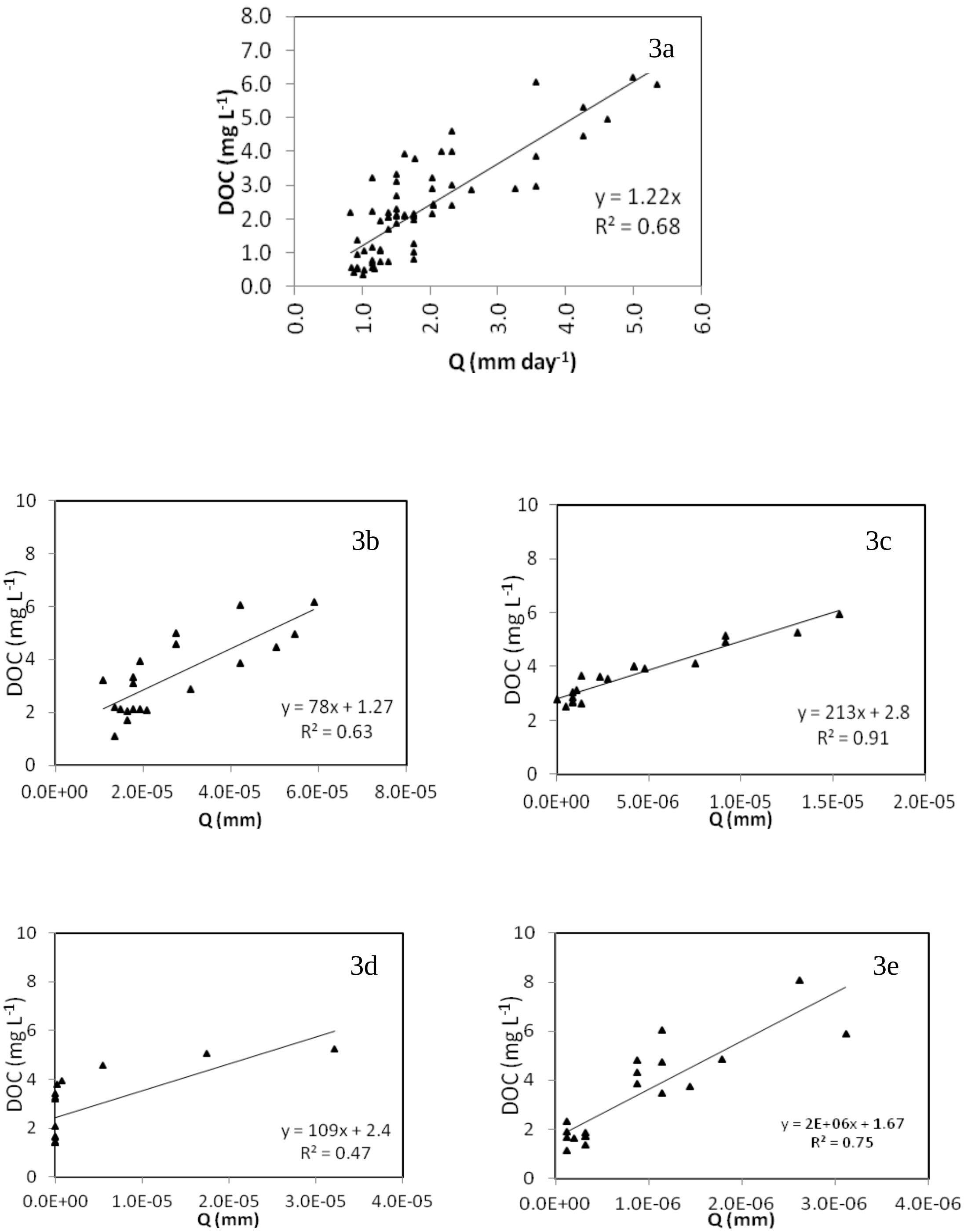

Figure 3.3. Enhanced type discharge solute concentration relationships on DOC for WS1 in wet season 2009 (3a) and for event data for all sites: 3b (WS1), 3c (WS2), 3d (WS3) and 3e (WS4).

watershed (Figure 3.4), dilution was observed over the whole period of monitoring for Ca (R^2 = 54), Mg (R^2 =43), and for sulfate-S (R^2 = 78), with correlations based on subtracting instrument detection limits (as per the methods section) and removal of <10 outliers. For instance, Ca, Mg and S are major constituents of the rock parent material of the site (IGN, 1990) and require a long process of mineralization. Rainfall events (in the short term) do not hasten these mineralization processes in this context, and the elements are thus diluted when streamflow increase as a result of rainfall inputs to the catchment (Germer et al., 2009). Ca, Mg and S showed higher correlation coefficients for event data (see equations in Figures 3.5a for Ca, 5b for Mg and 5c for SO_4-S). This dilution type of relationship with increased discharge exhibited by Mg, Ca, and SO_4-S are indicative of strong internal weathering association to geological features of the watershed.

As typically expected, Mg concentrations, although smaller than Ca, were tightly linked to Ca across catchments and seasons. In WS1, Ca:Mg ratio was 2:1, in WS2 was 8:1, in WS3 was 4:1 and in WS4 3.5:1; for an overall ratio close to 4:1 by mass (Table 3.4). Further cross correlation analysis among different chemical constituents in WS1, confirmed a Ca:Mg correlation coefficient (r^2) of 84.0 for long-term data (Table 3.6).

Hydrologically constant control

Hydrologically constant control refers to those in which chemical delivery changes in direct response to changes in water delivery. This relationship is generally expected for elements delivered through precipitation and that also do not have strong consumption or production in the watershed system. Chloride is the most common example for this type of chemical concentration-runoff relationship and is thus widely used to elucidate the contribution of new water (precipitation) to streamflow when Cl^-

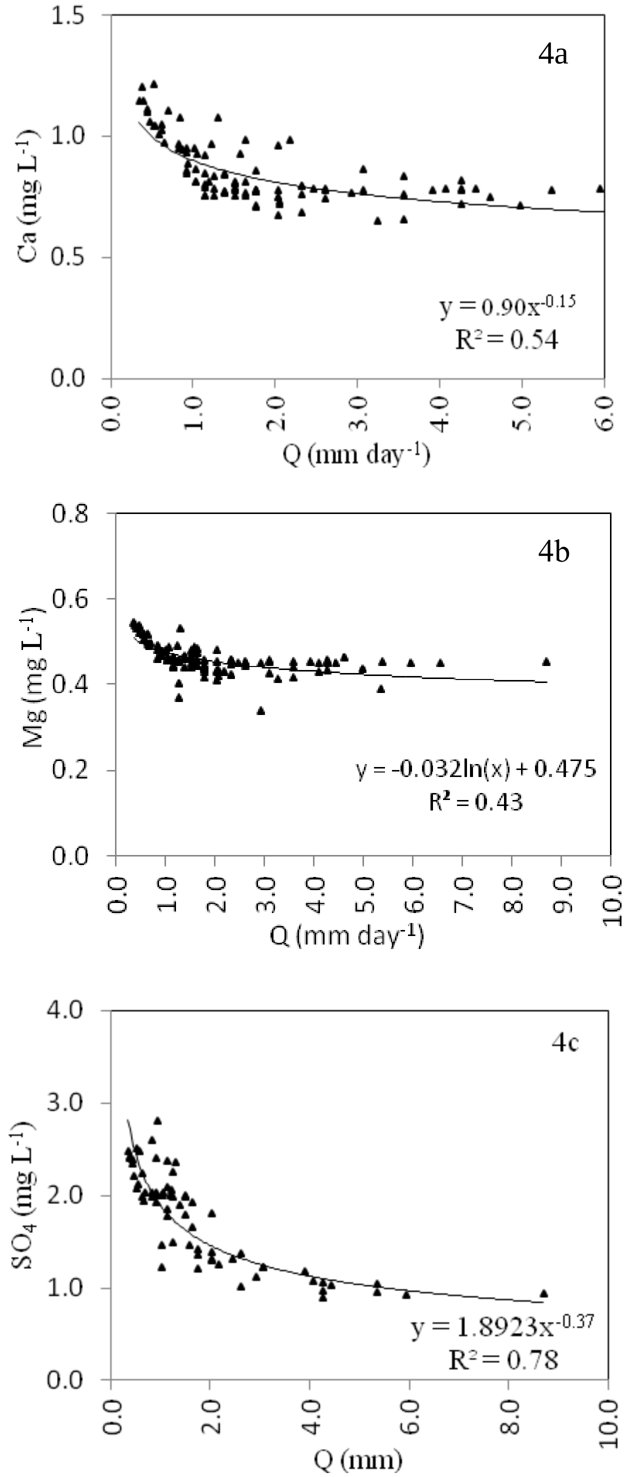

Figure 3.4. Dilution-type discharge solutes concentration relationships for WS1 for long-term data (4a for Ca, 4b for Mg and 4c for SO_4-S). To achieve a better relationship some outliers were removed as explained in materials and methods.

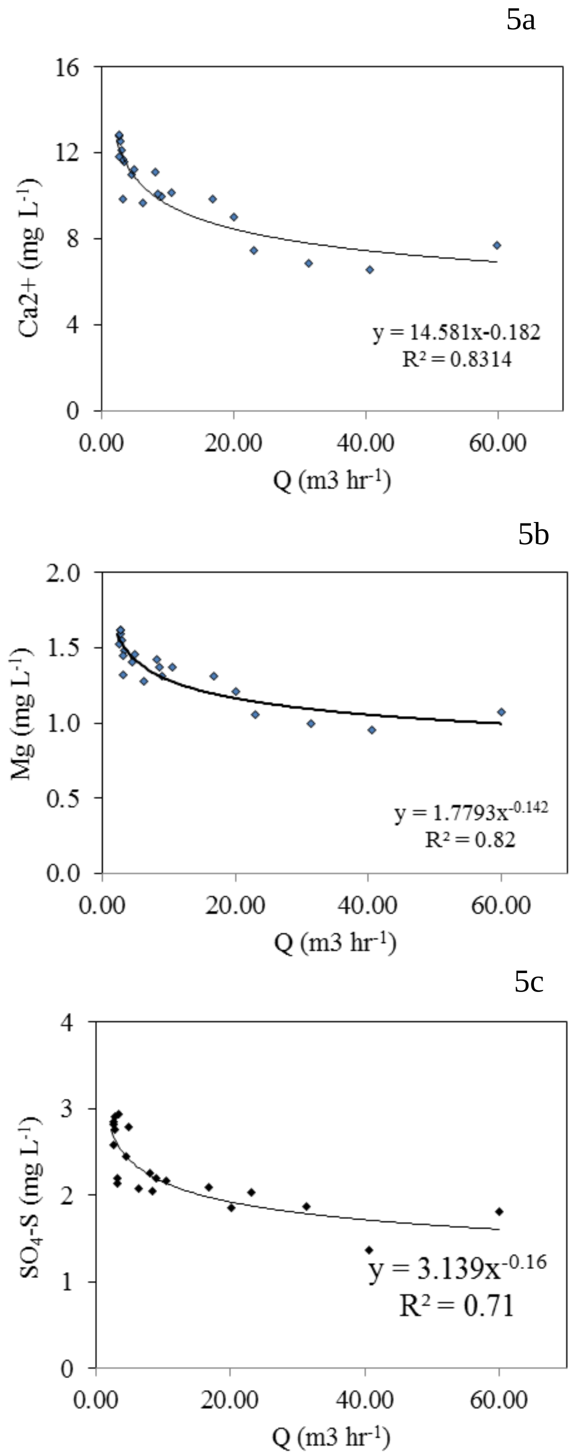

Figure 3.5. Dilution-type discharge solutes concentration relationships for WS1 during September 5-13 which includes a 33 mm rainfall event: 5a for Ca, 5b for Mg and 5c for SO_4-S.

Table 3.6. Correlation matrix of chemical constituents in WS1 catchment during the entire sampling period (May 2008-December 2009).

Element	TOC	Cl^-	NO_3-N	SO_4-S	Al^+	Ca^{2+}	K^+	Mg^{2+}	Na^+	SiO_2 -Si	Fe^{3+}	Total P
TOC	1.00											
Cl^-	-0.26	1.00										
NO_3-N	-0.12	-0.06	1.00									
SO_4-S	-0.38	0.34	0.19	1.00								
Al^+	-0.03	-0.26	-0.05	0.22	1.00							
Ca^{2+}	-0.02	-0.13	0.14	0.20	0.32	1.00						
K+	0.65	-0.05	-0.02	-0.09	-0.12	-0.04	1.00					
Mg^{2+}	-0.09	-0.11	0.09	0.39	0.43	0.84	0.00	1.00				
Na^+	0.35	0.19	0.06	0.11	-0.58	-0.01	0.55	0.07	1.00			
SiO_2 -Si	-0.08	0.08	-0.03	0.13	0.57	0.22	-0.20	0.33	-0.31	1.00		
Fe^{3+}	0.55	-0.08	-0.02	-0.05	0.31	0.21	0.54	0.22	0.13	0.08	1.00	
P	0.30	0.05	0.09	0.13	-0.41	-0.02	0.55	0.02	0.81	-0.26	0.16	1.00

concentrations in rainfall are significant and runoff occurs from permanently saturated areas and from area that become hydrological active during rainfall events. Hydrologically constant behavior was observed in the majority of elements: Na, SiO_2-Si, P and K. P and K exhibited high variability in concentration during low discharges, while Na and SiO_2-Si had more stable concentrations across different discharge conditions (Figure 3.6 and Table 3.4). In general, mean silica concentrations were somewhat similar between dry and wet seasons and across sites (Table 3.4), implying a weak dilution effect, which is also confirmed by the event data (Table 3.3). However silica also showed a relatively high standard deviation and coefficient of variability with event data (Table 3.4).

DISCUSSION

DOC trends

Cloudforest watersheds are generally rich in soil organic matter, with soil concentrations sometimes exceeding 10% (Martinez, 2007) and have high infiltration capacities (Hanson et al., 2004; Mendoza et al., 2002). Infiltrated water becomes enriched by contact with organic matter present when it passes through the forest canopy, the forest floor litter layer and the uppermost soil horizon. As can be observed in the event hydrographs across sites (Auxiliary material 1 and Figure 3.3), DOC concentrations show a relatively strong correlation with event-based streamflow with R^2 ranging from 0.47 to 0.91. This similitude in DOC release might indicate that same runoff mechanism were responsible for the observed discharge hydrograph which was

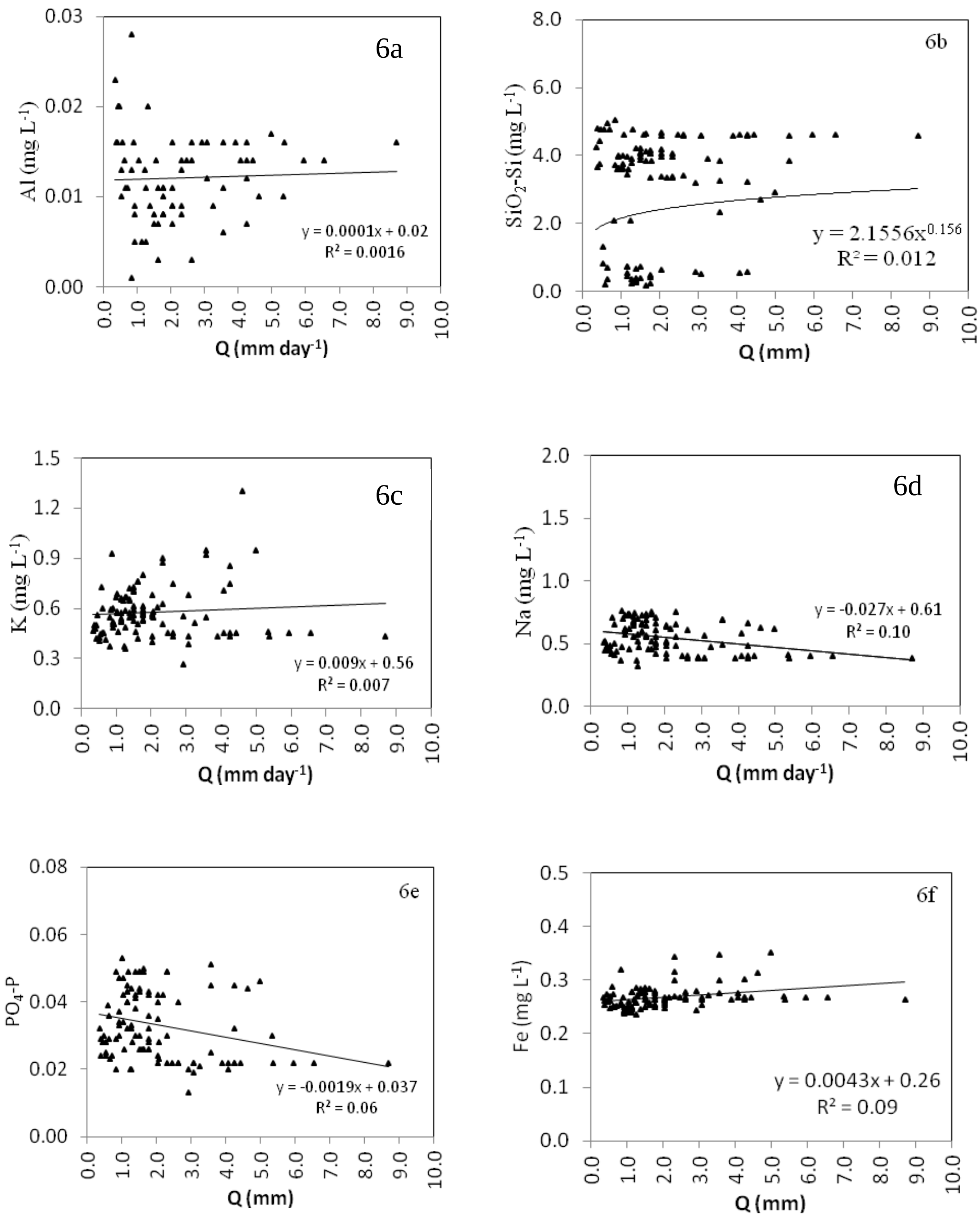

Figure 3.6. Constant-type discharge solutes concentration relationships for WS1 (long-term water samples, 2008-2010) 6a for Al, 4b SiO_2-Si, 4c for K, 4d for Na, 4e for PO_4-P and 4for Fe. To achieve a better relationship, some outliers were removed as explained in narrative.

composed of two type of source water: deep water percolation (baseflow) and a mixture of interflow, exposed bedrock and wet area runoff as it has been shown in the Ethiopian highlands (Steenhuis, 2009). Some studies found DOC peaking between runoff events, but other times peaking between precipitation events (Inamdar, et al., 2006). This mixture of results might indicate that DOC concentration is dependent on previous flushing events as the study of Inamdar et al. (2006) clearly shows. Since we had only one rainfall event we can't assure whether this behavior is recurrent. Similar findings were reported in forested watershed in the tropics (Johnson, et al., 2007, Salmon, et al., 2001). Regarding the DOC concentration in the stream water, our results are comparable to those found in similar studies across different forest ecosystems; similar DOC ranges (0.6-1.8 mg L^{-1}) were found in forested catchments of volcanic origin in Costa Rica (Newbold, et al., 1995) and at Hubbard Brook experimental forest (1.0 mg L^{-1}; Hobbie and Likens, 1973); while 2.1 mg L^{-1} was reported in a Chilean forested catchment (Salmon, et al., 2001). Similar behavior was found in a Brazilian Amazon headwater catchment, but with much greater concentration (>18.0 mg L^{-1}) than in our sites (Johnson, et al., 2006).

Ionic constituents

With regard to other streamwater constituents, our results were relatively similar to discharge-concentration relationships observed in other tropical rivers. For example, Lewis and Saunders, (1989) found strong dilution-type relationships for Na, Ca, Mg, SO_4 in the Orinoco River, Venezuela. Similar dilution relationships were found in the Chilean forest for Cl and Ca, but different for SO_4-S (Salmon et al., 2001) and for Cl and SO_4-S (Avila et al., 1992). This indicates that solute concentration relationships, especially for Ca and SO_4-S depend on both water residence time and the interaction with chemistry composition of parent material. This relationship might indicate source water from deep underground generally associated with ground water contributions (Salmon et al., 2001). A study in pristine tropical forests draining volcanic landscapes in Costa Rica (Pringle et al., 1990) found higher concentrations of Cl, Na and Mg, but similar concentrations of Ca and SO_4-S to our site. In general, solutes with strong links to parent material chemistry composition exhibit dilution-type relationship to discharge. According to preliminary application of PCA (not reported here), Ca and Mg are good candidates to explain the variability of streamflow data, and thus aid in assessing the contribution of groundwater.

What do event discharge concentrations tell us?

In general, as described earlier, the event data confirmed the presence of the three types of discharge-solutes concentration relationships. The enhancement behavior was clearly stronger with regard to DOC (event data) which had an even greater correlation r^2 ranging from 0.47 to 0.91 compared to 0.68 for long-term data. In our study some element concentrations increased with streamflow, indicating that some soil conditions must be met before concentration start to increase. Most other elements were congruent with long-term data.

This behavior of DOC strongly suggests that enrichment of organic carbon takes place as infiltrating water moves through the humus rich forest floor and rich organic matter topsoil layers. Al in contrast may be released during anaerobic conditions in the saturated areas and when perched water tables are formed on locations on the slopes. Then, as infiltrated water moves downward and laterally to the creeks, as interflow, might be responsible for DOC and Al increases as well (Figure 3.8: Auxiliary Material). According to preliminary PCA analysis (not reported here) DOC is a prime candidate to explain the variability of streamflow discharges during the event across watersheds (Appendix 3A.3), which in turn might explain one source of water in the stream and flowpaths (Interflow).

Constituents that are abundant in the bedrock (Ca, Mg, SiO_4-S and SiO2-Si) were greatly enriched in the stream water during baseflow conditions as compared to the soil solution. They also were heavily diluted as streamflow increases, indicating that their sources are deep percolating water (groundwater) which is not affected during normal stormflow events. The concentration of some of these elements is governed by dissolution of carbonates and oxidation of sulfides (Halloway and Dahlgren, 2001; Buker et al., 2010), greatly increasing the concentration of both cations and anions. Our data suggest that hydrologic pathways through the bedrock contribute considerably to stream solute concentrations during peakflow. According to PCA, Mg is the best candidate to represent groundwater contribution to streamflow, due to its stability in concentrations over time and flow conditions (Figure 3.4b).

A third component to explain the variability of the data would be one representing runoff contribution from stream channels and permanently saturated "wet areas". Since we did not measure any type of tracer that is nonreactive in this system, we cannot designate a solute or tracer to help in determining the contribution from wet areas to overall discharge. Chloride has been suggested in some studies, but chloride concentration in the soil of our site is greater than in the rainfall due to accumulation from previous wet deposition (Table 3.3).

CONCLUSIONS

1. The La Tigra Experimental Catchment exerts both qualitative and quantitative chemical changes in rainwater as it passes through the system. In general, precipitation chemistry is enriched threefold after it comes into contact with surface

soil layers and fivefold after it interacts with parent material and leaves the catchment as baseflow, indicating not only the strong influence of subsurface processes on water chemistry but also the dominant flow paths in all the sites.

2. Concentrations of some solutes differed between shallow and deep flow paths. Runoff during peak flow conditions -- which is generated by infiltration excess runoff from saturated areas and interflow -- resulted in higher concentrations of DOC. With our current data, it is impossible to determine how much discharge comes from saturated areas and how much from interflow. These two water sources are interconnected, making it impossible to accurately and unambiguously account for each separately.

3. Long-term biweekly sampling cannot capture the full gamut of discharge-solutes concentration relationships in this cloudforest ecosystem, especially for elements such as Fe and Al that appear only during peakflow. The principal reason is that peakflow usually occurs overnight in this watershed (Caballero et al., 2011), thus some elements (such those that become activated only during reduced soil conditions) are unlikely to be captured during routine biweekly sampling. Although this information comes from very limited measurements, it provides a first look into how the temporal variability in solute concentrations directly relates to subsurface flow paths in this forested watershed. As shown here, the combination of high frequency water sampling (event basis) with low frequency (biweekly basis) long-term sampling provides additional information that otherwise would not have been captured.

4. This study demonstrates that solute concentrations are affected by amount of water leaving the watershed. Those elements linked to rock mineralogy (Mg, Ca, SO_4-S, Na and SiO_2-Si) were significantly different (<0.0001) in the cloudforest and in the non-cloudforest which had different discharge rates, likely

leading to lower concentrations in the WS1 watershed with relatively greater flow per unit area. On the other hand, those element linked to plant-soil-water interaction in the upper area were not significantly different, meaning that their exports are not dependent on the cloudforest effect and residence time.

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SUPPORTING INFORMATION

Supplementary materials containing additional information of the watersheds and mentioned in the text are available as part of the online paper.

REFERENCES

Agudelo. N. (2010). (Personal communication). Plant species composition in la Tigra National Park.

Asano, Y., T. Uchida, Y. Mimasu, Y. and N. Ohte. (2009). Spatial patterns of stream solutes concentration in a steep mountain with an homogeneous landscape. Water Resources Research, Vol. 45 W10432, doi:10.1029/2008WR007466.

Avila, A., J. P. Ferran Rod and N. B. Colin. (1992). Storm solute behaviour in a montane Mediterranean forested catchment. Journal of Hydrology, 140, 143-161 143 Elsevier Science Publishers B.V., Amsterdam, 1992.

Bilby, R. E. and G. E. Likens. (1979). Effect of hydrologic fluctuations on the transport of fine particulate organic carbon in small streams. Limnol. Oceanogr., 24(1), 69-75.

Bishop, K.H., H. Grip and A. O'Neill (1990). The origins of acid runoff in a hillslope during storm events. Hydrol. 116:35-6.

Bruijnzeel, L. A. (1983). Evaluation of runoff sources in a forested basin in a wet monsoonal environment: a combined hydrological and hydrochemical approach. International Association of hydrological sciences Publication 140:165-174.

Bruijnzeel, L. A. (1991). Nutrient input-output budgets of tropical forest ecosystems: A review. Journal of Tropical Ecology, Vol. 7, No. 1 (Feb., 1991), pp. 1-24.

Bruijnzeel, L. A. (1990). Hydrology of Moist Tropical Forests and Effects of Conversion: A State of Knowledge Review. IHP-UNESCO Humid Tropical Programme, Paris, 224 pp.

Bruijnzeel, L. A. and J. Proctor. (1995). Hydrology and Biogeochemistry of Tropical Montane Cloudforests: What Do We Really Know?', in Hamilton, L. S., Juvik, J. O., and Scatena, F. N. (eds.), *Tropical Montane Cloudforests*, SpringerVerlag, Berlin, pp. 38–78.

Bruijnzeel, L. A. (1996). Predicting the Hydrological Impacts of Tropical Forest Conversion: The Need for Integrated Research, Keynote paper prepared for the ABRACOS Symposium on Amazonian.

Bruijnzeel, L. A. (2001). Hydrology of tropical montane cloudforests: A Reassessment. Land Use and Water Resources research. http://www.geo.vu.nl/users/trendy.

Bruijnzeel, L.A. (2006). Hydrological functions of tropical forests: not seeing the soil for the trees? Agriculture, Ecosystems and Environment 104: 185–228, (2004). Bruijnzeel, L.A.: Hydrological impacts of converting tropical montane cloudforest to pasture, with initial reference to northern Costa Rica. Final Technical Report DFID-FRP Project no. R7991. Vrije Universiteit Amsterdam and Forestry Research Progamme of the U.K. Department for International Development, Aylesford, 52 pp. (www.geo.vu.nl/~fiesta).

Bücker A., P. Crespo, H. G. Frede, K. B. Vache, F. Cisneros and L. Breuer. (2010). Identifying controls on water chemistry of tropical cloudforest catchments – combining descriptive approaches and multivariate analysis. Aquat Geochem 16(1): 127-149.

Caballero L.A., A. Rimmer and T. S. Steenhuis. (2011). Rainfall runoff relationships for a cloudforest watershed in Central America: Implications for water resource engineering. Available at (http://mc.manuscriptcentral.com/jawra).

Carpenter, R.H. (1954) Geology and ore deposits of the Rosario Mining District and the San Juancito mountains, Honduras, Central America. *GSA Bulletin*; January 1954; v. 65; no. 1; p. 23-38; DOI: 10.1130/0016-7606(1954)65 [23:GAODOT]2.0.CO;2.

Cavelier, J., and G. Golstein. (1989). Mist and fog interception in El fin cloudforest in Colombia and Venezuela. Journal of Tropical Ecology. 5:309:-322.

Caveleir J., M. Jaramillo, D. Soils and D. De Leon. (1997). Water balance and nutrient inputs in bulk precipitation in tropical montane cloudforest in Panama. Journal of Hydrology, 193 (1-4), pp. 83-96.

Cirmo, C. P. and J. J. McDonnell. (1997). Linking the hydrologic and biogeochemical controls of nitrogen transport in near-stream zones of temperate-forested catchments: A review, J. Hydrol., 199, 88– 120.

Charlier, J-B., P. Cattan, R. Moussa and M. Voltz. (2008). Hydrological behaviour and modeling of a volcanic tropical cultivated catchment. Hydrological processes, 22(22), 4355–4370.

Durand P., C. Neal, F. Lelong et al. (1991). Hydrochemical variations in Spruce, Beech and Grassland areas, Mont Lozere, Southern France. Journal of Hydrology, Volume: 129 Issue: 1-4, pages 57-70, Doi:10.1016/002-1694(91)90044-I.

Feddema J.J., K. W. Oleson, G. B. Bonan, L. O. Mearns, L. E. Buja, G. A. Meehl and W. M. Washington. (2005). The Importance of Land-Cover Change in Simulating Future Climates. Science 210 1674-1678.

Genereux, D. and C. Pingle. (1997). Chemical mixing model of streamflow generation at La Selva Biological Station. Costa Rica. Journal of hydrology, 199 (1997) 319-330.

Germer, S., C. Neill, T. Vetter, J. Chavez, A. V. Krusche and H. Elsenbeer. (2009). Implications of long-term land-use change for the hydrology and solute budgets of small catchments in Amazonia. Journal of Hydrology. Doi:10.1016/j.jhydrol.2008.11.013.

Halloway, J. M. and R. A. Dahlgren (2001). Seasonal and event-scale variation in solute concentration in four Sierra Nevada catchments. Journal of Hydrology 250 (2001) 106-121.

Hamilton, L.S. (1995). Mountain cloudforest conservation and research: a synopsis. Mountain Research and Development 15, 259–266.

Hanson, D.L., T. S. Steenhuis, M. F. Walter and J. Boll J. (2004). Effects of Soil Degradation and Management Practices on the Surface Water Dynamics in the Talgua River Watershed in Honduras. Land Degradation & Development. 15:367-381.

Hastenrath, S. (2002). The Intertropical Convergence Zone of the Eastern Pacific Revisited. International Journal of Climatology. 22: 347-356.

Hobbie J.E. and G. E. Likens. (1973). Output of phosphorous, dissolved organic carbon and fine particulate carbon from Hubbard Brook Watersheds. Limnology and Oceanography. Volume: 18 Issue 5 Pages 734-742.

Holder, C.D.: (2004). Rainfall interception and fog precipitation in a tropical montane cloudforest of Guatemala. Forest Ecology and Management 190 (2004) 373-384.

Holko, L. and A. Lepisto. (1997). “Modeling the hydrological behavior of a mountain catchment using TOPMODEL”. J. Hydrol., 196 (1/4), 361–377.

Inamdar, S. P., N. O. Leary, M. J. Mitchell and J. T. Riley. (2006). The impact of extreme events on solute exports from a glaciated forested watershed in western New York, USA. Hydrological Processes. *Hydrol. Process*., 20, 3423_3439.

Instituto Geográfico Nacional (IGN). (1990). Mapa Geológico Tegucigalpa, Honduras.

Johnson, M. S. J., C. E. Guimaraes and J. P. Novaes. (2006). DOC and DIC in flow paths of Amazonean catchments with hydrologically contrasting soils. Biogeochemestry 81:45-57.

Likens, G.E., E. S. Edgerton and J. N. Gallowey. (1983). The composition and deposition of organic carbon in precipitation. *Tellus*, 35B:16-24., 1983.

Lovett, G.M. (1984). Rates and mechanisms of cloud water deposition to a subalpine balsam fir forest. Atmospheric Environment 18 2, pp. 361–371.

Mendoza, G. and T. S. Steenhuis. (2002). Determination of hydraulic conductivities of hillsides with a hillside infiltrometer. Soil Science Society of America Journal; Sept/Oct 2002; 66, 5; Research Library.

Mann, P. (2007). Geology and tectonic development of the Caribbean plate boundary in Northern Central America. Special Paper 428, Page. 135. The Geological Society of America Inc.

MacDowell, W. H. and G. E. Likens. (1998). Origin, composition and flux of dissolve organic carbon concentration in the Hubbard Brook Valley. Ecological Monographs 58:177-195.

McDowell, W.H. and C. E. Asbury. (1994). Export of Carbon, nitrogen and major ions from three tropical montane watersheds. Limnology and Oceanography 39:111-125.

MacDowell, W. H. (1998). Internal nutrient fluxes in a Puerto Rican rain forest. Journal of Ecology (1998) 14:521-521-536.

Martínez, M. C. (2007). Potencial de captura de agua de los suelos de Capiro y Zapotillo, Guinope El Paraíso, Honduras. Proyecto de especial de graduación Escuela Agrícola Panamericana, Honduras.

Mulholland, P. J., G. V. Wilson, and P. M. Jardine (1990). Hydrogeochemical response of a forested watershed to storms: Effects of preferential flow along shallow and deep pathways, Water Resour. Res., 26, 3021–3036.

Moldan, B., and J. Cerny. (1994). Small catchment research. In Biogeochemistry of Small Catchments. Edited by B. Moldan and J Cerny. Scope 51. John Wiley and Sons.

Parker, G. G. (1983). Throughfall and steamflow in the forest nutrient cycle. Advances in Ecological Research. Volume: 13 Pages 57-133 Doi10:1016/S0065-2504(08)60108-7.

Perakis, S. S. and L. O. Hedin. (2002). Nitrogen loss from unpolluted South American forests mainly via dissolved organic compounds. Nature 415: 416-419.

Pringle, C. M., F. Triska and G. Browder. (1990). Spatial variation in basic chemistry of streams draining a volcanic landscape on Costa Rica's Caribbean slope * Hydrobiologia 206: 73-85.

Newbold, J.D., B. W. Sweeney, J. K. Jackson and L. A. Kaplan. (1995). Concentration and exports of solutes from six mountain streams in northwestern Costa Rica. J. Am. Benthol. Soc. (14) 1:21-37.

Reynolds, B., M. Hornung and P. A. Stevens. (1987). Solute budgets and denudation rate estimates for a mid-Wales catchment, Catena, 14, 13±23.

Salmon C. D., M. T. Walter, L. O. Hedin. and M. G. Brown. (2001). Hydrological controls on chemical export from an undisturbed old-growth Chilean forest. Journal of Hydrology 253(1-4) 69-80. doi:10.1016/S0022-1694(01)00447-4.

Lewis W. M. and J. F. Saunders. (1989). Concentration and transport of dissolve and suspended substances in Orinoco River. Biogeochemistry. Volume 7 Issue: 3 Pages 203-240 Doi: 10.1007/BF00004218.

Schellekens, J., F. N. Scatena, L. A. Bruijnzeel, A.I. J. M. van Dijk, M. A. Groen and R. J. P. van Hogezand. (2004). Stormflow generation in a small rainforest catchment in the Luquillo Exper imental Forest, Puerto Rico. Hydrol. Process. 18, 505–530. DOI: 10.1002/hyp.1335.

Schellekens, J. (2006). CQ-Flow: A distributed hydrological model for the prediction of impacts of land cover change, with special reference to the Rio Chiquito catchment, northwest Costa Rica. The Fiesta Project Final Technical Report, Universiteit Amsterdam.

Schmid, S. (2004). Water and Ion fluxes to a tropical montane cloudforest ecosystem in Costa Rica. Thesis Geographishes Institut der Universitat Bern.

.

Stadtmüller T. (1987). Cloudforests in the humid tropics. A bibliographic Review. Japan. The United Nations University. 82 p.

Stadtmüller, T. and N. Agudelo (1990). Amounts and variability of cloud moisture input in a tropical cloudforest. Int. Assoc. Hydrol. Sci. Publ. 193, 25–32.

Steenhuis, TS, A. S. Collick, Z. M. Easton, E. S. Leggesse, H. K. Bayabil, E. D. White, S. B. Awulachew, E. Adgo, A. A. Ahmed. (2009). Predicting Discharge and Erosion for the Abay (Blue Nile) with a Simple Model. Hydrological Processes 23: 3728–3737.

Veneklaas, E.J. and R. Van EK. (1990). Rainfall interception in two tropical montane rain forests, Colombia. Hydrological Processes, Vol. 4, 311-326.

Waterloo, M. J. (1994). Water and nutrient dynamics of *Pinus caribaea* plantation forests on degraded grassland soils in southwest Viti Levu, Fiji, Ph.D. dissertation, Vrije Universiteit, Amsterdam, The Netherlands.

Zadroga, F. (1981).The Hydrological Importance of a Mountain Cloudforest Area of Costa Rica. In R. Lal and W. Russell (eds.), Tropical Agricultural Hydrology, pp. 59–73. New York: John Wiley & Sons Ltd.

CHAPTER 4

EVALUATING THE HYDROLOGIC IMPACT OF CLOUD FORESTS USING A SEMI-DISTRIBUTED WATER BALANCE MODEL: LA TIGRA NATIONAL PARK, HONDURAS

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ABSTRACT

Water scarcity poses a major threat to food security and human health in Central America and is increasing due to deforestation, and population pressure. By simulating the major components of the water balance, the impacts of land management practices and climate change on water supply and water quality can be determined even with little measured data that are available in these regions. Four adjacent forested headwater catchments in La Tigra National Park, Honduras, ranging in size from 70 to 635 ha were instrumented and discharge measured over a one year period. A semi-distributed water balance model was developed to characterize the hydrology of the four catchments, one of which had primarily cloud forest cover. The water balance model simulated daily stream discharges well, with Nash Sutcliffe model efficiency (E) values ranging from 0.67 to 0.89. Analysis of calibrated model parameters showed that despite all watersheds having similar

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geologic substrata, the hydrological parameters the cloud forest had less plant-available water in the root zone and greater groundwater recharge, which resulted in watershed discharge on a per area basis four times greater than the other watersheds despite only relatively minor differences in annual rainfall. These results highlight the importance of cloud forests for sustained provision of clean, potable water, and the need to protect the areas from destruction, particularly in the populated areas of Central America.

KEY WORDS: Central America, rainfall-runoff; Thornthwaite-Mather; water balance model; cloud forest, monsoonal climate.

INTRODUCTION

Throughout Latin America water demand has increased in response to population growth, agricultural use, and industrial demand (PHO, 2001), while water supplies in the dry season are shrinking due to deforestation (San Martin, 2001; Barlow and Clarke, 2002; Bonell and Bruijnzeel, 2004; Bruijnzeel, 2004) and becoming more polluted due to inadequate waste treatment and increased use of agrichemicals (PHO, 2000, 2001). As a result, policymakers in Latin America are under increasing pressure to enact natural resource management policies to ensure a clean and adequate water supply. For instance, in Honduras, after a long debate and legislative process, two new water policy laws have been enacted: the Water Framework Law (Ley Marco del Sector Agua y Saneamiento) enacted in 2003 (Gonzalez de Asis et al., 2007; UNDP, 2010) and the General Water Law (Ley General de Aguas) enacted in 2009 (La Gaceta, 2009). In spite of these new policy frameworks it is widely recognized that the lack of science-based knowledge may hinder their effective application.

Developing relevant natural resources policies to effectively manage water resources is a complex process and requires input from experts, policy makers, regulators and stakeholders. Models of the water resource system can provide insight into the impacts of various scenarios such as climate change, landuse conversion, and increased demand. Unfortunately, most available hydrologic models such as the Soil and Water Assessment Tool model, MIKE-SHE, Hydrologic Engineering Center (HEC) models, and others have been developed in either North America or Europe where there are substantial hydrologic and climatic databases against which the models can be calibrated. However, in much of Latin America there are only sparse records available. Moreover, the rainfall-runoff relationships for monsoonal climates prevalent in Latin America greatly differ from the temperate climatic regimes in North America and Europe where most models were developed (Bruijnzeel, 2004; Araujo et al., 2008; Steenhuis et al., 2009). Despite recent advances in our understanding of tropical hydrology, application of models developed for one climatic regime to another remains problematic (Kovacs, 1984; Falkenmark and Chapman, 1993, Musiake, 2003, Peel et al., 2004 and Sivapalan, 2003).

Landscapes and land covers such as cloud forests, which are not found in temperate climates, occupy a key role in providing water to the lower and drier portions of watersheds in Latin America (Buytaert et al., 2005). Very few studies on the hydrologic impact of cloud forests have been carried out (Cavelier, et al., 1997 in Panama; Bruijnzeel, et al., 2006 and Schellekens, 2006 in Costa Rica). These studies have primarily dealt with the distribution of rainfall and canopy interception, and not on the distribution of surface runoff, baseflow, and deep percolation in cloud forests. Buytaert et al. (2006) carried out similar experimental work and compared the water balances of two small catchments in the páramo of Ecuador, one similar to cloud forests and one a disturbed agricultural system. In this study Buytaert et al. (2006) found that the natural vegetation maximized water retention in the soil by minimizing the plant-available water for evaporation. This made more water available for base and interflow in the cloud forest than in the disturbed system.

One model that has been specifically developed for cloud forests is the CQflow model (CQ). The main purpose of this model is to quantify the discharge from the Rio Chiquito catchment, Costa Rica and to evaluate the consequences of several land-use scenarios within the catchment for the Fiesta Project (Schellekens 2006). The CQ model is fully distributed and simulates (among other factors) fog interception. Recently, Buytaert and Beven (2011) applied Topmodel successfully to the páramo in the Ecuadorian Andes and found that saturated overland flow is a dominant hydrologic process. They also found that the exponential decline of conductivity with soil depth (which Topmodel assumes) was a reasonable characterization for subsurface flow. Another potential model that has been used in a wide variety of physiographic locations is the Thornthwaite Mather procedure (Thornthwaite 1948; Thornthwaite and Mather, 1955; Steenhuis and van der Molen, 1986), which was recently adapted for monsoon climates by separating the watershed into infiltration/recharge zones and runoff zones via saturation excess mechanisms. (Peranginangin et al., (2004); Steenhuis et al., 2009; Bayabil et al., 2010; Tesemma et al., 2010).

Our general objective is to increase the hydrologic knowledge base of cloud forest ecosystems in Latin America by presenting a comprehensive analysis of physical based watershed model parameters using data from a recent study that measured rainfall and discharge from a cloud forest and three surrounding watersheds in the La Tigra National Park in Honduras. The results of which are applicable to the improved planning of water resources. Differences in model parameters between cloud forests and the other, non-cloud forest watersheds are used to infer the effect of cloud forests on hydrology.

MATERIALS AND METHODS

Site Description

The La Tigra Experimental Catchment is located (87º 5’ W Long., 14º 10’ N Lat., WGS84 Datum) within La Tigra National Park, 12 miles north east of Tegucigalpa, central Honduras in the headwaters of the Choluteca River Basin, which drains into the Pacific Ocean (Figure 4.1). The experimental site is composed of four neighboring headwater catchments (WS1, WS2, WS3 and WS4, Figure 4.1) together comprising an area of 880 ha. The research area is characterized by steep slopes ranging from 20 to 30%. Stream channel mean slopes range from 14 to 21%. The general characteristics of each watershed are summarized in Table 4.1. Watersheds WS3 has withdrawals that serve as the water supply for the municipality of Valle de Angeles.

The study area has been protected since the 1940’s, first as forest reserve, and more recently as national park. Land cover is predominantly evergreen *Pinus oocarpa* at lower elevations up to 1500 m. Above 1800 m a

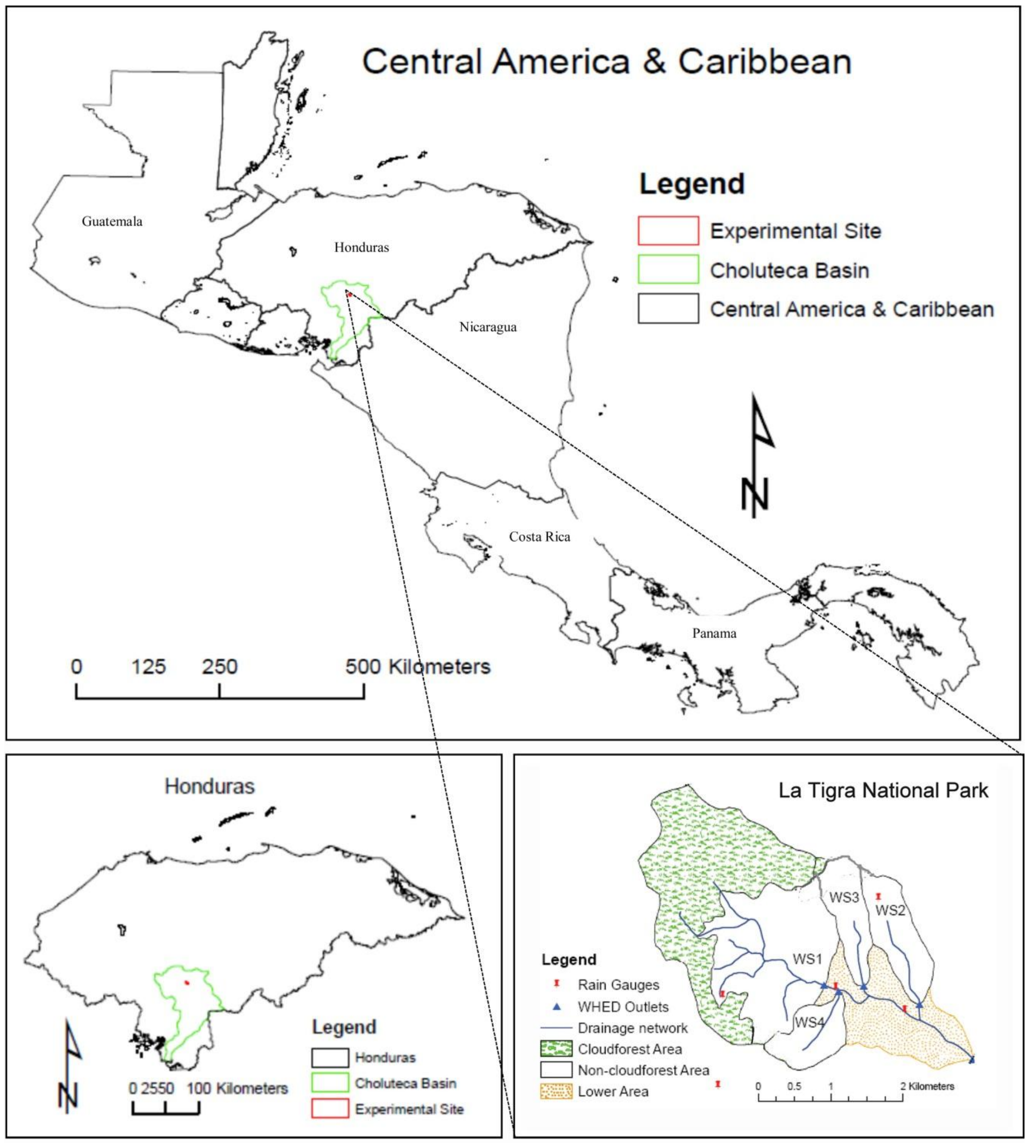


Figure 4.1. Study site, La Tigra Experimental Catchment, Honduras

Table 4.1. Characteristics of the four study catchments and their rivers in La Tigra National Park, Honduras, Central America.

	CATCHMENT			
	WS1	**WS2**	**WS3**	**WS4**
Catchment area (ha)	635	93	82	70
Cloudforest area %	58	0	4	0
Other forested %	41	100	96	84
Deforested %	1	0	0	16
Weir elevation at outlet (m)	1505	1374	1431	1486
Elevation range (m)	1505-2270	1374-1850	1431-2000	1486-1960
Mean elevation (m)	1905	1625	1730	1715
Mean slope (%)	22	20	27	30
Main stream channel length (m)	6600	1508	1105	994
Main stream channel slope (%)	18	14	18	21
Drainage density (km/km^2)	1.00	1.62	1.35	1.42
Mean annual temperature (ºC)	16-20	16-20	16-20	16-20
Mean annual precipitation (mm)	1085	1085	1085	1085
Mean annual discharge (mm)	520	-	-	-
Geology (bedrock formation)	Volcanic	Volcanic	Volcanic	Volcanic
Period of measurements	Apr 2008-Dec 2009	Apr 2008-Dec 2009	Wet season 2008-2009	Wet season 2008-2009
Type of stream	Perennial	Perennial	Intermittent	Intermittent

mixture of *Pinus maximinow* and various broadleaf plants (mainly *Quercus* species) are found. Between 1500 and 1800 there is a transition zone between the two vegetation covers. Cloud forest land cover was 58% for WS1, 0% for WS2, less than 4% for WS3 and 0% for WS4 (Table 4.1). A 2010 land use survey indicates that forest cover is predominant in all sites, except in WS4 where 16% is under agricultural cultivation (Table 4.1). The cloud forest is exposed to fog and lower temperatures, approximately 5-10º C (Bruijnzeel et al., 2006). Unpublished data from the Uyuca Mountains 14 km southeast indicate that temperatures drop approximately 6ºC per 1000 m of elevation gain (Agudelo, 2010: personal communication).

The climate is characteristic of monsoonal regions with very distinct dry and wet phases. The wet phase begins at the end of May or early June when the Intertropical Convergence Zone (ITCZ) becomes active, bringing warm moist clouds from the eastern Pacific to Central America and the Caribbean (Hastenrath, 2002). Annual precipitation averaged over the watershed is 1150 mm, with 90% of the rainfall falling from the end of May through October (Figure 4.A1 Supplementary Material). Lower elevations receive approximately 12% less rainfall than the cloud forest (Caballero et al, 2011)

Soils of the research catchments are Andisols of volcanic origin. A detailed soil map is not available for the study catchments, but based on geologic mapping, soils are underlain by silicate strata of medium coarse fragments of igneous, volcanic and calcareous rocks (IGN, 1956). The lower part of the LaTigra watershed has soils depths ranging from 0.3 to 0.9 meters, organic matter contents from 5 to 15 % (Lavaire and Fiallos, 2010). Soils have abundant pores, roots, and rock fragments of different sizes, all contributing to high permeabilities. Saturated hydraulic conductivities on disturbed soil samples ranged from 1.6 to 7 cm hr^{-1} (Lavaire and Fiallos, 2010) .A soil survey in a similar cloudforest ecosystem 30 kilometer southeast reported mean organic matter contents in excess of 7%, with values reaching 14% in the uppermost cloud forest areas. Bulk densities ranged from 0.4 to 1.0 gr cm^{-3} (Martinez, 2007). Martinez (2007) found soil depths ranging from 0.6 to 1.2 m and sometimes even greater in the upper part of the catchments.

Hydrologic data sets

Stream discharge was measured at each catchment outlet through a controlled concrete weir structure (Figure 4.A1 Supplemental Material). Water height were recorded on 10 minute intervals, and converted to volume using standardized rating curves (ISO, 1980; al. (2011).

Precipitation was measured by a network of four digital rain gauges located along the elevation gradient (1350-1850 m) to have a representative measurement of the average precipitation over the research sites. Potential evaporation was obtained from a nearby (20 km) digital weather station located in the Panamerican School of Agriculture (Zamorano University).

Rainfall-runoff model

We present only the conceptual Semi-Distributed Water Balance Model (SWB model), with the complete derivation from Tessema et al. (2010) in the Supplementary Material. The model is mathematically similar to Topmodel used by Buytaert and Beven (2011) (Walter et al., 2002). However, in Topmodel the entire watershed is underlain by a regional groundwater table that periodically intersects the soils surface generating runoff while in the

SWB model shallow perched water tables over a slowly permeable layer govern the formation of saturated runoff producing areas. In the SWB model the landscape is divided into two regions: well-drained hillslopes, and the flatter near stream areas that become saturated during the wet season due to upslope flow contributions. The hillslopes are further subdivided into two groups: degraded areas that have the hardpan exposed at the soil surface, and highly permeable soils above a restrictive layer at some depth. In the degraded areas (or rock outcrops) that have restricted infiltration, only a small amount of water can be stored before saturation excess surface runoff occurs. In contrast, on the highly permeable portion of the hillslopes, most of the water is transported as rapid subsurface flow (i.e., interflow over a restrictive layer) or as base flow that percolates from the soil profile to deeper subsoil and rock layers (Bayabil et al., 2010; Steenhuis, et al., 2009). The flatter areas that drain the surrounding hillslopes become runoff source areas when part of the profile is at or near saturation. Three separate water balances are calculated for each region. The water balance for the each of the three areas can be written as:

$$S_s(t) = S_s(t-\Delta t) + \left[P - E_a - R - P_{erc}\right]\Delta t \qquad (1)$$

where $S_s(t)$ is volume of plant available water in the soil profile above the restrictive layer (L), at time, t (T), $S_s(t\text{-}\Delta t)$ is the previous time step water storage (L), P is rainfall (L T^{-1}), E_a is actual evapotranspiration (L T^{-1}), R is saturation excess runoff (L T^{-1}), P_{erc} is percolation to the subsoil (L T^{-1}) and Δt is the time step (1 day in our case). Percolation from the infiltration zone occurs when the moisture inputs exceed field capacity. Surface runoff is produced when the soil is saturated, in this case equal to the amount of rainfall minus the water needed to saturate the soil profile. The actual evaporation from the soil, E_a is calculated with the Thornthwaite-Mather procedure (Thornthwaite 1948; Thornthwaite and Mather, 1955; Steenhuis and van der Molen, 1986), which assumes that evaporation decreases linearly with soil moisture content between field capacity (at which point E_a equals potential evaporation) and the wilting point at which E_a is zero. Once precipitation has infiltrated below the root zone there are two reservoirs, one for baseflow and one for interflow. The baseflow reservoir is associated with the groundwater in the near stream area and is simulated as linear reservoir with exponentially decreasing discharge. The interflow reservoir simulates water flowing down the slope over the restrictive layer and is a zero order reservoir (e.g., the discharge decreases linearly with reservoir volume).

Evaluation of model performance

To evaluate how well the predicted runoff matched observed values, we used the Nash-Sutcliffe model efficiency coefficient (E) or goodness-of-fit index (Nash-Sutcliffe, 1970), which is widely used to evaluate the predictive capacity of hydrologic models. In addition, the performance of the model during calibration was evaluated using the normalized root mean squared error (RMSE) and regression coefficient, R^2.

RESULTS

The hydrographs of the four watersheds in the La Tigra National Park were distinctly different (solid black lines in Figure 4.2). In WS1 (cloud forest, Figure 4.2a) the hydrograph is characterized by an initially steep receding limb

(October-December) becoming more and more shallow until the next rainfall season begins, usually in late May. Although smallest WS2 has streamflow throughout the year, the discharge is less than in WS1 and thus the portion of rain converted to streamflow is smaller (Figure 4.2b). In the two watersheds WS3 (Figure 4.2c) and WS4 (Figure 4.2d), there was only discharge in the wet, monsoon phase. In general, WS1 and WS2 have similar runoff responses during the wet season (Figures 4.2a and 4.2b). In WS3 and WS4, discharge drastically declines during the short rainless period between the wet seasons.

In the following sections, we use the SWB model parameters to investigate hydrological processes and how these parameters can explain differences in hydrological behavior among the four watersheds. The SWB model is a mathematical relationship between rainfall and evaporation (as input parameters) and the watershed discharge (as an output). Because the hydrograph is the output signal that integrates all processes that occur in the watershed, it is unlikely parameters resulting in a poor fit represent the physical processes occurring in the watershed. This approach, however, does not give insight into selecting from among potential mechanisms if they all fit the model with equal precision. Hence, when a good fit is obtained between observed and predicted outflow, we can assume that the hydrologic processes in the underlying model structure are valid. For example if the total discharge does not vary as a function of rainfall intensity, infiltration excess is likely not occurring in the watershed.

Model fitting

The semi-distributed SWB model applied here used the precipitation and potential evaporation as climatic input data to estimate the water balance.

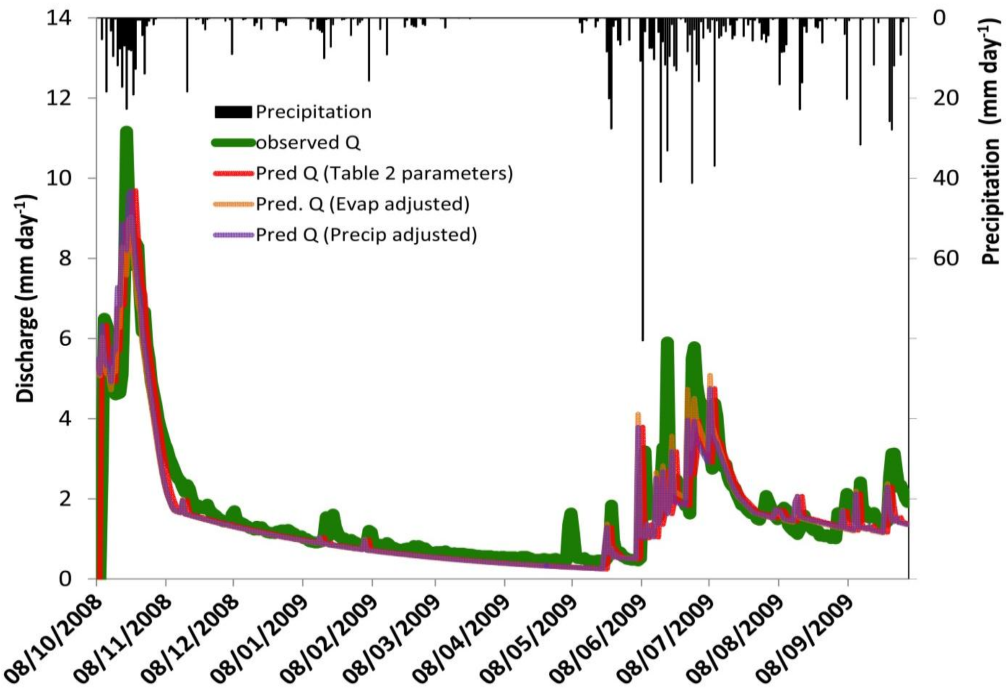


Figure 4.2a. Comparison of observed and predicted daily total streamflow discharges at WS1 catchment for various sets of input parameters listed Tables 4.2 and 4.3. For explanation of the legend see Table 4.3.

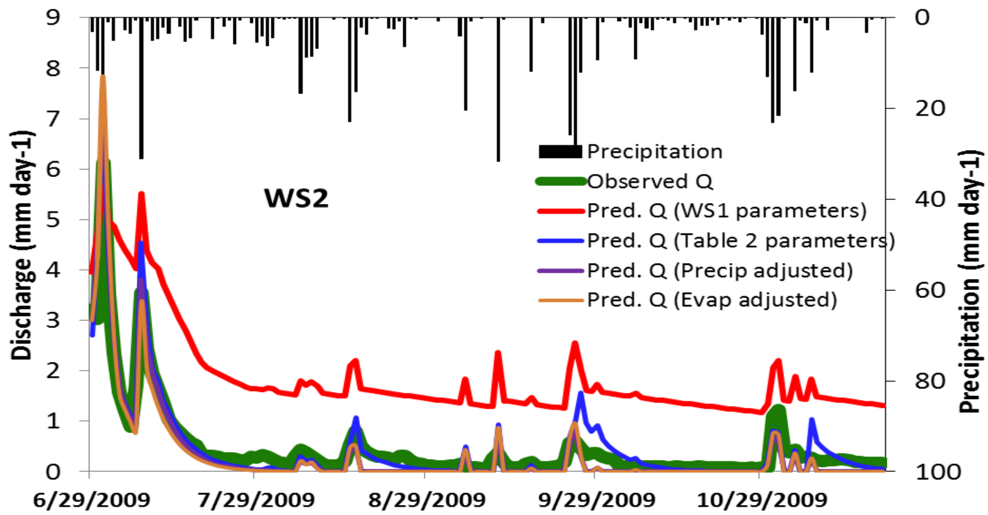


Figure 4.2b. Comparison of observed and predicted daily total stream flow discharges at WS2 catchment for various sets of input parameters listed Tables 4.2 and 4.3. For explanation of the legend see Table 4.3.

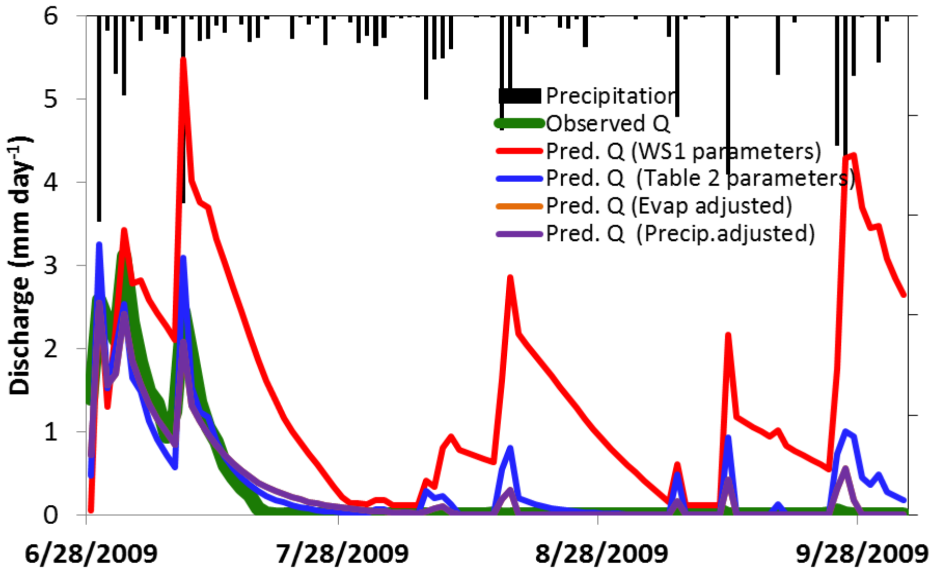


Figure 4.2c. Comparison of observed and predicted daily total streamflow discharges at WS3 catchment for various sets of input parameters listed Tables 4.2 and 4.3. For explanation of the legend see Table 4.3.

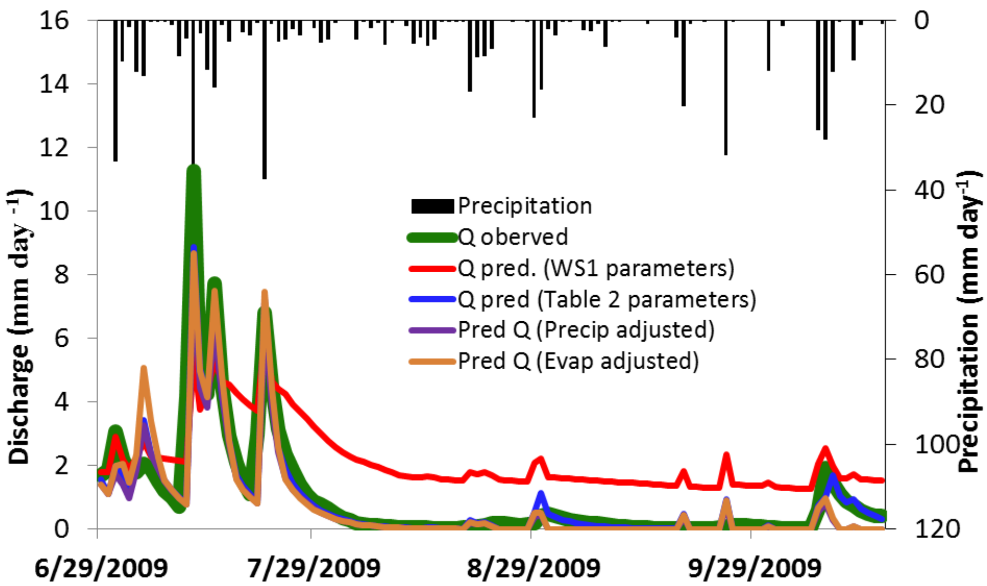


Figure 4.2d. Comparison of observed and predicted daily total streamflow discharges at WS4 catchment for various sets of input parameters listed Tables 4.2 and 4.3. For explanation of the legend see Table 4.3.

For precipitation (P), daily arithmetic averages obtained from either three or four digital rain gauges were used and thus daily precipitation model inputs were the same for all catchments (Figure 4.A3 in Supplementary material). Potential evaporation (PE) data from a nearby (20 km) weather station was used, which ranged from 1.3 to 5.6 mm d^{-1}, with an average of 3.6 mm d^{-1} over the measurement period a value quite similar to the 3.5 mm d^{-1} used in monsoonal climate of Ethiopia (Collick et al., 2009) and in the Caribbean (2.1-3.7 mm d^{-1}, Charlier et al., 2008). Other parameters needed to simulate the discharge included soil water storage capacity for the hillslopes and runoff contributing areas. The saturated areas were designated as those areas at the foot of the hills where saturation excess runoff is generated as well as areas of exposed bedrock (4.A4 see pictures in supplementary material). Saturation excess runoff in these highly conductive soils does not necessarily mean that the water table is at the surface. Only part of the soil profile needs be saturated for interflow to occur (Lyon et al., 2006). The hillslopes act as sources of the recharge to the aquifer and contribute flow to the saturated runoff generating areas via interflow from upslope areas.

We initially calibrated the model for WS1 and then validated the model for the other watersheds, but once the data were analyzed it became clear that although we could close the water balance for WS1, this was not possible for the other watersheds. For this reason, we first fitted the observed and predicted values for WS1 and then changed the fewest number of parameters to fit the other three watersheds using a simplified equifinality approach by varying those parameters within physically-justifiable ranges. We first adjusted the area that contributed runoff in order to fit the observed versus predicted values. Once we had the mass balance correctly fitted, the available water content of the soil was fitted, and subsequently a sensitivity analysis in which we varied the precipitation and potential evaporation was performed (Table 4.3).

Calibration for WS1

The predicted and observed streamflow for the WS1 watershed (cloud forest) for the period October 2008 through October 2009 is shown in Figure 4.2a as the dashed blue line. Average observed streamflow was 1.7 mm d^{-1} and model-predicted discharge was 1.6 mm d^{-1}; this best fit was obtained with only 4% of the watershed area contributing surface runoff, while the remaining 96% the watershed infiltrated precipitation and contributed subsurface flow (Table 4.2). Surface runoff was produced when the rainfall exceeded the calibrated maximum available rootzone water content of 5 mm (e.g. S_{max} = 5 mm, Table 4.2) in the soil, and thus any rain in excess of 5 mm produced runoff. The runoff areas are saturated areas caused by interflow from upslope (Harpold et al., 2010), or exposed bedrock. Thus 96% of the watershed had soils with high infiltration rates in which all rainfall infiltrated. Subsurface flow consists of slow (baseflow) and fast (interflow) components. Baseflow was simulated as a linear reservoir with a half life of 70 days, and interflow was simulated from a zero reservoir that drains, in 25 days (by calibration) after the reservoir fills up (Table 4.2). In the model, the groundwater reservoir fills first, and when the storage exceeds the equivalent of 200 mm over the whole watershed the zero order interflow reservoir fills (Table 4.2). The time that interflow stops is clearly visible in Figure 4.2a where on November 10th 2009, the rapid decline in discharge (i.e., interflow) stops, and the slope of the receding limb becomes much less steep (i.e., baseflow). The Nash-Sutcliffe model efficiency is 0.87, which is quite good for daily discharge predictions (Table 4.3a). The model

Table 4.2. Model input parameter values for surface flow, baseflow and interflow for the four catchments in the La Tigra National Park in Honduras.

Parameter	Watershed			
	WS1	WS2	WS3	WS4
Overland flow area	0.04	0.04	0.04	0.04
Area permeable hill slope (Af)	0.96	0.96	0.96	0.96
t* in days	20	3	3	3
$t_{1/2}$ (half life) in days	70	2.3	2.3	2.3
Maximum depth ground water reservoir (mm)	200	15	15	15
S_{max} overland flow area (mm)	5	5	5	5
S_{max} hill slope zone (mm)	20	200	200	200

Af = fraction of total area

Table 4.3. Comparison of observed versus model predicted daily discharge for the four watersheds WS1, WS2, WS3 and WS4.

Watershed	description	Ratio	Mean mm/day	St dev mm/day	RMSE mm d^{-1}	Nash Sutc	Linear regression		
							interc	slope	R^2
WS1	observed		1.74	1.60					
	Table 4.2 parameters		1.70	1.69	0.62	0.85	0.07	0.96	0.89
	WS2 parameters		0.54	1.59	1.64	-0.05	0.15	0.97	0.89
	Precipitation adjusted	0.90	1.58	1.64	0.54	0.89	0.27	0.93	0.90
	Evaporation adjusted	1.30	1.57	1.57	0.51	0.90	0.20	0.97	0.90
WS2	observed		0.47	0.80					
	WS1 parameters		1.88	0.96	1.48	-2.43	1.38	1.07	0.80
	Table 2 parameters		0.47	0.98	0.30	0.86	-0.08	1.18	0.93
	Precipitation adjusted	0.85	0.35	0.99	0.29	0.87	0.21	1.20	0.95
	Evaporation adjusted	1.40	0.33	0.97	0.31	0.85	-0.22	1.18	0.94
WS3	observed		0.31	0.70					
	WS1 parameters		1.44	0.72	1.45	-4.27	1.17	0.90	0.26
	Table 4.2 parameters		0.38	0.64	0.32	0.78	0.13	0.82	0.80
	Precipitation adjusted	0.60	0.31	0.70	0.21	0.91	0.07	0.78	0.94
	Evaporation adjusted	2.10	0.33	0.82	0.38	0.67	0.02	1.0	0.78
WS4	observed		0.98	1.72					
	WS1 parameters		2.16	1.06	1.57	0.16	1.67	0.51	0.67
	Table 4.2 parameters		0.87	1.51	0.54	0.90	0.05	0.84	0.91
	Precipitation adjusted	0.83	0.71	1.40	0.66	0.85	-0.04	0.77	0.89
	Evaporation adjusted	1.30	0.81	1.64	0.65	0.85	0.05	0.88	0.86

The Mean, Standard deviation (St Dev), root mean square error (RMSE), Nash Sutcliff Efficiency (Nash Sut) and the intercept (interc), slope and R^2 of the linear regression are given for various simulations. For "Table 4.2 parameters" the values listed in table 4.2 for the particle watershed are used. We also used the input values for Watershed WS1 for simulating the discharge of Watersheds WS2, WS3 and WS4 (WS1 parameters) and the WS2 parameters for simulating WS1 (WS2 parameters); Best fits were also obtained by setting the total contributing area to 1 while keeping the other model parameters in Table 4.2 for the particular watershed in Table 4.2 the same by multiplying the precipitation by a constant factor listed under "Ratio" (Precipitation adjusted) or by multiplying the potential evaporation by a constant factor (Evaporation adjusted). The value in "Ratio" was the best fit value.

predicted that an average of 1.4 mm d^{-1} of rainfall exited the watershed as subsurface flow, which was nearly equal to the 1.6 mm d^{-1} obtained from baseflow separation technique (Caballero et al., 2011).

Calibration for WS2, WS3 and WS4

For the three smaller watersheds, which had little or no cloud forest cover, we initially transferred the calibrated parameter set for WS1 (Table 4.2) but the fit was poor (Figure 4.2b, c, d and table 4.3) and tried to fit the hydrographs of each watershed by changing one parameter at a time until a good fit (R^2) was obtained. For every trial, we used the coefficient of determination (R^2) as a measure of good fit. The final model performance was also evaluated using the Nash Sutcliffe efficiency (E*) and the root mean square error (RMSE) (Table 4.3).

Model calibration was performance as follows: We first ensured that the water balances (observed and modeled) in each of the watersheds were as close as possible, by adjusting contributing areas of the hillslopes that provided water at the gage as runoff on the day of the rainfall, and sometime later for interflow and baseflow (Table 4.3). The remaining rainfall that not evaporated becomes interflow or baseflow down from the gage. By running the SWB model with varying contributing hillslope areas the water balance closed when the hillslope areas were fixed at 80% of the total area for WS2, 70% for WS4 and 30% for WS3 (Table 4.2). The fit between daily observed and predicted values was still poor and required adjusting the other model parameters (i.e, maximum depth of ground water reservoir and S_{max}). This was done first for WS2. Since the watersheds did not differ greatly except for the cloud forest cover, we kept the total of ground water and root zone storage for the hillslope area constant at approximately 220 mm (Table 4.2). In order to obtain a better fit we increased the maximum available rootzone water content, S_{max}, to 200 mm and reduced the maximum ground water storage to 15 mm, (Table 4.2) yielding a total storage of 215 mm for WS2 compared with 220 mm for SW1. Finally, to obtain good fits for the recession curves we adjusted decay constant for subsurface flows. Our experience in other monsoonal climates (Collick et al., 2009; Steenhuis et al., 2009) have been that the smaller the watershed, the faster the ground water outflow and this appeared to be true in this case as well. For the three small watersheds we used a half life of 5 days for the linear reservoir and 3 days for the zero reservoir to drain completely after a storm (Table 4.2), which were significantly less than the calibrated parameters for WS1 with a half life of 70 days for the baseflow reservoir and 25 days for the interflow reservoir to drain. The observed (black line) and the predicted curve (dashed red line) are depicted in Figure 4.2b. The Nash Sutcliffe efficiency for daily values was 0.76 and R^2 = 0.90 (Table 4.3)

Using the same parameter set as for WS2 with the adjusted contributing areas determent before (Table 4.2), the hydrographs for WS3 and WS4 were predicted. By comparing the observed and predicted outflow in Figures 4.2c and 2d, it is obvious that a relatively good fit was obtained with Nash Sutcliffe efficiencies of 0.90 for daily values and R^2 values of 0.90 for both watersheds (Table 4.3b). The two overland flow peaks predicted in streamflow at the end August and at the end of September in Figure 4.2c were predicted by the model but not observed. The small dam at the intake structure above the weir would have stored this small overland flow volume. This structure was not observed until the weir had been built and measuring the water intake was not possible due to funding limitations. In watershed WS4 a good fit was obtained when we assumed that 70% of hillslope area was contributing to streamflow (Figure 4.2d), resulting in an efficiency of 0.90 (Table 4.3).

Sensitivity analysis

In all four watersheds, the precipitation was assumed to be evenly distributed across the watershed and that the imbalance in the water balance was caused by deeper regional flows that bypassed the weirs. However, there is a possibility that the differences in the water balances are caused by the differences in rainfall. Therefore we repeated the calibrations by assuming the whole watershed is contributing flow (both runoff and baseflow) to the weir and varied the amount of rainfall by multiplying the average rainfall by a constant. We kept all other parameters the same as shown in Table 4.2. The best fit in Table 4.3 and Figure 4.2 (dotted blue line) was obtained by taking 85% of the average rainfall for WS2 and WS4 and 60% of the rainfall for WS3. Overall the measures of fit in are similar for the rainfall and contributing area adjustments (Table 4.3).

DISCUSSION

The four neighboring headwater catchments (WS1, WS2, WS3, and WS4, Figure 4.1) together comprise an area of 880 ha and are very similar geologically. All four watersheds are in a protected area and have never used for agriculture except for a small section of WS4. Despite that the runoff response varied greatly between WS1 and the three other smaller watersheds. The main difference in the watersheds was that WS1 has a large percentage of cloud forests, while the other watersheds located at lower elevations had little

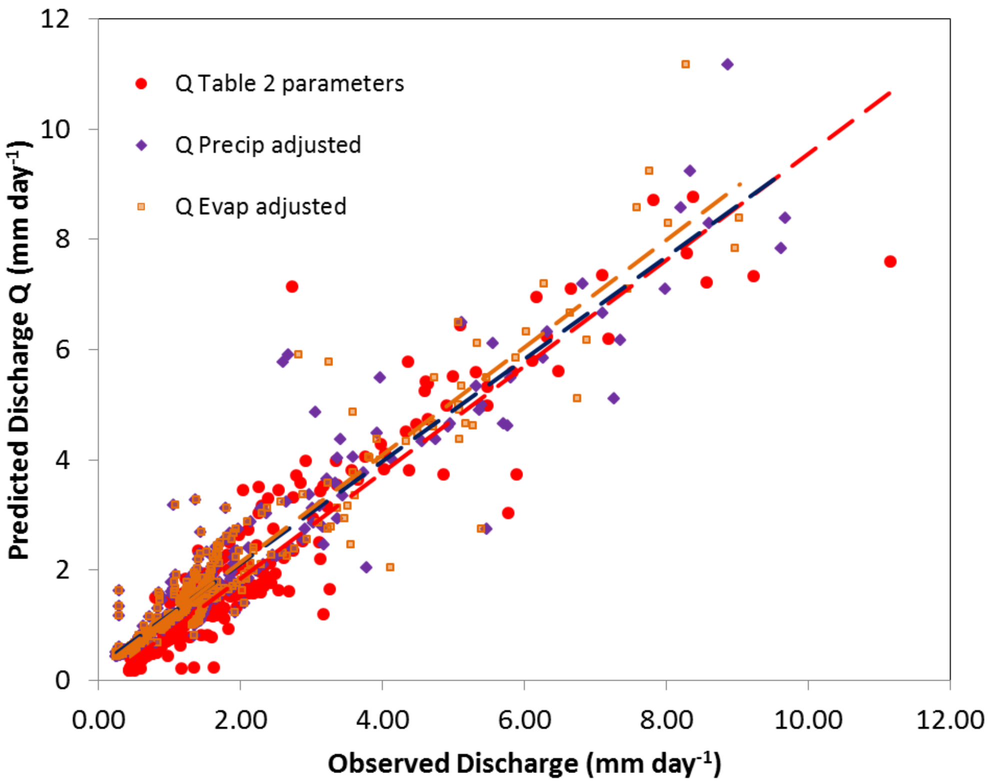


Figure 4.3a. Comparison of observed and predicted daily total streamflow discharges at WS1 catchment for various sets of input parameters listed Tables 4.2 and 4.3. The dashed lines are the linear regression lines and have the same color as the symbols. For explanation of both the legend and the value of the linear regression coefficients see Table 4.3.

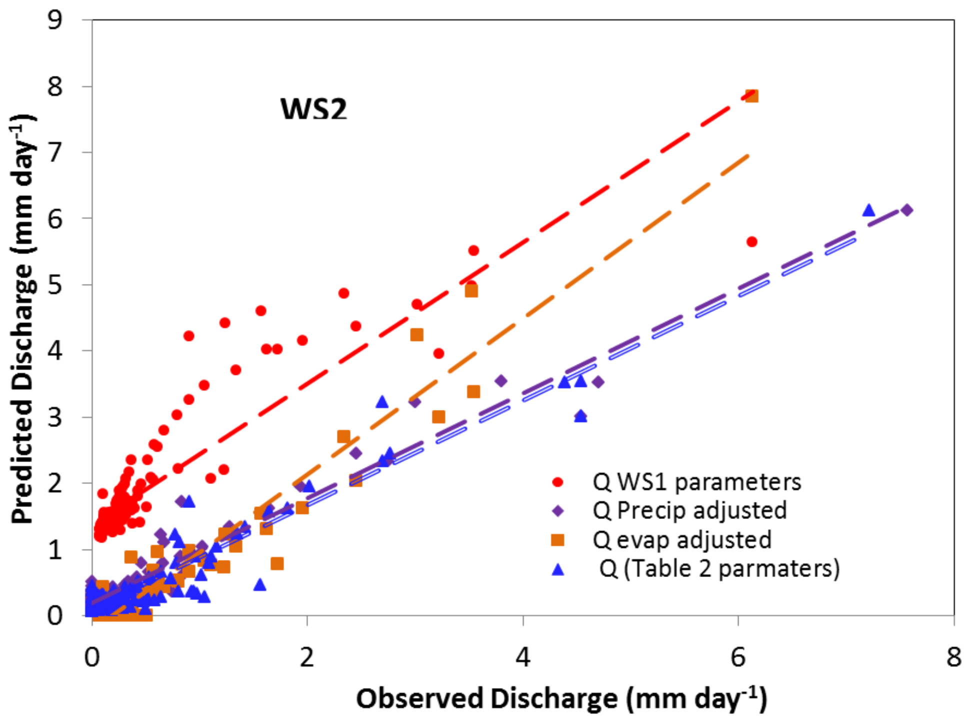


Figure 4.3b. Comparison of observed and predicted daily total streamflow discharges at WS2 catchment for the various sets of input parameters listed Tables 4.2 and 4.3. The dashed lines are the linear regression lines and have the same color as the symbols. For explanation of both the legend and the value of the linear regression coefficients see Table 4.3.

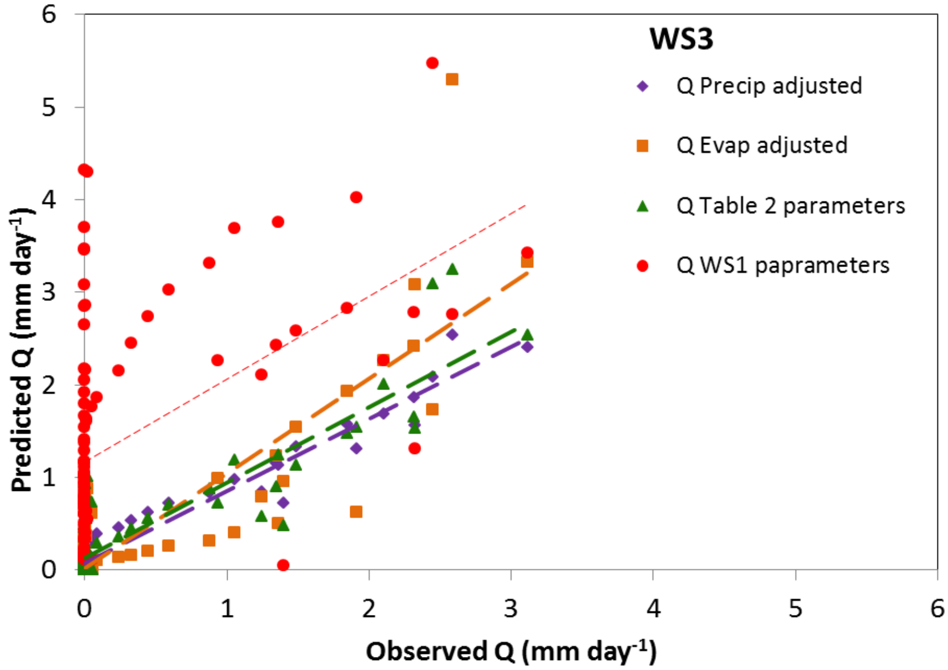


Figure 4.3c. Comparison of observed and predicted daily total streamflow discharges at WS3 catchment for the various sets of input parameters listed Tables 4.2 and 4.3. The dashed lines are the linear regression lines and have the same color as the symbols. For explanation of both the legend and the value of the linear regression coefficients see Table 4.3.

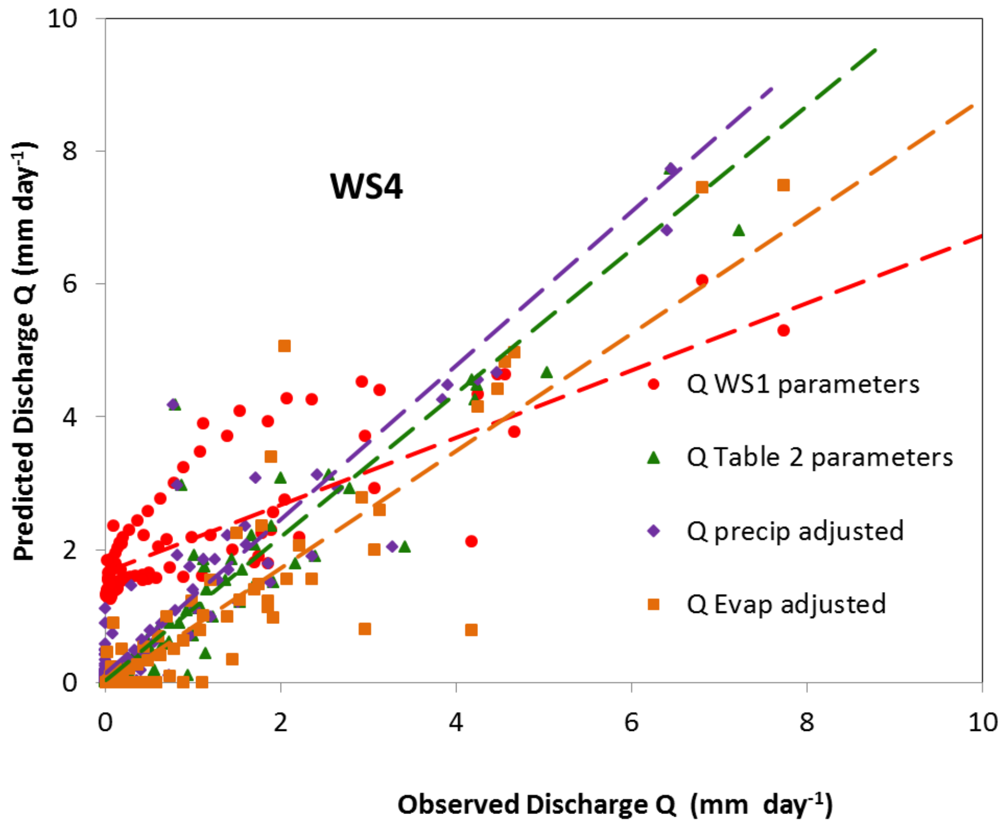


Figure 4.3d. Comparison of observed and predicted daily total streamflow discharges at WS4 catchment for the various sets of input parameters listed Tables 4.2 and 4.3. The dashed lines are the linear regression lines and have the same color as the symbols. For explanation of both the legend and the value of the linear regression coefficients see Table 4.3.

cloud forest cover (Table 4.1) In addition, there was a rainfall gradient with approximately 17% less rainfall at the outlet of the watershed at 1350 m elevation than upper gauge located at 1850 m or half way to the top of the watershed (2270 m).

Rainfall and contributing area

While the water balance closed for the larger (635 ha) WS1 watershed it could not be closed for the other three smaller watersheds. From other studies in cloud forest areas of Costa Rica and the island of Guadalupe, it is known that significant portions of water can bypass stream gauges when the watersheds are small and located in regions with volcanic soils (Schellekens, 2006; Charlier et al., 2008). In our study, there are two possible causes for the failure to close the water balance. First, and similar to the studies above, there is the possibility that not all the watershed area is contributing to the gage (Table 4.3b) and second precipitation amounts used in the model are not representative for the areas, as there was an increase in precipitation with elevation. Note this gradient was not incorporated into the model, as our primary purpose was to develop a model suitable for locations with sparse rainfall data.

Assuming the average measured rainfall was representative for the whole watershed, the water balance closed when in the smaller WS2 watershed 16% of the watershed area was not contributing; in WS3 this was 66% and WS4 26% (Table 4.2). These fractions of unaccounted water fall in the same range of the other cloud forest on volcanic soils in Costa Rica and on Guadalupe (Schellekens, 2006; Charlier et al., 2008).

Assuming that the whole watershed is contributing but the rainfall varies we found that we needed to multiply the precipitation by 0.85 for watersheds WS2 and WS4 and 0.60 for WS3 (Table 4.3). In contrast, for watershed WS1 we could not get a better fit by reducing the rainfall amount. Thus, while keeping the contributing area constant we could obtain the same fit for WS2 and WS4 by decreasing the rainfall by realistic amounts of approximately 15% for those watersheds that had significant forest coverage at lower elevations (Table 4.3). WS3 had the water supply system intercepting water before the weir so the reduction factor of 0.60 appears to be realistic as well.

Because we only have the integrated output signal from each watershed; it is not possible to determine whether reductions in the contributing area or reduction in rainfall can better account for the differences in resulting water balances. It is, in fact, likely that both factors are responsible for the unaccounted water.

Soil hydrologic parameters

The differences in soil hydrologic parameters between primarily-cloud forest WS1 and the other three watersheds were surprisingly high since all four watersheds are located in close proximity to each other. The difference in maximum soil storage of the root zone between the cloud forest watershed (WS1) and the other three watersheds was unexpected. To ensure that we did not have false model optima, we used the same rootzone storages for WS1 as for the other three watersheds (WS2, WS3 and WS4) by changing all input parameters including evaporation and precipitation. The only other good fit was when we decreased the potential evaporation to 60% of the observed value

(Table 4.3d) and increased the root zone storage, S_{max} to 40 mm (Figure 4.2a red dotted line). No other combination of parameter adjustments resulted in a reasonable model fit. This indicates that the maximum rootzone available water, S_{max}, for the cloud forest and the forest at lower elevations are distinctly different.

These results are in accordance with findings in the páramo by Buytaert et al. (2004, 2006) who measured decreased wilting points in the agricultural catchment after two year of cultivation compared to the could forest in a paired watershed study in the Ecuadorian Highlands. Decreases in wilting point translate in a greater amount of plant-available water, and Buytaert et al. (2005) observed even greater amount of plant-available water in laboratory measurements after the system was disturbed. In our case, as stated earlier, the S_{max} value is the maximum amount of water that can be extracted by plants in the root zone. By lowering the wilting point, more water could become available to the plant and hence provide a greater S_{max} value. Buytaert et al. (2004, 2005) found that the retention time for ground water and interflow decreased by an order of magnitude when the paramo was disturbed and subjected to drying; the wilting point also decreased significantly for the disturbed, agricultural paramo. Thus our results for the cloud forest are remarkable the same as for the paramo. Both are permanently wet and it might be the reason that they behave both the some and so much different than other systems that dry out at some time during the year.

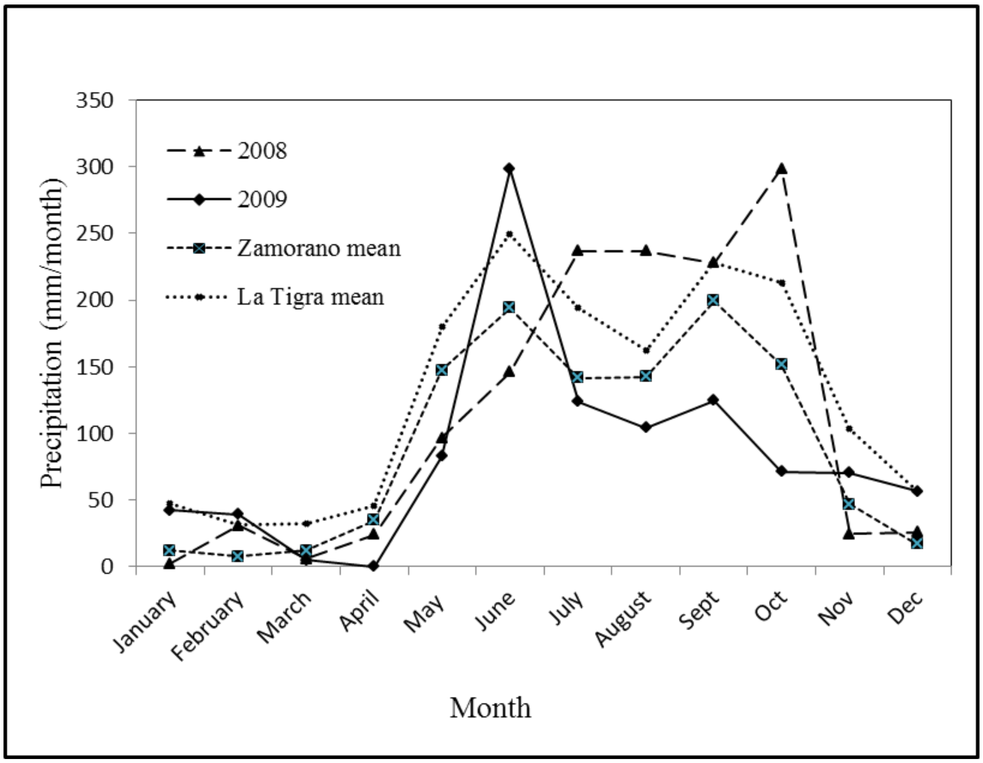
CONCLUSIONS

The aim of the study was to compare the hydrology of a cloud forest with forests at lower elevations using a simple water balance model suitable for these environments. Overall this relatively simple model fitted the observed outflow hydrographs well with relatively high Nash Sutcliff values for daily predicted values. Despite the similar climatic and geologic characteristics of the study catchments, the hydrology, as expressed by the model parameters, varied greatly between the cloud forest watershed and the other three watersheds in close proximity.

The cloud forest watershed had a distinctly smaller amount of plant-available water and greater groundwater storage, resulting in watershed discharges that were four times greater than those of the other watersheds, despite only relatively minor differences in annual rainfall amount. Despite limited data available to date, this modeling approach is a step forward in predicting water balances in cloud forests and forested areas in Central America, thus aiding in managing the ever growing water demand and scarce water supply resources which are threatened by both the loss of forest and pollution.

APPENDIX A

Precipitation, potential evaporation, weir location, sample of saturated areas and rainfall-runoff model description



Source: Zamorano University (1942-2009) and SANAA (1963-2008)

Figure A1. Monthly precipitation at experimental site compared to long-term average Zamorano weather station and La Tigra SANAA.

Figure A2. The weirs in the watershed.

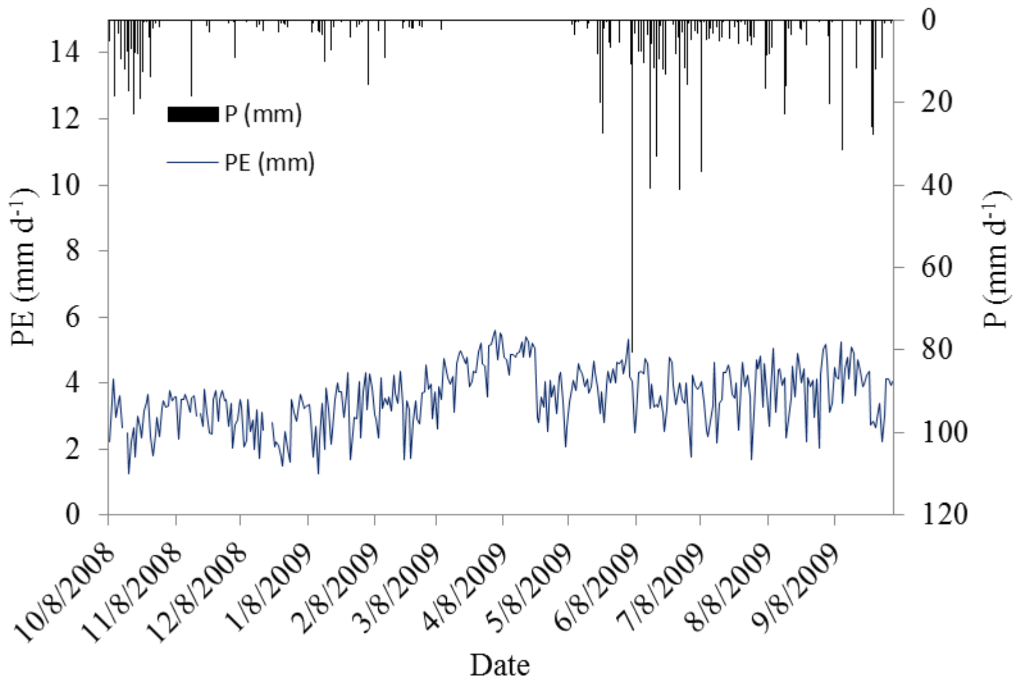


Figure A3. Daily average precipitation (P) and potential evaporation (PE) used in SWB model for the La Tigra National Park

Figure A4. Saturated area in the La Tigra National park. Spring house in the back collects the spring water.

Rainfall runoff model

(Taken nearly verbatim from auxiliary material in Tesemma et al., 2010)

The landscape is divided into two parts, the well-drained hillslopes, and the relatively flatter areas that become easily saturated during the rainfall season. The hillslopes are further divided into two parts that either are degraded or have highly permeable soils above a restricted layer at some depth. The degraded areas have the hardpan exposed at the soil surface. In these areas that have restricted infiltration, a small amount of water can be stored before saturation excess surface runoff occurs. On the highly permeable portion of the hillslopes most of the water is transported through subsurface as rapid subsurface flow (e.g., interflow over a restrictive layer) or base flow (percolated from the soil profile to deeper soil and rock layers). The flatter areas that drain the surrounding hill slopes become runoff source areas when saturated (Fig. A5 shows a schematic representation of a simplified hillslope). Three separate water balances are calculated. The water balance for the each of the three areas can be written as

$$S_s(t) = S_s(t-\Delta t) + \left[P - E_a - R - P_{erc}\right]\Delta t \qquad \text{(A1)}$$

Where P is rainfall (LT^{-1}), E_a the actual evapotranspiration (LT^{-1}), $S_s(t)$ is storage water in the soil profile at time t (L) above the restrictive layer, $S_s(t-\Delta t)$ is previous time step water storage (L), R is saturation excess runoff (LT^{-1}), P_{erc} is percolation to the subsoil (LT^{-1}) and Δt is the time step (10 days in our case). Percolation occurs on the non degraded hillslopes when the soil storage is more than field capacity. Surface runoff on the saturated bottom lands and degraded hill slopes occurs when they are saturated is equal the amount rainfall minus the water that is needed to fill up the soil to saturation.

When precipitation, P, is less than potential evaporation E_p, water is withdrawn from the soil system by soil evaporation and plant transpiration and the actual evapotranspiration is equal to

$$E_a = E_p \left[\frac{S_r(t)}{S_{r\max}} \right], \qquad \text{for } P < E_p \qquad \text{(A2)}$$

Where $S_r(t)$ is the soil moisture at time t for the root zone and $S_{r\ max}$ is the moisture content at field capacity for the permeable hillside and saturated moisture content for the runoff areas.

Since the soil moisture is less than $S_{r\ max}$ both R and P_{erc} are zero and Eq. A1 can be written in exponential form as (Steenhuis et al., 2009):

$$S_r(t) = S_r(t - \Delta t) \exp\left[\frac{(P - E_p)\Delta t}{S_{r\max}} \right], \qquad \text{for } P < E_p \qquad \text{(A3)}$$

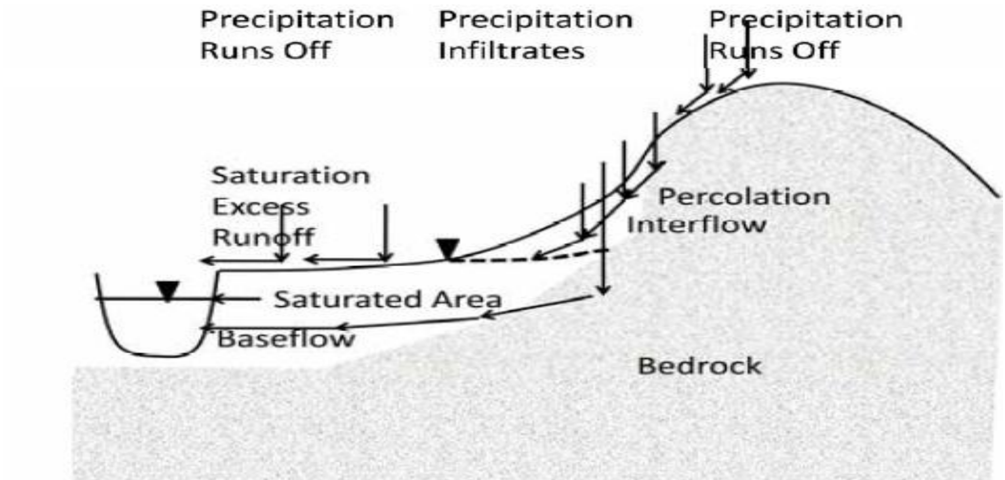


Figure A5. Schematic for saturation excess overland flow, infiltration, interflow and baseflow for a characteristic hill slopes in the Blue Nile Basin (after Steenhuis et al., 2009).

On the hillslopes, areas with high infiltration capacity the excess water (P_{erc}) becomes either interflow (Q_{if}) or baseflow (Q_{bf}) and is added to their respective reservoirs, the interflow reservoir ($S_{if)}$ and base flow reservoir (S_{bf}). Steenhuis et al. (2009) assumed that first the base flow reservoir is filled, and when full (at a storage S_{bfmax}) the interflow reservoir starts filling. The base flow reservoir acts as a linear reservoir and its outflow (Q_{bf}) when the storage is less than the maximum storage can be expressed as:

$$S_{bf}(t) = S_{bf}(t-\Delta t) + [P_{erc} - Q_{bf}(t-\Delta t)]\Delta t \qquad \text{(A4)}$$

$$Q_{bf}(t) = \frac{S_{bf}(t)\left[1-\exp[-\alpha\Delta t]\right]}{\Delta t} \qquad \text{(A5)}$$

where α is the reservoir coefficient (L^{-1}) and is equal to $0.69/t_{½}$. When baseflow storage (S_{bf}) is full, the baseflow can be calculated by setting

$S_{bf}(t)=S_{bfmax}$ in equation (A5). Equation (A4) reduces so that the water entering the reservoir is equal to what flows out calculated with equation (A5). After the base flow reservoir filled, the remaining percolation water fills up the interflow flow reservoir started from the hillslopes by gravity under these circumstances the flow decreases linearly (i.e., a zero order reservoir) after a recharge event. The total interflow at time t can be obtained by superimposing the fluxes for the individual events,

$$Q_{if}(t) = \sum_{\tau=1}^{\tau \le \tau^*} 2P^*_{erc}(t-\tau)\left[\frac{1}{\tau^*} - \frac{\tau}{\tau^{*2}}\right], \tau \le \tau^* \qquad \text{(A6)}$$

where τ^* is the duration of the period after the rainstorm until the interflow ceases, $Q_{if}(t)$ is the interflow at a time t, $P^*_{erc}(t-\tau)$ is the effective percolation on day t-τ. The effective percolation is defined as the total percolation minus the amount needed for refilling the baseflow aquifer. Refer to Steenhuis et al., (2009) for more details on the model development. References are in the main text.

REFERENCES

Agudelo, N. (2010). Temperature gradients in the Uyuca Mountain, Francisco Morazán, Honduras. (unpublished data).

Araujo, M., M. F. Costa, J. T. Aureliano and M. A. Silva, M.A. (2008). Mathematical modeling of hydrodynamics and water quality in a tropical reservoir, Northeast Brazil. Braz. J. Aquat Sci Technol., 12: 19-30.

Barlow, M. and T. Clarke. (2002). Blue Gold: The fight to stop the corporate theft of the world's fresh water. New York: The New Press, ISBN: 1-5684-816-6.

Bayabil, H. K., S. A. Tilahun, A. S. Collick, B. Yitaferu and T. S. Steenhuis. (2010). Are runoff processes ecologically or topographically driven in the (sub) humid Ethiopian highlands? The case of Maybar Watershed. Ecohydrology 3: 457-466.

Bonell, M. and L. A. Bruijnzeel. (2004). Forest, water and people in the humid tropics: past, present and future hydrological research for integrated land and water management. Cambridge University Press.

Bruijnzeel, L.A. (2004). Hydrological functions of tropical forests: not seeing the soil for the trees? Agriculture, Ecosystems and Environment 104: 185–228.

Bruijnzeel L.A. (2006). Hydrological impacts of converting tropical montane cloud forest to pasture, with initial reference to northern Costa Rica. DFID Project Report.

Buytaert, W., G. Wyseure, B. De Bievre and J. Deckers J. (2004). The use of the linear reservoir concept to quantify the impact of changes in land use on the hydrology of catchments in the Andes. Hydrologu and Earth Systems Science 8: 108-114.

Buytaert, W., G. Wyseure, B. De Bievre and J. Deckers (2005). The effect of land-use changes on the hydrological behavior of Histic Andosols in south Ecuador. Hydrological Processes 19: 3985–3997.

Buytaert, W., V. Iñiguez, R. Celleri, B. De Bièvre, G. Wyseure and J. Deckers. (2006). Analysis of the water balance of small páramo catchments in south Ecuador. In: J. Krecek and M. Haigh (eds.), Environmental Role of Wetlands in Headwaters, Springer. pp. 271 – 281.

Buytaert, W. and K. Beven. (2011). Models as multiple working hypotheses: hydrological simulation of tropical alpine wetlands. Hydrological Processes 25: 1784–1799.

Caballero L.A., A. Rimmer and T. S. Steenhuis. (2011). Rainfall runoff relationships for a cloud forest watershed in Central America: Implications for water resource engineering. Submitted 2011.

Cavelier, J., M. Jaramillo, D. Solis and D. DeLeon. (1997). Water balance and nutrient inputs in bulk precipitation in tropical montane cloud forest in Panama. Journal of Hydrology. 193: 83-96.

Charlier J. B., P. Cattan, R. Moussa and M. Voltz. (2008). Hydrological behavior and modeling of a volcanic tropical cultivated catchment. Hydrological Processes. Hydrol. Process. 22, 4355–4370. DOI: 10.1002/hyp.7040.

Collick, A. S., Z. M. Easton, T. Ashagrie, B. Biruk, S. Tilahun, E. Adgo, S. B.. Awulachew, G. Zeleke and T. S. Steenhuis. (2009). A simple semi-distributed water balance model for the Ethiopian highlands. Hydrological Processes 23: 3718–372.

Falkenmark M. and T. Chapman. (1993). Hidróloga comparada: Un enfoque ecológico a los recursos hídricos y de suelo. UNESCOCEDEX: Madrid.

Gonzales de Asis, M., D. O'Leary, P. Ljung and J. Butherworth. (2007). Improving Transparency, Integrity and Accountability in Water Supply and Sanitation. The International Bank for Reconstruction and Development. The World Bank. Washington D.C.

Harpold, A.A., S. W. Lyon, P. A. Troch and T. S. Steenhuis. (2010). The Hydrological effects of lateral preferential flow paths in a glaciated watershed in the Northeastern USA. Vadose Zone Journal 9: 397–414.

Hastenrath, S. (2002). The Intertropical Convergence Zone of the Eastern Pacific Revisited. International Journal of Climatology 22: 347-356.

IGN (Instituto Geográfico Nacional). (1956). Geología de Honduras 1:50,000 San Juan de Flores Quad.

International Organization of Standards (ISO). (1980). Water flow measurement in open channels using weirs and venturi flumes - Part 1: Thin plate weirs. 1438/1-1980(E). Available from Global Engineering Documents at http://global.ihs.com.

Kovacs G. (1984). Proposal to construct a coordinating matrix for comparative hydrology. Hydrological Sciences Journal-Journal des Sciences Hydrologiques 29: 435–443.

La Gaceta. (2009). Ley General de Aguas. Decreto No 181-2009. Congreso Nacional de Honduras. Gaceta No 32088 del lunes 14 de Diciembre.

Lavaire, T. and E. Fiallos. (2010). Potencial de captura de agua de los suelos de la microcuenca El Carrizal, Parque Nacional La Tigra, Francisco Morazán, Honduras. Proyecto de especial de graduación, Escuela Agrícola Panamericana, Honduras.

Lyon S.W., M. R. McHale, M.T. Walter and T. S. Steenhuis. (2006). The Impact of Runoff Generation Mechanisms on the Location of Critical Source Areas. Journal of the American Water Resources Association (JAWRA) 42: 793-804.

Martinez, M. C. (2008). Potencial de captura de agua de los suelos de Capiro y Zapotillo, Guinope, El Paraíso, Honduras. Proyecto de especial de graduación, Escuela Agrícola Panamericana, Honduras.

Musiake K. (2003). Hydrology and water resources in monsoon Asia: a consideration of the necessity of establishing a standing research community of hydrology and water resources in the Asia Pacific region. Hydrological Processes 17: 2701–2709, DOI: 10Ð1002/ hyp.1428.

Nash, J. E. and J. Sutcliffe. (1970). River flow forecasting through conceptual models, Part I. A discussion of principles. Journal of Hydrology 10, 282–290.

Peranginangin, N., R. Sakthivadivel, N. R. Scott, E. Kendy and T. S. Steenhuis. (2004). Water Accounting for conjunctive groundwater/surface water management: Case of the Singkarak-Ombilin River basin, Indonesia Journal of Hydrology: 292: 1-22.

PHO (Pan-American Health Organization and World Health Organization). (2000). Evaluación de los Servicios de Agua Potable y Saneamiento en las Américas. Available at: www.bvsde.ops-oms.org/eswww/eva2000/ Honduras/caliserv/ calidad.htm.

PHO (Pan-American Health Organization). (2001). Informe regional sobre la evaluación 2000 en la región de Las Américas. Agua Potable y Saneamiento, estado actual y perspectivas.

Peel, M.C., T. A. McMahon and B. L. Finlayson. (2004). Continental differences in the variability of annual runoff-update and reassessment. Journal of Hydrology 295: 185–197.

San Martin, O. (2001). Water Resources in Latin America and the Caribbean: Issues and Options. Inter-American Development Bank, Environment Division.

Sivapalan, M. (2003). Process complexity at hillslope scale, process simplicity at the watershed scale: Is there a connection? Hydrological Processes 17: 1037-1041.

Schellenkens, J. (2006). CQ-FLOW: A distributed hydrological model for the prediction of land-cover change, with special reference to the Rio Chiquito catchment, Northwest Costa Rica. The Fiesta Project, Vrije Universiteit, Amsterdam.

Steenhuis, T.S. and W. H. van der Molen. (1986). The Thornthwaite-Mather procedure as a simple engineering method to predict recharge. Journal of Hydrology 84: 221–229.

Steenhuis, T.S., A. S. Collick, Z. M. Easton, E. S. Leggesse, H. K. Bayabil, E. D. White, S. B. Awulachew, E. Adgo and A. A. Ahmed. (2009). Predicting Discharge and Erosion for the Abay (Blue Nile) with a Simple Model. Hydrological Processes 23: 3728–3737.

Tesemma, Z. K., Y. A. Mohamed and T. S. Steenhuis. (2010). Trends in rainfall and runoff in the Blue Nile Basin: 1964-2003. Hydrological Processes 25: 3747-3758.

Thornthwaite, C.W. (1948). An approach toward a rational classification of climate, Geogr. Rev. 38: 55–94.

Thornthwaite, C. W. and J. R. Mather. (1955). The water balance, Publ. Climatol. Lab. Climatol. Dresel Inst. Technol. 8: 1–104.

UNDP. (2010). Adaptation Fund Board Secretariat. Addressing Climate Change Risks on Water Resources in Honduras: Increased Systemic Resilience and Reduced Vulnerability of the Urban Poor. UNDP PIMS 4399; Atlas IDs-Proposal 00060323, Project 00075904, HND10).

USBR (U.S. Department of the Interior, Bureau of Reclamation). (1997). Water Measurement Manual. 3ed. Available at: http://www.usbr.gov/pmts/hydraulics_ lab/pubs/wmm/index.htm.

Walter M.T., T. S. Steenhuis, V. K. Mehta, D. Thongs, M. Zion and E. Schneiderman. (2002). Refined conceptualization of TOPMODEL for shallow subsurface flows. Hydrological Processes 16: 2041–2046.