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Runoff Generation In A Tropical Dry Forest Watershed: Processes, Patterns And Connectivity

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Graduate Program in Geology
A thesis submitted in partial fulfillment of the requirements for the degree in Doctor of Philosophy
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PROCESSES, PATTERNS AND CONNECTIVITY

Thesis format: Integrated Article

by

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Graduate Program in Geology

A thesis submitted in partial fulfilment
of the requirements for the degree of
Doctor of Philosophy

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Abstract

The lack of understanding regarding the controls that govern runoff generation in tropical dry forests represent a critical gap in the hillslope and catchment hydrology literature. Tropical dry forests account for approximately 42% of the global tropical forests, but represent less than 1% of the forest hydrology literature. Three complementary studies were undertaken in a small tropical dry forest watershed, Mexico, to assess the controls that govern the retention and release of a rainfall in the catchment as runoff. In the first study, the high soil surface hydraulic conductivities, absence of a water repellent surface and low rainfall intensities during the wet season allows most of the incoming rainfall to percolate through the near-surface soil layers, suggesting that runoff is generated through a subsurface flow mechanism. In the second study, it was found that two different thresholds were required for streamflow activation and stormflow generation. The long dry period depletes the stores of soil water. Only after the soil storage deficit in the upper metre is satisfied, is streamflow activated from the catchment. Once streamflow became persistent, the stormflow response was almost entirely governed by the rainfall event characteristics and not antecedent soil moisture conditions. The third study used a combination of isotopic, geochemical and hydrometric measurements to describe the water flow pathways, source areas and residence times of stream water in this catchment. It was shown that runoff produced during storm events were composed primarily of old water that likely originated from either deep subsurface soil layers or groundwater and the source areas expanded, likely through sub-basin connectivity, as catchment wetness increased through the wet season. Given the arid climate of the watershed and known hydrological literature regarding runoff generation in tropical forests, it was hypothesised that runoff in this catchment should be delivered from surface or near-surface sources. However, this dissertation has shown that the combination of deep, permeable soil on steep slopes have a stronger influence on runoff generation in this catchment than climate. These three studies have therefore demonstrated the importance of characterising the physical controls that govern runoff generation in forests that are data poor.

Keywords: Hydrology, Tropical forests, Runoff, Infiltration, Soil water repellency, Streamflow, Stable isotopes, Tracers

Co-Authorship Statement

I hereby declare that I am the sole author of this thesis, except where noted below. I understand that my thesis may be made electronically available to the public.

Exceptions to sole authorship:

For all chapters, Dr. Brian Branfireun acted as an advisor, editor and offered suggestions to the presentation and treatment of data in this thesis, and will be listed as a co-author on any subsequent publications originating from this dissertation. Each data chapter of this thesis has been submitted to a peer-reviewed journal, with chapters 2 and 3 already published. Minor typographical or editorial differences may exist between the version that appears in this thesis and the published versions.

Chapter 1

Chapter 1 utilises information and figures from *Farrick, K.K., and Branfireun, B.A., 2013. Left high and dry: a call to action for increased hydrological research in tropical dry forests. Hydrological Processes. 27, 3254-3262*. I was the lead author on this paper and did the majority of the writing and analysis. Dr. Branfireun contributed to the development of research questions and editing.

Chapter 2

Chapter 2 is published as *Farrick, K.K., and Branfireun, B.A., 2014. Infiltration and soil water dynamics in a tropical dry forest: it may be dry but definitely not arid. Hydrological processes. 28, 4377-4387*. I was the lead author on this paper and did the majority of the writing and analysis. Dr. Branfireun contributed to the development of research questions and editing.

Chapter 3

Chapter 3 is in press as *Farrick, K.K., and Branfireun, B.A., 2014. Soil water storage, rainfall and runoff relationships in a tropical dry forest catchment. Water Resources Research. 50, doi: 10.1002/2014WR016045*. The co-author on this paper is Brian A. Branfireun. I was the lead

author on this paper and did the majority of the writing and analysis. Dr. Branfireun contributed to the development of research questions and editing.

Chapter 4

Chapter 4 will be submitted to Journal of Hydrology. The co-author on this paper is Brian A. Branfireun. I was the lead author on this paper and did the majority of the writing and analysis. Dr. Branfireun contributed to the development of research questions and editing.

Acknowledgments

I am eternally grateful to my supervisor Dr. Brian Branfireun for all the great opportunities you have provided me over this five year journey. Your enthusiasm for the environment, science and attention to detail in our work has been a real inspiration over the years and I have truly learnt so much from you. I look forward to continued friendship in the future.

Special thanks to my comprehensive and defense committee members, Dr. Genevieve Ali, Dr. Irena Creed, Dr. Jason Gerhard, Dr. Rob Schincariol and Dr. Elizabeth Webb for their research advice and their invaluable suggestions on the final version of my dissertation.

I would like to thank all the amazing people I met and shaped my thinking at University of Toronto. Michael, Jason, Ben and Varun, those early conversations about our research and life will always stick with me. Ashley, Tom and Michelle, after transferring to Western, you made life in London that much easier. Calling you lab mates isn't enough, so to my friends and colleagues Cynthia, Filippo, Kaylin, James, Meghan, Stephanie, Tara and Yulia, thank you for your help while in the lab and all the great moments outside of it. To my roommate and adopted lab brother Aaron, I enjoyed our intellectual conservations regarding hydrology, fun and sometimes stressful moments as teaching assistants and always reflecting on our Trini pride.

I extend the most sincere gratitude to my colleagues and friends in Mexico. Michel and Gonzalo, thank you for the logistical support and making my stay there much easier. To my field assistants turned amigos, Luis, Alex, Jonny, Carlos and Jimmy, thank you for helping me with my Spanish, jokes and fun times in the field and all your hard work and sacrifice. To the Nava-Garibaldi family, you always made me feel welcome and part of your family, thank you for all the non-academic experiences and understanding what it is to be Mexicanada. I truly want to thank everyone in Mexico. Your heart and welcoming spirit are amazing, and always made the difficult times so much better.

To my parents, Feizal, Joy and Herb, thank you for all your love and support. You truly nurtured my love of science and encouraged me at every turn. To my grandparents for love and support.

To my siblings, Kyle, Kurt and Kamille for the years of stress relief and always asking me when I plan on finishing.

To my partner in life, Julie. You have been by my side through this entire experience. Thank you for your love, encouragement and unwavering support over the years and for putting up with my long summers away in Mexico. It has been quite the journey and I could not have asked for a more remarkable, intelligent and caring person to share it with.

Dedication

This is dedicated to my "mama" Miriam, although you weren't able to see me finish, I know you would be proud.

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Chapter 1

1.0 Introduction

1.1 Tropical dry forest distribution, major threats and hydrology

Tropical dry forests have lived in the shadow of their humid counterparts with respect to scientific research. Despite accounting for more than 42% of all tropical forests (Murphy and Lugo, 1995) and roughly 6% of the Earth's land surface, less than 15% of the literature on all tropical forest research has focused on tropical dry forests with the remainder highlighting work in tropical wet forests (Sánchez-Azofeifa *et al*., 2005; Santos *et al*., 2011). Many of these tropical dry forest regions are currently water-stressed and additional pressures from population growth, land use and future climate change will have significant implications for the future functioning of their natural and socioeconomic systems.

Tropical dry forests lie within the tropical zone, which extends from the equator to 23° in both the Northern and Southern hemispheres. They are broadly characterized as having a vegetation community typically dominated by deciduous to semi-deciduous trees, annual precipitation ranges from 250 – 2000 mm, average annual temperature greater than or equal to 17^oC, and an annual average ratio of potential evapotranspiration (PET) to precipitation (P) greater than 1 (Murphy and Lugo, 1995). The key defining feature of tropical dry forests is the occurrence of a distinct dry period that lasts between 3 and 7 months (Bullock *et al*., 1995). This ecosystem accounts for more than 42% of tropical forests and 19% of the world's total forest area (Murphy and Lugo, 1995; Miles *et al*., 2006). The majority of tropical dry forests occur in Central and South America (66.7%) with the remainder found in Asia (16.4%), Africa (13.1%) and small fragments in Oceania (3.8%). These forests are amongst the most diverse and complex in the world, displaying a high degree of endemism (Olson and Dinerstein, 1998; Suazo-Ortuño *et al*., 2008). This is especially true in the Mexico, where approximately 60% of the species found in that country are exclusive to these forests (Trejo and Dirzo, 2000).

Tropical dry forests are recognised as one of the world's most threatened terrestrial ecosystems with more than 97% of the existing area at risk from threats such as land use change, fragmentation, and climate change (Miles *et al*., 2006). The favourable climatic conditions of these regions have encouraged the rapid expansion of human settlements, with more than 42% of dry forests found alongside population densities greater than 250 people/km^2 (Miles *et al*., 2006). As a result of the high population density, more than 48.5% of tropical dry forests (Hoekstra *et al*., 2005) have been removed and converted to either agriculture or urban land uses. These changes have occurred in all of the tropical dry forest regions but are particularly severe in Asia and parts of Africa, resulting in the increased need to prioritise them for conservation (Miles *et al*., 2006).

Published research on the hydrology of tropical dry forests is rare, accounting for less than 1% of catchment hydrology literature. Despite this, limited information does exist regarding the water balance and runoff mechanisms observation in tropical dry forests. In many tropical dry forests the catchment scale water balance has typically been quantified. Evapotranspiration is the main source of water loss from tropical dry forests, on average accounting for 73-86% of the annual water loss (e.g. Sandström, 1996; Vose and Maass, 1999; Montenegro and Ragab, 2010). Discharge is often very low, representing less than 17% of the annual losses from the catchment (e.g. de Araújo and González Piedra, 2009; Montenegro and Ragab, 2010). Most of the studies in tropical dry forests which have examined runoff processes have generally focused on identifying the specific mechanism by which runoff is generated using either geochemical tracers (Sandström, 1996) or hydrometric analyses (McCarntey *et al*., 1998; Masiyandima *et al*., 2003; Mugabe *et al*., 2007) but not a combination of both.

Runoff is the process by which water flows over and within the soil substrate. There are three major mechanisms of runoff generation: Hortonian or infiltration excess overland flow (HOF), Saturated overland flow (SOF) and Subsurface stormflow (SSF). HOF typically occurs when rainfall intensity exceeds the infiltration capacity of the soil, resulting in flow over the surface. The generation of SOF is more complicated than HOF. The major requirement is that the soil must be saturated, either from above through precipitation or from below through a rising water table for flow to occur. Under saturated conditions additional precipitation reaching the soil

surface cannot infiltrate and flows as surface runoff. SSF can be generated under both saturated and unsaturated conditions. Under saturated conditions flow can occur from perched water tables or local groundwater mounds that lay close to the ground surface. Under unsaturated conditions, precipitation infiltrates the surface layer and moves laterally through the soil profile as bypass flow in macropores and pipes or as flow through the soil matrix.

In tropical dry forests, HOF and SOF are the main runoff generating mechanisms (Figure 1.1). These dominant mechanisms are quite different from humid tropical forests where runoff is often dominated by SOF, SSF and vertical runoff pathways. Despite the identification of the specific runoff generating mechanisms in tropical dry forests, much of these analyses remain preliminary and limited to less than six studies (Farrick and Branfireun, 2013). Therefore, there is a need for increased attention to processes that govern runoff generation particularly: 1. the surface controls on infiltration, 2. the specific thresholds of antecedent storage and rainfall, 3. isotopic and geochemical characterisation of water sources and connectivity across catchments.
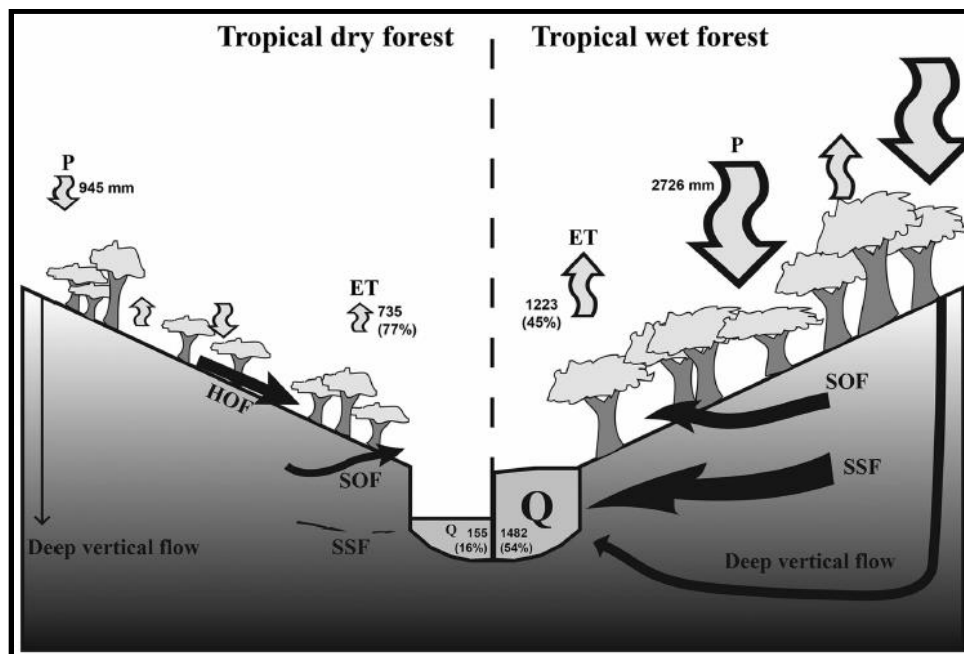
Figure 1.1 Comparison of the average annual water inputs and outputs and the dominant mechanisms of tropical dry and wet forests (from Farrick and Branfireun, 2013). Values in parentheses represent percentage of rainfall. Runoff arrow thickness represents the relative contribution to stormflow.

1.2 Knowledge gaps

1.2.1 Surface controls on infiltration

It is widely accepted that soil infiltration rates play a significant role in dictating the dominant hillslope runoff mechanism (Wilcox *et al*., 1997). Studies have shown that infiltration through the surface soil is mainly controlled by the interaction between the rainfall characteristics during a storm event and the physical properties at the soil surface, namely hydraulic conductivity (*K*) (Bonell and Williams, 1996; Martínez-Mena *et al*., 1998; Puigdefabregas *et al*., 1998). Of increasing importance is the recognition that the development of severe soil water repellency at the soil surface can result in a two to three time decrease in surface infiltration rates (Imeson *et al*., 1992; Martínez-Murillo and Ruiz-Sinoga, 2007). These relationships are particularly important in arid and semi-arid systems, where HOF is the primary form of runoff generation. While work in tropical dry forests catchments indicate that HOF does occur, these studies generally fail to examine the relationship between rainfall intensity and hydraulic conductivity and soil water repellency over a large spatial scale.

1.2.2 Threshold controls on runoff generation

The increased evidence supporting the non-linear, rainfall-runoff response has resulted in a shift in the focus of hillslope and catchment hydrologists to a greater emphasis on quantifying the hydrological thresholds and explaining their physical controls (Zehe and Sivapalan, 2009; Spence, 2010; Ali *et al*., 2013). It is clear that, as has been demonstrated in humid temperate, wet tropical and semi-arid catchments, that the exceedance of antecedent water storage or rainfall thresholds is often required for runoff generation (Spence and Woo, 2003; Cammeraat, 2004; Tromp-van Meerveld and McDonnell, 2006; Negishi *et al*, 2007; Oswald *et al*., 2011). While specific depths of rainfall are needed to generate substantial volumes of runoff, many studies suggest that the rainfall threshold is only exceeded after storage deficits are satisfied (Buttle *et al*., 2004, Fu *et al*., 2013). These observations have sparked debate in the hydrological community regarding the relative importance of storage versus precipitation thresholds. Recent studies have focused on combining both a depth equivalent of soil water storage and event rainfall to assess catchment scale runoff, which reflect both a storage and source area threshold

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(Detty and McGuire, 2010). To the best of my knowledge, no previous studies have examined the importance of storage and rainfall thresholds on streamflow generation in tropical dry forests. This information, will undoubtedly improve our ability to examine the impact of climatic and land use changes on runoff as well as provide an important metric for inter-catchment comparison (Ali *et al*., 2013).

1.2.3 Water flow pathways, source area contributions and residence times

While it is widely recognised that understanding the specific water flow pathways, source areas and residence times of stream water is essential for the management of surface and groundwater resources, these studies have mainly gone undescribed in tropical dry forests (Buttle and McDonnell 2004; Bonell and Bruijnzeel, 2005). Most of what is known regarding runoff pathways in tropical catchments stems from research in the wet tropics. Using either geochemical or isotopic tracers, runoff in wet tropical forests has been shown to be composed primarily of event water generated as SOF, return flow (RF) or shallow subsurface flow (Schellekens *et al*., 2004; Goller *et al*., 2005). While studies have focused on runoff processes in tropical forests characterised by low surface *K* or shallow impeding soil layers, research in tropical catchments with more permeable soils is severely lacking. Hydrological connectivity is regarded as one of the key controls in determining hillslope and catchment rainfall-runoff response and has been defined as the ability to transfer water from one part of a landscape to another (Bracken and Croke, 2007). The Hewlett and Hibbert (1967) Variable Source Area Concept (*VSA*) has shaped the hydrology community's concept of hillslope and catchment scale runoff response for over 40 years. The *VSA* considers that runoff from a catchment is a function of the upslope expansion of saturated subsurface areas, connecting the riparian zone to the hillslope. Recent work suggests that connectivity also occurs among distinct hydrological units across a hillslope (e.g. Tromp-van Meerveld and McDonnell 2006) or zero-order basins (e.g. Sidle *et al*., 2000) rather than evolving from near stream zones. Stream water residence time provides an excellent indication of the linkages among flow paths, water sources and storage in a catchment (McGuire and McDonnell, 2006). Although source area contributions and stream

water residence times have been well studied in humid temperate forests, to my knowledge these hydrological processes have not been examined in tropical dry forests.

1.3 General objectives

The research presented in this dissertation was carried out to improve our understanding of the controls that govern the translation of rainfall to runoff in tropical dry forests by addressing the research gaps presented in the previous section. Not only will the results help inform future predictions of runoff generation under changing climate and land use change but also add to our conceptual understanding of catchment hydrology. The general research objectives were as follow:

1. To test the hypothesis that the current knowledge regarding the relationship between soil surface hydraulic conductivity, soil water repellency and rainfall intensity in semi-arid systems is transferrable to tropical dry forests and to examine the relationship between rainfall intensity, soil surface hydraulic conductivity and soil water repellency.
2. To examine the soil water storage and hydrometeorological controls on streamflow activation and stormflow generation in a tropical dry forest catchment.
3. To identify the dominant flow pathways, water sources and residence times of streamflow. To investigate the dominant runoff flow pathways, source water contributions and residence times of streamflow in a tropical dry forest catchment.

1.4 Thesis organisation

This thesis has been prepared in the integrated article format and consists of three manuscripts related to the three main research objectives. The introduction (Chapter 1) provides an overall introduction to the thesis as whole. It provides background information regarding the distribution and threats to tropical dry forests, identifies the knowledge gaps that the research addresses and outlines the general objectives of the dissertation. The first manuscript (Chapter 2) investigates the controls on surface water infiltration over space and time. The second manuscript (Chapter 3) examined the soil water storage and hydrometeorological controls on annual streamflow

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activation and event scale stormflow generation. The third manuscript (Chapter 4) applies a combined isotopic, geochemical and hydrometric analysis to identify the primary water flow pathways, source area contributions to runoff and estimate the mean residence time of baseflow. The last chapter (Chapter 5) provides an overall summary and general conclusion of the work, and identifies future research directions.

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Chapter 2

2.0 Infiltration and soil water dynamics in a tropical dry forest: it may be dry but definitely not arid

2.1 Introduction

Tropical dry forests account for approximately 42% of the world's tropical forests (Murphy and Lugo, 1986; Miles *et al*., 2006). In Central America, the majority of these forests occur along the Pacific coast of Mexico and are considered to be the most biologically diverse tropical dry forests in the world (Olson and Dinerstein, 1998). Over the last 30 years, these forests have experienced a 12–16% decrease in area because of deforestation from urban encroachment, conversion to agriculture and fire (Miles *et al*., 2006). In Mexican tropical dry forest watersheds, the Intergovernmental Panel on Climate Change has forecasted increases in temperature and significant decreases in precipitation, which is expected to reduce the already limited runoff volumes generated (Bates *et al*., 2008), placing additional stress on groundwater resources in a country where water scarcity is its most important environmental challenge (Muñoz-Piña *et al*., 2008). Despite the importance of hydrology and water availability in this region, the literature on the hydrological processes in tropical dry forests is limited to a small number of short term investigations, in sharp contrast with the large body of literature on the hydrology of temperate and tropical humid forest watersheds (Farrick and Branfireun, 2013).

Tropical dry forests are characterized as having waxyleaved, drought-resistant vegetation, annual rainfall from 250–2000mm and an average annual temperature ≥ 17 °C (Murphy and Lugo, 1995; García-Oliva *et al*., 2003; Miles *et al*., 2006). The key defining feature of tropical dry forests is the occurrence of a distinct dry period that lasts between 3 and 7 months (Bullock *et al*., 1995). These climate and vegetation features are most similar to those of semi-arid regions; a 5–10 month dry period, annual rainfall from 235 to 805mm (Table 2.1), mean temperatures ≥ 16 °C and waxy, lipid-rich vegetation (Martínez-Mena *et al*., 1998; Descroix *et al*., 2001; Verheijen and Cammeraat, 2007; Zavala *et al*., 2009). As with semi-arid regions, the wide range of rainfall

and climate conditions may result in a variety of surface hydrological processes (Cerdà, 1998a). Despite the climatic range, the similarities between these two ecosystems suggest that the factors that govern hydrological processes in semi-arid areas such as vegetation type and annual precipitation regime may be transferrable to tropical dry forest catchments.

The rainfall–runoff relationship in semi-arid regions has been well studied, and it is widely accepted that runoff in semi-arid areas occurs primarily as infiltration excess (Hortonian) overland flow (HOF), where the rainfall rate exceeds the infiltration rate at the soil surface (Table 2.1). The generation of HOF is mainly controlled by the interaction between the rainfall characteristics during the storm event (Bonell and Williams 1986; Wilcox *et al*., 1997), slope, antecedent soil moisture (Ziadat and Taimeh, 2013) and the physical properties at the soil surface, of which hydraulic conductivity (Martínez-Mena *et al*., 1998; Puigdefabregas *et al*., 1998) and soil water repellency (Imeson *et al*., 1992; Cerdà and Doerr, 2007) are recognized as two of the most important controls.

The majority of precipitation events in semi-arid regions during the wet season are low-intensity and short-duration rainfalls (Martínez-Mena *et al*., 1998; Chamizo *et al*., 2012). Although low-intensity events dominate by frequency, high-intensity storm events are considered more important in runoff generation processes in semi-arid regions as more than 70% of the total annual rainfall can be delivered by infrequent, high-intensity, short-duration events (Wilcox *et al*., 1997; Martínez-Mena *et al*., 1998). Rainfall intensities as high as 120 mm/h have been recorded in semi-arid areas (e.g. Bonell and Williams, 1986; Wilcox *et al*., 1997), and in many instances, these rainfall intensities are one to two orders of magnitude greater than the infiltration rate at the soil surface (Table 2.1). Because they exceed infiltration, high rainfall intensities promote HOF and are therefore an important control in runoff generation in semi-arid regions.

In semi-arid systems, it is also well documented that low surface hydraulic conductivity (K) limits infiltration (Table 2.1). Many studies agree that the presence of soil surface crusts, stony ground cover, rocky outcrops and the lack of continuous vegetation produce low permeability surfaces that reduce infiltration (Wilcox *et al*., 1988; Puigdefabregas *et al*., 1998; Yair and Kossovsky, 2002). Surface crusts are formed mainly by the breakdown of soil aggregates by

raindrop impact or lateral surface flow. The aggregates then disperse across the surface and clog pore spaces, creating low K surfaces (Neave and Rayburg, 2007). Hydraulic conductivities have been shown to be two to five times lower for crusted versus non-crusted surfaces (Puigdefabregas *et al*., 1998; Eldridge *et al*., 2000; Descroix *et al*., 2001). The effect of stones and rock fragments on infiltration often depends on the size, orientation and position of the stones/rocks within the soil matrix (Poesen *et al*., 1994). In many cases, the presence of a discontinuous cover of loose, small stones at the soil surface increases surface roughness, which slows surface runoff and enhances infiltration along the contact between the stone and soil matrix (Poesen and Lavee, 1994; Poesen *et al*., 1994; Cerdà, 2001). Other work has shown that a 10–20% increase in the portion of embedded stones can reduce infiltration rates by an order of magnitude (Wilcox *et al*., 1988; Mayor *et al*., 2009). The low density of vegetation cover in many semi-arid regions generally promotes low levels of infiltration (Cerdà, 1996). Small areas of high infiltration two to three times higher than bare and crusted surfaces have been recorded under dense patches of vegetation (Nicolau *et al*., 1996; Cerdà, 1997b; Mayor *et al*., 2009). Higher plant density often reduces surface crust formation through greater soil stabilization (Lavee *et al*., 1998) and enhances infiltration through the formation of macropores and preferential flow pathways along the roots (Cerdà, 1997b; Calvo-Cases *et al*., 2003). Stone, crust and vegetation cover typically exhibit high spatial variability in semi-arid areas because of slope position and aspect (Cerdà, 1998b; Cerdà, 1999). These surface heterogeneities are responsible for the high spatial variability in infiltration and HOF observed in semi-arid regions (Cerdà, 1996, 1997b). It is important to note that during wetter months, higher soil moisture reduces infiltration rates over vegetated areas, producing more homogenous infiltration rates across the landscape (Cerdà, 1996, 1997a). Despite having high spatial variations in infiltration rates, there is still the general consensus that HOF dominates semi-arid landscapes as low K coupled with the high rainfall intensities promote HOF generation (Wilcox et al., 1997; Martínez-Mena et al., 1998).

Soil water repellency or soil hydrophobicity is the resistance of soils to surface wetting and is due to organic hydrophobic compounds being present in the soil matrix (Doerr *et al*., 2000). These organic, hydrophobic compounds are introduced into the soil by plant roots, fungal activity and the waxes and/or lipids derived from decomposing plant litter (Doerr *et al*., 2000).

Although these hydrophobic compounds are always present in the soil, repellent conditions are activated under low soil moisture conditions, making soil moisture a particularly important control on the establishment of repellency. Low soil moisture conditions lead to the concentration of hydrophobic compounds and establishment of repellent surfaces that persist as long as such moisture conditions are maintained (Leighton-Boyce *et al.*, 2005; Verheijen and Cammeraat, 2007). Semi-arid regions are therefore particularly prone to high levels of soil water repellency due to generally low soil moisture and the presence of waxy-leaved, drought-resistant vegetation (Cerdà and Doerr, 2007; Verheijen and Cammeraat, 2007). Furthermore, frequent fires in these regions further enhance repellency as hydrophobic compounds may become more concentrated in the surface soils after burning (Doerr *et al.*, 2000).

In semi-arid regions, soil water repellency has been shown to affect infiltration and, as a consequence, runoff generation. Under high levels of repellency, Imeson *et al.* (1992) recorded a reduction in infiltration from 58.2 to 28.7 mm/h, whereas Martínez-Murillo and Ruiz-Sinoga (2007) recorded a reduction from 36 to 28 mm/h. The reduced infiltration can increase HOF production by an order of magnitude on repellent versus non-repellent soils (Martínez-Murillo and Ruiz-Sinoga, 2007). Similar to semi-arid areas, the hydrology of tropical dry forest catchments may be controlled by soil water repellency due to the presence of a waxy and lipid-rich litter layer (García-Oliva *et al.*, 2003) and prolonged dry periods, which reduces soil water content to the point of activating soil water repellency.

Rainfall intensity, surface soil hydraulic conductivity and soil water repellency are important controls that govern infiltration rates and magnitudes and ultimately runoff generation in semi-arid environments. We have no such insights into the hydrology of tropical dry forest catchments. However, similarities in climate and abundance of waxy-leaved, drought-resistant vegetation suggest that the factors that govern infiltration and the runoff generation mechanisms in semi-arid areas are transferrable to Mexican tropical dry forest catchments. To test this hypothesis, we will:

1. characterize soil surface hydraulic conductivity and soil water repellency over space and time in two dominant tropical dry forest types (mixed deciduous and pine-oak);

2. characterize the frequency and nature of the shallow soil water response to rainfall input in two dominant tropical dry forest types (mixed deciduous and pine-oak); and

3. consider rainfall event characteristics and timing in light of the measured hydraulic conductivity, soilwater repellency and soil water response to evaluate if HOF is a dominant runoff mechanism on tropical dry forest hillslopes.

Table 2.1 The annual precipitation, maximum rainfall intensity, hydraulic conductivity and primary runoff mechanism of the examined semi-arid watersheds.

Source	Site	Annual precipitation (mm)	Maximum rainfall intensity (mm/h)	Hydraulic conductivity (mm/h)	Primary runoff mechanism
Osborn and Renard (1970)	Walnut Gulch, USA	292	>12.7	30.8	HOF
Bonell and Williams (1986)	Torrens creek, Australia	552	120	25.8	HOF
Hussein (1996)	Mosul City, Iraq	333	60	46 – 37	HOF
Nicolau *et al*. (1996)	Rambla Honda, Spain	300	18	116 – 15	HOF SOF
Sandström (1996)	Harra, Tanzania	807	-	-	HOF
Solé-Benet *et al*. (1997)	Tabernas, Spain	235	85.2	44.5	HOF
Wilcox *et al*. (1997)	New Mexico, USA	500	120	2.7	HOF
Bergkamp (1998)	Castilla la Mancha, Spain	400	55	-	HOF SOF
Lavee *et al*. (1998)	Mishor Adumin, Israel	260	-	10.7	HOF

Martínez-Mena *et al*. **(1998)**	Chicamo, Spain	298	110	8.2 – 4.9	HOF SOF
Puigdefabregas *et al*. **(1998)**	Rambla Honda, Spain	300	38	72 – 42	HOF SOF
Descroix *et al*. **(2001)**	Western Sierra Madre, Mexico	450	-	14.8	HOF
Calvo-Cases *et al*. **(2003)**	Alicante, Spain	387 – 474	55 – 27	32.7 – 32.9	HOF SOF
Bartley *et al*. **(2006)**	Weany Creek, Australia	450 – 650	132 – 60	–	HOF
Wilcox *et al*. **(2008)**	Sonora, USA	550	-	190	SSF
Mayor *et al*. **(2009)**	Ventos, Spain	275	64	50 – 33	HOF SOF
Montenegro and Ragab (2010)	Mimoso, Brazil	650	-	11.3	HOF
Chamizo *et al*. **(2012)**	El Cautivo, Spain	235	57	13.3 – 8.8	HOF SOF
Liu *et al*. **(2012)**	Upper Wei River basin, China	512	-	112 – 20	SSF HOF

2.2 Study site

The study was conducted in a 55-ha sub-watershed of the Lake Zapotlán watershed, approximately 100 km south-southwest of Guadalajara, Jalisco, Mexico and 5 km northeast of Ciudad Guzmán, Jalisco, Mexico (19.44°N 103.26°W) (Figure 2.1).

The climate is tropical savannah (Köppen-Geiger: Aw) with a distinct wet and dry season (Peel *et al*., 2007). The average annual rainfall (1982–2003) is 813 mm, of which 95% falls between June to September (Ortiz-Jiménez et al., 2005). Mean annual temperature (1982–2003) is 19.6 ℃ with maximum temperatures occurring in July (Ortiz-Jiménez *et al*., 2005). The catchment is dominated by two distinct forest types. The highly heterogeneous mixed deciduous forest (dominated by *Carpinus caroliniana* and *Mimosa adenantheroides* but with a complex mix of understorey and herbaceous vegetation) occurs at elevations between 1600 and 1800m above sea level (masl) (Figure 2.1). The pine-oak forest (almost exclusively *Pinus montezumae* and *Quercus laeta* with a largely unvegetated understorey) occurs exclusively at elevations greater than 1800 masl. Prior to our investigation, soil cores were collected from the upper 50 cm of the soil profile at both forests and were analysed for particle size distribution using the hydrometer technique (Bouyoucos, 1962), bulk density (ρb) and total porosity (*n*). Total porosity was calculated on the basis of the relationship between ρb and particle density (ρp) (Dingman, 2002). The texture in the upper 50 cm of soil varies between forest types. Soil at the deciduous forest is characterized as sandy clays, whereas sandy loams are present at the pine forest (Table 2.2). The portion of stone fragments in the surface soil was low, representing less than 5% of the total soil volume.
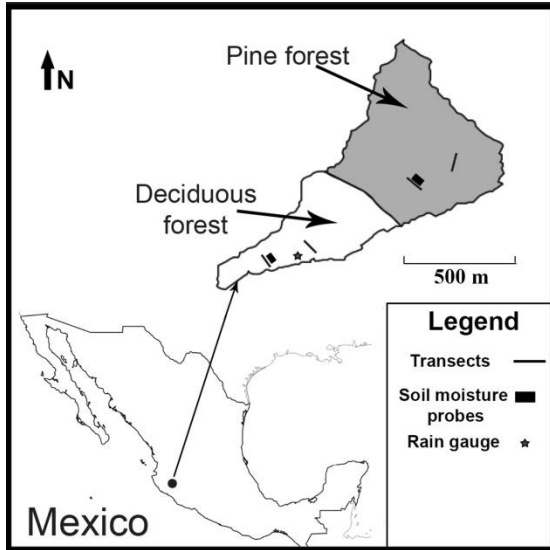
Figure 2.1 Map showing the location of the study watershed, instrumentation and sampling transects.

Two 100-m-long transects were established in each forest type and instrumented with a rain gauge and soil moisture probes from July 2010 to June 2011 (Figure 2.1). Near-saturated hydraulic conductivity and soil water repellency measurements were made at 10m intervals along each transect during this period. These sites were selected as they represented the typical vegetation and slope conditions in the watershed.

Table 2.2 General soil physical properties at different depths at the deciduous and pine forest. Values in parentheses indicate standard deviation

Location	Depth (cm)	Sand (%)	Silt (%)	Clay (%)	Soil type	Bulk density (g/cm^3)	Porosity
Deciduous	0 – 5	56	20	24	Sandy clay-loam	0.82 (0.26)	0.69 (0.10)
	5 – 10	45	17	38	Clay loam	1.10 (0.20)	0.58 (0.03)
	10 – 20	44	18	38	Clay loam	1.15 (0.07)	0.56 (0.16)
	20 – 30	38	19	43	Clay	1.18 (0.21)	0.55 (0.11)
	Average	46	18.5	35.5	Sandy Clay	1.06	0.60
Pine	0 – 5	64.5	21.5	14	Sandy loam	0.91 (0.16)	0.66 (0.06)
	5 – 10	75	12.5	12.5	Sandy loam	1.03 (0.10)	0.60 (0.08)
	10 – 20	50	24	26	Sandy clay-loam	1.14 (0.40)	0.57 (0.10)
	30 – 45	46	36	18	Loam	0.99 (0.27)	0.62 (0.07)
	Average	59	23.5	17.5	Sandy loam	1.02	0.61

2.3 Methods

2.3.1 Rainfall measurements

Open-field precipitation was measured using a Texas Electronics Inc. TE525M-L tipping bucket rain gauge from July 2010 to June 2011 at the deciduous forest site (Figure 2.2). The depth (mm), duration (h) and intensity (mm/h) were calculated for individual storm events over the measurement period.

2.3.2 Soil surface hydraulic conductivity

Near-saturated hydraulic conductivity was measured in July and August 2010 using a Decagon Devices Inc. Mini disk infiltrometer. The mini disk infiltrometer may not be the ideal tool for near-saturated hydraulic conductivity measurements because of the small disk diameter (4.5 cm) and small area of influence; however, conductivity results from this device are accepted in the literature (González-Pelayo *et al*., 2010; Ronayne *et al*., 2012). This device was used because of its portability and low water volumes required for each test, both necessitated by the remote and mountainous sampling locations, which would render the use of more sophisticated devices such as a Guelph Permeameter impractical. At each sampling location, duplicate measurements of hydraulic conductivity were made within 50 cm of the other, under a pressure head of -2 cm. Despite rugged terrain and private property access issues, a total of 80 measurements were made in these forests. At each location, prior to measurement, the soil surface was cleared of large fragments of loose coarse organic litter and the surface smoothed to allow complete contact between the infiltrometer and soil surface. Where the surface was uneven, a thin layer of fine silica sand was placed underneath the stainless steel disk to produce a smooth surface and improve contact with the soil. The infiltrometer was held in place using a ring stand and clamp. Tests were run until a steady state infiltration rate was achieved.

The hydraulic conductivity (K) of the soil was calculated according to Zhang (1997).

$$K = \frac{C_1}{A_2} \tag{2.1}$$

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where C_1 is the slope of the curve of the cumulative infiltration vs. the square root of time (t), and A_2 is a dimensionless coefficient. The value of A_2 depends on the suction disk diameter and the soil texture of the sample location. A_2 is calculated from:

$$A_2 = \frac{11.65(n^{0.1}-1)\exp[2.92(n-1.9)\alpha h_o]}{(\alpha r_o)^{0.91}} \qquad (2.2)$$

where n and α are the van Genuchten parameters for a specific soil texture (Carsel and Parrish, 1988), r_0 is the disk radius, and h_0 is the suction at the disk surface.

2.3.3 Soil water repellency

Soil water repellency was measured in August 2010, December 2010, May 2011 and June 2011 using the molarity of an ethanol drop test (Watson and Letey, 1970). This test utilizes the known surface tensions of standardized solutions of ethanol in water. Droplets (~0.5 ml) with increasing ethanol concentrations (0%, 3%, 5%, 8.5%, 13%, 24% and 36% ethanol in deionized water) were applied to the bare soil surface until the droplet infiltrated in <5 s (Watson and Letey, 1970). The results from the ethanol drop tests were categorized into simple hydrophobicity classes (Table 2.3). As there is no standard for the presentation of hydrophobicity data, we have elected to present our results in terms of simple hydrophobicity classes. To allow for comparison with other hydrophobicity studies, the ethanol concentrations were converted to its equivalent water drop penetration time (Table 2.3).

Table 2.3 Ethanol drop test and Water drop penetration time after Doerr (1998)

	Ethanol drop test			Water drop penetration time	
Class	Ethanol %	Soil water repellency	Class	Penetration time	Soil water repellency
1	0	Very hydrophilic	1	0-5 s	Wettable
2	3	Hydrophilic	2	5-60 s	Slightly water repellent
3	5	Slightly hydrophilic	3	60-600 s	Strongly water repellent
4	8.5	Moderately hydrophobic	4	600s-1 h	Severely water repellent
5	13	Strongly hydrophobic	5	1-3 h	Extremely water repellent
6	24	Very strongly hydrophobic	6	3-6 h	Extremely water repellent
7	36	Extremely hydrophobic	7	>6 h	Extremely water repellent

Before infiltration and repellency measurements were made, soil samples adjacent to measurement plots were taken to a 5 cm depth and were used to determine the volumetric water content (VWC) and soil textural properties (Bouyoucos, 1962).

2.3.4 Soil water content measurements and soil water response

Volumetric water content was measured at two soil pits at each of the forest transects using Campbell Scientific Inc. CS615 water content reflectometers from July 2010 to June 2011. These probes were inserted horizontally in the pits at depths of 10, 20 and 30 cm at the deciduous forest and 10, 20 and 45 cm below the surface at the pine forests. The reflectometers were calibrated in the lab using soil extracted from the same area as per Stenger *et al*. (2005). Total depth equivalent of soil water (mm) was calculated as the VWC (%) from each layer multiplied by the thickness of the measurement layers centred about each measurement (mm).

Soil water infiltration and percolation through the soil pits were characterized for 62 storm events, from 1 July 2010 to 30 June 2011. Events ≥2mm were selected for characterization as they produced a measurable change in shallow soil moisture. Infiltration and percolation rates were calculated using the lag time between rainfall input and the change in depth equivalent of

soil water at each water content reflectometer probe, for both forest sites. The total event increase in soil water was calculated as the difference between the soil water prior to the start of the event and the peak increase in soil water during the storm event.

2.4 Results and discussion

2.4.1 Rainfall

A total of 765mmof rainfall was measured between 1 July 2010 and 30 June 2011, 7.9% below the long-term annual average of 831mm (Figure 2.2). A total of 49% of the rainfall fell from June to July during the early wet season, whereas 50% was delivered from August to September during the late wet season (Figure 2.2). Only one rainfall event was recorded during the dry period. This 9mm event occurred during the early dry season (~1% of the annual precipitation) (Figure 2.2).
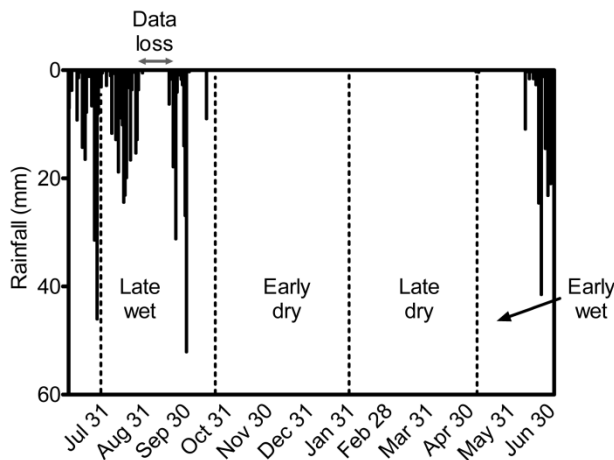
Figure 2.2 Daily rainfall from July 2010 to June 2011. The grey arrow indicates rain gauge error and period of data loss.

Over this 1 year period, there were 62 discrete rainfall events. By using frequency distribution analysis, the events were categorised into 25th, 75th and 95th percentiles. Rainfall events ≤18mm occurred 47 times (75% of all events) but only accounted for 39% of the total rainfall, whereas storm events >18mm occurred 16 times (25% of all events) and accounted for 61% of the total annual rainfall during the study period. Storms were generally of short duration, with 75% being ≤6 h (Figure 2.3). Extreme, long-duration events of 23–40 h occurred and accounted

for 5% of all storms (Figure 2.3). Rainfall intensity ranged between 0.2 and 26.1 mm/h, but only 5% of the events exceeded 17 mm/h (Figure 2.3). These relatively higher intensity events only contributed to 12% of the total annual rainfall. Storms with intensities ≤4.2 mm/h represent 75% of the events and constituted 64% of the annual rainfall. The high frequency of low intensity events can be attributed to convective storms of local atmospheric origin in the region (García-Oliva *et al*., 1995). Over an 8-year period, García-Oliva *et al*. (1995) showed that only 3% of storm events had intensities that were greater than 24 mm/h and attributed this to the low frequency of cyclonic storm systems.
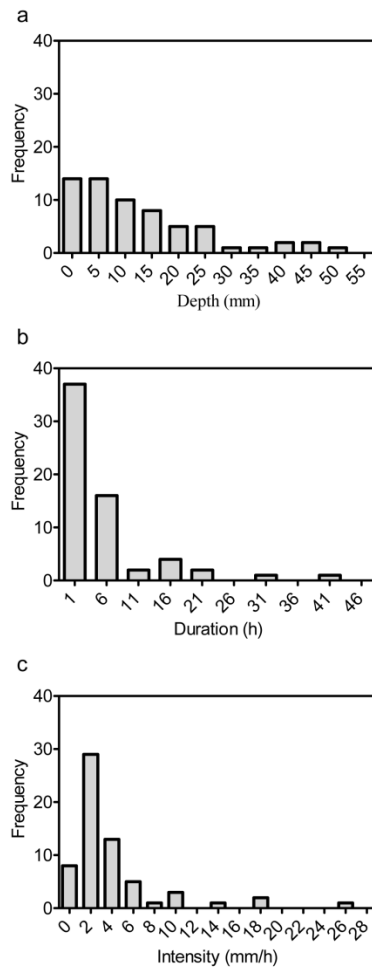
Figure 2.3 Frequency distribution of a) depth (mm), b) duration (h) and c) intensity (mm/h) for 62 storm events

The rainfall characteristics at this research site are similar to those of many semi-arid regions, dominated by short-duration, low-intensity events during the wet season (Martínez-Mena *et al*., 1998; Chamizo *et al*., 2012). However, two important differences are noted. Firstly, the

maximum recorded rainfall intensity is only 26.1 mm/h and was substantially lower than other maximum intensities in the semi-arid literature (Table 2.1). Secondly, the contribution of high-intensity events to the total annual rainfall is only 12% as compared with 70% for semi-arid catchments (Wilcox *et al*., 1997; Martínez-Mena *et al*., 1998).

2.4.2 Soil surface hydraulic conductivity

At the deciduous forest, *K* was spatially variable across both transects (Figure 2.4). Hydraulic conductivity ranged from 24.4 to 164.0 mm/h at transect one and 21.0 to 144.1 mm/h at transect two. The mean *K* was 68.7 ± a standard deviation of 49.4 mm/h, with a median value of 50.4 mm/h. Hydraulic conductivity at all measurement locations was greater than the 75th percentile of rainfall intensity, whereas three of the 20 measurements had *K* that was less than the maximum rainfall intensity (Figure 2.4). Given that *K* exceeds the rainfall intensity of 75% of the storm events, infiltration would not be limited during most of the wet period, and HOF is not generated during these events.

Hydraulic conductivity at the pine forest was also heterogeneous (Figure 2.4). The *K* values ranged from 9.2 to 53.9 mm/h at transect one and 11.8 to 58.8 mm/h at transect two; with a mean value of *K* was 25.0 ± a standard deviation of 17.0 mm/h, which was two times lower than the deciduous forest. The relationship between *K* and rainfall intensity was similar to the deciduous forest. Hydraulic conductivity values at all measurement points exceeded the 75th percentile of rainfall intensity (Figure 2.4). Unlike the deciduous forest, there were a greater number of measurement locations where *K* was lower than the maximum rainfall intensity (Figure 2.4). These 11 points indicate that under the highest intensity, infiltration would be limited over much of the surface, and HOF can be generated.

The mean *K* values of 68.0 mm/h at the deciduous forest and 25.0 mm/h at the pine forest were greater than *K* measured in many semi-arid areas (Table 2.1). The higher *K* at our site is likely a result of the higher density of vegetation and understorey growth than in semi-arid catchments. Lavee *et al*. (1998) observed that infiltration increased from arid to semi-arid to sub-humid zones, attributing this to greater vegetation densities. A higher plant density often enhances

infiltration through macropore formation and preferential flow along root channels (Bergkamp 1998; Calvo-Cases *et al*., 2003). Greater plant density also improves soil stability and reduces surface crust formation as a result of greater organic matter input (Wilcox *et al*., 1988; Lavee *et al*., 1998).
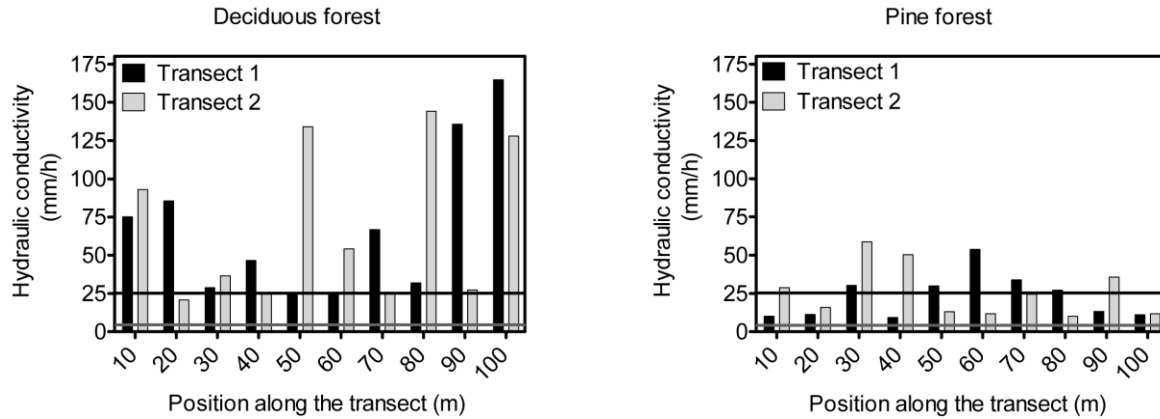
Figure 2.4 Spatial distribution of surface hydraulic conductivity (mm/h) along two transects at a) the deciduous forest and b) the pine forest. The black horizontal line indicates the maximum rainfall intensity. The grey horizontal line indicates the 75th percentile of rainfall intensity

Our results show that unlike many semi-arid regions, *K* does not limit infiltration across most of the ground surface. In many semi-arid regions, rainfall intensity exceeds *K* producing HOF (Table 2.1). At our research site, 90% of the storm events did not exceed *K* at the deciduous forest and 85% did not exceed *K* at the pine forest, indicating that HOF would not occur through most of the wet season. At locations where the rainfall intensity exceeds *K*, HOF will be generated. However, this surface flow will be discontinuous across the hillslope as it would re-infiltrate at high *K* surfaces downslope. This response is more like that of humid temperate forests, where relatively higher *K* and low-intensity rainfall promote infiltration and the production of more vertical and lateral subsurface flows (Bonell, 1993).

2.4.3 Soil water repellency

Soil water repellency at the deciduous forest showed high temporal variability (Table 2.4). Fifty days after the start of rainfall (late wet), the soil surface was non-repellent with a mean water drop penetration time of 0 s. When measured 54 (early dry) and 177 (late dry) days after the last rainfall event, the penetration time increased to mean values of 3.8 and 2.6 h (extremely

repellent) for each of the sample dates (Table 2.4). Repellent conditions persisted through the periods without rainfall and were reduced to a mean penetration time of 0.6 s when measured 14 days (Early wet) after the onset of rainfall (Table 2.4). Soil water repellency showed little spatial variability during each measurement period. All of the sampled areas were non-repellent during both wet periods. When measured during the early dry period, extreme water repellent conditions were present at 18 of the 20 sampling locations whereas the other two locations exhibited slight water repellency. During late dry period, 17 of the 20 locations were extremely repellent whereas the other three locations were slightly repellent (Table 2.4).

The temporal distribution of soil water repellency at the pine forest did not differ substantially from the deciduous forest (Table 2.4). Both wet periods had penetration times of 0 s. The water drop penetration time increased to 3.5 h during the early dry period and was reduced to 2.5 h during the late dry period. The spatial distribution of hydrophobic conditions showed low variability. There was no variability during both wet periods, with 100% of the sampled area being non-repellent. Likewise, 100% of the sample area was extremely repellent in the early dry period. During the late dry period, 17 of the 20 sampling locations were extremely repellent with the remaining three locations slightly repellent.

Water drop penetration times at the research site are similar to other semi-arid regions, dominated by long penetration times and extreme repellency (Martínez-Murillo and Ruiz-Sinoga, 2007; Zavala *et al*., 2009). Extreme levels of soil water repellency are produced under dry conditions and organic matter input. These conditions were present and persisted during both dry periods, where soil moisture is low (Figure 2.5) and litterfall is highest (García-Oliva *et al*., 2003). The extreme water repellent conditions at the end of the dry period suggest that there would be a reduction in infiltration at the start of the wet season. However, the effect of repellency on infiltration appears to be minimal as there are clear, albeit small increases in soil moisture 10 cm below the soil surface at both forests (Figure 2.5).

The absence of repellency during the wet season is unlike those of many semi-arid regions, where there is no substantial difference in repellency between dry and wet seasons (Crockford *et al*., 1991; Verheijen and Cammeraat, 2007; Zavala *et al*., 2009). Crockford *et al*. (1991) and

Zavala *et al*. (2009) attributed this to the short duration and sporadic nature of rainfall, which prevent the flushing of hydrophobic compounds from the soil. At our site, the longest period without rainfall was 3 days, indicating that there will be significant flushing of hydrophobic compounds, which prevents the re-establishment of soil repellency. The absence of soil water repellency during the wet season indicates that infiltration will not be impeded and HOF is unlikely to occur.

Table 2.4 Water drop penetration time at each individual plot for the Late wet (august 2010), Early dry (December 2010), Late dry (May 2011) and Early wet (June 2011) seasons. s = seconds, h = hours

Plot	Deciduous forest				Pine forest			
Season	Late wet	Early dry	Late dry	Early wet	Late wet	Early dry	Late dry	Early wet
1	0 s	4.5 h	3.5 h	0 s	0 s	3.5 h	3.5 h	0 s
2	0 s	4.5 h	3.5 h	0 s	0 s	2.5 h	2.5 h	0 s
3	0 s	2.3 h	0.3 h	0 s	0 s	5.3 h	2.5 h	0 s
4	0 s	4.3 h	4.5 h	0 s	0 s	4.3 h	2.3 h	0 s
5	0 s	3.5 h	3.5 h	1 s	0 s	4.3 h	2.5 h	0 s
6	0 s	5.3 h	2.3 h	0 s	0 s	3.5 h	1.3 h	0 s
7	0 s	3.5 h	4.3 h	0 s	0 s	3.5 h	2.2 h	0 s
8	0 s	4.5 h	0.3 h	4 s	0 s	4.5 h	1.3 h	0 s
9	0 s	3.0 h	1.3 h	1 s	0 s	0.6 h	4.5 h	0 s
10	0 s	3.5 h	3.5 h	0 s	0 s	3.5 h	2.3 h	0 s
Mean	0 s	3.8 h	2.6 h	0.6 s	0 s	3.5 h	2.5 h	0 s

2.4.4 Soil water response to rainfall inputs

At the deciduous forest, the response to rainfall between 10 and 30 cm below the surface was monitored during the wet period (Figure 2.5). Events ≥2mm produced an increase in VWC at the 10 and 20 cm soil layers, whereas events ≥7mm were needed to produce a response at the 30 cm layer. During these storm events, water rapidly percolated through the profile. Mean percolation rates of 288, 240 and 221mm/h were calculated for the 10, 20 and 30 cm layers respectively.
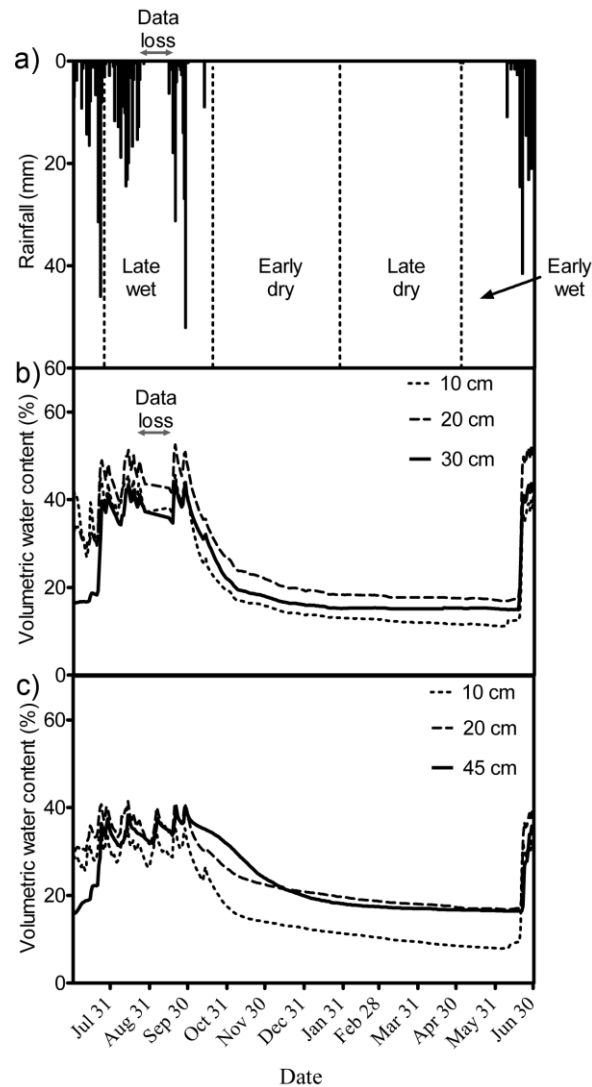
Figure 2.5 a) Daily rainfall and daily volumetric water content at b) the deciduous forest and c) the pine forest. The grey arrow indicates logger error and period of data loss

At the pine forest, the 10 cm layer responded to events ≥2 mm, whereas the 20 and 45 cm layers responded to events greater than 3 and 9mm respectively. Percolation through the profile was rapid during rainfall events and decreased with depth with a mean value of 206 mm/h at 10 cm, 170 mm/h at 20 cm and 103 mm/h at 45 cm. The response to rainfall was slower and less frequent than the deciduous forests. This is likely due to the lower K at the pine forest and rainfall interception by the litter layer, which in other pine forests, can retain up to 1.7 mm of rainfall (Putuhena and Cordery, 1996).
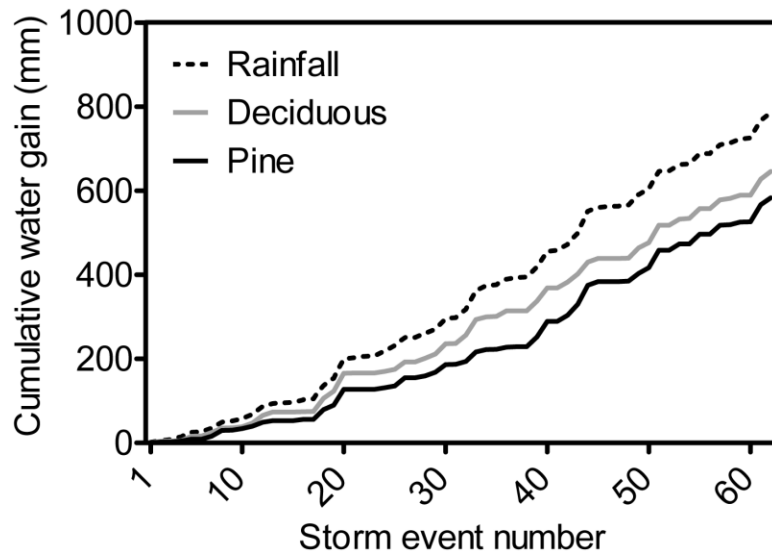
Figure 2.6 Cumulative gain in rainfall and event-level soil water at the deciduous and pine forest for 62 events

The increase in the depth equivalent soil water for the shallow soil profile at both forests had a strong positive relationship to the depth of rainfall. This increase is the difference between the soil water prior to the start of the event and the peak soil water during the storm event and represents the gain before water is lost to evaporation or deep percolation. This change, barring interception, saturation or overland flow should equal the depth of rainfall. The event-level increase in soil water at the deciduous forest was between 11% and 100% of the incoming rainfall, whereas at the pine forest ranged from 8% to 100% of the incoming rainfall. Over the 62 storm events, the event-level increase in soil water resulted in a cumulative gain of 645 mm or 84% of the 765mm of rainfall at the deciduous forest and 583mm or 76% of the total storm inputs at the pine forest (Figure 2.6). The high re-infiltration of rainfall through the shallow soil is unlike semi-arid regions where only 44% of rainfall is redistributed (Bergkamp *et al*., 1999). The greater redistribution of water at our site can be attributed to higher density of vegetation, which slows surface runoff and enhances infiltration (Cerdà, 1997b; Bergkamp *et al*., 1999).

2.5 Conclusion

Despite having similar climate and vegetation regimes, the hydrological controls that govern runoff generation in semi-arid catchments are not transferable to tropical dry forests. Our results show that extreme levels of soil surface water repellency develop but do not persist during the wet season and K was greater than the rainfall intensity of more than 75% of storm events. These conditions promoted the rapid infiltration and percolation of water through the upper 30 cm of soil, indicating that subsurface flow, not infiltration excess overland flow, is the dominant runoff process in this landscape. If we are to improve our understanding of runoff generation in tropical dry forests, then a better grasp on the mechanisms and controls on subsurface flow processes is essential. In order to do so, we need to increase the spatial extent of hydraulic conductivity measurements and examine changes in hydraulic conductivity at different soil depths. Furthermore, a greater characterisation of hydraulic conductivity – soil moisture curves and direct measurements of soil water fluxes are essential in improving the understanding of unsaturated zone soil-water dynamics.

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Chapter 3

3.0 Soil water storage, rainfall and runoff relationships in tropical dry forest catchment

3.1 Introduction

Tropical dry forests account for approximately 19% of the total world forested area and 42% of the global tropical forests (Murphy and Lugo, 1986). These forests are widespread in Central and South America, particularly along the Pacific coast of México where more than 40% of the vegetation is tropical dry forest (Sánchez-Azofeifa et al., 2009). The key defining features of tropical dry forests are a distinct 3 – 7 month dry period and an average ratio of annual potential evapotranspiration to rainfall >1 (Murphy and Lugo, 1995; Miles et al., 2006) indicating an arid and water limited climate (Budyko, 1974) and intermittent streamflow, where streams remain dry for 6 – 10 months and streamflow is activated during the short but intense wet season (Vose and Maas, 1999; Mugabe et al., 2007).

Most research in tropical dry forests has focused on quantifying the catchment scale water balance (Lugo et al., 1978; de Araújo and González Piedra, 2009; Montenegro and Ragab, 2010), while other work has focused on identifying the primary forms of runoff generation (Sandström, 1996; Masiyandima et al., 2003; Mugabe et al., 2007). In tropical dry forests, runoff is strongly controlled by the hydraulic properties of the surface and shallow subsurface soils. It has been shown that runoff is dominated by infiltration excess (Hortonian) overland flow (HOF), saturation excess overland flow (SOF) (Sandström, 1996; McCartney et al., 1998; Mugabe et al., 2007) and limited subsurface stormflow (Masiyandima et al., 2003). These studies provide a foundation of knowledge concerning the mechanisms by which runoff is generated. Building upon this knowledge, it is now possible to address questions about the specific thresholds needed to initiate runoff. It has clearly been demonstrated in both humid and arid temperate catchments that specific thresholds of antecedent storage or precipitation are required to initiate runoff (Ali

et al., 2013); however, debate still exists as to the relative importance of storage versus precipitation thresholds (McDonnell, 2013).

A non-linear relationship between antecedent soil water storage and runoff generation has been observed in many catchments. Threshold shallow soil moisture contents must first be reached before there is an abrupt increase in streamflow and generation of large amounts of stormflow (Western and Grayson, 1998; James and Roulet, 2007; Penna *et al*., 2011b). The soil moisture content threshold reflects changes in the storage deficits and the overall wetness of the catchment. Above the threshold, storage deficits are low and connectivity between the hillslope and stream channel occurs through lateral flow processes (Buttle *et al*., 2004; Tromp-van Meerveld and McDonnell, 2006b; Oswald *et al*., 2011; Camporese *et al*., 2014). Recent studies have focused on combining both a depth equivalent antecedent soil water content and event precipitation to assess stormflow generation at the catchment scale. Detty and McGuire (2010) showed that an increase in the sum of antecedent soil water content and event precipitation from 302 to 332 mm increased stormflow from 0.3 to 11 mm or a 97% increase in the volume of stormflow. It has been suggested that the combined antecedent soil water and precipitation threshold indicate a storage threshold amount, but also a source area threshold, where above the threshold there is increased hillslope-stream connectivity across increasing the size of the catchment contributing area (Detty and McGuire, 2010; Fu *et al*., 2013).

Notwithstanding the importance of non-linear storage and runoff relationships, threshold rainfall has been shown to be an important control of runoff across humid temperate forests (Spence and Woo, 2003; Kim *et al*., 2005), humid tropical forests (Negishi *et al*, 2007) and semi-arid catchments (Cammeraat, 2004). Rainfall-runoff thresholds have been identified at the small plot (Nicolau *et al*., 1996), hillslope (Tromp-van Meerveld and McDonnell, 2006a) and catchment scale (Cammeraat, 2004; Fu *et al*., 2013). The threshold depth of rainfall required to generate runoff typically increases from the plot to catchment scale, representing not only the greater complexity in the interactions among rainfall, soil, vegetation and topography, but also changes in the dominant runoff mechanisms (Nicolau *et al*. 1996; Buttle *et al*., 2004; Cammeraat, 2004; Tromp-van Meerveld and McDonnell, 2006a). Ali *et al*. (2013) suggested that in order to avoid the inherent problems with scaling up non-linear rainfall-runoff responses, it may prove more

useful to characterise rainfall thresholds at the catchment scale. Although research characterizing the threshold rainfall needed to generate runoff at the catchment scale is limited, most studies indicate that this threshold is only exceeded after storage deficits are satisfied, thereby connecting hillslopes to streams (Buttle *et al*., 2004; McGuire and McDonnell, 2010; Fu *et al*., 2013).

Because the storage reservoirs are depleted during the $3 - 7$ month long dry season, tropical dry forests can be used to test the relative importance of storage versus precipitation thresholds. Understanding non-linear threshold relationships at the catchment scale is critical in model development, which often suffers from scaling issues (Zehe and Sivapalan, 2009; Penna *et al*., 2011a). Furthermore, because non-linear stormflow behaviours are observed across many catchments, thresholds present a uniform metric e.g. for inter-catchment comparison (Ali *et al*., 2013). In this study we investigated the relationship between soil water storage, rainfall and runoff in a Mexican tropical dry forest catchment with the goal of improving our understanding on the controls that govern streamflow generation. The specific objectives of this work were to:

1. Identify the soil water storage and hydrometeorological controls on streamflow activation after the dry season in a tropical dry forest catchment

2. Determine if the dominant controls on seasonal streamflow activation are also the primary controls on stormflow runoff generation at the event scale.

3. Use the rainfall-runoff relationship and lag to peakflow to gain insight into the dominant runoff mechanism(s) in a tropical dry forest catchment.

3.2 Study site

The study was conducted in a 3.15 km^2 catchment in the Lake Zapotlán watershed, approximately 100 km south-southwest of Guadalajara, Jalisco, Mexico; 5 km northeast of Ciudad Guzman, Jalisco, Mexico (19º46N 103º27W – 19º47N 103º25W) (Figure 3.1). The climate is Tropical Savannah (Köppen-Geiger: Aw) with a distinct wet and dry season (Peel *et al*., 2007). The average annual precipitation (1972 – 2003) is 813 mm, of which 95% falls between June to September (Ortiz-Jiménez *et al*., 2005). Rainfall is dominated by short duration,

low intensity storm events (Farrick and Branfireun, 2014). The strong wet-dry seasonality results in intermittent streamflow production from the catchment. Mean annual temperature is 19.6ºC with maximum temperatures occurring in July (Ortiz-Jiménez *et al*., 2005). The annual average ratio of potential evapotranspiration to rainfall in the region is 2.5 (Farrick and Branfireun, 2013), which indicates an arid and water limited climate (Budyko, 1974).

Elevation ranges from 1557 metres above sea level (masl) at the primary outflow channel to 2170 masl at the headwater sub-basin. The catchment is steep with slopes ranging from 30º to over 40º. The study area is underlain by Pleistocene volcanic basaltic andesite and volcanic fine tuff. The channel width ranges from <0.20 m in the headwater sub-basins to 1.0 – 1.5 m at the primary outflow channel. The stream channels are deeply incised and steep, with a narrow riparian zone (0.2 – 1 m). The soil is classified as a chromic cambisol with andic properties of volcanic origin (Gómez-Tagle, 2009). The soil at the hillslope is deep, often >1 m. Soil textures are mainly loams and sandy soils and vary from sandy loams in the O and upper A horizons to loams and sandy-clay loams at depths below 50 cm (Farrick and Branfireun, 2014). The surface hydraulic conductivity (K) is highly variable, ranging from 9 – 164 mm h^{-1} (Farrick and Branfireun, 2014). Bulk density and total porosity of the upper 100 cm of soil range from 0.91 – 1.14 g cm^{-3} and 0.57 – 0.66 respectively (Farrick and Branfireun, 2014).
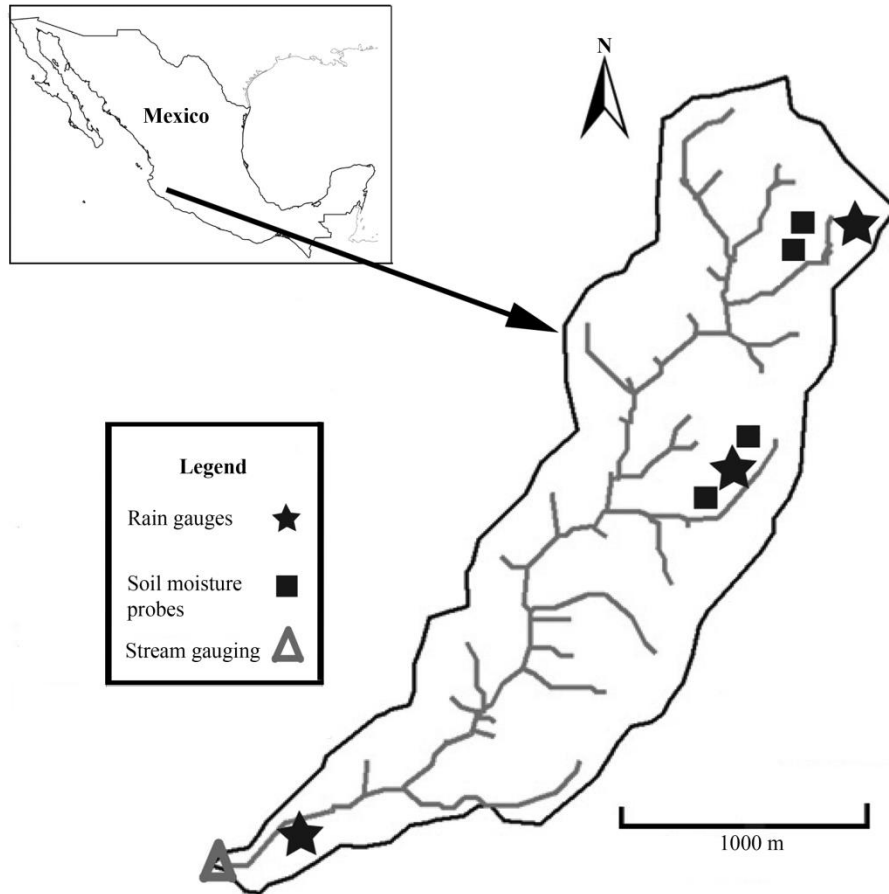
Figure 3.1 Location of study catchment and hydrological instrumentation

The catchment is dominated by two distinct forest types. The pine-oak forest (almost exclusively *Pinus montezumae, Quercus laeta*) occurs at elevations greater than 1800 masl and occupies 82% of the catchment area. The highly heterogeneous mixed deciduous forest (dominated by *Carpinus caroliniana*, *Mimosa adenantheroides*, with a complex mix of understorey and herbaceous vegetation) occurs at elevations between 1600 – 1800 masl and covers 13% of the catchment. Land development in the catchment is low with fragmented agricultural plots and roads occupying 4 and 1% of the catchment respectively.

3.3 Methods

3.3.1 Hydrometric measurements

Open field precipitation was measured from 1 May to 24 September 2012 at 10 minute intervals using a Texas Electronics tipping bucket rain gauge installed at three locations across the catchment at 1600, 1800 and 1950 masl (Figure 3.1). Rain storm events were defined as periods of rainfall greater than 1 mm, separated by 6 h (Penna *et al*., 2011b). A total of 56 storm events were identified during the study period.

Volumetric water content (VWC cm^3 cm^{-3}) was measured at four soil pits installed across the catchment using Campbell Scientific Inc. CS615 Water Content Reflectometers from 1 May to 24 September 2012 (Figure 3.1). Two of these pits were located at convex hillslopes, approximately 60 m upslope of the stream, while the other two pits were installed at concave hillslopes approximately 20 m upslope of the stream. We did not measure VWC in the small riparian areas as research in other steep catchments with incised streams and narrow riparian zones shows that contributions from the near stream area to runoff is often very low (Sidle *et al*., 2000; McGuire and McDonnell, 2010). The reflectometers were inserted horizontally in the soil pits at depths of 10, 30, 50 and 100 cm below the surface at all four locations. The reflectometers were calibrated in the laboratory using soil extracted from the same area following the technique of Stenger *et al*. (2005). The error of calibrated reflectometers was low with a standard deviation of $\pm$ 3%. The depth equivalent soil water (mm) between the 10 cm and 100 cm soil layers at each soil pit was calculated as,

$$Depth\ equivalent\ soil\ water = \sum_{i=1}^{N} [VWC_i \times D_i] \qquad (3.1)$$

where i is the index representing the different soil layers, N is the number of instrumented soil layers (3), VWC_i is the average of the volumetric water content values bounding each layer (e.g. VWC_1 uses the mean VWC from the probes 10 and 30 cm below the surface) and D_i is the thickness of the soil between two reflectometer probes ($D_1 - D_3$: 200 mm, 200 mm and 500 mm). Using the depth equivalent soil water from the two pits at the convex hillslope, we calculated the average value for the convex hillslope. The same method was used for the two soil pits at the concave hillslope.

The drainable porosity at each instrumented soil layer was calculated by subtracting the *VWC* at field capacity from the total porosity. Total porosity was determined from earlier work in these forests (Farrick and Branfireun, 2014), while field capacity was estimated as the *VWC* at which the rate of decline during drainage conditions became insignificant (Oswald *et al*., 2011).

The stream water level at the primary outflow channel (Figure 3.1) was recorded at 10-minute intervals, using a 0.8 mm resolution Odyssey capacitance water level logger (Dataflow Systems Ltd.). Modification of the channel occurred before the onset of this current study, where large rocks and boulders were removed and an artificial wall was constructed and backfilled to produce a quiescent pool and a small dam structure that had a clear free fall on the downstream side. Discharge was calculated from the water level using the end-depth method (Jain *et al*., 2007). This method was selected as the stream fit the criteria required to accurately measure discharge: free fall where the drop is greater than the stream stage, rectangular, smooth channel without rocks or boulders. Discharge was calculated as:

$$Q = C \sqrt{g \ bh^{3/2}} \qquad\qquad (3.2)$$

where Q is the discharge (m^3/s), C is the coefficient of discharge, g is the acceleration due to gravity, b is the channel width (m) and h is the water level (m). A value of 1.66 was used for the coefficient of discharge. We confirmed the accuracy of the estimated discharge by conducting manual discharge measurements using a stopwatch and buckets under flow conditions, ranging from 5.42×10^{-3} to 1.46×10^{-2} m^3/s. However, due to the largest events occurring at night when access to the site was restricted, we were unable to capture the peakflow of larger storm events. Without measuring these flow conditions, there is a degree of uncertainty with respect to the discharge and the quickflow volumes produced during the largest storms.

3.3.2 Graphical hydrograph separation

Storm runoff events were defined as the period from the initial rise in discharge from a local minimum in the hydrograph to the next local minimum and were separated into quick flow (*QF*) and delayed flow (*DF*) volumes using the local minimum method (Sloto and Crouse, 1996). Quick flow, *DF* and total event runoff (*R*), all in mm, were calculated as the sum of the 10

minute values (mm) over the selected event period. The gross event rainfall depth (P) was calculated as the sum of the 10 minute values for the duration of the storm event. The hydrologic behaviour of the catchment was examined during the study period using the ratio of total runoff to rainfall (R/P) and the ratio of quick flow to rainfall (QF/P).

The lag time between storm onset and peak streamflow (T_{rise}) was calculated as the time difference (hours) between the start of rainfall and peakflow (Mosley, 1979). The lag time between storm onset and peak soil moisture response at the 10 to 100 cm layers was calculated as the time difference (hours) between the start of rainfall and peak in VWC. The lag time was determined for all rainfall-stormflow events observed during the wet season.

3.3.3 Antecedent wetness calculations

The effect of antecedent conditions on the catchment runoff response was assessed using three different measures of antecedent wetness: 1. antecedent precipitation (mm), 2. antecedent soil water (ASW) (mm) and 3. the sum of the antecedent soil water and event rainfall ($ASW + P$) (mm). Antecedent precipitation was calculated as the cumulative rainfall (mm) seven days prior to the start of the runoff producing storm event (James and Roulet, 2009). Antecedent soil water was calculated as the depth equivalent soil water (mm) before the onset of a storm event. The antecedent soil water and event rainfall index was calculated as the sum of the antecedent soil water prior to the storm event and the gross event rainfall depth (Detty and McGuire, 2010). In order to investigate the threshold behaviour of the catchment, we examined the relationship between these antecedent wetness indices and quick flow for the 21 events recorded after streamflow had commenced.

Piecewise regression analysis (PRA) was used to examine the threshold behaviour of QF versus the combined $ASW + P$. Piecewise regression models are broken-stick models, where two or more lines unite at an unknown point, known as break-points. These break-points represent the threshold in relationships (Toms and Lesperance, 2003). These models have successfully been used to determine the breakpoint in storage-discharge relationships in boreal catchments (Oswald *et al.*, 2011). The piecewise regression analysis was performed using WinBUGS1.4, an

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interactive Windows based program for Bayesian analysis of complex statistical analysis (Lunn *et al*., 2000).

3.4 Results

3.4.1 Hydrometeorological conditions

The total rainfall from May to September, 2012 was 599 mm which was 24% below the long-term seasonal average of 789 mm for the same period (Figure 3.2). One hundred and nineteen millimetres or 20% of the total rainfall was discharged as streamflow during May – September. The remaining 80% or 440 mm was distributed among evapotranspiration, ground water recharge or change in unsaturated soil storage.

From May 17 to June 10, a total of 15 mm of rainfall was recorded. During this period a stable mean *VWC* of 13.8% between the 10 and 100 cm soil layers from all four pits was recorded (Figure 3.2). The size and frequency of storm events increased after June 10. A total of 176 mm of rainfall was recorded from June 10 to July 7. During this transition phase or wetting up period, there was a progressive increase in *VWC* from the 10 to 100 cm layer, with the *VWC* reaching field capacity at 10, 30 and 50 cm layers (Figure 3.2). Streamflow was absent for most of this period and was only activated after the *VWC* at 100 cm below the surface increased to a threshold value of 23% at the convex hillslope and 29% at the concave hillslope, a mean of 26% from both slopes, which was near field capacity (Figure 3.2). The activation of streamflow occurred after a cumulative input of 191 mm of rainfall over 52 days.
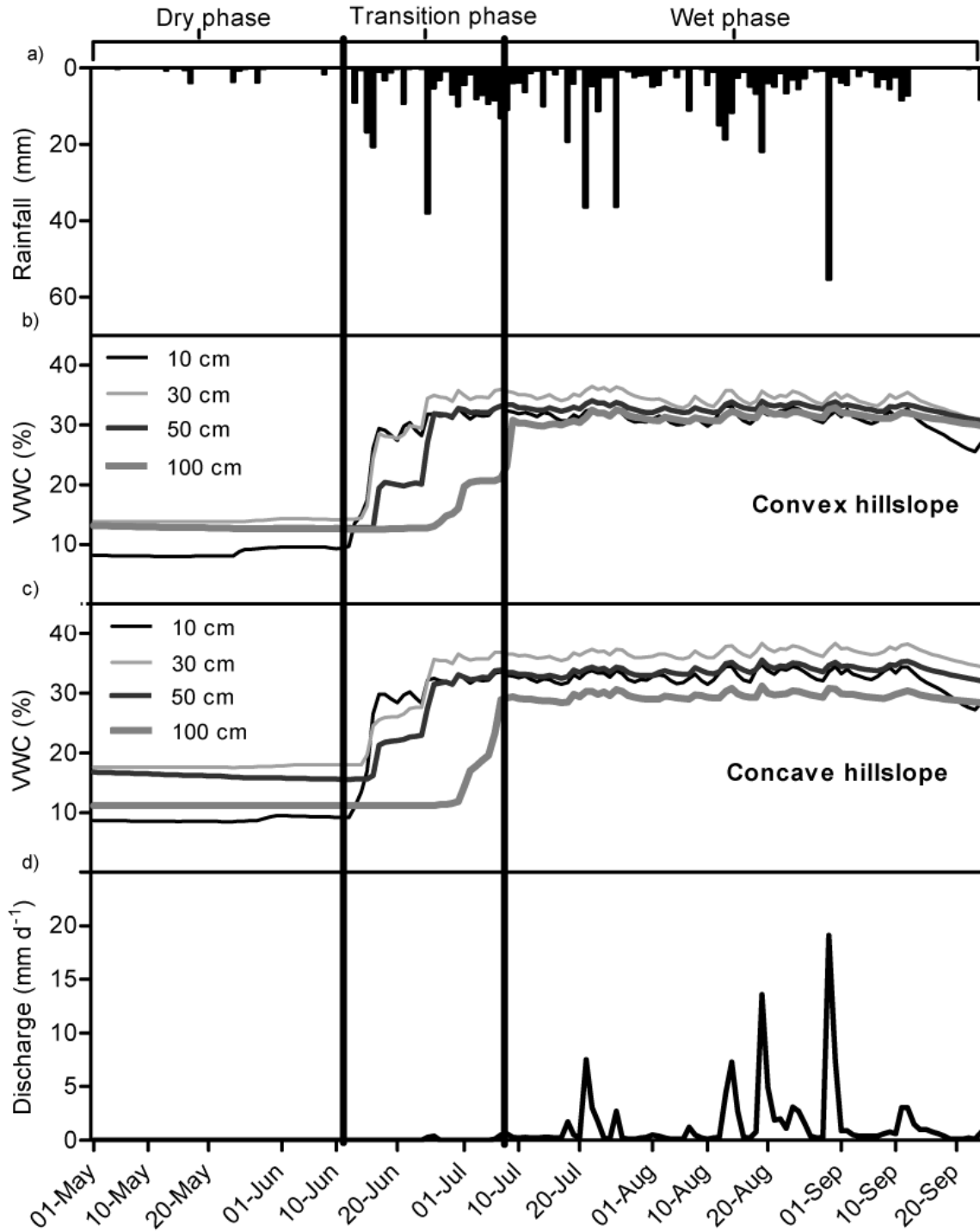
Figure 3.2 Daily variation in a) rainfall, b) convex hillslope volumetric water content, c) concave hillslope volumetric water content and d) discharge from May to September, 2012

More importantly, a storage deficit of 162 mm of soil water, calculated as the change in the mean

depth equivalent soil water (mm) between the 10 and 100 cm soil layers from all four soil pits

from June 10 (start of the transition phase) to July 7 (streamflow activation), was satisfied before streamflow was activated (Figure 3.3). While the cumulative soil water remained below the cumulative precipitation for most of study period, from June 25 and June 29, the cumulative soil water was larger than rainfall. Total precipitation between the onset of streamflow on July 7 and the end of the study on Sep 24 was 408 mm, of which 118 mm or 29% was discharged as streamflow.
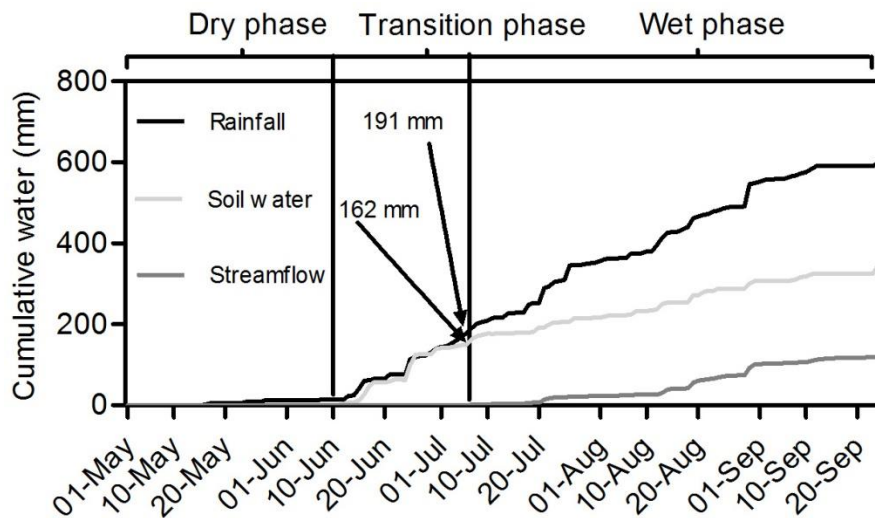
Figure 3.3 Cumulative rainfall, soil water and streamflow from May to September, 2012

3.4.2 Rainfall – runoff relationships

A series of 21 storm events were monitored during the wet phase after streamflow had commenced. Rainfall depths during these storms ranged from 2.2 to 58.6 mm, with 75% of the events being ≤15 mm. The mean rainfall intensities ranged from 0.5 to 25.1 mm h^{-1} with 95% events <20 mm h^{-1}. Rainfall durations ranged from 0.7 to 22.5 hours during the measurement period (Table 3.1). During the wet period, storm runoff during an event ranged from 1 to 35.5 mm and showed a statistically significant linear relationship with P, ($r^2 = 0.76$; $p < 0.0001$; $R = 0.5949P$-3.693) (Table 3.1). A minimum P threshold of 4.1 mm was needed to generate a 1 mm increase in runoff. The quick flow or stormflow component of the hydrograph ranged from 0.2 to 32.2 mm and was strongly influenced by the event rainfall depth, increasing linearly with P ($r^2 = 0.84$; $p < 0.0001$). Quick flow showed high variability in the relationship with precipitation when

P was greater than 14 mm (Figure 3.4a). The mean *QF/P* was 0.26 and ranged from 0.04 to 0.72 (Table 3.1) and had a linear relationship with *P* ($r^2 = 0.40$; $p < 0.0001$), which was statistically significant and showed low scatter throughout the relationship when *P* was less than 14 mm (Figure 3.4b). The frequency distribution of the portion of rainfall delivered as *QF* indicates that for 75% of the storm events *QF/P* was less than 0.36.
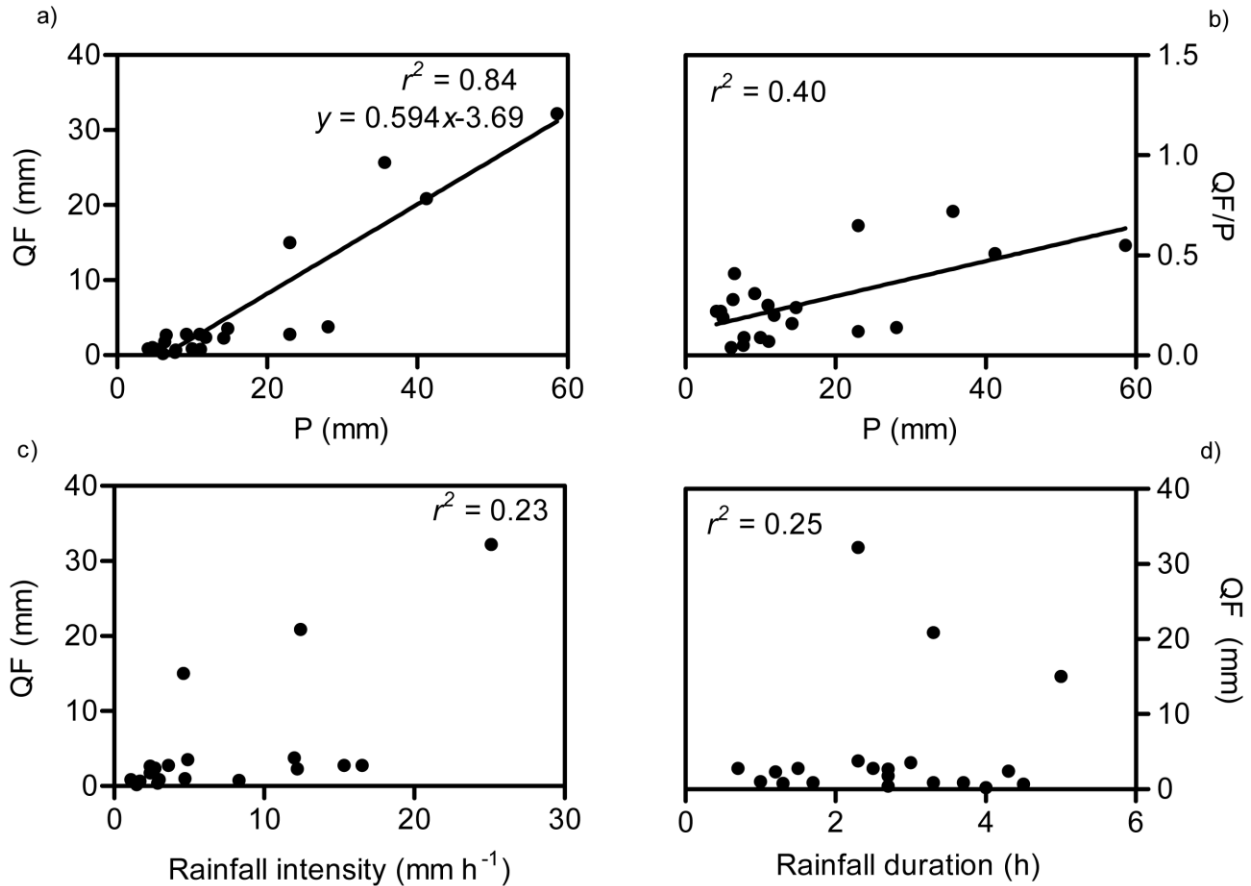
Figure 3.4 Relationship between a) event quick flow (QF) and event rainfall (P), b) runoff coefficient (QF/P) and P, c) QF and rainfall intensity (mm^{-1}) and d) QF and rainfall duration (hours)

Quickflow had a weak positive relationship with rainfall intensity and generally increased with increased rainfall intensity, although the linear relationship ($r^2 = 0.23$, $p < 0.03$) was poor during the wet phase (Figure 3.4c). Similarly, rainfall duration had a weak influence on *QF*. There was a non-significant linear increase in *QF* ($r^2 = 0.25$; $p < 0.02$) with increased duration (Figure 3.4d).

Table 3.1 Storm event characteristics from July to September, 2012. P = event rainfall; R = total event runoff; QF = event quick flow; QF/P = runoff coefficient. Above threshold events represent storms that generate more than 4 mm of QF

Storm Date	Event number	Rainfall characteristics			Stormflow response			Antecedent wetness conditions	
		P (mm)	Rainfall intensity (mm h^{-1})	Duration (h)	R (mm)	QF (mm)	QF/P	Antecedent soil water (mm)	Antecedent soil water + P (mm)
Below Threshold Events									
8 Jul	1	10.0	3.0	3.3	1.0	0.9	0.09	261	271
18 Jul	2	23.0	15.3	1.5	3.0	2.8	0.12	254	277
21 Jul	4	28.1	12.0	2.3	6.1	3.8	0.14	261	289
23 Jul	3	14.2	12.2	1.2	2.5	2.3	0.16	267	281
1 Aug	5	5.0	3.0	1.7	1.3	0.9	0.19	260	265
7 Aug	6	11.8	2.7	4.3	2.9	2.4	0.20	257	269
18 Aug	7	4.7	4.7	1.0	1.3	1.0	0.22	266	270
21 Aug	8	6.5	2.4	2.7	4.8	2.7	0.41	268	274
23 Aug	9	4.1	1.1	3.7	2.2	0.9	0.22	271	275
23 Aug	10	6.1	1.5	4.0	2.6	0.2	0.04	272	278
25 Aug	11	6.3	2.4	2.7	4.8	1.8	0.28	273	279
1 Sep	12	7.8	1.7	4.5	2.5	0.7	0.09	268	275
9 Sep	13	7.7	2.9	2.7	1.9	0.4	0.05	267	274
11 Sep	14	11.0	16.5	0.7	7.0	2.8	0.25	270	281
12 Sep	15	9.2	3.6	2.5	7.0	2.8	0.31	273	282
24 Sep	16	11.1	8.3	1.3	1.7	0.8	0.07	244	255
Mean		10.4	5.8	2.5	3.3	1.7	0.18	264	275
Above Threshold Events									
21 Jul	17	14.7	4.9	3.0	7.9	4.0	0.24	279	294
26 Jul	18	41.2	12.4	3.3	21.6	20.9	0.51	265	306
13 Aug	19	35.6	1.6	22.5	29.2	25.7	0.72	260	295
19 Aug	20	23.0	4.6	5.0	24.0	15.0	0.65	277	300
30 Aug	21	58.6	25.1	2.3	35.5	32.2	0.55	261	319
Mean		34.6	9.7	7.2	23.6	19.5	0.53	268	303

The mean lag time between storm onset and peak streamflow was 3.8 h and ranged from 1.5 to 6.5 h. Events where $T_{rise} \geq 2.5$ hours accounted for 75% of the events. The gross event rainfall had little control on T_{rise}, with a Spearman's correlation coefficient that was not statistically significant ($r_s = -0.52$; $p < 0.02$). Events with long lag times had relatively small volumes of rainfall in comparison to events with short lag times. Rainfall intensity had a stronger forcing on T_{rise} than P, with a Spearman's correlation coefficient that was statistically significant ($r_s = -0.83$; $p < 0.0001$) (Figure 3.5).
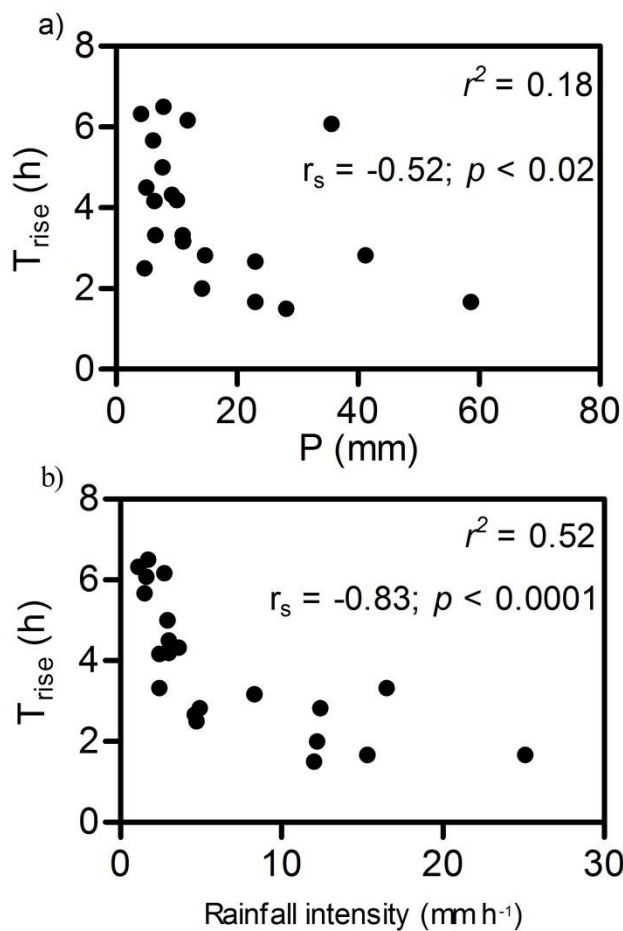
Figure 3.5 Relationship between a) the lag time in peakflow (T_{rise}) and event rainfall (P) and b) T_{rise} and rainfall intensity. r^2 represents the coefficient of determination for the linear regression. r_s represents the Spearman's correlation coefficient.

3.4.3 Antecedent wetness controls on stormflow

The depth equivalent antecedent soil water calculated between the 10 and 100 cm soil layers from all soil pits ranged from 244 – 279 mm with a mean of 265±8 mm (standard deviation) during the wet phase. The influence of the antecedent soil water content on stormflow generation was not significant ($r^2 = 0.001$; $p < 0.89$) and no threshold response was observed. There were two exceptions to this; July 18 and August 19, where two storm events of 23 mm produced substantially different stormflow responses. The July 18 event, with 254 mm of *ASW*, produced relatively small amounts of *QF* (2.8 mm) and *QF/P* (0.12). In contrast, the August 19 event with 277 mm of *ASW* and produced *QF* (15 mm) and *QF/P* (0.65) that were 5 times greater than for the July 18 event (Table 3.1).

When the antecedent soil water content was summed with the event rainfall, a strong linear relationship was observed with *QF* ($r^2 = 0.84$; $p < 0.0001$). Although the linear relationship between *QF* and *ASW* + *P* was strong, it did not differ from that recorded for *QF* vs. *P* (Figure 3.4a). Using the piecewise regression analysis (PRA), we were able to calculate a threshold response for the *QF* and *ASW* + *P* relationship at the convex and concave hillslope. The efficiency criteria of the PRA model including the 95% confidence interval of the minimum and maximum quickflow below (CI_1) and above (CI_2) the breakpoint and the root mean square error (RMSE) of the predicted quickflow are summarized in Table 3.2. The *QF* predicted from the PRA fit well to the observed values. The uncertainty in the predicted *QF* was low as indicated by RMSE <2 mm and the narrow interval of CI_1 and CI_2 (Table 3.2). The uncertainty in the predicted *QF* at the convex hillslopes was less than at the concave hillslopes; however, the difference was not substantial. Likewise, the difference in the threshold breakpoint between sites varied by only 4 mm (Table 3.2; Figure 3.6a-b). We therefore used the mean value from all slope locations to show a breakpoint in the non-linear relationship between *QF* and *ASW* + *P* at 289 mm ± a standard error of 2.3 mm (Figure 3.6c). Below this threshold, events produced less than 2.8 mm of stormflow and a mean *QF/P* of 0.18, while above threshold events produced more *QF* (3.5 – 32.2 mm) with a mean *QF/P* of 0.53 (Table 3.1). The volume of stormflow produced by all events above the threshold (97 mm) was 78% of the total stormflow generated over the wet season (124 mm) from July 7 to September 24. The total stormflow above the threshold also

represented a substantially larger fraction of the total precipitation after activation (23%) than the stormflow below the threshold (6%).
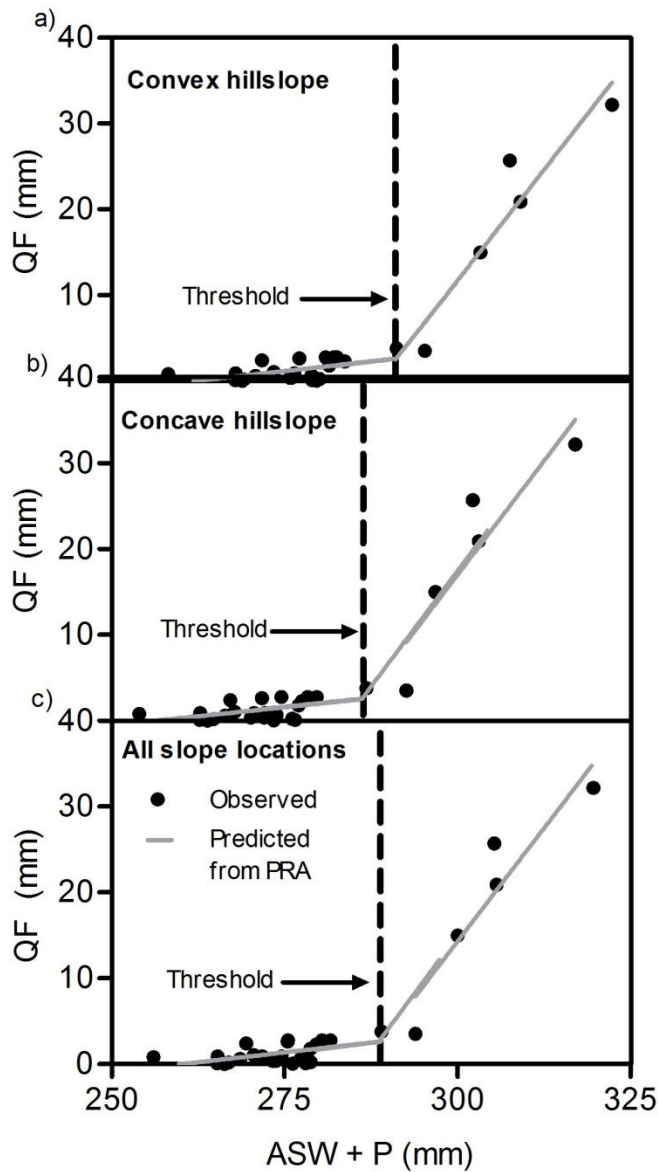
Figure 3.6 Threshold relationship between the quickflow (QF) measured at the primary outflow and a) antecedent water + event rainfall (P) at the convex hillslope, b) antecedent water + P at the concave hillslope and c) mean antecedent water + P from all slope locations. Black circles represent the observed QF and the grey line represents the predicted response from the Piecewise Regression Analysis (PRA)

The mean lag time between storm onset and peak rise in soil moisture at all the instrumented soil layers was shorter for storm events above the 289 mm $ASW + P$ threshold (2.3 h) than for storm

events below this threshold (5.7 h). This was particularly evident at the 50 and 100 cm soil layers. At the 50 cm layer the mean lag time above the threshold (2.5 h) was nearly two times shorter than below the threshold (6.2 h). Likewise, the lag time for storms above the threshold at the 100 cm layer (2.8 h) was four times shorter than for storms below this threshold (11.4 h). The mean lag time between storm onset and peak streamflow for rainfall events above the threshold was 3.2 h, which was 40 and 24 minutes slower than the lag times recorded at the 50 and 100 cm soil layers respectively.

The antecedent precipitation, represented by the cumulative rainfall seven days prior to the storm event, ranged from 2.7 to 79 mm with a mean value of 46.5 mm. Antecedent precipitation had little effect on both R ($r^2 = 0.01$; $p < 0.50$) and QF ($r^2 = 0.01$; $p < 0.65$).

Table 3.2 Assessment of the Piecewise Regression Analysis (PRA) model efficiency used to produce the non-linear response between quickflow and the sum of antecedent soil water and event rainfall

Location	Breakpoint (mm)	β_1	QF_1	95% CI_1	β_2	QF_2	95% CI_2	RMSE (mm)
Convex hillslope	291	0.0798	1.046 2.477	0.273, 1.809 1.159, 4.032	1.175	6.475 34.74	3.853, 9.360 31.61, 37.79	1.595
Concave hillslope	287	0.0723	1.004 2.582	0.068, 1.913 1.078, 4.483	1.220	9.076 35.20	6.154, 12.00 31.39, 39.00	1.900
All slope locations	289	0.0804	1.044 2.511	0.205, 1.860 1.106, 4.243	1.205	7.680 35.08	4.873, 10.61 31.61, 38.45	1.720

The breakpoints derived from the PRA are displayed as the sum of antecedent soil water and event rainfall (mm). β_1 is the slope below the breakpoint. β_2 is the slope above the breakpoint. The minimum and maximum quickflow below (QF_1) and above (QF_2) the breakpoint. The 95% confidence interval of the minimum and maximum quickflow below (CI_1) and above (CI_2) the breakpoint. RMSE is the root mean square error of the modelled quickflow.

3.5 Discussion

3.5.1 Streamflow activation in tropical dry forest

Threshold responses in runoff generation to *VWC* have been observed at other catchments including dry rangeland (Western and Grayson, 1998), temperate humid forest (James and Roulet, 2007; 2009), and steep alpine (Penna *et al.*, 2011b) catchments. A mean threshold *VWC* of 26% at the 100 cm soil layer was necessary to activate streamflow from the catchment in the present study. This threshold *VWC* was similar to 23% recorded by James and Roulet (2007) but substantially lower than the threshold *VWC* of 41 – 46% observed by Western and Grayson (1998) and Penna *et al.* (2011b). Below the *VWC* threshold, streamflow was not generated; however, the increase in soil moisture at the 10, 30 and 50 cm soil layers in the hillslopes indicates that vertical flow processes were active. Although vertical processes dominated during the 52 day transition phase, from June 25 to June 29 the cumulative soil water was larger than cumulative rainfall (Figure 3.3) indicating small amounts of lateral flow in the 10 to 50 cm soil layers, towards the pits. Above the *VWC* threshold, streamflow activation signals the occurrence of lateral flow from the hillslope. Grayson *et al.* (1997) described this *change of state* as a switch in the dominant direction of soil water movement from vertical flow under dry antecedent conditions to lateral flow under wet antecedent conditions. The threshold response from soil pits at the convex and concave hillslopes was highly synchronous, suggesting that hydrological connectivity was achieved across the catchment, connecting hillslopes to streams.

Our current study showed that a response in the *VWC* at deeper soil layers (100 cm) was necessary for streamflow activation. This response was unlike the shallow near-surface *VWC* threshold observed in temperate humid catchments (Western and Grayson, 1998; James and Roulet, 2007; Penna *et al.*, 2011b). These differences likely reflect the variability in the soil properties which affect the mobile soil water that is available for drainage and flow, namely the drainable porosity. Field studies in temperate humid catchments with shallow soils show that drainable porosity often quickly decreases with depth due to changes in bulk density or soil texture (Weiler *et al.*, 2004; 2005). The decrease in drainable porosity with depth allows a small

input of rainfall to elicit a rapid rise in the water table which increases the potential for shallow lateral subsurface flow (Uchida *et al*., 2006). With the shallow soil water near its field capacity, an increase in *VWC* satisfies the field capacity, initiating drainage of mobile soil water and transport along the shallow saturated lateral flow pathways (Penna *et al*., 2011b). Unlike those humid temperate forests, high drainable porosities (0.29 – 0.22) were observed through the profile at all four soil pits in our study site. Although the *VWC* between the 10 and 50 cm soil layers were near or at field capacity, the increase in *VWC* during rainfall inputs remained below the moisture content at saturation, suggesting that transient saturation, which supports saturated lateral flow and the near surface threshold response, does not occur.

With the *VWC* at the 10, 30 and 50 cm soil layers satisfying the field capacity, the excess water can readily drain from these layers. Vertical drainage continues through the profile until the *VWC* at the 100 cm layer is brought to field capacity. Once the unsaturated storage between the 10 and 100 cm soil layers is "filled", the drainable water is "spilled" laterally downslope, activating streamflow. Unlike the saturated fill and spill mechanism observed in many catchments (Spence and Woo, 2003; Tromp van Meerveld and McDonnell, 2006b), the increase in *VWC* at the 100 cm layer was only marginally higher than the field capacity, suggesting that flow was largely unsaturated.

Although most work in steep catchments with incised stream channels and narrow riparian zones indicates that contributions to runoff from near-stream areas are very low (Sidle *et al*., 2000; McGuire and McDonnell, 2010), without direct measurement of riparian groundwater or soil moisture near the stream, we are unable to exclude the possibility that runoff may be generated from near-stream variable source areas as well. In the year prior to our study, shallow wells (1.5 – 2 m) were installed near the location of the soil pits; water tables were not observed to develop at these depths. While transient subsurface saturation was not observed at any of the four soil pits, we are unable to exclude the possibility of its formation as water tables have been shown to be 5 – 10 m below the ground surface in very wet Mexican tropical montane cloud forests (Muñoz-Villers and McDonnell, 2012). Furthermore, with measurement errors in *VWC* and uncertainties in the porosity, it is possible that saturation may have occurred. We suggest that

future work should examine the role of near-stream areas and groundwater in controlling streamflow activation.

The 191 mm of cumulative rainfall prior to streamflow activation recorded at our site falls within the 64-545 mm range observed at other tropical dry forest catchments (Ewel and Whitmore, 1973; Masiyandima *et al*., 2003; Mugabe *et al*., 2007). Although the cumulative rainfall has been used to assess the timing of streamflow activation in tropical dry forests, it is not the most accurate method, as the interannual variation in the frequency, depth, intensity and duration of storm events has been shown to directly affect the rate at which storage deficits are satisfied and consequently the amount of rainfall needed to activate streamflow (Mugabe *et al*., 2007).

The use of the soil water deficit as a metric of streamflow activation accounts for the annual rainfall variability by assuming the deficit is a consistent value that does not vary yearly. By the end of the dry phase the lowest values of *VWC* were recorded (Figure 3.2), indicating the maximum storage deficit. The seven month dry phase is part of the annual cycle in this region (Ortiz-Jiménez *et al*., 2005; Farrick and Branfireun, 2014) and we expect these low and stable *VWC* and maximum soil water deficit to be achieved annually. While the soil storage deficit approach has been shown to be applicable at our current research site, we suggest that this method be tested in conjunction with other alternative hypotheses such as the variable source area at other dry forest catchments, before the storage deficit method can be recommended as the primary method of determining the storage deficit in dry forests across the tropics.

Using the cumulative increase in soil water we were able to estimate the soil storage deficit. Because streamflow was not activated until water percolated 100 cm below the surface we assumed that water loss during this period was likely due to losses by canopy and litter layer interception, deep recharge, lateral flow or evapotranspiration. However, without direct measurements of these water balance components, we are unable to identify the water losses during the wetting up period.

3.5.2 Controls on stormflow runoff generation

Once the storage threshold needed to activate streamflow was satisfied, the stormflow response of individual events was predominantly controlled by the size of the storm event. During the course of the wet season, the *VWC* at all the instrumented soil layers remained near or at field capacity. Under these conditions, the mobile water content that is available for drainage and flow is generally limited by the depth of rainfall and not storage. This is evident in the breakpoint of the non-linear relationship between *QF* and *ASW* + *P*. Above the breakpoint, rainfall inputs greater than 14 mm triggered a rapid and large increase in *VWC* above field capacity. The increased *VWC* was still below total porosity (saturation), suggesting that conditions remained unsaturated. The large input of mobile water was able to displace soil water during drainage. Below the breakpoint, rainfall inputs were on average 24 mm lower than storm events above the breakpoint. Because the increase in *VWC* was smaller, the subsequent displacement and discharge of soil water as quickflow was substantially lower than above the breakpoint. While much of the catchment hydrological literature shows that increasing stormflow and *QF/P* above these threshold breakpoints represents an increase in hydrological connectivity due to the upslope expansion of subsurface saturated areas (Kim *et al*., 2005; Detty and McGuire, 2010; McGuire and McDonnell, 2010), the unsaturated conditions maintained during storm events at our catchment suggest that the increase in *QF/P* from 0.17 below the breakpoint to 0.53 above the breakpoint reflects an increase in the rapid subsurface flow through preferential pathways or the greater displacement of soil water.

The precise mechanism of this displacement and discharge during stormflow and streamflow activation remains unclear. However, similar work by Torres *et al*. (1998), in a catchment with steep slopes and highly permeable soils that remained unsaturated, suggest that rainfall inputs over generally wet soil produces a pressure wave that generates a rapid response and displacement of water in the unsaturated zone. While we do not have the field pressure head data needed to indicate pressure wave translation, the rapid increase in *VWC* above field capacity but still below saturation provides the conditions necessary to cause pressure wave translations through the unsaturated zone. Furthermore, because the maximum rise in soil moisture at the 100 cm soil layer occurred before the peak in streamflow, it supports the rapid pressure wave movement through the soil matrix.

Although it has been shown that the soil moisture data supports the unsaturated flow of water through the soil matrix, it is important to consider a preferential flow mechanism at our catchment, as it widely recognized that subsurface flow can occur through preferential flow pathways that bypass the soil matrix (e.g. Weiler *et al*., 2005; McGuire and McDonnell, 2010). Other work in steep catchments suggests that under higher antecedent wetness and larger storm events, the number of interconnected lateral preferential flow pathways increase across the hillslope, essentially improving connection between upslope areas and the stream (Sidle *et al*., 1995; 2000; 2001). Rapid flow through preferential pathways is supported by observations in other dry (Sandström, 1996) and humid tropical forests (Elsenbeer and Lack, 1996; Negishi *et al*, 2007), which show that large portions of stormflow is generated through soil pipes and macropores. While large macropores and soil pipes were not directly observed, it is likely that streamflow activation after the dry season and stormflow may occur as rapid flow through preferential pathways. Future work in this catchment should test the hypotheses of pressure wave translation and preferential flow.

The minimum rainfall threshold of 4.1 mm needed to generate a measurable increase in runoff was similar to that observed by Fu *et al*. (2013). They attributed the streamflow for events below the minimum threshold to direct precipitation on the stream channel and infiltration-excess overland flow (HOF) over steep slopes. Although steep sloped areas exist at our site, the high surface *K* recorded in these forests (Farrick and Branfireun, 2014) were 2 – 3 times higher than the maximum rainfall intensity during the study period, indicating that HOF is absent over these slopes. While the majority of the catchment is characterised by incised streams and steep slopes, in very limited sections of the catchment, the slope is gentle and the near-stream area wider. The low runoff coefficients observed for events below the rainfall threshold suggests that runoff may be generated from these near-stream areas. We do not have the hydrochemical and isotopic tracer data to support this assumption or the soil moisture results needed to show that rapid saturated overland flow from near-stream areas occurred. In spite of these data deficiencies, studies from other research catchments show that low runoff coefficients indicate stormflow produced from near-stream areas under small rainfall events, (Sidle *et al*., 2000; McGlynn and McDonnell, 2003; Penna *et al*., 2011b).

Earlier work in these forests, which focused on characterizing the surface controls on infiltration, suggests that since high surface hydraulic conductivities allowed more than 70% of the rainfall to percolate through the soil, shallow subsurface flow was the primary runoff generating mechanism across the hillslope (Farrick and Branfireun, 2014). However, because observations at plots and hillslopes are not easily scaled up, the relationship between stormflow and rainfall intensity and the mean lag time to peakflow can provide much insight with regard to the primary runoff generating mechanism at the catchment scale. At catchments where HOF is dominant, a strong, positive linear relationship ($r^2 > 0.9$; $p < 0.0002$) exists between stormflow and rainfall intensity (Martínez-Mena *et al.*, 1998; Cammeraat, 2004). The weak stormflow–rainfall intensity relationship ($r^2 = 0.23$, $p < 0.03$) observed at our site indicates that stormflow was generated by subsurface flow. The lag to peakflow – catchment size relationship developed by Dunne (1973) indicates than an HOF dominated catchment of 3.15 km^2 would produce a mean lag time of 0.52 hours or 31 minutes. This is 7.3 times faster than the mean lag time of 3.8 hours recorded at our site, indicating that HOF was not the dominant runoff mechanism. While these results support a subsurface stormflow generation mechanism, we present the lag time comparisons with caution as topographic (i.e. slope, flow path length, flow path gradient) and other morphological features will strongly influence the streamwater transit time (McGuire *et al.*, 2005).

The lag to peakflow showed a negative relationship with rainfall intensity ($r^2 = 0.52$; $p < 0.0002$). The shorter lag in peakflow observed under higher intensity rainfall is likely the result of the rapid entry and mobilisation of water through the catchment. In these forests the surface hydraulic conductivity can be $2 - 7$ times greater than the maximum rainfall intensity (Farrick and Branfireun, 2014). Because infiltration through the surface is not limited, we can assume that the rate of flow through the soil and discharge to the stream is primarily controlled by the rainfall intensity. The exact mechanism of the transfer is unknown; however, subsurface flow as a result of pressure wave transmission (Torres *et al.*, 1998) or the mixing and discharge of stored water through macropores and large cracks (McDonnell, 1990; Buttle and Turcotte, 1999) has been shown to be strongly influenced by intensity.

The hydrometric evidence provided in our current study indicates that runoff is generated as subsurface stormflow. This response is unlike those observed in most catchments with similar hydroclimatic regimes, where runoff is dominated by HOF (Sandström, 1996; McCartney *et al*., 1998) or SOF (Masiyandima *et al*., 2003; Mugabe *et al*., 2007). Given that the rainfall distribution and extended dry periods are similar among these dry forest catchments, the difference in runoff mechanisms may be attributed to the high soil surface infiltration and percolation rates found in our catchment. While the response at our catchment differs from most studies published to date for dry forest regions, the highly permeable soils of volcanic origin that characterises our catchment are distributed worldwide, with more than 60% located in tropical countries (Takahashi and Shoji, 2002) suggesting that the observed runoff response may occur over much of the tropics.

3.5.3 Streamflow activation and stormflow generation under future climate change

The streamflow produced from our research catchment and other tropical dry forest regions are in many cases the main source of water for agricultural systems and many wetlands and small lake systems (Farrick and Branfireun, 2013). Therefore, understanding the impact of future climate change on streamflow production is imperative. Using a regional climate model for Mexico and Central America, Karmalkar *et al*. (2011) indicate that over the next 30 to 50 years, these regions may experience a 13 to 27% decrease in wet season rainfall. Given that stormflow in our catchment was strongly controlled by the depth of rainfall, the projected decrease in precipitation will likely result in a substantial reduction in the volume of stormflow in this catchment.

From the PRA, strong linear relationships were observed between quickflow and *ASW + P* both below and above the breakpoint. Using these two linear relationships it is possible to demonstrate how changes in rainfall will alter the runoff generated. As this study was conducted in a year with 24% less rainfall, the derived linear relationship may not account for the changing frequency and size of storm events, which can alter stormflow generation. We suggest that

stormflow from this catchment be monitored over multiple years in order to account for climatic variability, before examining the long-term impacts of climate change.

3.6 Conclusion

We examined the processes that govern streamflow activation and stormflow generation in a tropical dry forest catchment, México. Our results show that two different controls were responsible for streamflow activation and stormflow generation. Unsaturated soil water storage was the main control on streamflow activation, while the gross event rainfall depth was the dominant control on stormflow generation. The change in the dominant control from unsaturated storage to rainfall depth suggests that once streamflow is activated, the storage deficit become low enough that less rainfall goes into storage and more is being translated to runoff. These results stress the importance of using a combined storage-rainfall threshold approach when examining stormflow generation at the catchment scale. The subsurface stormflow runoff mechanism observed during this study is unlike those observed in most arid and tropical wet forests, where runoff is dominated by infiltration excess overland flow and saturated overland flow. This illustrates the importance of characterising the specific runoff generating mechanism for a given catchment.

Our findings have important implications with regards to the ecological and human systems that are supported by these dry forest catchments. Runoff produced from this and other dry forest catchments is the primary water source to lake and wetland systems and is important for agriculture through direct extraction and shallow ground water recharge. The expected reduction in stormflow volume under the projected change in rainfall will reduce the supply of water and jeopardise the functioning of these systems. These results are therefore important to the mitigation and adaptive strategies needed for these regions and should strongly be looked at by land managers and policy developers.

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Chapter 4

4.0 Flow pathways, source water contributions and water residence times in a Mexican tropical dry forest catchment

4.1 Introduction

Most of our current understanding of runoff generation processes in tropical systems has been produced from research in lowland and montane catchments of the humid tropics (Bonell and Bruijnzeel, 2005; Levia *et al*., 2011; Farrick and Branfireun, 2013). While it is generally recognised that rapid flow processes dominate runoff in forested tropical catchments, the specific water flow pathways, source areas and residence times of stream water often remain unclear (Buttle and McDonnell 2004; Bonell and Bruijnzeel, 2005). This is especially true for tropical dry forests, where most research has often focused on quantifying the catchment scale water balance (de Araújo and González Piedra, 2009; Montenegro and Ragab, 2010). Understanding the water flow pathways in a catchment is necessary for the management of surface and groundwater resources. This is particularly important in tropical dry forests where land use change (Miles *et al*., 2006) coupled with the projected decrease in precipitation (Bates *et al*., 2008) are expected to reduce the already limited streamflow observed in these catchments (Farrick and Branfireun, 2014b).

In most humid tropical forest catchments, runoff is characterised by the rapid translation of rainfall to runoff. Stormflow has been shown to be composed of 40 to 81% event water (Schellekens *et al*., 2004; Goller *et al*., 2005), most of which is translated downslope as saturation-excess overland flow (SOF) (Elsenbeer *et al*., 1994; 1995b), return flow (RF) through soil pipes (Schellekens *et al*., 2004; Negishi *et al*., 2007) or through shallow lateral pathways near the soil surface (Schellekens *et al*., 2004; Goller *et al*., 2005). These studies show that the hydraulic properties of the shallow subsurface soil, often determines the dominant runoff mechanism. Shallow confining soil layers with low hydraulic conductivities (*K*) impede vertical

flow through the highly permeable surface soils, leading to shallow subsurface and SOF generation (Bonell and Gilmour, 1978; Elsenbeer and Vertessy, 2000; Godsey *et al*., 2004).

In the semi-arid tropics, geochemical tracer studies typically show that storm runoff can be composed of up to 75% event water (Sandström, 1996; Hughes *et al*., 2007; Ribolzi *et al*., 2007). However, unlike the humid tropics, the low surface K in the semi-arid tropics often limit infiltration, resulting in most runoff being generated as infiltration-excess overland flow (HOF) (Bonell and Williams, 1986).

Most concepts regarding hydrological connectivity and variable source areas have originated primarily from research in steep humid temperate forest catchments (Bracken *et al*., 2013). These studies often show that connectivity between the riparian zone and hillslope, is needed to generate substantial amounts of subsurface flow (Bracken and Croke, 2007) and the relative contribution from either source often varies on a seasonal (Ocampo *et al*., 2006; McGuire and McDonnell, 2010) or event basis (McGlynn *et al*., 2003; Subagyono *et al*., 2005). In geographical regions where a distinct dry-wet season occurs, hillslope and riparian areas often remain hydrologically disconnected for extended periods of the year. As rainfall and antecedent wetness increases, the upslope expansion of saturated subsurface areas, often through a rise in the riparian and hillslope water table, connects these two landscape units (Ocampo *et al*., 2006). The improved connectivity results in a shift in dominant source areas from the riparian zone to the hillslope (Ocampo *et al*., 2006; Jencso *et al*., 2009). In wetter temperate catchments, with a more even annual rainfall distribution, hillslope – riparian connectivity is affected by the size of the storm event. Under small rainfall inputs, connectivity is low and most runoff is generated from the riparian zone, while large rainfall events improve connectivity with most runoff generated from the hillslope (McGlynn *et al*., 2003; Subagyono *et al*., 2005). Although hillslope – riparian connectivity is important for stormflow generation in temperate forests, in humid tropical forests where SOF and RF are the dominant mechanisms, hydrological connectivity develops by surface drainage expansion. Zimmerman *et al*. (2014) showed that as antecedent wetness increased, SOF was generated at progressively higher upslope positions, which drained into ephemeral channels, essentially expanding the size of the source area contributions to runoff.

Isotopic residence time analyses have emerged in the last two decades as an important tool that can provide insights into hillslope runoff processes. As stream water residence time is strongly influenced by topographic (McGuire *et al*., 2005) and internal catchment features such as soil depth and subsurface geology (Soulsby *et al*., 2006; Katsuyama *et al*., 2010), it provides an excellent indication of the coupling among the flow paths, water sources and storage in a catchment (McGuire and McDonnell, 2006). Although the use of isotopic residence time analyses in catchment-scale hydrology has increased, it has generally been limited to humid temperate catchments (McGuire and McDonnell, 2006). Though Buttle and McDonnell (2004) suggest the use of residence time techniques in tropical forests catchments as a means to improve the understanding of the translation of rainfall to stream water, application of these techniques have been limited to very few studies (e.g. Crespo *et al*., 2012).

In this study we report the research on rainfall-runoff response of a steep, tropical dry forest catchment with highly permeable soils. High hydraulic conductivities, high soil porosities and soil moisture response in deep soil layers suggested that runoff in this catchment is generated as subsurface flow through the displacement of stored water in the near-saturated or saturated zone (Farrick and Branfireun, 2014a; b). In order to test this hypothesis, the objective of this work is to use a combined hydrometric, isotopic and geochemical approach to examine the source areas of stream water, dominant flow pathways, and the timing of the translation of rainfall into runoff.

4.2 Study area

The study was conducted in a 3.15 km^2 catchment in the lake Zapotlán watershed, approximately 100 km south-southwest of Guadalajara, Jalisco, Mexico; 5 km northeast of Ciudad Guzman, Jalisco, Mexico (19ºN 103ºW) (Figure 4.1). The climate is Tropical Savannah (Köppen-Geiger: Aw) with a distinct wet and dry season (Peel *et al*., 2007). The average annual precipitation (1972 – 2003) is 813 mm, of which 95% falls between June to September (Ortiz-Jiménez *et al*., 2005). Rainfall is dominated by short duration, low intensity storm events (Farrick and Branfireun, 2014a). The strong wet-dry seasonality results in intermittent streamflow production

from the catchment, with most flow occurring from July to October (Farrick and Branfireun, 2014b). Mean annual temperature is 19.6ºC with maximum temperatures occurring in July (Ortiz-Jiménez *et al*., 2005).
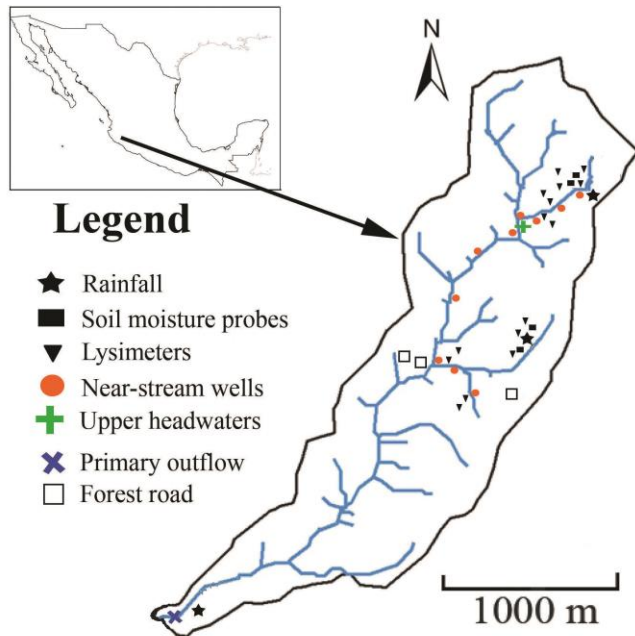
Figure 4.1 Location of the study site and isotopic and geochemical water sampling locations across the catchment

Elevation ranges from 1557 metres above sea level (masl) at the primary outflow channel to 2170 masl at the headwater sub-basin. The catchment is steep with slopes ranging from 18º to over 52º. The study area is underlain by Pleistocene andesitic basalt-basaltic andesite and volcanic fine tuff. The channel width ranges from <0.20 m in the headwater sub-basins to 1.0 – 1.5 m at the primary outflow channel. The stream channels are deeply incised and steep with a 0.2 – 1 m wide riparian areas. The bedrock along the incised channels is weathered and highly fractured. The soil is classified as chromic cambisols with andic properties of volcanic origin (Gómez-Tagle, 2008). The soil at the hillslope is deep and were >1 m deep except for on or near limestone bedrock outcrops, which are common. Soil textures are mainly loams and sandy soils and vary from sandy loams in the O and upper A horizons to loams and sandy-clay loams (>40% sand) at depths below 50 cm (Farrick and Branfireun, 2014a). The surface hydraulic conductivity is highly variable, ranging from 9 – 164 mm h^{-1} (Farrick and Branfireun, 2014a). Bulk density in

the upper 100 cm of soil ranges from 0.91 – 1.1 g cm^3 (Farrick and Branfireun, 2014a) while the total porosity and drainable porosity range from 0.57 – 0.60 and 0.22 – 0.29 respectively (Farrick and Branfireun, 2014b).

The catchment is dominated by two distinct forest types. A highly heterogeneous mixed deciduous forest (dominated by *Carpinus caroliniana*, *Mimosa adenantheroides*, but with a complex mix of understorey and herbaceous vegetation) occurs at elevations between 1600 – 1800 masl and covers 13% of the catchment. A pine-oak forest (almost exclusively *Pinus montezumae, Quercus laeta*) occurs at elevations greater than 1800 masl and occupies 82% of the catchment. Land development in the catchment is low with fragmented agricultural plots and unpaved roads occupying 4 and 1% of the catchment area respectively.

4.3 Methods

4.3.1 Hydrometeorological measurements

Open field precipitation was measured from 1 May to 24 September 2012 at 10 minute intervals using a Texas Electronics tipping bucket rain gauge installed at three locations across the catchment at 1600, 1800 and 1950 masl (Figure 4.1).

Volumetric water content (VWC) was measured at four soil pits located along two hillslopes using Campbell Scientific Inc. CS615 Water Content Reflectometers from 1 May to 24 September 2012 (Figure 4.1). The reflectometers were inserted horizontally in the soil pits at depths of 10, 30, 50 and 100 cm below the surface at all four locations. The reflectometers were calibrated in the laboratory using soil extracted from the same area following the technique of Stenger *et al*. (2005). Antecedent soil water was calculated as the depth equivalent of soil water (mm) prior to storm event using the methods of Farrick and Branfireun (2014b). Antecedent precipitation was calculated as the sum of rainfall 30 days prior to the storm event.

The stream water level at the primary outflow channel (Figure 4.1) was recorded at 10-minute intervals, using a 0.8 mm resolution odyssey capacitance water level logger (Dataflow Systems

Ltd.). Modification of the channel at the outlet of the catchment occurred at some point in the past; large rocks and boulders were removed and an artificial wall constructed and backfilled to produce a small impoundment and free fall structure. Discharge was calculated from the water level using the end-depth method (Jain *et al*., 2007). This method was selected as the stream fit the criteria required to accurately measure discharge: free fall where the drop is greater than the stream stage, rectangular, smooth channel without rocks or boulders. Discharge was calculated as:

$$Q = C\sqrt{g}\ bh^{3/2} \hspace{4cm} (4.1)$$

Where Q is the discharge (m^3/s), C is the coefficient of discharge, g is the acceleration due to gravity, b is the channel width (m) and h is the water level (m). We confirmed the accuracy of the discharge measurements at lower flows by conducting manual stage-discharge relationships using a stopwatch and buckets.

Storm runoff events were defined as the period from the initial rise in discharge from a local minimum in the hydrograph to the next local minimum and were separated into quick flow (QF) and delayed flow (DF) volumes using the local minimum method (Sloto and Crouse, 1996). The gross event rainfall depth (P) was calculated as the sum of the 10 minute values over the duration of the storm event. The hydrologic behaviour of the catchment was examined during the study period using the ratio of quick flow to rainfall (QF/P).

The lag time between storm onset and peak streamflow (T_{rise}) was calculated as the time difference (hours) between the start of rainfall and peakflow (Mosley, 1979). The lag time between storm onset and peak soil moisture response at the 10, to 100 cm layers was calculated as the time difference (hours) between the start of rainfall and the peak increase in VWC. The lag time between storm onset and peak near-stream groundwater was calculated as the time difference between storm onset and the peak rise in water table. The lag time was determined for all rainfall runoff events during the wet season.

4.3.2 Isotope and geochemical water sampling

4.3.2.1 Rain and baseflow sampling for residence time analysis

Bulk rainfall and baseflow under non-storm conditions were collected weekly from July to September, 2012. Rainfall was sampled from three 20-L HDPE buckets with a 20 cm diameter funnel from the upper headwater basin (2000 masl), half basin mark (1800 masl) and the primary outflow channel (1600 masl). To prevent changes to the isotopic ratio of rainwater as a result of isotopic fractionation due to evaporation, a 2 cm thick mineral oil layer was added to the bucket. Baseflow was collected on the same day and the same locations as bulk rainfall samples.

4.3.2.2 Geochemical sampling from lysimeters, wells, seeps and baseflow

Water samples were collected every two weeks from various soil water lysimeters, near-stream wells and seepage from exposed hillslopes along the forest roads (Figure 4.1). Zero-tension lysimeters were installed at five convex and five concave hillslopes at depths of 10, 30, and 50 cm below the soil surface. Lysimeters were also installed at a depth of 100 cm but failed to collect water, likely as a result of poor contact between the lysimeter and soil. The lysimeters were constructed from a 20 x 10 x 5 cm, high density polyethylene (HDPE) container, which was inserted horizontally in the soil and drained into a 1000 ml HDPE bottle. Near-stream sampling wells were constructed from a 5 cm (inner diameter), slotted PVC tube and screened (250 μm Nitex®) along the entire length. The wells were installed approximately 1 m from stream, 40 – 60 cm below the surface. Sample water was collected from the lysimeters and wells using a hand operated suction pump. Baseflow samples were collected from the upper basin and primary outflow, usually weekly.

4.3.2.3 Rainfall-runoff event sampling

The hydrological response of the catchment to storm events was examined during four storms, representing a range of antecedent wetness and rainfall characteristics. During these events

samples of rainfall and stream water was intensively collected for stable isotope (δ^2H and δ^{18}O) and geochemical (major anions and cations) analysis.

Composite rainfall samples were collected for each event in a 20 L high-density polyethylene (HDPE) bucket with a 20 cm diameter funnel placed near the tipping bucket gauges (Figure 4.1). Stream water at the primary outflow channel (1600 masl) was collected during storm events using an automatic water sampler (Model 6700, Teledyne ISCO, Inc). The auto-sampler was programmed to start sampling 1 h before the onset of rainfall to include a sample of pre-event baseflow. Stormflow was sampled at constant sampling intervals (30 to 120 min depending on the magnitude of the event) and collected in an individual 1000 ml glass bottle at each sampling interval. Water samples were collected from the field within 24 h of an event.

4.3.3 Isotopic and geochemical storage and analysis

All isotope samples were stored in a 20 ml HDPE vials with displacement caps, while all geochemical samples were collected in 100 ml HDPE bottles and refrigerated until they were filtered within 48 h of collection. Isotopic samples were analysed for δ^{18}O and δ^2H using cavity ring-down spectroscopy (L2120-*i*, Picarro, Inc.). The isotope values are reported in permil (‰) relative to the Vienna Standard Mean Ocean Water (VSMOW). The precision of the δ^{18}O and δ^2H measurements was 0.1‰ and 0.5‰ respectively. All geochemical samples were vacuum filtered with of 0.45μm nylon filter within 48 hours into a 60 ml HDPE (Wilde *et al*., 2004.) and frozen until laboratory analysis. Samples were analysed for dissolved organic carbon (DOC) and a suite of anions and cations. Ions were measured using Ion Chromatography at the Biotron analytical services laboratory, Western University, using a Dionex ICS-3000 (anions) and Dionex ICS-1600 (cations). Dissolved organic carbon was measured using a Picarro iTOC.

4.3.4 Residence time modelling

The mean residence time of stream water was estimated using the sine wave approach, using the assumption that water fluxes in the catchment are under a steady state. This method was selected due to the short length and coarse frequency of spatial and temporal tracer sampling (Tekleab *et al*., 2014). The seasonal trends in the rainfall and stream water were modelled using periodic

regression analysis (Bliss, 1970) to fit the sine wave curve to the annual δ^{18}O variations in rainfall and stream water. The predicted δ^{18}O can be calculated as:

$$\delta^{18}O = X + A[\cos(ct - \theta)] \tag{4.2}$$

Where δ^{18}O is the modelled δ^{18}O (‰) composition, X is the weighted mean annual measured δ^{18}O (‰), A is the annual amplitude of the measured δ^{18}O in rainfall, c is the radial frequency constant (0.051502 rad d^{-1}), t is the time in days after the start of the sampling period and θ is the phase lag of predicted δ^{18}O in radians. The radial frequency constant was modified from the original 0.0174214 rad d^{-1} designed for a 365 day flow cycle to 0.051502 rad d^{-1} to fit the 122 day flow period observed in this catchment. Sine wave models fitted to the rainfall and stream water δ^{18}O variations were used and the mean residence time (T) was calculated as:

$$T = c^{-1}\left[\left(\frac{A_{Z2}}{A_{Z1}}\right)^{-2} - 1\right]^{0.5} \tag{4.3}$$

Where A_{Z2} is the amplitude of stream water δ^{18}O, A_{Z1} is the amplitude of rainfall δ^{18}O and c is the radial frequency of annual fluctuations defined in equation (4.2). The mean transit time was also calculated by this method, substituting δ^{2}H for δ^{18}O. The overall performance of the sine wave model was evaluated using the root mean square error (RMSE).

4.3.5 Isotopic hydrograph separation

A one tracer, two component hydrograph separation (Sklash and Farvolden, 1979) was conducted to partition the storm runoff into pre-event (water stored in the catchment prior to the storm event) and event (direct water input into the catchment) water sources. The technique involves a mass balance approach using δ^{18}O or δ^{2}H as a tracer and can be described using the following mixing equations:

$$Q_t = Q_p - Q_e \tag{4.4}$$

$$Q_t C_t = Q_p C_p + Q_e C_e \tag{4.5}$$

Where Q_t, Q_p and Q_e represent current streamflow, pre-event and event water volumes, respectively and C_t, C_p and C_e are the corresponding concentrations of δ^{18}O or δ^{2}H isotopes (‰ VSMOW). The tracer concentration of the base flow one to two hours prior to the storm event was used to represent C_p (Sklash and Farvolden, 1979). The C_e was determined as the mean

isotopic composition of the storm rainfall. The contributions of event and pre-event water to total runoff can be determined by combining equations 4.4 and 4.5 as:

$$Q_p = Q_T \left(\frac{C_t - C_e}{C_p - C_e} \right)$$
(4.6)

The uncertainty associated with the calculated fractions of event and pre-event water was evaluated using the technique of Genereux (1998).

4.3.6 Topographic analysis

A 15 x 15 m digital elevation model was used to calculate the topographic features of the catchments using System for Automated Geoscientific Analyses (SAGA). The stream network and catchment area was calculated using the multiple flow direction algorithm in SAGA. The computed stream network and catchment elevation data was used to delineate the sub-basin area of the upper headwater basin and half basin mark. Other topographic attributes such as topographic index, slope and flow path length were computed and used as metrics of internal catchment form. These values were correlated with the mean residence at each catchment to examine possible relationships. The results of these analyses are presented in Table 4.1.

Table 4.1 Catchment characteristics of the stream water sampling locations used in residence time analysis

Sampling location	Elevation (masl)	Catchment area (km^2)	Mean slope (deg)	Maximum slope (deg)	Mean flowpath length (m)
Upper headwater basin	2000	0.35	15	37	226
Half-basin mark	1800	1.41	18	43	201
Primary outflow	1600	3.15	24	52	157

4.4 Results

4.4.1 Seasonal hydrometeorological conditions

The total rainfall from July to September, 2012 was 429 mm. The highest monthly rainfall was recorded in July (212 mm) and August (196 mm) and the lowest was recorded in September (50 mm). The total seasonal streamflow was 119 mm or 27% of the wet season rainfall (Figure 4.2). The soil moisture at all depths remained near or at field capacity over the wet season, with a mean daily VWC over the upper 100 cm of soil from all pits of 32.7±0.9%. The mean daily near-stream water table position at all locations during the wet season was 0.63±0.13 m below the surface. Surface saturation in the near-stream zone only occurred during the largest storm event (58.6 mm) of the wet season, with groundwater levels rising above the surface. However, surface saturation was only recorded in four of the ten wells.

Over the wet season, from July to September, 21 storm events produced runoff volumes greater than 1 mm. The mean QF/P was 0.26 and ranged from 0.04 to 0.78. The stormflow hydrographs were generally flashy with a rapid rise and recession. The mean lag time between storm onset and peak streamflow (T_{rise}) was 3.8 h and ranged from 1.5 to 6.5 h.

The lag time in the response between storm onset and peak soil moisture response varied according the depth of the measured soil layer. Soil moisture at 10 cm and 30 cm peaked before streamflow, with mean lag times of 1.9 and 3.4 h respectively. The response at the 50 and 100 cm layers were more variable. The mean lag time at the 50 cm layer was 5.0±3.7 h, while the mean lag time at the 100 cm layer was 8.4±8.0 h, which was two times slower than the T_{rise}. The lag time at the 100 cm layer was strongly influenced by the depth of rainfall. Events greater than 14 mm produced a mean lag time of 2.8 h at the 100 cm, which was 24 minutes faster than the corresponding T_{rise} (3.2 h). Events less than 14 mm had a mean lag time of 10.8 h, two times slower than the corresponding T_{rise} (4.0 h).
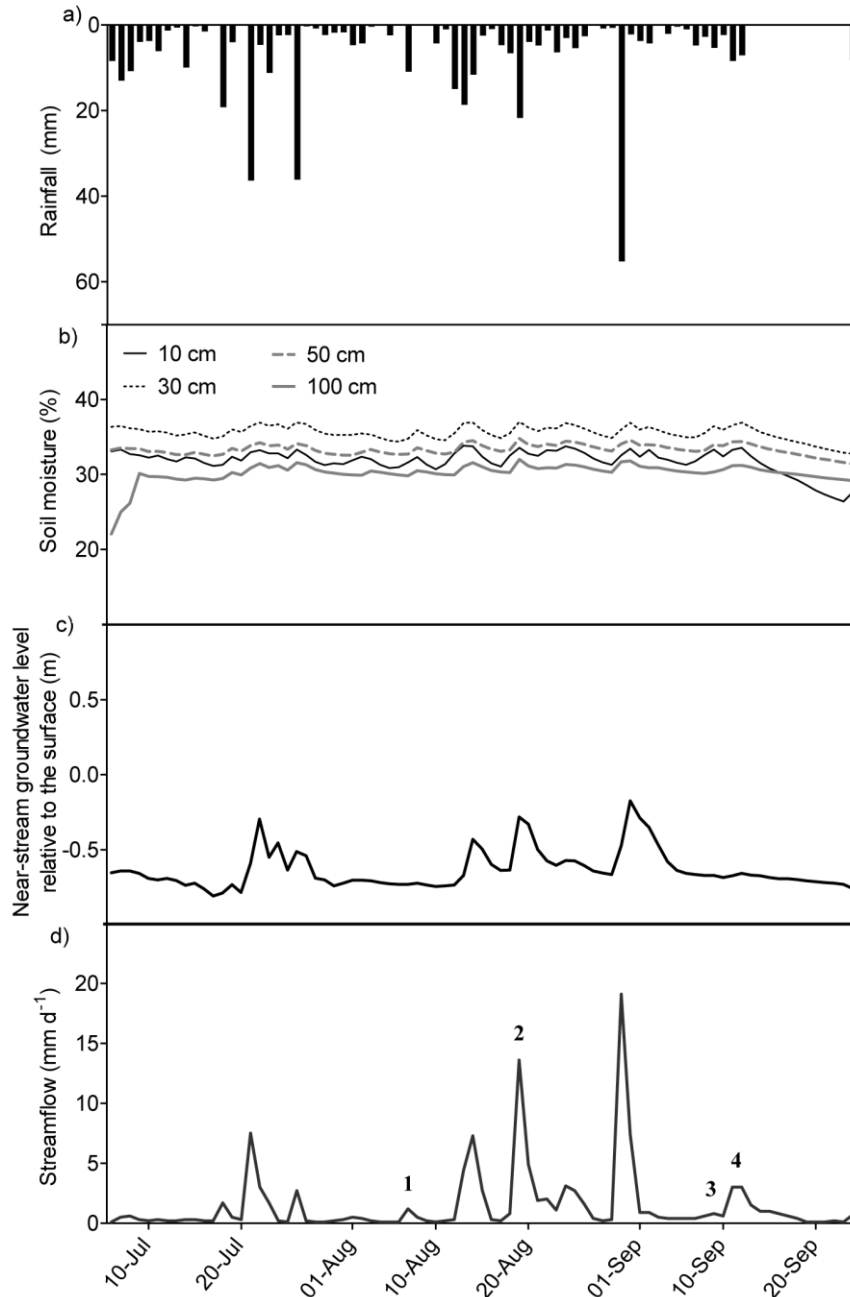
Figure 4.2 Daily change in a) rainfall, b) soil moisture, c) near-stream groundwater and d) streamflow from July to September, 2012. The numbers on figure 4.2a represent the storm events sampled for water isotopes and geochemistry.

The mean lag time between storm onset and peak rise in near-stream groundwater was 3.6±2.6 h. Unlike the 50 and 100 cm soil moisture response, the lag time in the response of the near-stream groundwater were not strongly affected by the depth of rainfall. The mean lag time for events

greater than 14 mm were 3.3 h, which were slightly slower than the 3.0 h response recorded for storms less than 14 mm.

4.4.2. Stream water residence time

The seasonal isotopic ratios from rainfall and stream water plotted on the local meteoric water line (LMWL) and showed no evidence of evaporative enrichment (Figure 4.3). The weighted mean value of $\delta^{18}O$ in rainfall over the entire catchment was -8.7‰ and the arithmetic mean was -8.1‰. Rainfall ranged between -14.4‰ to -4.7‰ for $\delta^{18}O$, while δ^2H values ranged from - 103.7‰ to -23.8‰. Rainfall samples were increasingly depleted in both $\delta^{18}O$ and δ^2H as the altitude of the sampling location increased. Mean values of -7.4‰ for $\delta^{18}O$ and -48.3‰ for δ^2H were measured at 1600 masl and decreased to values of -8.9‰ for $\delta^{18}O$ and -57.7‰ for δ^2H at 2100 masl. The isotopic composition of stream water was less variable than the rainfall. The mean weighted value of stream water at the primary outflow was -9.3‰ for $\delta^{18}O$ and ranged from -9.9‰ to -8.3‰ for $\delta^{18}O$. Streamflow samples showed a similar depletion in the isotopic signature with increasing altitude. Mean arithmetic values of -8.9‰ for $\delta^{18}O$ and -60.4‰ for δ^2H were measured at the primary outflow (1600 masl) and decreased to values of -9.4‰ for $\delta^{18}O$ and -64.1‰ for δ^2H at the upper headwater basin (2000 masl).
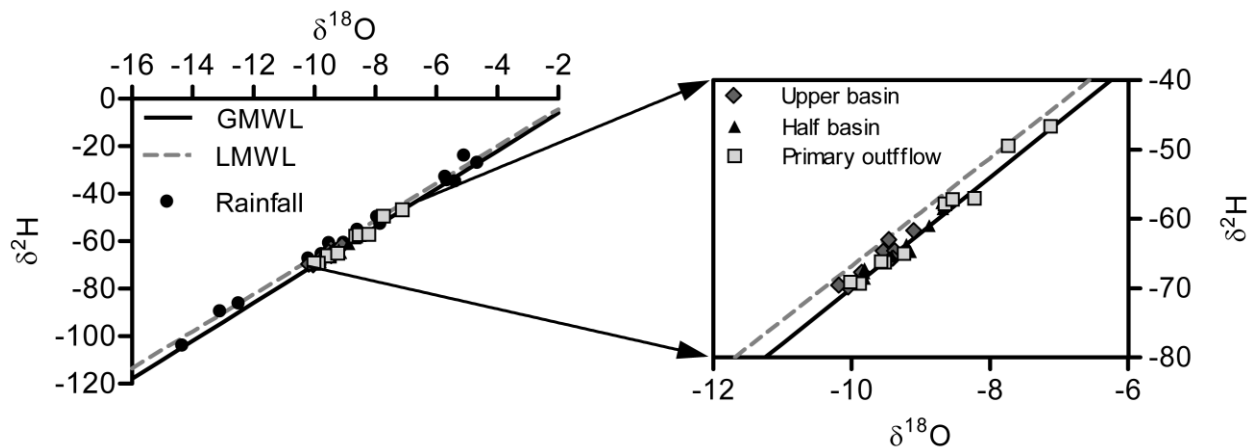
Figure 4.3 δ^2H and $\delta^{18}O$ signatures of rainfall and stream water. The insert shows the isotopic signatures of stream water from the upper basin, half basin mark and primary outflow.

The stream water isotope composition strongly followed the pattern observed for rainfall albeit with a strongly damped signature (Figure 4.4). The modelled $\delta^{18}O$ in stream water at the primary outflow, half basin mark and upper headwater basin fit well to the observed isotope values, with RMSE of 3.1-3.4‰ for rainfall and 0.5-1.0‰ for stream water (Figure 4.4). The mean stream water residence time at the primary outflow calculated from the $\delta^{18}O$ sine wave curve was estimated at 52 days. The mean residence times at the half basin mark and upper headwater basin were longer at 105 and 110 days respectively. The estimated residence time using δ^2H did not differ significantly from $\delta^{18}O$ with times of 48, 91 and 116 days estimated at the primary, half basin mark and upper headwater basin respectively.
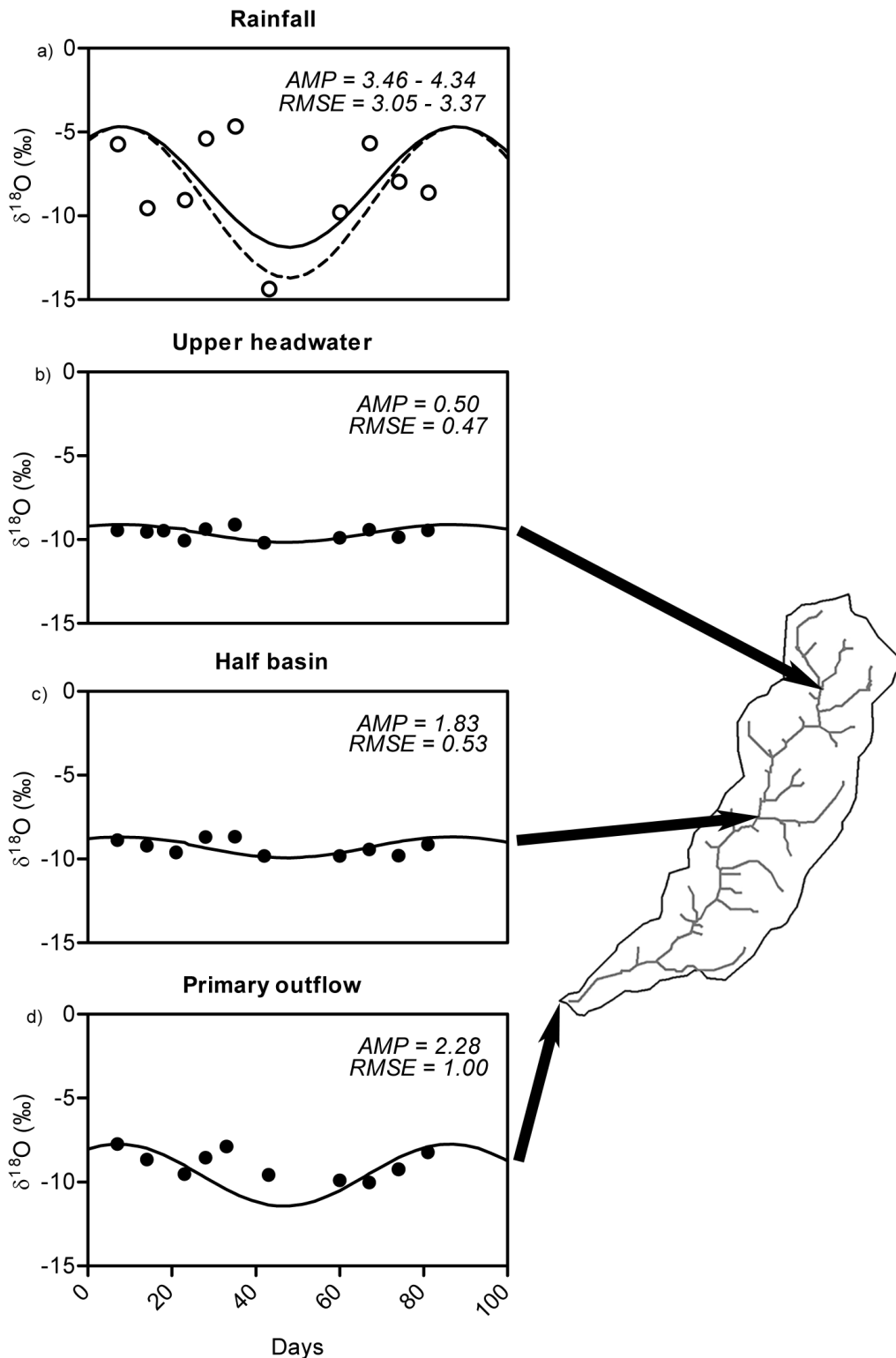
Figure 4.4 Sine wave regression models for $\delta^{18}O$ in rainfall (the solid line represents the weighted rainfall and the dashed line represent the actual rainfall), and streamwater at b) upper headwater basin, c) half basin mark and d) primary outflow. The amplitude (AMP) and root mean square error (RMSE) of the modelled isotopic signature is included

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The mean residence time had a strong negative relationship with catchment area, with residence time decreasing with increasing catchment area ($r^2 = 0.90$). Mean residence time had a similarly negative relationship with mean catchment slope ($r^2 = 0.91$) and strong positive relationship with the mean flow path length ($r^2 = 0.79$). The mean topographic index did not vary substantially among the catchments (8.4 to 8.6) and strongly influenced the mean residence time at each sampling location ($r^2 = 0.80$).

4.4.3. Water geochemistry

The geochemistry of stream water varied seasonally and as well as between locations. The concentrations of Ca^{2+}, Mg^{2+}, Na^+ and K^+ at both the upper basin and primary outflow fluctuated over the sampling period, but showed a general increase as the wet season progressed from July to September (Figure 4.5). The concentrations of the major ions at the primary outflow were 2 to 3 times higher than baseflow at the upper basin, but exhibited similar ratios (Table 4.2). Piper diagrams derived from the cation-anion distribution indicate that baseflow from both stream sources fall on a line trending from $Ca-HCO_3$ type waters (Figure 4.6).
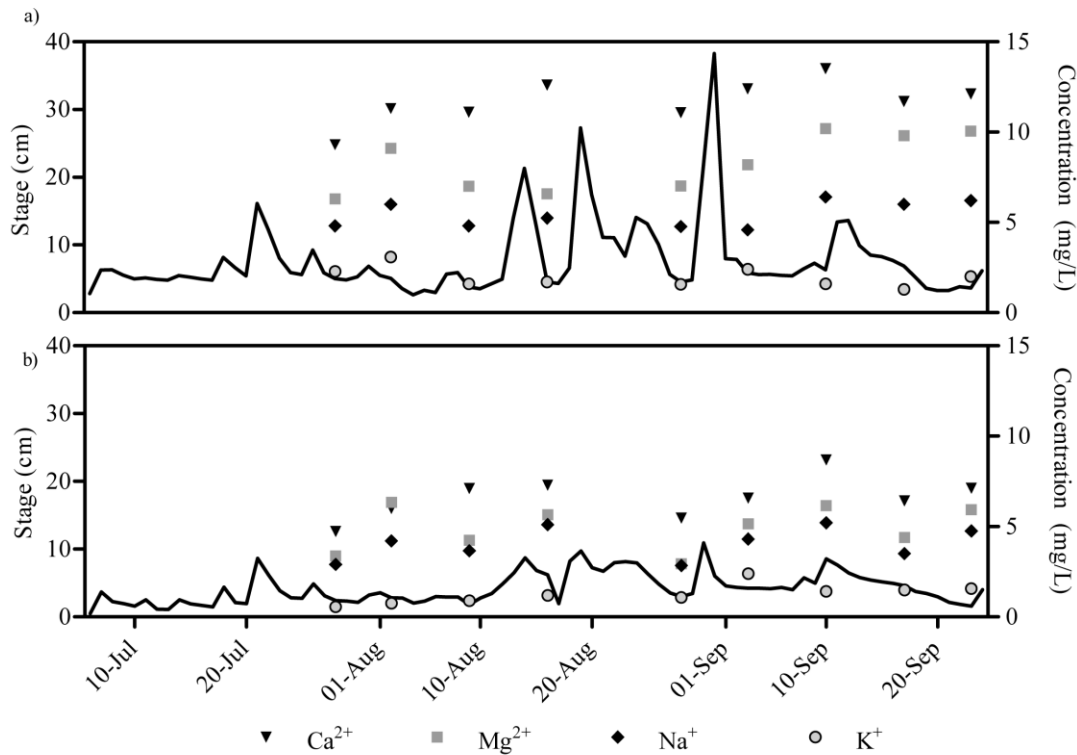
Figure 4.5 Seasonal variation in calcium, magnesium, sodium and potassium concentrations baseflow at a) the primary outflow and b) upper headwater basin

The concentrations of Ca^{2+}, Mg^+ and K^+ in the near-stream groundwater were greater than the concentrations recorded from the baseflow at both stream sources; however, Na^+ concentrations from near-stream groundwater were less than baseflow. Most near-stream groundwater were dominated by carbonate mineral dissolution, falling along a similar Ca-HCO_3 facies trend lines as baseflow (Figure 4.6). Soil water from the 10 cm lysimeter had the highest DOC (range from 10 to 52 mg/l) and decreased as the wet season progressed (45.9 – 15.8 mg/l). Concentrations of Ca^{2+} and K^+ were also highest in water collected from the upper 10 cm of soil and decreased with sampling depth (Table 4.2).
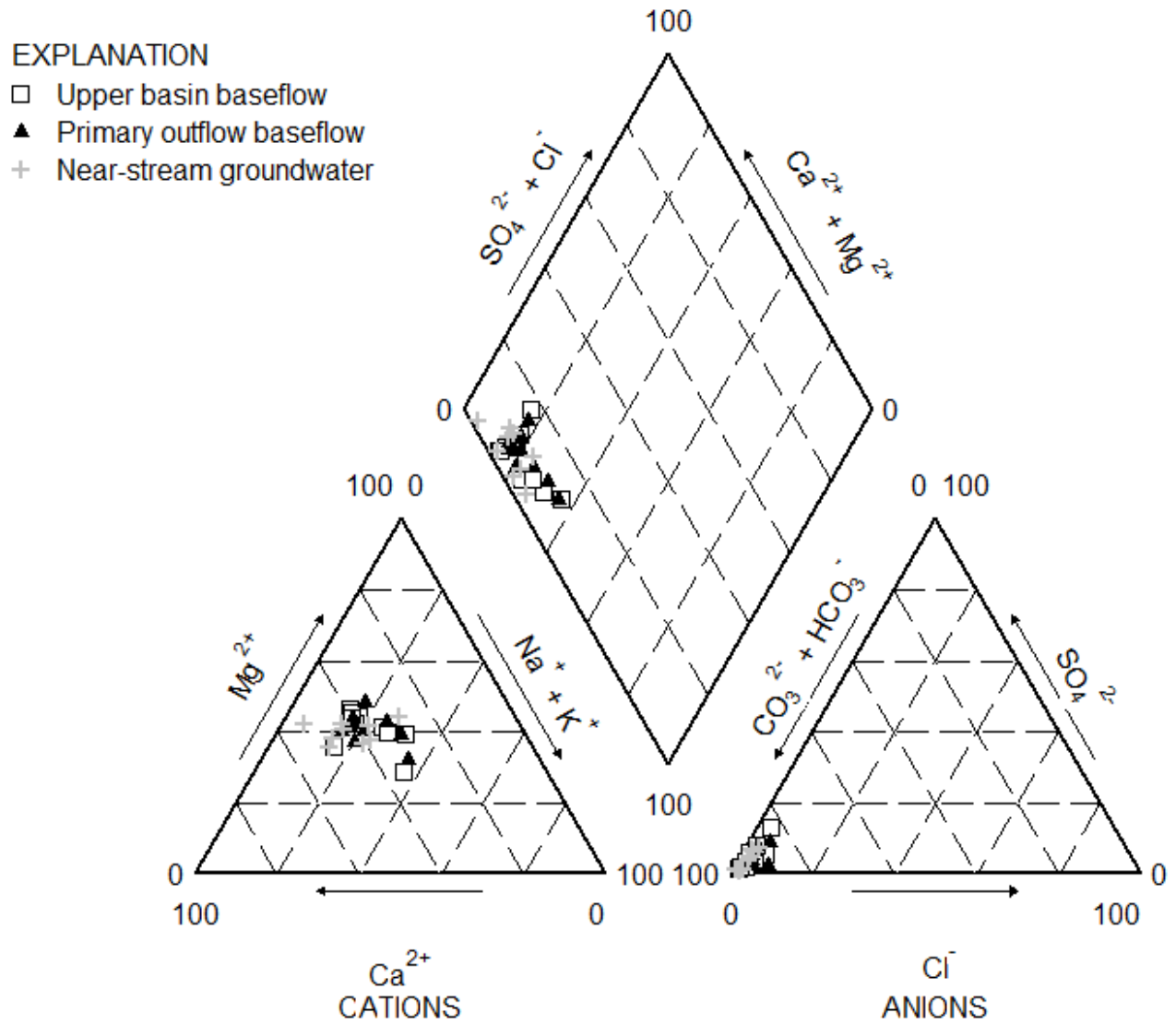
Figure 4.6 Piper plot of the major ion chemistry from the baseflow and near-stream groundwater from the study catchment

Table 4.2 Mean concentrations of the major cations and anions of rainfall, soil water, seepage from the forest road, near-stream groundwater and baseflow from the upper headwater basin and primary outflow. Values in parentheses are the standard deviation

Component	Ca^{2+} (mg/l)	Mg^{2+} (mg/l)	Na^+ (mg/l)	K^+ (mg/l)	Cl^- (mg/l)	SO_4^{2-} (mg/l)	DOC (mg/l)
Rainfall	0.51 (0.20)	0.15 (0.06)	0.20 (0.02)	0.25 (0.12)	0.17 (0.05)	0.69 (0.21)	0.79 (0.32)
Lysimeter (10 cm)	20.2 (10.9)	3.75 (1.26)	2.45 (1.40)	5.90 (0.43)	1.07 (0.96)	4.93 (3.40)	28.0 (15.9)
Lysimeter (30 cm)	5.86 (3.54)	4.37 (1.94)	2.56 (1.57)	3.00 (0.86)	1.28 (0.50)	7.59 (4.01)	14.5 (3.95)
Lysimeter (50 cm)	6.49 (2.51)	4.44 (1.97)	3.27 (1.80)	2.85 (1.01)	0.94 (0.40)	8.48 (4.25)	6.23 (1.50)
Seep from the forest road	5.50 (1.43)	4.28 (1.11)	4.93 (1.42)	0.95 (0.19)	0.98 (0.11)	4.27 (0.64)	4.20 (1.52)
Near-stream groundwater	12.6 (3.42)	6.55 (2.05)	3.89 (1.18)	2.88 (1.05)	1.55 (0.50)	2.93 (1.86)	9.85 (8.30)
Baseflow (upper headwater basin)	6.57 (1.21)	4.91 (1.24)	4.03 (0.88)	1.26 (0.54)	1.83 (0.21)	2.03 (0.50)	3.16 (1.43)
Baseflow (primary outflow)	11.9 (0.83)	8.30 (1.48)	5.58 (0.73)	1.90 (0.60)	2.85 (0.93)	7.96 (1.77)	4.24 (1.24)

4.4.4 Isotopic hydrograph separation and water geochemistry during stormflow

Of the 21 storm events, seven were sampled for hydrograph separation. However, three events were discarded because they failed to capture the entire stream hydrograph. The event characteristics of the four storm events investigated in detail are summarised in Table 4.3. Storms 1, 3 and 4 had rainfall inputs <12 mm and QF/P ratios that ranged from 0.13 to 0.25. These storms were marked by increasing antecedent soil water. The first storm (August 7) had the driest antecedent soil water conditions (287 mm) and the smallest pre-event contributions (72% for $\delta^{18}O$). Antecedent soil water conditions increased for the two remaining small storms. Although pre-event contributions for storm 3 on September 9th (95% for $\delta^{18}O$) and storm 4 on September 11th (79% for $\delta^{18}O$) were higher than storm 1, pre-event contributions did not show a proportional increase with antecedent soil water. For these small storm events, most of the event water contributions (18% for $\delta^{18}O$), occurred during the recession limb (Figure 4.7a; 4.7c; 4.7d). Storm 2 (August 19) was generated under 23 mm of rainfall and produced a QF/P ratio of 0.65. The highest seasonal antecedent soil moisture over the upper 100 cm of soil was recorded prior to storm 2 (Table 4.3, Figure 4.7). Despite the rapid rise of the hydrograph, stormflow generated under this 23 mm event was overwhelmingly dominated by pre-event water (97% for $\delta^{18}O$) over the entire hydrograph (Figure 4.7b).
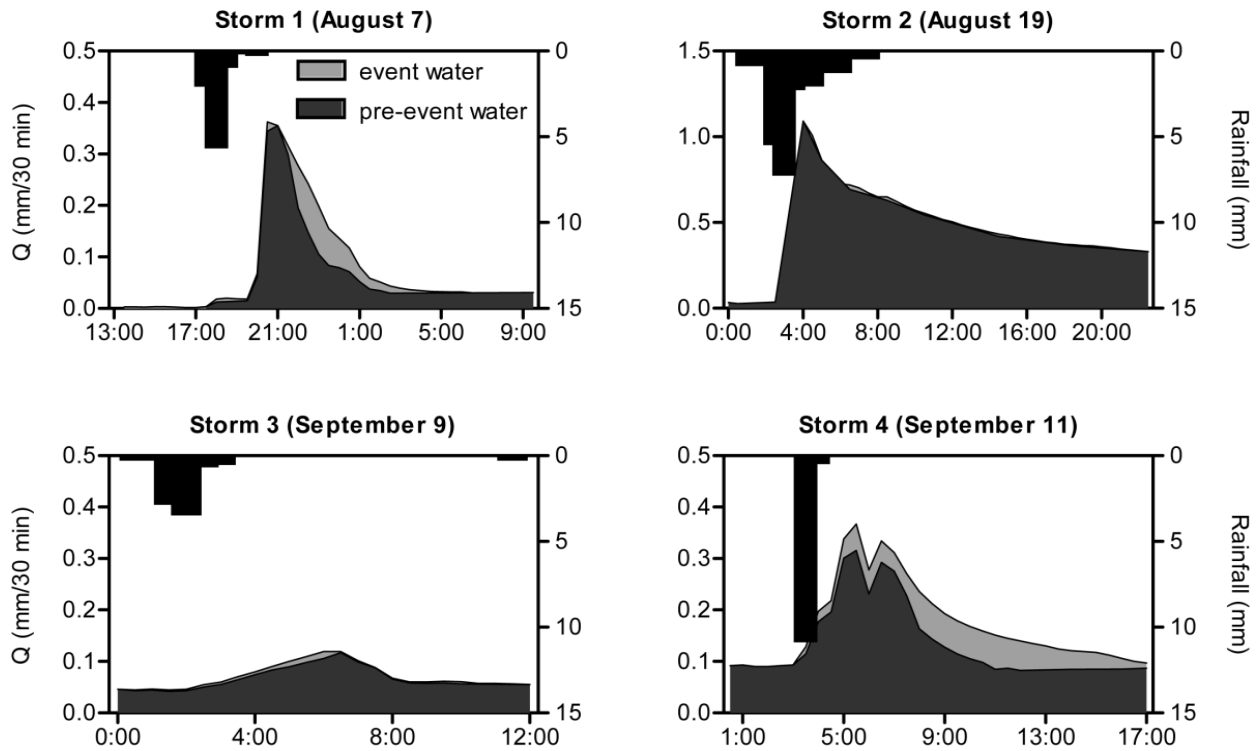
Figure 4.7 The portioning of stormflow into its event and pre-event water sources using a one-tracer two component hydrograph separation analysis with $\delta^{18}O$ as the tracer. Note that storm 2 is plotted with a different y-axis scale because of the significantly higher discharge.

Table 4.3 Storm event characteristics of the four monitored stormflow events

	Storm 1	**Storm 2**	**Storm 3**	**Storm 4**
Date	7 Aug	19 Aug	9 Sep	11 Sep
P **(mm)**	12	23	8	11
Rainfall intensity (mm/h)	2.7	4.6	2.9	16.5
Rainfall duration (h)	4.3	5.0	2.7	0.7
QF **(mm)**	2.4	15	1	2.8
QF/P	0.20	0.65	0.13	0.25
Peak discharge (m^3/h)	0.61	1.64	0.19	0.57
Antecedent soil water (mm)	287	310	299	302
30 day antecedent precipitation (mm)	173	188	192	196

We examined the behaviour of the stream water geochemistry over the course of the four storm events and describe the changes in the cation concentration during stormflow:

1. Storm 1 (August 7). The concentration of K^+ increased during the initial increase in storm runoff from to 1.60 mg/l to 3.28 mg/l, which was most similar to the concentration at the 10 cm lysimeter. A similar increase in Ca^{2+} and Mg^{2+} were observed during this period (Figure 4.8a). After the initial peak, the concentrations of K^+, Ca^{2+} and Mg^{2+} decreased during the rising limb to concentrations most similar to that of baseflow from the upper headwater basin. The recession limb was marked by a small increase in the concentration of Ca^{2+}, Mg^{2+} and Na^+. The concentrations of Ca^{2+} during the recession limb was similar to that of baseflow from the primary outflow and near-stream groundwater, while , Mg^{2+} and Na^+ continued to reflect the concentrations from upper basin baseflow. Potassium remained relatively consistent during the recession limb (Figure 4.8a).

2. Storm 2 (August 19). The concentration of Ca^{2+}, Mg^{2+}, Na^+ and K^+ decreased from 12.6, 6.62, 5.89 and 1.74 mg/l to 10.8, 5.30, 5.11 and 1.55 mg/l respectively at peakflow (Figure 4.8b). These concentrations during the rising limb were most similar to the baseflow concentrations from the upper headwaters recorded three days earlier. The concentrations of K^+, Mg^+ to Na^+ ions gradually increased nine hours into the recession limb, still reflecting concentrations from the upper basin. Ca^{2+} behaved similarly to the other ions (Figure 4.8b); however, the increase in Ca^{2+} during the recession limb was greater (2.7 mg/l), occurring more rapidly than the other ions. Calcium concentration during storm runoff recession reflected the concentrations from the primary outflow (Figure 4.8b).

3. Storm 3 (September 9). The concentration of Mg^{2+}, Na^+ and K^+ were relatively consistent over both the rising and recession limbs at 9.45 to 10.69 mg/l for Mg^{2+}, 6.25 to 6.76 mg/l for Na^+ and 1.56 to 1.98 mg/l for K^+ (Figure 4.8c). The mean concentration of Mg^{2+} and Na^+ during this storm was more concentrated than storms one and two. Calcium concentrations increased during the rising limb, reaching a maximum value of 15.6 mg/l one hour before peak stormflow, which was most similar to concentrations from the near-stream groundwater and primary outflow baseflow (Figure 4.8c). Calcium concentrations remained high during peakflow and quickly decreased during the recession limb.

4. Storm 4 (September 11). Mg^{2+}, Na^+ and K^+ concentrations remained relatively stable over the duration of the storm event (Figure 4.8d). Only during the recession limb was a decrease in Mg^{2+} (3 mg/l) and Na^+ (1.15 mg/l) recorded. Calcium increased from 11.7 to

14.9 mg/l during the first peak in stormflow to values similar to primary outflow baseflow and near-stream groundwater (Table 4.2). Calcium concentrations decreased during the initial stormflow recession, but rapidly increased 2.1 mg/l during the second peak (Figure 4.8d). High Ca^{2+} concentrations were maintained two hours into the recession limb and then rapidly decreased to baseflow concentrations observed at the start of the storm event.
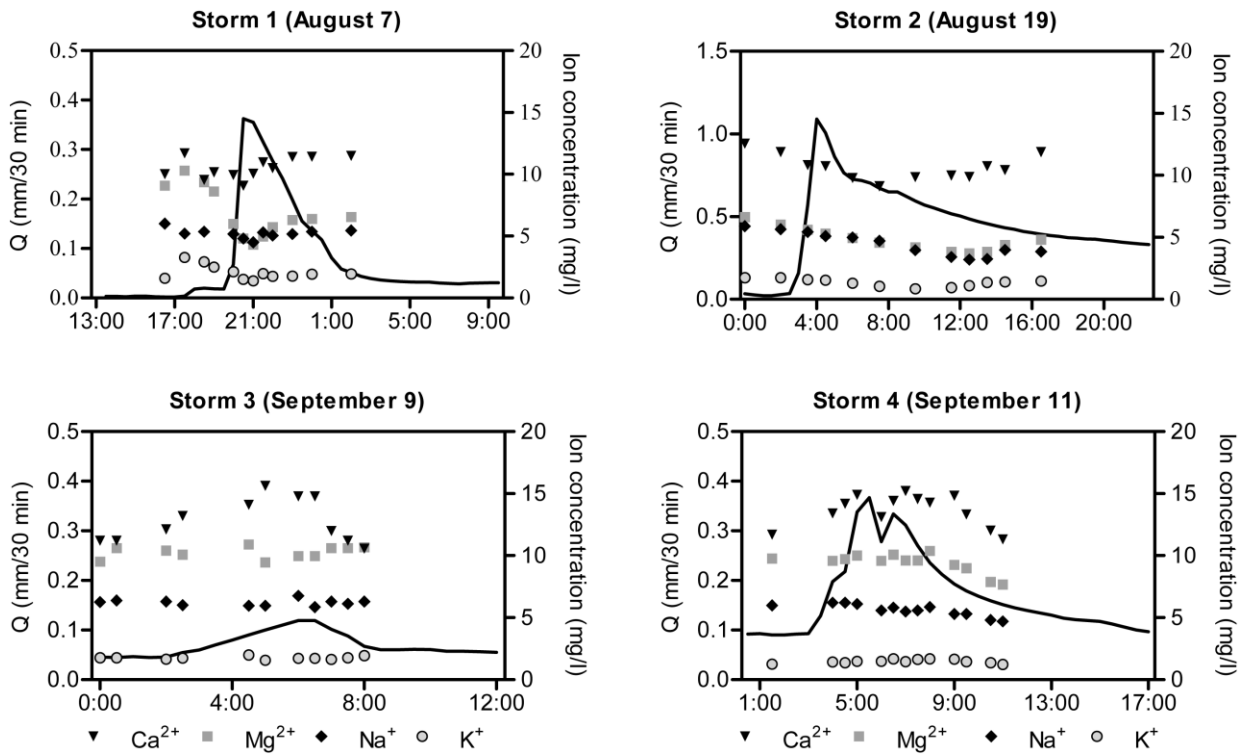
Figure 4.8 Concentrations of calcium, magnesium, sodium and potassium in the storm water for the storm events on a) August 7, b) August 19, c) September 9 and d) September 11. Note that storm 2 is plotted on a different scale

4.4.5. Soil moisture and near-stream groundwater response to rainfall inputs

We examined the soil moisture and near-stream groundwater response to rainfall during the four monitored events. Because of the general rapid response at the 10 and 30 cm layers we focused on characterising the response at the 50 and 100 cm layers, which showed greater variability in the response. The peak in soil moisture at the 50 cm layer for storms one and three occurred four

and two hours into the recession limb respectively. Although storm four (September 11) also had similar rainfall input, soil moisture at 50 cm layer peaked one hour before peak stormflow. At the 100 cm layer, the maximum increase in soil moisture occurred 10, 23 and 0.5 hours after peak stormflow for storms one, three and four respectively. A 1 – 2% increase in soil moisture was recorded for these events. For storms one, three and four, a 4 – 7 cm increase in groundwater was recorded before peak stormflow.

The response of the soil moisture and near-stream groundwater recorded during storm two (August 19) was the opposite of that observed under the three smaller storms. The maximum increase in soil moisture occurred 44 and 40 minutes before peak stormflow at the 50 and 100 cm layers respectively. During this 23 mm event, a 36 cm increase in the near-stream groundwater was recorded during the recession limb.

4.5 Discussion

4.5.1. Stream water residence time across the catchment

To our knowledge, this study represents one of the first reported estimates of stream water residence time in tropical dry forests. The calculated residence times in our study were shorter than most studies, reflecting the shorter subsurface contact time and intermittent nature of streamflow at our site. An earlier study investigating the storage and rainfall controls on streamflow activation indicated that 191 mm of rainfall over a 25 day period was needed to produce consistent discharge from the catchment (Farrick and Branfireun, 2014b). The 52 – 110 day stream water transit time was longer than this 25 day wetting up period, supporting the hypothesis by Farrick and Branfireun (2014b) that streamflow in this catchment is produced from the displacement of water stored in deeper soil layers.

The estimated stream residence time was not uniform across the catchment. The decrease in the length of the residence time from the upper basin to the primary outflow suggests that a different runoff delivery mechanism may exist at lower elevations. The geochemistry of stream water from both the upper basin and primary outflow presented in the Piper plots falls along the same

Ca-HCO$_3$ facies (Figure 4.6). Piper plots provides a useful tool in characterising source water, as they are designed to show the essential chemical character of water according to the relative concentrations of the dissolved ions (Piper, 1944). These geochemical observations suggest that a similar subsurface water source contributes to baseflow across the catchment, which is also well mixed.

It is important to note that the residence time not only reveals information about the source areas, but often reflects the length of contact time of mobile water with those source areas (DeWalle *et al*., 1997; Wolock *et al*., 1997). The difference in stream water residence time likely reflects changes in topographic features across the catchment. The catchment area upstream of the upper basin sampling location was characterised by a mean slope of 15º and a mean flow path length of 226 m, while the lower half of the catchment was marked by an increase in slope (24º) and substantial reduction in flow path length (157 m). The increase in slope and reduction in flow path length effectively increases discharge of subsurface water from the hillslope to the stream, resulting in the shorter stream residence time at the primary outflow. These findings are similar to work by McGuire *et al*. (2005) who demonstrated that the mean residence time in a steep humid forest catchment was strongly controlled by the internal form and structure of the catchment and not the basin area.

While important, it is unlikely that surface topography is the only factor governing residence time distributions in our catchment. In other studies in steep forested catchments, greater bedrock permeability allows a more rapid translation of stored subsurface water to the stream, resulting in shorter residence times (Asano *et al*., 2003; Katsuyama *et al*., 2010). Highly fractured bedrock embedded into the sidewalls of the incised stream channels have been observed in the lower half of the basin. The extent of the fractured rock upslope from the stream channel is currently unknown; however, if this fractured rock does indeed extend upslope, it may allow the more rapid transfer of subsurface water from the hillslope to the stream, further explaining the reduced length of the stream residence time at the primary outflow. Future research in this catchment should therefore focus on the impacts of additional catchment features such as bedrock permeability and soil depth on residence time. Stream water samples should also be collected at

higher spatial resolutions in order to better identify where substantial change in the catchment hydrology occurs.

Our results are presented with caution given the short sampling period of the study, which may underestimate the residence time (Muñoz-Villers and McDonnell, 2012). Furthermore, even though the application of the sine wave method is often appropriate for studies such as ours where the frequency of sampling is coarse and the length of the study is short (Tekleab *et al*., 2014), this approach works under the assumption of a steady state system, which McGuire and McDonnell (2006) state is almost always violated. We recognise that given the intermittent nature of streamflow in our catchment, a steady state is not achieved over most of the year. To reconcile this we assumed that steady state was achieved once streamflow became consistent and then modified the radial frequency constant to fit the 122 day streamflow cycle. The estimated residence time while preliminary seems to fit the model well (Figure 4.3) with the root mean square errors falling within the range of other studies which used the sine wave approach (e.g. DeWalle *et al*., 1997; Soulsby *et al*., 2000).

4.5.2. Subsurface stormflow in a tropical dry forest

The rapid rise and recession of the stormflow hydrograph observed in our tropical dry forest catchment was typical of the response observed in most tropical forest catchments (Elsenbeer *et al*., 1995a; Goller *et al*., 2005). Despite this rapid response, stormflow was overwhelmingly dominated by pre-event water (75 – 98%). The high pre-event contributions recorded from our catchment challenges the long held observation that stormflow in humid and semi-arid tropical catchments are composed of small (30 – 40%) fractions of pre-event water. In humid and semi-arid tropical catchments, the rapid translation of rainfall to runoff through surface and near-surface pathways reduces the mixing and displacement of old water stored in groundwater or deeper subsurface soil layers, resulting in low pre-event water in stormflow (Schellekens *et al*., 2004; Goller *et al*., 2005). Given the strong relationship between high event water and shallow flow pathways, the large pre-event water contributions in stormflow at our catchment suggests that storm runoff is likely generated from deep subsurface flow pathways.

In most geochemical studies of runoff generation in tropical forest catchments, a large and rapid increase of K^+ is typically observed during peak stormflow (Elsenbeer *et al*., 1995b; Sandström, 1996). The high K^+ is often the result of flow over surface or through shallow subsurface pathways, which is enriched with potassium from litter/organic matter (Elsenbeer *et al*., 1995a; Schellekens *et al*., 2004). The highest K^+ recorded in the lysimeters within the upper 10 cm of soil at our research site suggests a similar leaching from litter and organic matter. However, the enriched K^+ from this soil layer was not observed in the water sampled over the majority of the stormflow hydrographs, suggesting that runoff was not a result of flow over the surface or through shallow subsurface pathways. The depleted K^+ in stormflow was most similar to the mean concentrations of K^+ recorded in baseflow from the upper headwater basin and primary outflow channel. Likewise the Ca^{2+}, Mg^{2+} and Na^+ concentrations recorded during stormflow generally reflected the ion chemistry of the near-stream groundwater and baseflow from the primary outflow and upper headwater basin. As baseflow is defined as the portion of flow that originates from delayed subsurface flow or groundwater (Tallaksen, 1995), we suggest that the strong baseflow signature in stormflow originates from near-saturated subsurface soil water or transient groundwater sources. This hypothesis is supported by the geochemistry of the underlying bedrock. The andesitic-basalt/basaltic-andesite that characterises this region is composed of K^+, Na^+, Mg^{2+} and Ca^{2+} ratios of 1:4:5:7 (Luhr and Carmichael, 1980; 1981). The baseflow from the upper basin and primary outflow and stormflow show similar mean geochemical ratios of 1:3:5:7. Despite these results, we lack the hillslope groundwater tracer and hydrometric data to support direct groundwater contributions and suggest that future work in this catchment should test for groundwater contributions to runoff.

Previous work in this catchment has suggested that given the rapid increase in soil moisture above field capacity but below saturation during storm events, that storm runoff may be generated as mass or pressure wave translation through the unsaturated zone (Farrick and Branfireun, 2014b). Torres *et al*. (1998) suggests that when soil is in the near-zero head pressure range, rainfall inputs can elicit a small increase in pressure head and a large increase in hydraulic conductivity. The resulting increase in the hydraulic gradient and hydraulic conductivity generates pressure waves which produces a rapid response, displacement and discharges water in the saturated zone. For storms two (August 19) and four (September 11), the rapid increase in

soil moisture at the 50 and 100 cm soil layers, coupled with large contributions of pre-event water, low K^+ in storm water and Ca^{2+}, Mg^{2+} and Na^+ concentrations reflective of near-stream groundwater and baseflow sources, suggest that the displacement of stored water from the saturated zone occurs during these larger storm events.

Although high pre-event contributions and Ca^{2+}, Mg^{2+}, Na^+ and K^+ reflective of near-stream groundwater and baseflow were also recorded during storms one (August 7) and three (September 9), the soil moisture response at depths below 50 cm lagged behind peak streamflow, suggesting a different runoff mechanism. In a catchment with soils of similar volcanic origin, Muñoz-Villers and McDonnell (2012) showed that despite a delayed response in the soil moisture at depths below 70 cm, runoff was composed of 72 – 99% pre-event water. Muñoz-Villers and McDonnell (2012) attributed the high pre-event to vertical preferential flow in near-stream areas, which bypasses the soil matrix and displaces near-saturated soil water or hillslope groundwater. Such a mechanism may occur for these two events as the maximum increase in near-stream groundwater occurred before peak stormflow. However, despite this rapid increase, the geochemistry reflective of near-stream sources was not observed until the recession limb (Storm 1) or just before peakflow (Storm 3), suggesting delayed contributions. It is important to note that these sampling wells were located in the upper half of the catchment (Figure 4.1) and the delayed contributions are likely a result of transport times from these locations.

Caution must be exercised when attributing runoff generation to direct contributions from the near-stream area. Chanat and Horberger (2003) suggest that substantial hillslope and groundwater contributions may be masked by the higher ion concentrations in the near-stream or riparian zone. In our current study, this masking effect likely exists during storm one (August 7). Water collected at the 50 cm lysimeter had Ca^{2+} and Mg^{2+} values two times lower than the near-stream water (Table 4.2). Even if this hillslope water was discharged during the rising limb, the higher Ca^{2+} and Mg^{2+} in the near-stream water may mask hillslope contributions.

4.5.3. The influence of catchment wetness on source area contribution

We found that while stormflow was composed of water with Ca^{2+}, Mg^{2+}, Na^+ and K^+ concentrations most similar to baseflow, near-stream groundwater and soil water from the 50 cm lysimeter, seasonal variations in the contributing area of the catchment was observed. During storms one (August 7) and two (August 19) the rising limb was characterised by Ca^{2+} concentrations from the upper basin, while Mg^{2+} concentrations observed over the entire hydrograph was characteristic of baseflow from the upper basin. By September, runoff from storms three (September 9) and four (September 11) were dominated by water with higher concentrations of Ca^{2+} and Mg^{2+} which were most similar to baseflow at the primary outflow. Stream water at the primary outflow should represent a mixture of water draining from all the sub-basins within the catchment. If this is indeed the case, then storm runoff that reflects the geochemistry of baseflow from the primary outflow likely represents contributions from across the entire catchment.

The changing seasonal geochemical signal in storm runoff from the upper basin to the primary outflow suggests that as the wet season progressed and catchment wetness (represented by the 30 day precipitation, Table 4.3) increases, the proportion of the catchment which contributes to stormflow also increases. In a humid temperate catchment with a similar dry-wet seasonality and narrow riparian zones, Sidle *et al*. (2000) showed as the rainy season progressed and catchment wetness increased, there was an increase in the number of linked zero-order basins. This connectivity was brought about through the expansion of preferential flow networks. While we do not have the hydrometric data to show the upslope expansion subsurface saturation, our geochemical data suggests that increasing contributions from near-saturated subsurface soil water or groundwater sources to streamflow may occur as catchment wetness increased. From July to September, there was an increase in Ca^{2+} and Mg^{2+} in baseflow at the primary outflow by 2.8 and 3.8 mg/l respectively. As these are the dominant minerals in the underlying bedrock, it is reasonable to assume that the increasing concentrations represent greater contributions from saturated subsurface areas as the wet season progressed.

This study provides preliminary evidence of seasonal changes in hydrological connectivity. However, without characterising the geochemistry of a combination of large and small storm events in July and September and the current absence of soil and groundwater end-members from the lower half of the basin, we are unable to completely identify and compare changes in source area contributions over the entire wet season. Therefore future studies in this catchment should investigate this hypothesis of seasonal hydrological connectivity, ensuring a full range of storms are captured.

4.6 Conclusion

We examined the runoff generation mechanisms in a tropical dry forest catchment, Mexico. We found that over the four monitored storm events, runoff was dominated by pre-event water. The geochemistry of these storms strongly reflected the baseflow and deep subsurface soil water source waters in the catchment. The combined isotope-geochemical tracer and hydrometric analysis suggest that despite the rapid rise and recession of the storm hydrograph, shallow flow processes do not control the runoff response in this catchment. Although the runoff response at our catchment is unlike that of most arid and humid tropical forests, it is very similar to the limited work describing runoff generation in tropical catchments of highly permeable soils of similar volcanic origin.

Although preliminary, this study provides evidence that where a strong dry-wet seasonality occurs, hydrological connectivity is seasonally and not event driven. The sub-basins at higher elevations are important water sources to runoff, particularly during the early part of the wet season, when most runoff originated from the headwaters. These findings have important implications with regards to land management in tropical dry forest catchments. Much of the current extent tropical dry forest in Mexico and Central and South America is under threat of land use change, mainly due to agricultural conversion. Decision and policy makers are often faced with the task of selecting the appropriate area for development. Our current research suggests that development in the headwater sub-basins should be avoided given the potentially large contributions to runoff.

4.7 References

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Chapter 5

5.0 Summary and conclusions

5.1 General summary

Understanding the controls that govern the storage and discharge of rainfall as runoff has never been more important, as many forested catchments are coming under increasing stress due to climate and human induced changes. These changes are expected to have a strong impact on the quantity of water discharged from the catchment. Studies in humid temperate, wet tropical and dry, semi-arid forested catchments have shown that the relationship between rainfall intensity and hydraulic conductivity and the development of soil water repellency at the soil surface play a significant role in dictating infiltration rates and magnitudes and ultimately runoff generation. The rate by which rainfall infiltrates and percolates through the unsaturated soil layers to the saturated zone is often essential in satisfying storage deficits. Only after these deficits are satisfied is a threshold relationship between rainfall and discharge developed. In many cases the rainfall threshold reflects the level of hydrological connectivity that is achieved across the catchment. In humid temperate forests where subsurface flow is dominant, large rainfall events are often needed to connect hillslope and riparian areas, thereby generating substantial volumes of runoff is often generated in the catchment. In wet tropical forests and semi-arid systems where runoff is generated as overland flow, connectivity does not occur through hillslope-riparian linkages, but rather through the expansion of ephemeral stream channels. While infiltration characteristics, storage-rainfall thresholds and hydrological connectivity are well researched in other forested catchments, they have generally remained undescribed in tropical dry forests. The overall goal of this dissertation was to improve our understanding of the controls on runoff generation and streamflow response in a tropical dry forest catchment.

The climatic conditions and waxy leaved, drought-resistant vegetation typical of tropical dry forest ecoregions are most similar to those that characterise semi-arid regions. In semi-arid systems, low hydraulic conductivities and extreme levels of water repellency limit infiltration, resulting in infiltration-excess overland flow. Given that little is known about the surface

hydrology of tropical dry forests, chapter two tested the hypothesis that the controls on runoff in semi-arid systems are transferable to tropical dry forest hillslopes. The results showed that the low rainfall intensities, high surface hydraulic conductivities and lack of repellent surfaces during the wet season resulted in more than 70% of annual rainfall percolating through the upper 30 cm of soil. The main conclusion from chapter two was that in spite of similar climate and vegetation regimes, hydrological knowledge from semi-arid catchments is not transferable to this tropical dry forest.

The finding that infiltration was not limited by the physical surface properties suggested that subsurface flow, not infiltration-excess overland flow is the dominant process in this catchment. These observations provided the motivation to examine the subsurface soil storage controls on runoff. Despite the importance of satisfying storage deficits before runoff can be generated, much debate still exists regarding the relative importance of storage versus precipitation threshold controls on runoff generation. Chapter three investigated the soil water storage and hydrometeorological controls on streamflow activation and stormflow generation. Because the soil storage reservoir in the study site is depleted during the seven month long dry period, streamflow remained absent through the dry season and the early wet season, and was only activated after soil storage deficits over the upper 100 cm were satisfied. Interestingly, once streamflow was activated, storage had little influence on storm runoff. When the depth equivalent soil water prior to a storm event (proxy of storage) was summed with the event rainfall depth, a threshold response was observed with stormflow. Above this threshold, the stormflow response and magnitude was almost entirely governed by rainfall event characteristics. These results demonstrate that over the course of the wet season in tropical dry forests the dominant control on runoff generation changed from unsaturated soil storage to the depth of rainfall. This change in the dominant control suggests that after streamflow is activated, the storage deficit becomes low enough that less rainfall goes into storage and more is being translated to runoff. Overall, chapter three suggests that unless the storage threshold is reached, rainfall has little control on runoff generation.

While chapter three showed the importance of threshold storage and precipitation in controlling runoff, it did little to indicate where the water flow pathways and sources areas of runoff

originated and how these sources may change with time. In chapter four, a combination of isotopic and geochemical tracers and hydrometric information was used to examine these flow pathways and source area contributions to runoff. The results from chapter four suggest that runoff originates from deep subsurface layers. Whether these water sources originated from the capillary fringe or hillslope groundwater still remain unknown; however, the geochemistry does suggest substantial contributions from the saturated zone. Like the other catchments with steep slopes and narrow riparian zones, connectivity appeared to be driven by linkages among sub-basins across the catchment rather than the hillslope-riparian connections that characterises most catchments. The residence time analysis of stream baseflow proved to be a powerful tool in identifying potential changes in the topography and internal form of the catchment.

The findings from this dissertation have important implications regarding the potential changes to runoff volumes and the specific generating mechanism at this site. Given that stormflow is strongly controlled by the depth of rainfall, the decrease in the total annual precipitation expected for this region (Karmalkar *et al.,* 2011) will likely result in lower volumes of runoff generated. Furthermore the increase in temperature coupled with lower rainfall will produce a drier climate, likely extending the number of days needed to activate streamflow. These climatic shifts may not affect the subsurface mechanism at this site, as the high hydraulic conductivity at the soil surface will not impede infiltration, even under small increases in rainfall intensity. Under an alternate scenario of increased annual rainfall with events of greater intensity, streamflow will be activated earlier and there will be an increase in total volume of runoff. The greater input of rainfall may impact the runoff generating mechanism, as deeper soil layers, particularly at the base of the hillslope, may become saturated resulting in the development of SOF.

5.2 Concluding remarks

In the classic approach to classifying catchment scale runoff generation, Dunne (1983) highlighted the various environmental controls on the different runoff mechanisms (Figure 5.1). Dunne's approach, while based on multiple field investigations, only provides a qualitative assessment of the controls based on climate, vegetation, topography and soils. While it makes

sense conceptually to lump poorly studied catchments according to the parameters provided in the Dunne diagram, the results produced from this dissertation highlight the importance of a quantitative characterisation of the controls on runoff generation in a given catchment. If the Dunne diagram is used, then infiltration-excess overland flow would be expected given the arid climate and xerophytic vegetation. However, chapters two, three and four indicate that subsurface stormflow is the dominant mechanism, thereby suggesting that the deep, permeable soil and steep slopes and narrow valley bottoms (topography) exert a stronger control on runoff in this catchment than climate or vegetation (Figure 5.1).
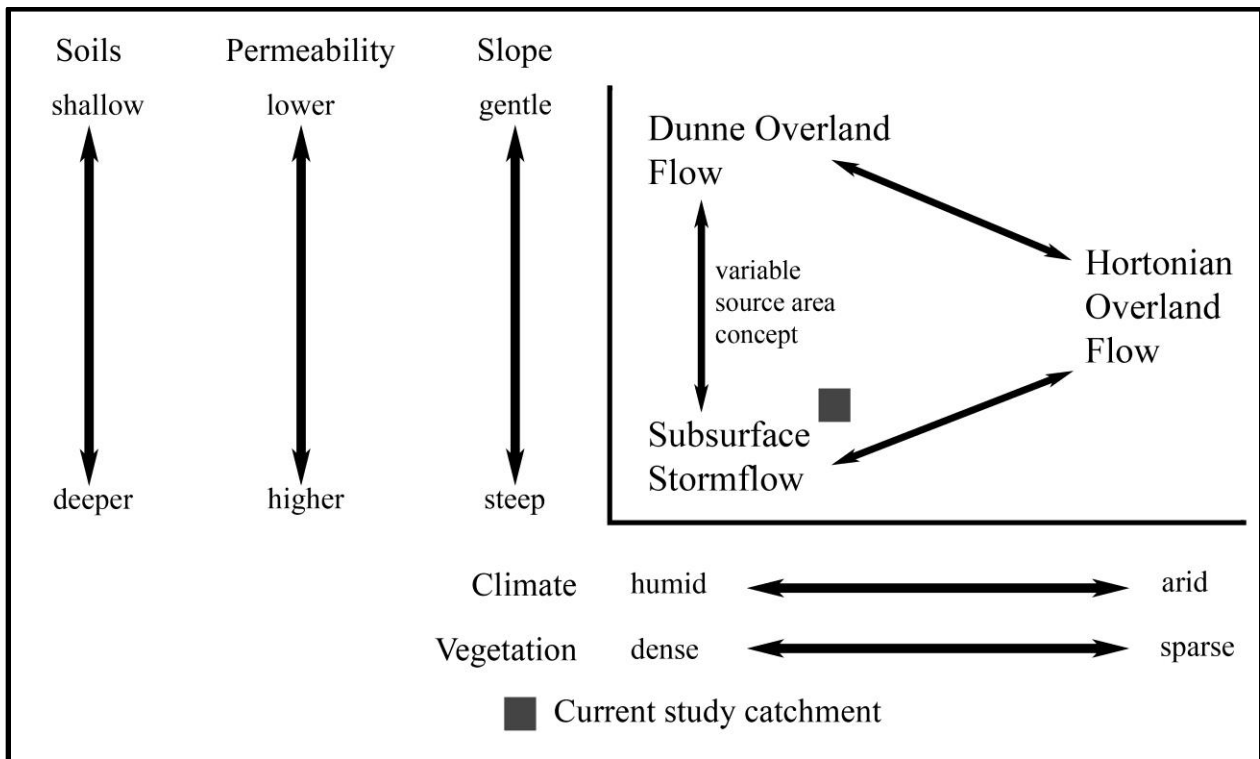

Figure 5.1 Modified version of the Dunne diagram illustrating the environmental controls on the different runoff generating mechanism (after Dunne, 1983). The runoff generating mechanism from the current study catchment is plotted as the grey square.

Unlike the Dunne diagram, which attempts to classify catchments based on qualitative features, it has been suggested that a catchment scale threshold runoff response provides a more quantitative unit of catchment classification and inter-catchment comparison. The shape of the non-linear storage-discharge relationship often reflects the underlying mechanisms controlling the retention and release of rainfall in a catchment. In humid temperate catchments where large

storage deficits are satisfied before subsurface stormflow is generated, the non-linear storage-discharge relationship often takes on a "hockey stick" shape. A similar hockey stick shape is produced from the antecedent soil water-rainfall and runoff response in chapter three. This shape and the isotope and geochemistry data (chapter four) indicates a subsurface flow mechanism in this tropical dry forest and strongly supports the suggestion made by Ali *et al*. (2013) that the threshold response provides an improved metric of catchment classification. Not only does identifying the threshold help in conceptualising how the runoff response may vary under a range of conditions, the linear relationship above the threshold also supports the development of algorithms needed for predictive models in these catchments.

5.3 Recommendations for future research

The research presented in this thesis posed and answered a number of fundamental questions regarding the controls on runoff generation in a tropical dry forest; yet a number of questions remain to be answered with respect to annual variations in runoff response and the specific source area contributions to runoff in this catchment. The following suggestions will continue to improve our understanding of the hydrology of tropical dry forests:

1. Conduct similar investigations over a multi-year time period. Chapters three and four examined the threshold response and water flow pathways in a year in which precipitation was 24% below the annual average. This leads to the questions regarding the impact of interannual variability on the runoff response, specifically in a year with above average rainfall. Characterising the threshold runoff response over a greater range of rainfall frequencies, intensities and magnitudes will improve the threshold response derived from the piecewise regression analysis, thereby improving our ability to assess how climatic shifts will affect the runoff response in this catchment (Ali *et al*., 2013).

2. Explore the effectiveness of the soil water storage – rainfall threshold approach at other tropical dry forest site. The technique used in chapter three provides a simple and easily repeatable approach. However, unless it is tested at other dry forest sites, the approach cannot be recommended for catchment wide adoption.

3. Extend the soil moisture measurement locations for threshold analysis to include near-stream areas. Although most work in steep catchments with incised stream channels and narrow riparian zones suggest that contributions from near-stream areas are very low. However, without direct measurement, the possibility of contributions to streamflow activation cannot be excluded.

4. Map the occurrence and extent of hillslope groundwater across the catchment, using network of wells and pressure transducers. Multi-level well should be installed to map vertical flow gradients to provide direct evidence of flow pathways during storm events.

5. Explore the direct contributions of subsurface water at depths below one metre to stormflow using isotopic and geochemical tracers. Such exploration would confirm if the high pre-event water contributions and strong baseflow geochemical signature recorded during stormflow (chapter four) originated from deep subsurface soil layers or transient groundwater. This would further our process based understanding of runoff generation processes in tropical dry forests.

6. Application of more appropriate isotopic tracers to improve the estimate of the mean stream water residence time. The use of stable isotope tracers (δ^2H and δ^{18}O) while suitable have been shown to underestimate the age of baseflow in residence time analyses. Tritium based characterisation of the baseflow reveals larger proportions of older water and would like improve the estimation of residence time in this catchment (Stewart *et al*., 2010; 2012).

7. Examine the partitioning of water among the major tree species and dominant forest types across the catchment. Canopy interception, transpiration and the rooting depth have strong impact on the partitioning of incoming rainfall and the pool of stored soil water. While these processes have been shown to strongly impact the amount of water available for runoff in wet tropical forests (Bonell and Bruijnzeel, 2005) other work has suggested that two pools of water exists, a mobile pool for runoff and a less mobile one for transpiration (Goldsmith *et al*., 2011). Investigating these processes in tropical dry forests will improve our understanding of water flow in these catchments.

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Appendices

Appendix 1. Change in the soil surface hydraulic conductivity (mm/h) at variable pore pressures at four locations at a) the deciduous forest and b) pine forest.
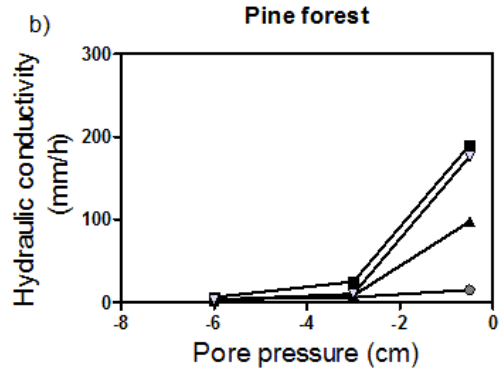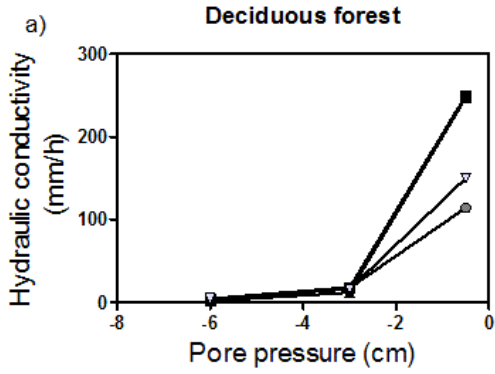
Appendix 2. Example of the graphical hydrograph separation using the local minimum method. The hydrographs are separated by the dashed line, connecting the local minima. The line separates the quickflow (black area) from the baseflow (grey area).
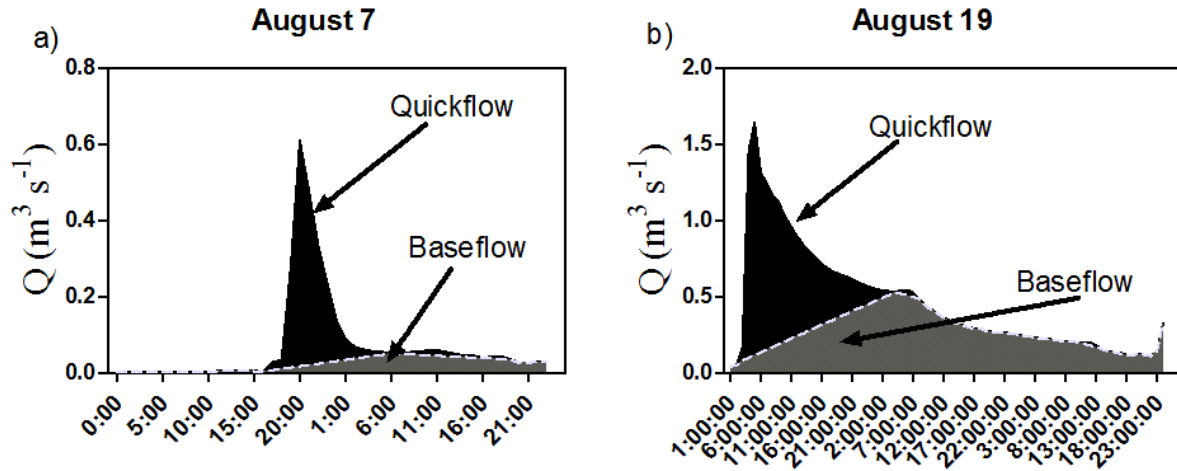
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