

THE EFFECT OF WATERSHED STRUCTURE AND CLIMATE ON STREAMFLOW
RESPONSE, HYDROLOGIC MEMORY, AND RUNOFF SOURCE AREAS

by

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ABSTRACT

Watershed-scale hydrology research has long focused on understanding how various feedbacks in the soil-vegetation-atmosphere continuum affect streamflow. With this dissertation I sought to contribute to our understanding of how watershed characteristics (e.g. topography and vegetation) and climate affect various aspects of watershed hydrology, such as streamflow response times, watershed memory, and runoff source areas. Specifically, I was interested in 1) how watershed structure and climate affect inter- and intra-watershed variability in hydrologic response times, 2) how past precipitation and watershed memory affect runoff response on time scales of months to years, and 3) how runoff source areas vary through time. I approached these challenges/questions through a combination of spatially and temporally intensive and extensive observations synthesized as a) application of a simple lumped model to distill complex watershed behavior into comparable metrics across nested watersheds, b) empirical analysis of long-term hydroclimatic data sets to investigate the effect of watershed memory on the hydrologic response of watersheds, and c) the development of a parsimonious but fully distributed hydrologic rainfall-runoff model to characterize the effect of topographically driven lateral water redistribution and water uptake by vegetation on landscape scale hydrologic connectivity. We demonstrated that 1) differences in response times between watersheds were caused by differences in watershed structure while differences in response times between years were a function of maximum snow accumulation; 2) we found strong influences of past precipitation on runoff from monthly to annual time scales; 3) runoff source areas were highly variable over the course of two water years and exhibited hysteretic spatial behavior over the course of the snow melt seasons. This dissertation contributed new hydrologic understanding of how watershed properties (topography, geology, vegetation etc.), climatic variability, and the interactions between them affect hydrologic response at the watershed scale.

1. INTRODUCTION

Watershed hydrology in its broadest sense focuses on the interactions of the hydrosphere, the biosphere, and the atmosphere. Hydrologic research today studies water movement in the soil-plant-atmosphere continuum and builds on the legacy of ~150 years of physical hydrology that began in the middle of the 19th century in Europe with experiments focused on how forest cover influences precipitation amounts, affects groundwater levels, and modifies streamflow. Mostly observation and empirical data analysis formed the basis for what we know today about how water moves through the landscape.

One of the first physical hydrologists with a correct understanding of the water cycle was arguably Leonardo da Vinci [*Pfister et al.*, 2009]). While many of today's fundamental hydrological principles were developed in the centuries after da Vinci (e.g. Darcy, Manning etc.), our knowledge of watershed hydrology increased significantly with the inception of designated research areas and the advent of the paired watershed approach approximately 100 years ago at Wagon Wheel Gap in Colorado [*Bates and Henry*, 1928]. One of the most well known hydrological experiments that informed us about watershed-scale plant water use of different tree species was started at the Coweeta Hydrologic Laboratory in the 1930s and 1940s [*Hoover*, 1943; *Swank and Douglass*, 1974]. The series of long-term experiments on forest manipulation at Wagon Wheel Gap and Coweeta showed early on that removing vegetation in the Rocky Mountain west does not have the same effect on watershed runoff as removing vegetation in the humid

southeast. The percentage increase in water yield at Coweeta was much greater than at Wagon Wheel Gap, suggesting that the timing and amount of water inputs (spring melt in Colorado vs. somewhat homogenous precipitation over the course of the year in North Carolina) in combination with general climate and different physical watershed characteristics can exert a strong control over how much water is taken up by vegetation, stored in the soil, and transported to the stream.

Hoover and Hursh [1943] summarize the benefits of small experimental watersheds as follows:

“Few watersheds are uniform within themselves as regards topography, vegetation, and soil-profile, but small watersheds show less variability in these properties and are, therefore, most useful in the study of contributions of tributaries to runoff of larger streams.”

In the following decades, small watershed research areas were created worldwide and the research conducted at these sites set the stage for today’s research watershed hydrologists. Despite marked progress, many hydrological challenges remain. In the tradition of small watershed studies we address specific questions in important research fields for today’s hydrologists that harken back to the history of forest and watershed hydrology and include: (1) How does watershed structure and climate influence hydrologic response? (2) How do lateral water redistribution and vegetation processes mediate spatial patterns of water storage and connectivity of upland portions of watersheds to the stream network? (3) How do past climate and hydrologic conditions influence the hydrologic behavior of the present?

In forested headwater catchments, metrics such as topography, geology, vegetation, and soil characteristics have long been identified as a first order control on soil moisture (e.g. *Burt and Butcher* [1985]; *Western et al.* [1999]), hydrologic response (e.g. *Dunne and Black* [1970] ; *Jencso et al.* [2009]) and biogeochemical processes (e.g. *Riveros-Iregui and McGlynn* [2009]; *Pacific et al.* [2011]). Likewise, climatic variability has been documented to affect hydrologic response between and within watersheds (e.g. *Budyko* [1974]; *Montgomery et al.* [1997]; *Nijssen et al.* [2001]; *Jones et al.* [2012]). Despite this effort, the combined effects of landscape and vegetation structure, and the superimposed effect of climatic variability on water redistribution and hydrologic response remain poorly understood. This is partially due to the complex interactions between climatic variability, watershed storage, landscape water redistribution, and largely unknown memory effects.

A parsimonious way to quickly assess the lumped hydrologic response of a watershed is through the use of transfer function models. Transfer function models are essentially quantitative filters that represent the translation of a measured input (precipitation or snowmelt) into an output (most often streamflow). The convolution of an effective input (the fraction of input that actually becomes runoff) and the transfer function (for example a gamma or exponential function typically parameterized via inverse methods) results in runoff. Transfer function models were first applied to hydrologic analyses in the middle of the 20th century (e.g. *Sherman* [1932]; *Dooge* [1959]) and have since been applied to a variety of hydrologic problems such as solute transport in groundwater (e.g. *Maloszewski and Zuber* [1982]) and rivers (e.g. *Wallis et al.* [1989]), and residence time modeling and hydrograph separation with the aid of

tracers (e.g. *Weiler et al.* [2003]; *Hrachowitz et al.* [2009]). Although transfer function models *per se* are not physically based, the parameters of the individual transfer functions can often be physically interpretable, for example as watershed or storm specific response time distributions.

The Tenderfoot Creek Experimental Forest (TCEF) in central Montana (see following text for detailed description) is an ideal place to study the influence of watershed characteristics on hydrologic response. With streamflow and precipitation data for seven subwatersheds dating back nearly 20 years and an extensive monitoring system for many hydrologic variables, new research can build on a strong foundation of accumulated hydrologic knowledge. Previous research highlighted that watershed structure exerted a strong control on a variety of hydrological and biogeochemical processes, such as shallow groundwater dynamics [*Jencso et al.*, 2009; *Jencso et al.*, 2010; *Jencso and McGlynn*, 2011; *Emanuel et al.*, 2014], soil respiration [*Riveros-Iregui et al.*, 2007; *Pacific et al.*, 2008; *Pacific et al.*, 2009; *Riveros-Iregui and McGlynn*, 2009; *Riveros-Iregui et al.*, 2011; *Riveros-Iregui et al.*, 2012], export of dissolved organic carbon [*Pacific et al.*, 2010], net ecosystem productivity [*Emanuel et al.*, 2011], observation-informed hydrologic modeling [*Smith et al.*, 2013; *Smith et al.*, 2014], groundwater-surface water interactions [*Payn et al.*, 2009; 2012; *Kelleher et al.*, 2013; *Patil et al.*, 2013; *Ward et al.*, 2013], vegetation water stress [*Emanuel et al.*, 2010], beetle induced tree mortality [*Kaiser et al.*, 2013], and snow accumulation [*Woods et al.*, 2006]. While TCEF offered ideal conditions for watershed intercomparison mainly in regards with watershed structure, we were not able to compare the effect of different vegetation types on hydrologic response since TCEF consists only of coniferous

vegetation. Furthermore, due to the nature of snowmelt-dominated systems with just one main runoff even per year, it is not possible to study the more dynamic rainfall-runoff relationships that wetter, rainfall-dominated systems offer.

Variability in hydrologic response within and among watersheds has been attributed not only to watershed physical and biological properties (e.g. *Hewlett and Hibbert* [1967]; *Sidle et al.* [1995]; *Western et al.* [1999]; *Jencso et al.* [2009]) but also to climatic influences (e.g. *Budyko* [1974]; *Arora* [2002]; *Jones et al.* [2012]). However, the effect of variability in precipitation on hydrologic response of a watershed on monthly to annual time scales remains largely unknown and while understanding of the spatial variability of runoff and its influences (often large scales) has increased, influences on the variability of hydrologic response at one location through time has received less attention, especially the consideration of past precipitation on time scales exceeding one month.

Some of the earliest and simplest methods to account for past precipitation include antecedent precipitation or wetness indices (API) that consider the amount of precipitation before some point in time, e.g. a runoff event. APIs have been most often applied on a storm basis (e.g. *Linsley et al.* [1949]; *Sittner et al.* [1969]; *Fedora and Beschta* [1989]; *Sidle et al.* [2000]; *Kim et al.* [2005]) and rarely take into account time periods longer than a few weeks or months prior to the hydrologic event in question. Recent studies demonstrated the longer-term effect of carry-over storage on the predictions on annual water balance in a variety of geographic locations, such as Australia [*Jothityangkoon and Sivapalan*, 2009], an Amazonian watershed [*Tomasella et al.*, 2008], and central Nebraska [*Istanbulluoglu et al.*, 2012]. This is not surprising, as it has been known and acknowledged for decades that water stored in the soil can sustain

baseflow over long periods of time (e.g. *Hewlett and Hibbert* [1963]). Still, in many hydrologic applications the storage term is assumed to be stationary, with no or negligible changes in storage occurring from one year to the next (e.g. *Budyko* [1974]; *Eagleson* [1978]). Studies from across the world determining that evapotranspiration was somewhat constant between years (e.g. *Stoy et al.* [2006]; *Kosugi and Katsuyama* [2007]; *Ohta et al.* [2008]; *Oishi et al.* [2010]) cast further doubt on the practice of assuming negligible storage changes. However, attempts to actually quantify watershed storage are rare and often limited to (large-scale) modeling (e.g. *Werth and Guntner* [2010]) and large-scale remote sensing studies (e.g. *Syed et al.* [2008]; *Reager and Famiglietti* [2013]) with long temporal and low spatial resolution, which then take into account all storage components, including open bodies of water. Attempts to quantify soil water storage through soil properties are problematic as they are mostly interpolated from point measurements and even in low-topography terrain associated with large errors [*Proulx-McInnis et al.*, 2013]. Using the long-term data set from the Coweeta Hydrologic Laboratory (see below for detailed site description) we quantified the effect of watershed memory on hydrologic response and present a simple framework to approximate lumped watershed storage from the long-term water balance of a watershed.

While lumped approaches (both modeling as well as empirical) offer insights into the general functioning of watersheds, they lack the detail that distributed modeling applications can offer to inform about the spatiotemporal distribution of water stored in a watershed and, more importantly, the runoff source areas. The formation of areas in a watershed that subsequently contribute to runoff was first identified nearly four decades ago and has been known as variable source area concept (e.g. *Hewlett and Hibbert*

[1963], *Tsukamoto* [1963; 1966], *Hewlett and Hibbert* [1967]). Prior to the discovery of rapid subsurface flow, researchers attempted to quantify areas that contributed to watershed runoff by Hortonian or infiltration excess overland flow (e.g. *Betson* [1964]; *Ragan* [1968]; *Dickinson and Whiteley* [1970]). However, over time it became evident that the major runoff generation mechanism in the variable source area concept is subsurface storm flow [*Hewlett and Hibbert*, 1967; *Hewlett*, 1974,], but also saturation excess overland flow [*Dunne and Black*, 1970] that lead to a rapid hydrologic response in the stream. The variable source area concept quickly became incorporated into hydrologic models (e.g. TOPMODEL [*Beven and Kirkby*, 1979]). However, *McDonnell* [2003] pointed out that since the inception of the variable source area concept our understanding of runoff processes has grown to a point where it may be necessary to revisit at the least the implementation of the variable source area into hydrologic models, as often other runoff generation mechanisms may be a more suitable descriptor of local hydrology. Nonetheless, the variable source area remains an enticing concept, especially with the expansion and contraction of areas that are contributing to runoff. It is important to note that active areas are by no means only limited to the near-stream regions. While *Hewlett* [1974] was very specific about the distinction in the variable source area concept between surface and subsurface flow, his definition of subsurface flow essentially allows for inclusion of all near-stream and distal runoff generation mechanisms that deliver water to the stream via subsurface flowpaths.

More recently the hydrologic community begun using hydrologic connectivity as a metric / mechanism to describe runoff source areas. While there is no one single definition of what constitutes this hydrologic connection, and definitions range from

rather loose ecological [*Pringle*, 2003] to more strict, hydrologic definitions [*Jencso et al.*, 2009], it can generally be considered as spatially and temporally variable conditions that allow for significant flux of water between landscape positions (i.e. from the hillslope to the stream). In return, the establishment of a hydrologic connection—or activation of runoff source areas—is often associated with threshold behavior at the hillslope (e.g. *McGlynn and McDonnell* [2003a]; *McGlynn and McDonnell* [2003b]; *Tromp-van Meerveld and McDonnell* [2006]; *Lehmann et al.* [2007]; *McGuire and McDonnell* [2010]) and watershed scale (e.g. *Spence* [2000]; *Walter et al.* [2000]; *Western et al.* [2001]; *Stieglitz et al.* [2003]; *Zehe et al.* [2005]; *Jencso et al.* [2009]; *Jencso and McGlynn* [2011]; *Smith et al.* [2013]). While the exceedance of thresholds may lead to the activation of individual landscape positions, it is necessary to distinguish between “active areas” and “contributing areas” [*Ambroise*, 2004]. Landscape positions may be active in the sense that a threshold has been exceeded, but if positions downhill along a flowpath are not exceeding the same threshold, the active area is effectively not contributing to runoff. “Contributing areas” are thus “active areas” that have a continuous hydrologic/hydraulic connection to the stream network [*Ambroise*, 2004].

Despite increased research on thresholds, the prediction and mapping of contributing areas has mostly been limited to saturated areas via remote sensing (e.g. *Verhoest et al.* [1998]), modeling applications (e.g. *Beven* [1979]; *Frankenberger et al.* [1999]), or field observations (e.g. *Dunne and Black* [1970]; *Anderson and Burt* [1978]). However, a subsurface hydrologic connection can be established well before an area becomes surface saturated, for example through saturated throughflow, macropore flow, or other preferential pathways (e.g. *Sidle et al.* [1995]; *McGlynn et al.* [2002]). Determination of

threshold behavior at the watershed scale and the ability to map areas with higher probability of hydrologic connectivity to a stream network is crucial for answering questions related to water quantity and quality and runoff source area attribution.

The scope of this dissertation is to elucidate factors governing hydrologic response at the watershed scale. With each chapter I iteratively add complexity in both space and time to the analysis, i.e. from general landscape and climate controls on hydrologic response to storage effects and watershed memory to the spatiotemporal evolution of watershed runoff source areas. With this dissertation I sought to address the following questions:

- 1) What is the role of landscape structure in determining inter-watershed differences in mean response time?
- 2) Can intra-watershed (i.e., annual) variability in mean response time be explained by climate variability?
- 3) How do variability in precipitation and system memory modulate hydrologic response across monthly, seasonal, and annual time scales?
- 4) What is the influence of storage state on the annual water balance and how different are storage dynamics under coniferous and deciduous vegetation?
- 5) What is the combined effect of topographically driven lateral water redistribution and water uptake by vegetation on runoff source areas and the extent of active and contributing areas through space and time?

We approached these challenges/questions through a combination of spatially and temporally intensive and extensive observations synthesized as (1) application of a simple lumped model to distill complex watershed behavior into comparable metrics

across nested watersheds, (2) empirical analysis of long-term hydroclimatic data sets to investigate the effect of watershed memory on the hydrologic response of watersheds, and (3) the development of a parsimonious but fully distributed ecohydrologic rainfall-runoff model to characterize the effect of topographically driven lateral water redistribution and water uptake by vegetation on landscape scale hydrologic connectivity.

Study Sites

This research was conducted at two different sites in the United States: (1) The Tenderfoot Creek Experimental Forest (TCEF) in central Montana, and (2) the Coweeta Hydrologic Laboratory in southwestern North Carolina (see Figure 1.1 for locations within the contiguous United States).

Tenderfoot Creek Experimental Forest (TCEF), Montana

TCEF is located in the Little Belt Mountains in central Montana. Tenderfoot Creek is a tributary of the Smith River, which itself is a tributary of the Missouri. The greater watershed area encompasses 2300 ha and consists of seven nested watersheds ranging in size from 318 ha to 555 ha (Figure 1.1 a). Elevations range from 1992 m to 2426 m. Within the experimental forest, slope generally increases in a downstream direction. The riparian areas become more narrow and the streams more incised toward the watershed outlets. Streamflow was measured with Parshall-flumes and H-flumes depending on the location. Two Natural Resources Conservation Service (NRCS) Snow Telemetry (SNOTEL) sites are located in TCEF: the Stringer Creek SNOTEL is located near the Stringer Creek and lower Tenderfoot Creek watershed outlets, and the Onion Park

SNOTEL is located in the eastern part of TCEF at an intermediate elevation (see Figure 1.1 for locations). Evapotranspiration was measured at an eddy covariance tower located in the upper portion of the Stringer Creek watershed.

The predominant vegetation species is lodgepole pine (*Pinus contorta*) but subalpine fir (*Abies lasiocarpa*), Engelmann spruce (*Picea engelmannii*), and whitebark pine (*Pinus albicaulis*) also occur in the forest. Two of the subwatersheds were partially clearcut between 1999 and 2001, and approximately 50% of basal area were removed during the treatments [Hardy *et al.*, 2006] (Figure 1.1 b). The affected areas comprise 32% to 45% of the watershed areas. Stands at TCEF are single-aged and double-aged [McCaughey *et al.*, 2006], building a moderately complex mosaic of stand polygons ranging in size from tens of ha to almost 300 ha [Barrett, 1993]. Fire history dates back as far as the late 16th century, but most stands regenerated after fires between 1726 and 1873. The vast majority of trees are between 120 and 260 years old [Barrett, 1993].

There are four surficial geologic layers present at TCEF, with Flathead sandstone being the dominant one. The steeper parts near the main outlet are mostly underlain by Granite gneiss. Biotite Hornblende Quartz Monzonite and shale can be found in the higher elevations [Reynolds and Brandt, 2006].

Soils at TCEF are loamy Typic Cryochrepts on the hillslopes and clayey Aquic Cryoboralfs in riparian areas and treeless parks [Holdorf, 1981]. Average soil depths on the hillslopes were found to be approximately 1 m, based on the installation of >180 shallow groundwater wells [Jencso *et al.*, 2009]. Riparian soils are typically slightly deeper.

Climate at TCEF can be described as continental with approximately 880 mm annual

precipitation recorded across the 1961-1990 base period [Farnes *et al.*, 1995]. Typically 75% of the precipitation falls as snow. Peak discharge occurs between late April and late June.

Coweeta Hydrologic Laboratory, North Carolina

The Coweeta Hydrologic Laboratory is located in southwestern North Carolina in the Nantahala Mountain Range of the southern Appalachian Mountains (lat. 35°03'N, long. 83°26'W, Figure 1.2). Coweeta was founded in the 1930s “as a testing ground for certain theories in forest hydrology” [Swank and Crossley Jr., 1988]. Out of 32 weirs that were installed since the 1930s, streamflow is currently recorded in 16 watersheds. Watersheds range in size from 3 ha to 760 ha. The slopes are generally steep with average watershed slopes sometimes exceeding 30°. Soils are generally deep but vary with elevation, with the lower elevation watersheds possessing soils deeper than 3-4 m, and the higher elevation watersheds exhibit soil depths of up to 2 m. Several watersheds at Coweeta were converted from mixed deciduous forest to eastern white pine (*Pinus strobus*) from the late 1930s to the 1950s (two of them are shown in Figure 1.2 with red outlines).

Climate at Coweeta can be classified as Marine, Humid Temperate to Humid Subtropical [Swift *et al.*, 1988]. Average annual precipitation was 1791 mm near the basin outlet for the period 1937-2011. Annual air temperature measured at the same location was 12.6°C.

Dissertation Organization

The following chapters present the research conducted as part of this dissertation. While each chapter's content is technically independent of the other chapters, the work logically builds on itself to provide insight into the drivers of hydrologic response at the watershed scale. The methods used range from the application of a parsimonious, lumped rainfall-runoff model to empirical data analysis and finally the creation of a fully spatially distributed watershed model.

Chapter 2 (*Landscape structure and climate influences on hydrologic response*) evaluates the effect of various topographic and climatic metrics on the rate of water delivery to the outlet. We utilized a simple transfer function rainfall-runoff model (modified from TRANSEP [Weiler *et al.*, 2003]) to quantify watershed scale mean response times (MRT) across seven adjacent subwatersheds in the Tenderfoot Creek Experimental Forest (TCEF, Montana). MRT is a measure of the time required to discharge an amount of water equal to a precipitation input. We hypothesized that MRT is affected by watershed structural metrics as well as variability in climatic conditions. We examined runoff and precipitation data from seven watersheds (five pristine, two with silvicultural treatments) over 12 years of record to learn: What drives the intra- and inter-watershed variability of hydrologic response (streamflow)? The analyses indicated strong relationships in the unharvested watersheds between MRT and landscape metrics such as watershed slope, distance from creek, convergence (measure of terrain dissection), the underlain geologic strata, and tree height. Furthermore it highlights that even subtle differences in both watershed structure and climatic inputs can lead to

significant variability in hydrologic response.

Chapter 3 (*Watershed memory at the Coweeta Hydrologic Laboratory: The effect of past precipitation and storage on hydrologic response*) explored the effect of climatic variability on hydrologic response and how watershed hydrologic memory can be propagated through time across contrasting vegetation and watershed structure within Coweeta. Even though antecedent precipitation has been acknowledged as a major factor in runoff generation, most research has focused on time scales shorter than 30 days prior to some precipitation or runoff event. Long term (monthly to annual time scales) effects of past precipitation on runoff remain poorly described. This chapter assesses the influence of past precipitation (i.e. watershed memory) on runoff at various temporal scales at the Coweeta Hydrologic Laboratory using long-term precipitation and runoff data from five watersheds. Among those watersheds are two sets of paired low-elevation watersheds with different aspects, and a steeper high-elevation watershed. Each watershed pair consists of a coniferous and a deciduous watershed, allowing us to investigate the effect of different vegetation types on watershed memory. We also approximated annual watershed storage and linked variability in storage to variability in annual runoff.

Chapter 4 (*The combined effects of topography and vegetation on watershed connectivity*) builds upon the previous chapter by discretizing watershed storage in space and time. Whereas Chapter 3 treated watershed storage as a lumped annual value due to limitations of the water balance approach and general data availability, Chapter 4 assesses spatially explicit storage at high temporal resolution. We developed a parsimonious but fully distributed rainfall-runoff model and calculated watershed storage

to inform understanding on the evolution of watershed connectivity—as mediated by topography and vegetation—across two water years in the Lower Stringer Creek subwatershed at the Tenderfoot Experimental Forest. The questions we seek to answer revolve around the spatiotemporal evolution of watershed connectivity, i.e. the establishment of a hydraulic connection between different parts of the landscape and the stream network.

Chapter 5 presents a brief summary of the main results of the previous chapters, explains the implications of this research and gives recommendations for future research.

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Figures



Figure 1.1: Location of the two experimental watersheds in the contiguous United States

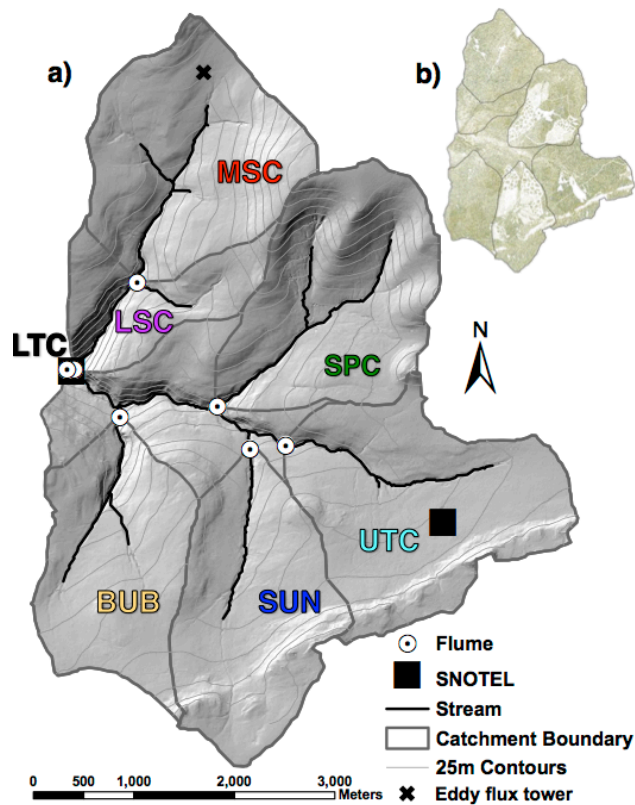


Figure 1.2: Tenderfoot Creek Experimental Forest with the seven subwatersheds, the locations of flumes, SNOTEL sites, and eddy flux tower (a); silvicultural treatments in SUN and SPC (b)

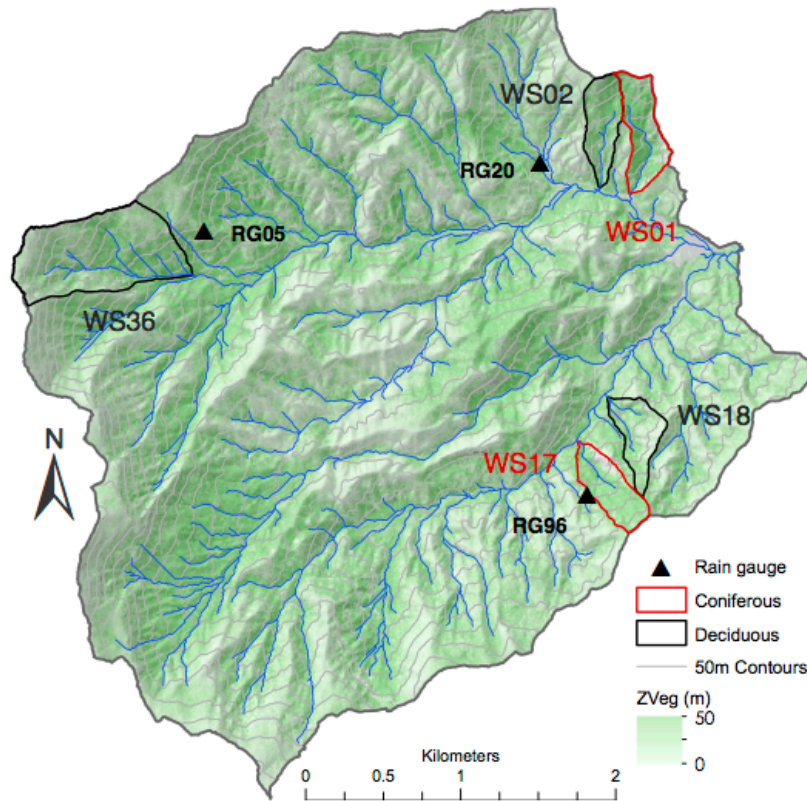


Figure 1.3: The Coweeta Hydrologic Laboratory, highlighted are the five watersheds and three rain gauges used for the analyses in Chapter 3

2. LANDSCAPE STRUCTURE AND CLIMATE INFLUENCES ON HYDROLOGIC RESPONSE

Contribution of Authors

First Author: Fabian Nippgen

Contributions: Co-developed the project, analyzed the data, created figures, wrote the manuscript.

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Abstract

Climate variability and catchment structure (topography, geology, vegetation) have a significant influence on the timing and quantity of water discharged from mountainous catchments. How these factors combine to influence runoff dynamics is poorly understood. In this study we linked differences in hydrologic response across catchments and across years to metrics of landscape structure and climate using a simple transfer function rainfall-runoff modeling approach. A transfer function represents the internal catchment properties that convert a measured input (rainfall/snowmelt) into an output (streamflow). We examined modeled mean response time, defined as the average time that it takes for a water input to leave the catchment outlet from the moment it reaches the ground surface. We combined 12 years of precipitation and streamflow data from seven catchments in the Tenderfoot Creek Experimental Forest (Little Belt Mountains, southwestern Montana) with landscape analyses to quantify the first order controls on mean response times. Differences between responses across the seven catchments were related to the spatial variability in catchment structure (e.g. slope, flowpath lengths, tree height). Annual variability was largely a function of maximum snow water equivalent. Catchment averaged runoff ratios exhibited strong correlations with mean response time while annually averaged runoff ratios were not related to climatic metrics. These results suggest that runoff ratios in snowmelt-dominated systems are mainly controlled by topography and not by climatic variability. This approach provides a simple tool for assessing differences in hydrologic response across diverse watersheds and climate conditions.

Introduction

Catchment structure (topography, geology, vegetation and soil patterns) has long been recognized as a first order control for soil moisture [*Burt and Butcher*, 1985; *Western et al.*, 1999], hydrologic response [*Dunne and Black*, 1970; *Jencso et al.*, 2009; *Quinn et al.*, 1991] and biogeochemical processes [*Pacific et al.*, 2011; *Riveros-Iregui and McGlynn*, 2009] in forested mountain landscapes. At the same time, the impact of climatic variability on hydrologic processes and water resources has been intensively studied across various temporal and spatial scales (e.g. *Montgomery et al.* [1997], *Nijssen et al.* [2001], *Niehoff et al.* [2002]). However, the extent to which the intersection of landscape structure and climatic variability influences hydrologic response and the timing of streamflow remains largely unknown. This raises the question: Do climate variability and landscape characteristics explain differences in the hydrologic responses among catchments?

A parsimonious way to assess the response of catchments is through the use of transfer function models, which often utilize only rainfall and catchment runoff. Transfer functions are quantitative filters that describe the translation of a measured input (e.g. snowmelt/rainfall) into an output (e.g. runoff). In hydrology, transfer function models have been used to model solute transport through porous media [*Beven and Young*, 1988; *Jury*, 1982; *Maloszewski and Zuber*, 1982], river solute transport [*Wallis et al.*, 1989] and tracer residence time modeling and hydrograph separation [*Hrachowitz et al.*, 2009; *McGuire et al.*, 2002; *Weiler et al.*, 2003]. Rainfall-runoff applications of transfer functions date back to the unit hydrograph and its adaptations [*Dooge*, 1959; *Sherman*,

1932] where an amount of excess rainfall (i.e. effective rainfall) is convoluted with the unit hydrograph and thus translated into catchment runoff. Transfer function approaches usually assume linear behavior between rainfall amount and the resulting runoff, but they can contain non-linear modules, as for the calculation of effective precipitation [*Jakeman and Hornberger, 1993*]. Transfer functions allow for qualitative and quantitative analysis of the integrated processes in catchments that translate inputs signals into output signals, subsuming complex catchment characteristics into one parsimonious function. Runoff transfer functions may differ among different catchments, and even for a single catchment the transfer function may differ through time.

In this study we use a simple transfer function rainfall-runoff model (modified from TRANSEP, see *Weiler et al. [2003]*) to estimate mean response times. Mean hydrologic response time is a measure of how long it takes to discharge an amount of water equal to an effective precipitation input. Mean response time should not be confused with mean residence time (or mean transit time), terms that are often used in tracer based catchment modeling. Though seemingly similar, they are two very different metrics of catchment hydrology. Mean response time as it is used in this paper, refers to the transmission of an input magnitude or pressure response, while mean residence or transit time refers to the transmission of the water molecules themselves and not the pressure response. A mean residence time describes how long an individual water input (or molecule) remains in the catchment, while mean response time does not contain information about the age of water.

Using this simple transfer function model, we examine how catchment structural characteristics and climatic variability affect the hydrologic response of mountain

catchments. We examined runoff and precipitation data from seven adjacent catchments located in the mountains of central Montana over a 12-year period in conjunction with ten landscape metrics derived from high resolution LIDAR (Light Detecting And Ranging) data.

Focus on adjacent catchments reduces confounding factors typically encountered when comparing catchments across regions. The close proximity and similar elevation range of the catchments also limit hydrologic process differences and rainfall/snow variability. Differences in transfer functions and the resulting mean response times are therefore signatures that we expect to be related to distinct catchment characteristics including structure and climate. We address the following questions in this study:

1. What is the role of landscape structure in determining inter-catchment differences in mean response time?
2. Can intra-catchment (i.e. annual) variability in mean response time be explained by climate variability?

Methods

Site Description

The Tenderfoot Creek Experimental Forest (TCEF) is located in the Lewis and Clark National Forest, which is part of the Little Belt Mountains in the Northern Rocky Mountains of central Montana (lat. 56.55°N, long. 110.52°W). The total area of TCEF encompasses 2300 ha and includes seven gauged catchments, Bubbling Creek (BUB), Lower Stringer Creek (LSC), Middle Stringer Creek (MSC), Lower Tenderfoot Creek

(LTC), Spring Park Creek (SPC), Sun Creek (SUN) and Upper Tenderfoot Creek (UTC), with subcatchments ranging in size from 318 ha to 554 ha (Figure 2.1). Tenderfoot Creek is a headwater of the Smith River that drains into the Missouri River. In all subcatchments the slope generally increases downstream with more constrained and smaller riparian areas in a downstream direction. Rock outcrops and steep talus slopes are most frequent on lower portions of the catchment.

The dominant tree species is lodgepole pine (*Pinus contorta*), though the forest also includes subalpine fir (*Abies lasiocarpa*), Engelmann spruce (*Picea engelmannii*) and whitebark pine (*Pinus albicaulis*). Shrubs and grasses dominate the riparian areas, which make up about 14% of the study area. Two of the subcatchments, SPC and SUN, were partially clearcut between 1999 and 2001 to investigate the effect of different silvicultural treatments on water yield and sediment export. Each watershed included two different types of treatments: In one treatment the forest was thinned and in the other treatment individual groups of trees were left standing with clearcut corridors between them; both treatment types removed approximately 50% of the tree basal area with total treated area of ~127 ha in SPC (~32% of the catchment area) and ~162 ha in SUN (~45% of the catchment area).

Soils at TCEF are loamy Typic Cryochrepts located along hillslope positions and clayey Aquic Cryoboralfs in riparian zones and parks [Holdorf, 1981]. Watershed soils are 0.5 to 2 m deep. Average soil depths were found to be approximately 1 m, based on the installation of >180 shallow groundwater wells and soil pits located on hillslopes and in riparian areas [Jencso *et al.*, 2009].

The most dominant geologic layer is Flathead Sandstone. Granite gneiss underlies the sandstone and is prevalent towards the bottom of the watersheds. The highest elevations of LSC and SPC are underlain by biotite hornblende quartz monzonite, which lies over a layer of shale (Wolsey formation). The sandstone portions of TCEF tend to have a gentler slope whereas the other formations generally coincide with steeper slopes.

Climate at TCEF is predominantly continental. For the 1961-1990 base period annual precipitation was about 880 mm [*Farnes P.E.*, 1995]. Approximately 70% of the precipitation volume falls as snow. Discharge usually peaks between mid April and the end of May. Stream discharge peaks are generated by snowmelt and rain on snow events during the spring melt period. Discharge typically reaches its minimum in late September or October.

Landscape Analysis

Topographic variables were derived from 1 m resolution airborne laser swath mapping (ALSM) data. The ALSM based digital elevation model (DEM) was resampled to 10 m, a resolution high enough to capture the actual topography but coarse enough not to introduce errant behavior due to micro topographical features such as fallen trees or individual boulders (see *Jencso et al.* [2009]).

We calculated catchment-wide means and medians for 10 different landscape variables (Table 2.1). Topographic variables were derived from the 10m DEM, whereas tree height (TH) was calculated by subtracting the last return LIDAR coverage (bare earth) from the first returns (elevation of canopy) (see *Emanuel et al.* [2010]). Tree

heights of up to 2 m were then resampled to 0 m to avoid noise caused by riparian vegetation and clear-cut regrowth. Two sample non-parametric Kolmogorov-Smirnov tests were performed on pairs of watersheds to test for significant differences between entire distributions of landscape variables, while Kruskal-Wallis tests were performed on all seven watersheds to test for significant differences in the medians of landscape variable distributions.

To assess interdependency among the landscape metrics, we also calculated a cross-correlation matrix for Lower Tenderfoot Creek (LTC) with Pearson's correlation coefficients for all landscape metrics with distinct values for each grid cell. Since TH was calculated based on the original 1 m resolution LIDAR data, the remaining variables were resampled from 10m resolution to 1 m resolution in order to correlate them with the 1 m TH coverage. The correlations are based on more than 20,000,000 grid cells for each landscape variable.

Hydrologic Measurements

Precipitation and discharge data were available for years 1996 – 2002 and 2005 – 2009. The model simulation period was from March 1st to the last day of February of the following year instead of the conventional water year (e.g. year 1996 runs from March 1, 1996 to Feb 28, 1997). This time period was selected to match the characteristic timing of annual snowmelt and the event-based nature of transfer function models, including a short “warm up” period and a longer time following the annual event. Discharge for the years 1996 through 2005 was made available by the USDA Forest Service. Discharge for

the period 1996 through 2001 was available in daily time steps. For the years 2002 and 2005, stage was recorded every 15 minutes at H-Flumes or Parshall Flumes located at the catchment outlets using a high-precision potentiometer (Model 3540, Bourns Inc., Riverside, CA) and later aggregated to a daily time series. The USDA obtained missing data for individual catchments and years by regression with other catchments. Data for the entire melt period was missing for the following watersheds and years: UTC 1997, SUN 2000 and UTC 2000. Small portions (~15%) of the falling limb of the hydrograph were missing for SPC 2001, BUB 2001 and LTC 2001, small portions (~10%) of the rising limb of the hydrograph were missing for SUN 1997, MSC 1998 and LTC 1998. For the years 2006 through 2010, stage was recorded every 30 minutes at the same flumes using capacitance rods with ± 1 mm resolution (TruTrack Inc., Christchurch, NZ) and later aggregated to a daily time series. Precipitation and snowmelt were measured at two Natural Resources Conservation Service (NRCS) SNOTEL stations, one at Onion Park (2259 m) near the headwaters of UTC and the other at Stringer Creek near the bottom of the Experimental Forest next to the LSC flume (1996 m). The calculations of liquid water input (comprising snowmelt and rainfall) at the surface are based on four assumptions: (i) when both cumulative precipitation and SWE increase, neither melt nor liquid precipitation is occurring, (ii) when SWE decreases and cumulative precipitation stays the same, melt is occurring, (iii) when cumulative precipitation increases and SWE decreases, both melt and rain are occurring and finally (iv) when SWE is zero and cumulative precipitation increases, rain is occurring. We then subtracted SWE from cumulative precipitation. This can be expressed mathematically as

$$\frac{dI}{dt} = \begin{cases} 0, \frac{dP}{dt} > 0 \cap \frac{dS}{dt} > 0 \\ -\frac{dS}{dt}, \frac{dP}{dt} = 0 \cap \frac{dS}{dt} < 0 \\ \frac{dP}{dt} - \frac{dS}{dt}, \frac{dP}{dt} > 0 \cap \frac{dS}{dt} < 0 \\ \frac{dP}{dt}, \frac{dP}{dt} > 0 \cap S = 0 \end{cases} \quad (1)$$

where I is the amount of liquid water input to the ground, P is cumulative precipitation (both liquid and solid), S is snow water equivalent and t the time step. Note that a reduction of S results in the release of melt water from the snow pack.

This results in a time series of cumulative liquid input, the numerical differentiation of which yields an input per unit time. Precipitation and snowmelt data were aggregated from 6 hour time steps to a daily time series. Based on observed differences in precipitation and SWE of up to 20 % between the two SNOTEL stations, we weighted and combined the input of the two SNOTEL sites by elevation. We calculated a linear regression for the elevation range between the Stringer and Onion Park SNOTEL sites and assigned each DEM grid cell a precipitation value based on the regression. All elevations >2259 m were assigned the Onion Park precipitation. Since TRANSEP is a lumped model the semi-distributed precipitation was then averaged to a single value per time step for each catchment. In addition total SWE was calculated for the Onion Park and Stringer Creek SNOTEL sites for each of the 12 years. Maximum precipitation and snowmelt intensities were calculated with a moving average from one to 21 days.

Uncertainties Associated with the Use of Two SNOTEL Sites

Calculating response times to determine differences between the catchments requires exact input time series for each of the catchments. An important assumption for the analysis is that differences in snow accumulation, the timing of melt and the magnitude of melt rates are adequately reflected by the elevation weighting employed based on the two SNOTEL sites. This is a strong but necessary assumption due to the limitation to two precipitation measurement sites. Multiple factors could play a role in determining SWE characteristics across the landscape. However, we believe that the elevation weighted input adequately describes the snow accumulation and melt timing differences between the catchments. *Smith and Marshall* [2010] quantified uncertainties for a predictive runoff model for TCEF and found that elevation had a greater impact on both runoff and melt than aspect. Personal observations of the authors over the course of six winters corroborate the assumption that elevation has a greater impact on snow accumulation than aspect. In addition, the SNOTEL sites are usually located on flat ground, not ‘favoring’ a certain aspect. We acknowledge that aspect could lead to slight differences in the onset of snowmelt. However, considering the duration of the whole melt period this initial uncertainty is likely negligible. The general spatial variability of precipitation is –in our opinion- of less concern in contrast to convective summer rain storm events, which can be small in extent and therefore highly localized. Frontal snow storms at TCEF are usually of a greater spatial extent and easily cover the full extent of the experimental forest.

Hydrologic Model

We used a simple rainfall-runoff transfer function model, a module of the TRANSfer function hydrograph SEParation model (TRANSEP) developed by *Weiler et al.* [2003]. A transfer function translates a measured input (effective daily rainfall + snowmelt) into an output (daily discharge) (Figure 2.2).

The rainfall-runoff module in TRANSEP includes a non-linear loss function, $s(t)$, to first transform measured precipitation into effective precipitation [*Jakeman and Hornberger, 1993*]:

$$s(t) = b_1 p(t) + (1 - b_2^{-1})s(t - \Delta t) \quad (2)$$

$$s(t = 0) = b_3 \quad (3)$$

$$p_{eff}(t) = p(t)s(t) \quad (4)$$

where $s(t)$ is the antecedent precipitation index, b_1 maintains the water balance that the total effective precipitation equals the total runoff, b_2 determines the rate at which the watershed dries out, b_3 sets the initial state of catchment wetness at the beginning of the time series, p_{eff} the effective precipitation and $p(t)$ is the measured rainfall and snowmelt input at time t .

We applied the gamma distribution transfer function to daily data

$$g(\tau) = \frac{\tau^{\alpha-1}}{\beta^\alpha \Gamma(\alpha)} \exp\left(-\frac{\tau}{\beta}\right) \quad (5)$$

where $g(\tau)$ is the value of the transfer function at a given lag time between rainfall and runoff, $\Gamma(\alpha)$ is the gamma function, α is the shape parameter and β a scale parameter.

Multiplying α and β yields the mean value of the gamma distribution [Kirchner *et al.*, 2000], which can be interpreted as the mean response time. Therefore mean response time is defined as the average time that it takes for a given amount of water input (liquid precipitation and/or snowmelt) to leave the catchment through the outlet. The two-parameter gamma function was selected for our analysis as it is more flexible than an exponential function [Hrachowitz *et al.*, 2010] but does not require an additional splitting parameter as, for example, the two-parallel-linear-reservoir transfer function.

The parameters in equations 2 and 5 were calibrated with 150,000 Monte Carlo simulations. Convolution of the transfer function (2) and the effective precipitation (5) results in runoff:

$$Q(t) = \int_0^t g(\tau) p_{eff}(t - \tau) d\tau \quad (6)$$

where $Q(t)$ is the runoff at time t .

Correlations between TRANSEP calculated mean response times and the catchment-averaged landscape and annual climatic variables were calculated as Pearson correlation coefficients. The means and medians of the landscape variables were tested for significant correlation with the catchment averages of mean response times (each catchment averaged over the 12 years of simulation). In addition, maximum annual SWE (mm per year) and maximum input (mm per time step) were tested for correlation with the annual averages of mean response time (each year averaged over the seven catchments) in order to distinguish between landscape controls on catchment hydrologic response and the influence of climatic fluctuations on intra-catchment variability. We also calculated the runoff ratios for each catchment (averaged over 12 years) and each

year (averaged over seven catchments) and tested for correlation with the corresponding mean response times.

Results

Catchment Hydrology

Annual elevation-weighted precipitation for LTC over the 12 years of simulation period averaged 790 mm and was 110 mm less compared to the 1961 – 1990 base period. The highest amount of precipitation and melt water entered the catchments in 1997 with an average of 960 mm among all catchments; 2001 had the least amount of input with an average of only 621 mm (Figure 2.3a). The timing and intensity of inputs varied from year to year but the average range among catchments was only 17 mm per year (BUB 788 mm, UTC 805 mm). The inter-annual variability in precipitation was much greater and hence more important than the inter-catchment variability of the precipitation and melt inputs among the seven catchments.

Runoff peaks varied strongly from year to year, with up to 23 days between the annual maximum peaks and peak magnitudes ranging from 4.2 to 15.1 mm/day at LTC. Single peaks were the exception, and multiple peaks generally occurred between mid April and late May (Figure 2.3b).

Catchment Transfer Functions

Model results indicate strong differences in transfer functions (Figure 2.4a). The greatest differences in catchment response occurred over the first 10 to 15 day portion of

the transfer function (Figure 2.4b), with fractions of the input leaving the catchments (response fraction) within the first 24 hours ranging from 10.3 % (LTC) to 4.4 % (SPC) with an average of 7.6 % over all seven catchments. Response fractions for all catchments declined after the first day and averaged 3.6 % on the 10th day after an input and ranged from 4.3 % in BUB to 3.0 % in SPC. Sixty days after the input event, average runoff contribution decreased to 0.1 % of the original input with variations among catchments from 0.05 % for LTC to 0.25 % for SPC (Figure 2.4c). After 132 days, nearly all inputs left the catchments with runoff comprising less than 0.01 % per day of original input. On average, a single precipitation or snowmelt input sustained runoff in the streams for about 132 days. The shortest complete transfer function was LSC (96 days), while LTC and SUN both maintained runoff for 153 days. Catchments with greater early response proportions do not necessarily maintain higher response through the season. For example SPC relatively low early response fractions but greater response than SUN by day six and greater than the other catchments by day 14. Whereas the daily SUN water contributions after 14 days exceed those of the other catchments, it still lags behind in total amount of water that has left the outlet in the remaining catchments and remains lower for the later parts of the transfer function. The cumulative transfer functions (Figure 2.4d) show that on average 20.6 % of an input has left the catchments as streamflow after only three days, with individual catchments ranging from 12.8 % (SPC) to 25.7 % (LTC). After nine days an average of 50 % of the effective precipitation has left the catchments, with a maximum of 59.9 % for LSC and a minimum of 37 % for SPC. After 35 days an average of 90 % of the input has left the catchments, ranging from 96.1 % in LSC to 82.9 % in SUN. After 85 days 99 % of the original input has left the catchments.

Averaged over the 12 years of simulation, mean response time for the seven catchments ranged from 9.56 days for LSC to 18.67 days for SPC, with the average of all catchments being 13.72 days (Table 2.2). Average Nash-Sutcliffe efficiencies across catchments ranged from 0.80 to 0.90.

Annual Transfer Functions

The averaged transfer functions for each year of the 12-year simulation period (averaged over seven catchments for each year) exhibit strong differences (Figure 2.5a for averaged transfer functions and Figure 2.5b for the averaged cumulative transfer functions). The input was greatest in 1997 and smallest in 2001 (highlighted in Figure 2.5 as solid black line and dashed black line respectively). The mean response times for each year averaged over the seven catchments ranged from 7.16 days in 2006 to 24.95 days in 2001 (Table 2.3). Average Nash-Sutcliffe efficiencies across years ranged from 0.72 to 0.92.

Landscape Analysis

The means and medians for slope (SLP), distance from creek (DFC), gradient to creek (GTC) and hillslope power (HP) for all seven catchments indicate an identical ranking for the catchments (i.e. the smallest mean corresponds to the smallest median and the largest mean to the largest median (Table 2.4)). Means and medians for convergence (CON) differ in sign, denoting slight convergence in the median but small divergence in the mean. The strongest departure between mean and median was evident for local input

(LI), with the mean exceeding the median by up to an order of magnitude. Maps for eight of the ten topographic parameters are shown in Figure 2.6.

Both the Kolmogorov-Smirnov and the Kruskal-Wallis tests resulted in p -values $\ll 0.01$ and highlight the significant difference for all the landscape metrics between the catchments.

The strongest cross-correlations between the landscape metrics for LTC are correlations between SLP and GTC ($r_p = 0.64$), INS and SLP ($r_p = -0.58$), GTC and DFC ($r_p = -0.43$) and SLP and DFC ($r_p = -0.30$). The metric with the weakest correlations was tree height (TH) with no correlation stronger than $r_p = \pm 0.08$. The cross-correlation matrix is shown in Figure 2.7.

Mean Response Time Correlations with Topographic Parameters

There were no significant correlations between the landscape variables and modeled mean catchment response times, with the exception of mean convergence with $r_p = -0.94$ (Table 2.5). However, when the two catchments subjected to silvicultural harvesting (SUN and SPC) were omitted from the correlation, several significant relationships emerged. We excluded SUN and SPC from the correlation analysis because the silvicultural treatments and harvesting of more than 30% of the catchment areas have wide-ranging effects on hydrological processes [Bosch and Hewlett, 1982; Stednick, 1996]. We focus our analysis of landscape variables on the five remaining catchments and discuss the behavior of SUN and SPC in detail in the following section.

The correlations for both the mean and median for the majority of the landscape parameters were equally strong (Table 2.5). However, both LI and RB showed strong differences in mean and median in the correlations with mean response time, with the medians showing the stronger/significant correlations. The distributions for both LI and RB are highly positively skewed. We thus focus on the medians because they are more representative of the distributions of LI and RB than the arithmetic mean, which is more susceptible to extreme outliers and skewness.

The landscape variables plotted versus the mean response times of the different catchments indicate strong correlations for multiple landscape metrics (Figures 2.8 and 2.9). Although excluded from the correlation analysis, SPC and SUN are displayed in Figure 2.8 for completeness.

Mean Response Time Correlations with Climatic Variables

Maximum elevation weighted SWE for LTC ranged from 282 mm in 2001 to 467 mm in 1997 (Table 2.6). The correlation with the annually averaged mean response times was significant ($r_p = -0.82$) the greater the snowpack was during a given year, the faster the mean response time averaged among all seven catchments (Figure 2.10). From the input intensities for temporal windows ranging from one to 21 days, the period with the greatest intensity over a 13-day moving window exhibited the strongest and most significant correlation with mean annual response time ($r_p = -0.76$). Maximum input intensities over 13 days ranged from 16.3 mm/day in 2001 and 1998 to 23.1 mm/day in 1997, resulting in 211.9 mm in 13 days in 2001 and 1998 to 300.3 mm in 1997.

Mean Response Time Correlations With Runoff Ratios

There was a strong, significant correlation between mean response time and the runoff ratio for each catchment (Figure 2.11) with $r_p < -0.96$ for both the modeled and measured runoff ratios, omitting SUN and SPC. Larger runoff ratios were generally associated with shorter mean response times. However, with SUN and SPC included the correlation decreased to $r_p = -0.74$ for observed runoff ratios and $r_p = -0.79$ for simulated runoff ratios. In contrast, there was no significant correlation for averaged annual response times and annual runoff ratios ($r_p = 0.41$ for the observed runoff ratios and $r_p = -0.42$ for the simulated runoff ratios) (Figure 2.11b).

Discussion

The influence of topographic structure, geology, vegetation distribution and climate on hydrologic response is of intense interest. However, the independent and combined influences of climatic variability and landscape characteristics on stream response remain poorly understood. We calculated mean response times, the average time it takes for a liquid water input from the moment it hits the ground surface to leave the catchment through the outlet, for both catchment averages and annual averages for seven catchments and 12 years. We analyzed the seven TCEF subcatchments for a total of nine landscape structure metrics that can all be derived from airborne LIDAR data, including tree height distributions and potential solar insolation. We also calculated fractions of

geologic strata types for each catchment. In addition to stationary topographic metrics we analyzed non-stationary climatic metrics, derived from two NRCS SNOTEL sites within TCEF. These data allowed us to independently examine the influence of both topography and climate on modeled mean response time. The following discussion focuses on the five more pristine catchments BUB, LSC, LTC, MSC and UTC. The two catchments, which were subjected to silvicultural treatments, SUN and SPC, will be discussed separately.

What are the Interdependencies of the Selected Landscape Metrics?

Comparing the effect of different landscape metrics on hydrologic response raises the question about redundancies in the chosen metrics, i.e. whether two seemingly different metrics are effectively expressing the same phenomenon due to strong cross-correlation. To date, a few studies focusing on landscape evolution have addressed cross-correlations between catchment characteristics, most notably the effect of vegetation on landscape metrics such as drainage density [*Istanbulluoglu and Bras, 2005*] and the effect of drainage networks structure on vegetation distribution and plant water stress [*Caylor et al., 2004*]. To our knowledge no studies have calculated grid cell based correlations of LIDAR derived topographic metrics. We investigated metric interdependencies by calculating cross-correlation between metrics for which values existed for each cell.

The correlation matrix calculated for LTC shows no correlations stronger than $r_p = \pm 0.64$ and only four of the 21 correlations are stronger than $r_p = \pm 0.30$ (Figure 2.6). However, some redundancy in the chosen landscape metrics is apparent with the most

obvious being SLP (the slope angle of an individual cell based on the surrounding eight cells) and GTC (the gradient integrated over the flow path from a particular cell to the stream). Based on these definitions SLP and GTC are not necessarily correlated with each other (e.g. a low angle cell sits on top of a steep flow path). However, Figure 2.6 indicates that the steepest local slopes of LTC correspond with the steepest gradients. This also explains the negative correlations between DFC-GTC and DFC-SLP: in general slope and gradients at TCEF increase toward the outlet of each catchment and are therefore correlated with DFC.

The negative correlation between SLP and INS is a result of how slope and insolation interact. Insolation is a function of aspect and slope. In general south facing slopes have a higher insolation than north facing slopes. However, on south facing hillslopes a steeper slope increases solar insolation, whereas a steeper slope on north facing areas leads to a decrease in insolation.

Florinsky and Kuryakova [1996] reported correlations between vegetation type and topographic metrics, especially those affecting solar insolation. However, no correlations between TH and landscape metrics were observed at TCEF.

Due to the lack of strong cross-correlations between the landscape variables we can assume that potential significant correlations between landscape variables and mean response times are the result of individual but interacting effects, rather than a manifestation of one effect (one variable) in several variables.

How does topography explain the observed
differences in average catchment response times?

For a majority of the landscape metrics considered in this study, we found strong linear relationships with mean response time (Figures 2.8 and 2.9). SLP and GTC both exhibited a strong negative correlation with mean response time. The similar strength of the correlations is a result of the strong correlation of the two metrics with each other. These results confirm the greater the slope or the gradient to creek the faster the mean response time.

Conversely, greater DFC increases the mean response time, which is consistent with geomorphic instantaneous unit hydrograph theory [*Gupta et al.*, 1980; *Rodriguez-Iturbe and Valdes*, 1979]. Simply stated, the greater the distances that water or input signals must travel, the longer it takes them to reach the catchment outlet. The positive relationship between DFC and mean response time is likely amplified by the slight negative correlation between SLP/GTC and DFC: short flow paths tend to have steeper gradients while longer flow paths are often associated with gentler gradients (Figures 2.6 and 2.7).

The inverse relationship found for LI (positive correlation with response time) and CON (negative correlation with response time) seems counterintuitive at first since both metrics are a measure of flow accumulation. The results for local input are consistent with the findings of [*Jencso et al.*, 2009] who found a strong relationship between LI size and the duration of a hydrologic hillslope-riparian area-stream hydrologic connection. *Jencso et al.* [2010; 2009] reported that a larger LI does not necessarily mean that more water gets transported from the hillslope to the streams at any

given time but rather that larger LI hillslopes contribute for a longer period of time than smaller size LIs. It is the longevity of the hillslope-stream connection that is mediated by LI and not how fast water is being delivered to the stream. Note that LI is the UAA size for grid cells bordering the stream, whereas CON was calculated for every grid cell in each catchment, resulting in a metric that is more suitable to assess overall catchment concavity than LI, which describes overall hillslope size and connectivity potential for near-stream cells. Comparing CON and LI we found a significant negative correlation between the means and medians of the two metrics at the catchment scale, with $r_p = -0.98$: the larger the mean and median local input to the stream, the smaller the overall catchment convergence. A higher catchment convergence means that all catchment grid cells have more neighboring cells draining into them than catchments with a smaller convergence. CON can hence be considered a measure of catchment complexity or dissection, with less convergent catchments being less dissected, resulting in larger hillslope UAAs while more convergent catchments exhibit a higher degree of topographic dissection. For instance, the most convergent catchments at TCEF are also the steepest catchments. Topographic dissection increases at TCEF in a downstream direction. UTC, at the headwaters of TCEF, consists of large, relatively planar hillslopes with long flow path lengths and can be considered relatively simple in terms of topographic dissection. However, in a downstream direction, slopes increase, the stream becomes more incised, the riparian areas narrow, flow path lengths become shorter, local input decreases, and convergence increases leading to greater flow landscape dissection.

Another metric describing the potential for water accumulation and the associated potential energy is hillslope power (HP). In contrast to CON, HP is highly related to LI.

In addition to upslope area, HP incorporates slope and therefore the energy gradient associated with this water. Therefore, HP can be related to the potential rate of water movement. This may only apply to saturated soils and not necessarily to unsaturated soils, where additional forces (e.g. matric potential) influence water movement. However, soils at TCEF are saturated or near saturation for periods during snowmelt [*Jencso et al.*, 2009]. The correlation between HP and mean response time is negative, a higher HP was correlated with a lower mean response time (similar to SLP and GTC).

The final topographic metric, riparian buffering (RB), is also related to LI in that it is the dimensionless ratio of riparian area and the corresponding hillslope area that drains into it (riparian area / hillslope area). Its correlation with mean response time is negative ($r_p = -0.81$ for the median) but below the significance threshold, the correlation for the mean is even less strong ($r_p = 0.07$, Table 2.5). A negative correlation indicates that higher riparian buffering potentials are associated with shorter mean response times, a result that seems counterintuitive. One would expect a greater potential for volumetric buffering as indicated by *McGlynn and McDonnell* [2003], *McGlynn and Seibert* [2003] and *Jencso et al.* [2010] with more riparian area in relation to hillslope area. In fact, *Jencso et al.* [2010] observed a logarithmic relationship between riparian buffering potential and the turnover half-life time for the groundwater stored in riparian areas (the time it takes to replace 50% of riparian water with water from an adjacent hillslope). Based on the findings of *Jencso et al.* [2010], a positive correlation between RB and mean response time would have been expected. However, RB is a good example why it is important to differentiate between mean residence time (replacement of riparian water by

hillslope water) and mean response time (displacement of water in the saturated riparian zone by hillslope water).

Do Vegetation and Energy Availability have an Impact on Mean Response Times?

Vegetation affects watershed hydrology in many ways and at all times during the hydrologic year. Two important hydrologic processes affected by vegetation in snow-dominated catchments during the winter include snow interception and subsequent sublimation from the canopy [*Hedstrom and Pomeroy, 1998; Pomeroy et al., 2002; Pomeroy et al., 1998*]. During the growing season, vegetation acts as loss mechanism for soil moisture because of evapotranspiration [*Rodriguez-Iturbe et al., 1999*]. However, for snow-dominated systems, interception and sublimation are likely the dominant processes that affect mean response time at the annual scale.

Our results indicate that tree height (TH) and mean response time are positively correlated ($r_p = 0.97$), with a greater TH leading to a longer mean response time (Table 2.5 and Figure 2.8). TH was calculated by incorporating all grid cells, including clear cuts, thins, parks, and cells absent of trees in forests, which reduced the estimated catchment average tree height considerably but is a better surrogate for “tree-biomass” for a given catchment in a lodgepole pine dominated landscape. Mean response time can be affected by vegetation and TH in various ways, most notably in the accumulation and melt of snow in snow dominated systems. Forests often accumulate less snow [*Woods et al., 2006*], but they hold the snow longer than clearcut areas and parks [*Winkler et al., 2005*]. This behavior is attributed to attenuated energy input to the subcanopy snowpack

and enhanced canopy interception and sublimation. In hydrologic models attenuation of shortwave radiation by forest canopies is often calculated by incorporating LAI (e.g. *Wigmosta et al.* [1994]), with a greater LAI leading to less shortwave radiation penetrating the canopy, resulting in less energy to melt the subcanopy snowpack. LAI estimates in coniferous forests are often made through allometric relationships with tree height. *Keane et al.* [2005] established tree height-LAI relationships for TCEF for ground based LAI estimation methods and *Jensen et al.* [2008] found a linear relationship between LIDAR derived tree height and LAI for coniferous forests (among others consisting of pine and spruce species) in northern Idaho. Based on the relationship found by *Jensen et al.* [2008] and *Keane et al.* [2005], TH corresponds to a greater LAI and therefore increased potential to reduce incoming shortwave radiation received by the snowpack in forested areas, thus slowing down snowmelt and mean response time. The water input to the system as measured by the SNOTEL sites is potentially biased in that the SNOTEL stations are located in open areas and typically melt out earlier than forested sites due to differences in snowpack energy balances [*Woods et al.*, 2006]. Personal observations by the authors at TCEF during the last eight years corroborate that the forests hold snow longer than open areas. Therefore, the TH-mean response time relationship could be influenced by delayed melt in forests as compared to the SNOTEL sites.

In addition to the potential effect of TH on snowpack energy balances, topographically mediated potential solar insolation also varies across the 7 catchments. Potential solar insolation (INS) is a function of aspect, slope, and geographic latitude and describes the incoming shortwave radiation for each grid cell accumulated over one year.

In snow dominated catchments increased solar radiation can decrease snow accumulation by enhancing both sublimation of the snowpack and vegetation intercepted snowfall [Jost *et al.*, 2007; Lopez-Moreno and Stahli, 2008]. Therefore, INS can influence both the magnitude and timing of snowmelt. In general north facing slopes receive less incoming radiation, accumulate more snow, and release this water later in the melt period [Murray and Buttle, 2003]. This leads to a dampened yet sustained input through time, which increases mean response time. South facing catchments would exhibit the opposite behavior. Most catchments, however, contain a wide range of aspect and slope. Therefore, mixtures of catchment landscape attributes can effectively work in opposition thereby muting the effect of INS. As expected, the correlations between mean response time and mean and median INS are negative (higher INS leads to shorter mean response times) but with $r_p = -0.39$ for the mean and $r_p = -0.53$ for the median respectively they are both below the 95% significance threshold. INS is the potential solar insolation on the ground surface and does not incorporate any attenuating factors, i.e. vegetation. The effective insolation that reaches the ground surface may differ significantly from the calculated potential solar insolation because of interception by the vegetation. The strong correlation between TH and mean response time highlights the importance of vegetation on hydrologic response and also explains why solar radiation input to the ground surface as single metric might be insufficient for prediction of hydrologic response.

What is the Role of Geology on Hydrologic Response?

Bedrock geology can influence water chemistry (e.g. *Gardner and McGlynn* [2009], *Newton et al.* [1987] and *Gardner et al.* [in press]) and the routing of water downslope and downstream (e.g. *Freer et al.* [2002] and *Onda et al.* [2001]).

We evaluated the influence of geologic strata at TCEF to quantify the potential effect of different types of bedrock on mean response time. Strong, significant correlations between fraction of surficial bedrock geology and mean response time were observed for three of the four geologic layers at TCEF, with sandstone exhibiting the highest correlation, $r_p = 0.95$ (Table 2.5 and Figure 2.9). In contrast to the sandstone layer, correlations between mean response times and the shale and monzonite layers were negative. While sandstone is a sedimentary rock with relatively high hydraulic conductivities, both shale and monzonite (an igneous intrusive rock), have hydraulic conductivities several orders of magnitude lower than sandstone in unfractured conditions [*Freeze and Cherry*, 1979], likely leading to less recharge into deeper groundwater reservoirs. Potentially elevated seepage occurring into the sandstone relative to the shale and monzonite may reenter the shallow soil system and/or stream network further downslope. Because of longer travel times in the bedrock layer than in the (saturated) soil system, baseflow contributions in catchments with greater sandstone fractions may be higher and mean response times longer. *Jencso et al.* [in review] found that the intersection of hillslopes greater than 0.5 ha with the sandstone layer extended flow

duration curves for TCEF catchments. This is consistent with our findings that the sandstone fraction in a catchment is positively correlated with mean response time.

Can Intra-Catchment (i.e. Inter-Annual) Variability be Explained by Climate Forcing?

The transfer functions for both catchment averages over time (Figure 2.4) and annual averages of mean response time across catchments (Figure 2.5) exhibited considerable variability. While catchment averages of mean response time can be explained by topographic characteristics of each catchment (Figures 2.8 and 2.9), the question remains: what causes the differences observed in response time between the individual years (Figure 2.5)? Higher rainfall intensities generally lead to faster hydrologic response [Montgomery *et al.*, 1997], and antecedent moisture is often a key factor for low intensity events [Castillo *et al.*, 2003; Niehoff *et al.*, 2002]. At TCEF, higher input intensities that are typically a combination of snowmelt and rainfall, lead to faster annual mean response times (Table 2.6). We evaluated input intensities for temporal windows ranging from one to 21 days. We selected the period with the greatest intensity for each temporal window and correlated them with catchment mean response times. For these input intensities, the correlation was strongest for the 13 day window ($r_p = -0.76$, Table 2.6). However, somewhat surprisingly the simple metric of maximum annual snow water equivalent (SWE) exhibited a slightly stronger correlation ($r_p = -0.82$, Table 2.6). A power law regression between maximum annual SWE and mean response time indicated that that 75 % of the variability in annual mean response time could be explained by maximum annual SWE alone (Figure 2.10).

Since mean response time was calculated for each input signal it is likely that the size of the snowpack affects response times by shifting soils from unsaturated to saturated, increasing hydraulic conductivity, increasing the saturated thickness (higher water tables), and increasing hydraulic gradients toward the stream. Greater SWE could also lead to a greater number of consecutive melt inputs to the ground (i.e. days with melt), increasing soil moisture for the next input and maintaining higher perched water tables above the bedrock-soil interface. This would suggest that both the input intensities and the duration of melt events have an effect on the rate of hydrologic response in snow-dominated systems.

What is the Relationship Between Mean Response Times and Runoff Ratios in the Context of Topographic and Climatic Variability?

In our study, runoff ratios exhibited different behavior when they were calculated for catchment averages or annual averages. There was a strong negative relationship between runoff ratios and mean response time for catchment averages (Figure 2.11a). However, there was no visible pattern for annual runoff ratios versus annual mean response time (Figure 2.11b), suggesting that runoff ratios in a snowmelt dominated system like TCEF are determined by topography and not by climatic variability. This result is corroborated by non-significant relationships between maximum annual SWE and annual runoff ratios (Figure 2.12a) as well as annual precipitation amounts and annual runoff ratios (Figure 2.12b). This means that in general steeper catchments will have a higher runoff ratio compared to flatter watersheds but the ratio is independent of the amount of snowmelt. These results are counter to previous research, which indicated

runoff ratios largely depended on local climate, i.e. the availability of energy and precipitation [Arora, 2002; Budyko, 1974; Donohue *et al.*, 2007; Hewlett and Hibbert, 1967; Pike, 1964; Zhang *et al.*, 2001].

We can assume that independent of the actual input magnitudes the catchments wet up fairly quickly due to the large melt input in even the low precipitation years. SWE magnitudes may therefore be secondary to topographic influences in determining runoff ratios. However, this does not contradict our finding that SWE impacts mean response times. While response time is controlled by input intensities over longer time periods, annual runoff ratios appear to be mostly independent of max SWE and max input intensities due to the large magnitude of snowmelt events and the effect of ET and other melt/precipitation characteristics. These results suggest that even during small melt events, the velocity of subsurface flow depends mostly on the topography of each catchment, especially slope (i.e. Darcy's law), assuming homogenous soils with homogenous hydraulic conductivities across TCEF.

What Explains the Different Behavior of SUN and SPC?

We excluded both the SUN and SPC catchments initially because of the silvicultural treatments in 2001/2002 and the potential treatment impact on the hydrologic cycle. Many studies have focused on the effect of silvicultural treatments and disturbances on hydrologic response [Bosch and Hewlett, 1982; Cheng, 1989; DeBano, 2000; Miller, 1984; Swank and Douglass, 1974] and most studies found increased water yield and increased storm flow following clear cutting.

At TCEF, both SUN and SPC exhibited considerably longer mean response times than the other catchments and do not fit on any of the regressions shown in Figure 2.8. The harvests occurred in 2001 and 2002. We compared the mean response times of all catchments for both the pre and post logging periods (Table 2.7). The mean response times of SPC and SUN were longer than the mean response times of the remaining catchments for both the pre and post logging period, but more closely approached the response times of the other catchments in the post logging period. In general, mean response times decreased from the pre to post logging period for all catchments, which was likely caused by larger snowpacks in the post logging period, but the absolute decline is greatest for SPC and SUN. Therefore, patch cuts and thinned stands increased the responsiveness of the two managed catchments, however, they remained less responsive (longer mean response times) than the other catchments and less responsive than the landscape metric regressions would predict.

Longer mean response times in SPC and SUN suggest that factors other than those derived from the digital elevation model, vegetation, and fractional geology must influence catchment runoff behavior in these two catchments. LIDAR data capture surface topography in great detail but are unable to detect subsurface topography. *Payn et al.* [(in review)] have shown that subsurface contributing area in TCEF may differ from the contributing area derived from LIDAR data during low flow conditions. Also the fraction of geologic strata for each catchment does not reveal information about the dip of the layers, the thickness of the layers, nor the location of geologic transitions relative to runoff generation processes.

SPC covers the full range of geologic variability in TCEF with four different geologic layers. The transition from Wolsey shale to Flathead sandstone in SPC coincides with a noticeable break in slope (Figures 2.6a and 2.6h). The forested and clearcut hillslopes transition into a wide, open park (meadow). This park area is one of the wettest sites in TCEF with water tables close to the surface even in the late summer when most of TCEF has already dried down. *Payn et al.* [(in review)] analyzed spatial patterns of stream baseflow generation at TCEF and found steady or increasing contributions relative to the overall LTC runoff for the spring visible at the shale – sandstone interface during the summer dry down period in 2006. *Payn et al.* [(in review)] suggest extensive storage in the SPC ridge area, which would sustain a higher runoff in the summer baseflow period and lead to longer mean response times. Interestingly, an equal transition in MSC/LSC from shale to sandstone does not produce visible springs or wet areas, suggesting differences in the stratigraphic transition between the two catchments.

On the other hand, SUN comprises mostly sandstone with a small fraction of granite gneiss and does not exhibit noticeable transitions that would favor the formation of springs. The most notable geologic feature in SUN is a fault line (Quartzite ridge fault) that dissects SUN east to west in approximately equal portions. Runoff generated on the catchment portions above the fault could infiltrate along the fault into deeper geologic layers and reenter the system further downstream as return flow, thus increasing hydrologic response times. The fault also runs through BUB and UTC. However, the areas above the fault in these two catchments are significantly less than in SUN.

The tailing behavior of the transfer functions of SPC and SUN also suggests higher baseflow contributions as compared to the remaining catchments, which may be a

combined result of sustained groundwater contributions and reduced evapotranspiration (Figure 2.4c).

Conclusions

The use of runoff transfer functions provides a simple tool for assessing the hydrologic response of catchments and relating differences in response to landscape structure and climate. Based on our correlations using modeled mean response time and variables of landscape structure and climate we conclude:

- 1) Differences in mean response time among five of the TCEF catchments were caused by differences in landscape structure. There were no strong cross-correlations among the TCEF landscape variables, suggesting that the observed correlations are not just the effect of one or two variables superimposed onto the others. However, it is intriguing that metrics that are seemingly independent from each other show equally strong correlations with mean response time. This raises questions of how independent catchment characteristics are in fact related and form during landscape evolution.
- 2) Vegetation has an effect on mean response time most likely due to reduced snow accumulation, the shading of the subcanopy snowpack and the attenuation of solar radiation used to melt the snowpack.
- 3) The Sandstone layer at TCEF showed the strongest correlation with mean response time. Variable hydraulic conductivities of the geologic strata at TCEF

potentially lead to differing recharge rates into deeper groundwater layers, which also affected mean response time.

- 4) Inter-annual variability in mean response time was a result of variable climatic conditions, i.e. the maximum amount of annual snow water equivalent and differences in melt intensities
- 5) The two apparent outliers SPC and SUN suggest controls on hydrologic response in these watersheds other than metrics of surface topography, i.e. metrics than can be derived from digital elevation models. Geology was likely responsible for this unexpected behavior. However, we were unable to ultimately answer the question of why the two catchments show hydrologic response that does not fall in line with the relationships established for the other five catchments.
- 6) For these snow dominated systems runoff ratios were largely a function of topography. This means that in general steeper catchments had higher runoff ratios compared to flatter watersheds. The runoff ratio was independent of the amount of snowmelt.

Some of the correlations shown in this study are very strong and suggest that one single metric explains catchment runoff behavior (e.g. SLP and CON with an $r_p = -1.00$).

However, we emphasize that catchment heterogeneity is too complex to reduce the governing factors of hydrologic response to individual landscape structural metrics. It is rather the interaction of several metrics that ultimately determines how a watershed responds to a precipitation input.

We tested the transfer function rainfall-runoff modeling framework in a snowmelt-dominated system of the northern Rocky Mountains. Future applications of the

model and analysis approach could involve comparisons of catchments in other places to test whether the relationships we established at TCEF also hold for different landscapes, climatic settings and scales.

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Tables

Table 2.1: Landscape metrics used in the analysis, their abbreviations and definitions

Landscape metric	Abbrev.	Definition
Slope (°)	SLP	Slope of a grid cell based on the greatest elevation difference with its eight neighboring cells.
Convergence (%)	CON	Convergence or divergence of a grid cell using the surrounding eight cells; 100 % convergence means all surrounding grid cells to flow into the center cell, 100 % divergence being all surrounding cells facing away from the center cell. After [Moore <i>et al.</i> , 2001]
Distance from creek (m)	DFC	The distance from a grid cell to the stream along a flow path
Gradient to creek (-)	GTC	The gradient of a flow path
Local input (ha)	LI	Upslope accumulated area (UAA) of a grid cell bordering the stream. [Grabs <i>et al.</i> , 2010; Seibert and McGlynn, 2007]
Hillslope power (-)	HP	UAA of a grid cell times the slope of the contributing area
Riparian buffering (-)	RP	Ratio of riparian area to hillslope area [B L McGlynn and Seibert, 2003]
Tree height (m)	TH	Average pixel elevation (tree height) above ground surface including the silvicultural treatment areas, calculated by subtracting the last return from the first return LIDAR coverage
Potential Solar Insolation (kWhm ⁻² y ⁻¹)	INS	Potential annual solar insolation calculated for each grid cell. [Böhner and Antonic, 2009]
Geology	GEO	Different geologic strata [Reynolds and Brandt, 2006]

Table 2.2: Modeled mean response times (days) for the seven TCEF catchments, averaged over the 12 simulated years

Watershed	BUB	LSC	LTC	MSC	SPC	SUN	UTC
Days	13.23	9.56	11.00	10.72	18.67	18.27	14.56

Table 2.3: Modeled mean response times (days) for each individual year, averaged over the seven catchments

Year	1996	1997	1998	1999	2000	2001	2002	2003	2004	2005	2006	2007
Days	12.66	9.76	24.73	13.22	13.81	24.95	17.44	11.95	7.16	10.44	10.13	8.34

Table 2.4: Topographic parameters for each catchment (means and medians with percent of catchment area for geology). The median for TH was 0 and is not shown here.

	BUB	LSC	LTC	MSC	SPC	SUN	UTC
SLP (°)							
<i>Mean</i>	6.5	9.5	8.0	8.6	9.5	6.5	5.4
<i>Median</i>	5.3	8.3	6.3	8.1	8.6	5.1	4.0
CON (%)							
<i>Mean</i>	-0.020	-0.016	-0.014	-0.013	-0.029	-0.032	-0.026
<i>Median</i>	0.344	0.601	0.511	0.550	0.712	0.256	0.231
DFC (m)							
<i>Mean</i>	575	452	545	491	440	691	714
<i>Median</i>	515	396	466	439	394	642	614
GTC (-)							
<i>Mean</i>	0.135	0.168	0.142	0.140	0.153	0.104	0.088
<i>Median</i>	0.104	0.149	0.119	0.140	0.145	0.094	0.072
LI (m ²)							
<i>Mean</i>	6511	7280	7343	8480	6645	10546	10058
<i>Median</i>	3345	2357	2403	2367	1695	3326	3510
HP (-)							
<i>Mean</i>	517	595	547	594	552	563	450
<i>Median</i>	215	313	248	310	238	221	191
RB (-)							
<i>Mean</i>	0.69	0.045	0.465	0.053	0.115	0.044	0.043
<i>Median</i>	0.022	0.028	0.028	0.031	0.043	0.024	0.024
TH (m)							
<i>Mean</i>	3.65	3.01	3.03	3.10	2.51	2.41	3.83
INS (kWh/year)							
<i>Mean</i>	3392	3428	3418	3442	3463	3401	3427
<i>Median</i>	3419	3461	3443	3461	3487	3430	3450
GEO (% catchment area)							
<i>Sandstone</i>	92.9	56.2	70.3	54.7	31.2	91.4	97.6
<i>Wolsey shale</i>	0	19	9.8	19.9	26.7	0	2.2
<i>Monzonite</i>	0	15.3	10.8	23.8	40.4	0	0

Table 2.5: Correlations between the modeled mean response time averaged over the five catchments without SPC and SUN and the topographic parameters. Significance level for a two-tailed test with $n = 7$ and $\alpha = 0.01$ equals $r_p = \pm 0.87$ and $\alpha = 0.05$ equals $r_p = \pm 0.75$. Significance level for a two tailed test with $n = 5$ and $\alpha = 0.05$ (bold) equals $r_p = \pm 0.88$ and $\alpha = 0.01$ (underlined) equals $r_p = \pm 0.96$.

	All catchments ($n = 7$)			W/o SUN and SPC ($n = 5$)		
SLP	Mean	Median		Mean	Median	
r_p	-0.25	-0.23		<u>-1.00</u>	<u>-0.95</u>	
CON	Mean	Median		Mean	Median	
r_p	-0.94	-0.20		-0.87	<u>-1.00</u>	
DFC	Mean	Median		Mean	Median	
r_p	0.34	0.42		0.95	<u>0.98</u>	
GTC	Mean	Median		Mean	Median	
r_p	-0.43	-0.32		-0.91	<u>-0.97</u>	
LI	Mean	Median		Mean	Median	
r_p	0.30	0.10		0.46	<u>0.96</u>	
HP	Mean	Median		Mean	Median	
r_p	-0.28	-0.64		-0.95	-0.93	
RB	Mean	Median		Mean	Median	
r_p	0.46	0.27		0.07	-0.81	
TH	Mean			Mean		
r_p	-0.45			<u>0.97</u>		
INS	Mean	Median		Mean	Median	
r_p	0.11	0.10		-0.39	-0.53	
GEO	Shale	Sandst.	Monz.	Shale	Sandst.	Monz.
r_p	-0.13	0.04	0.11	-0.89	0.95	-0.83

Table 2.6: Maximum SWE elevation weighted for LTC and maximum calculated input intensities over 13 consecutive days and r_p with the averaged response time for each year. Significance level for $n = 12$ and $\alpha = 0.01$ equals $r_p = \pm 0.71$.

Year	Max SWE Elevation weighted (mm)	Max. input intensity over 13 days (mm/day)
1996	387	21.0
1997	467	23.1
1998	306	16.3
1999	319	19.2
2000	364	16.9
2001	282	16.3
2002	337	17.6
2005	375	17.2
2006	446	22.8
2007	352	20.5
2008	395	18.2
2009	408	21.5
Correlations with mean response time	-0.82	-0.76

Table 2.7: Comparison of mean response times for the entire simulation period and the pre and post logging period

	BUB	LSC	MSC	LTC	SPC	SUN	UTC	Average
1996-2009	13.23	9.56	11.00	10.72	18.66	18.27	14.56	13.72
Pre logging	15.52	10.58	14.76	13.64	23.34	22.60	15.21	16.52
Post logging	10.93	8.55	7.23	7.81	13.99	13.94	13.92	10.91

Figures

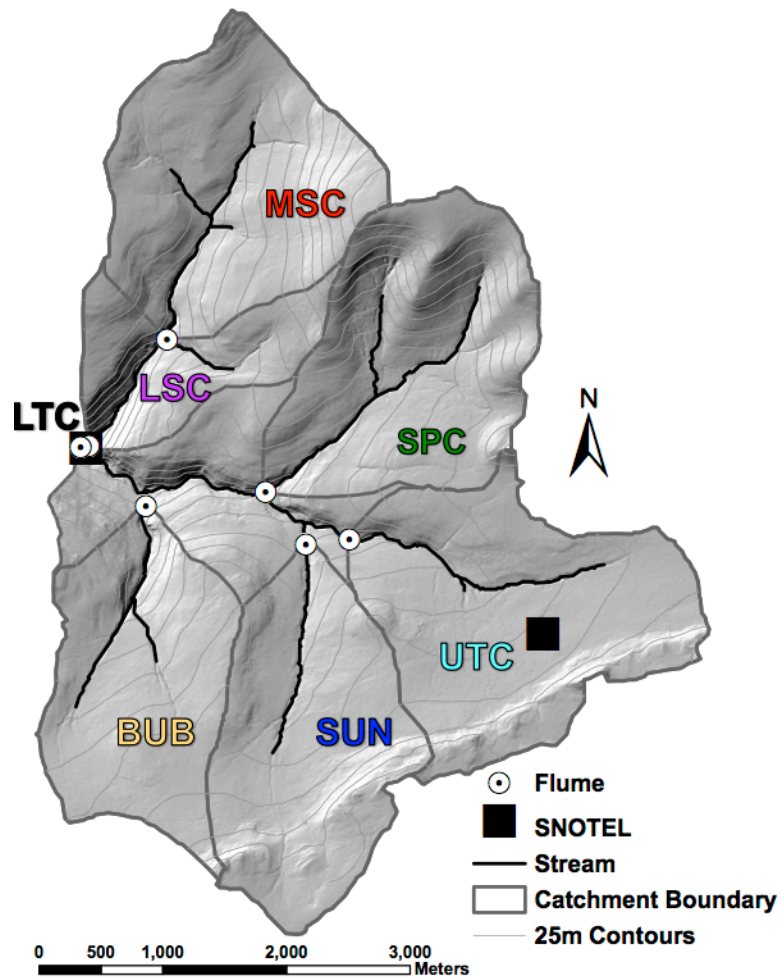


Figure 2.1: Tenderfoot Creek Experimental Forest with two SNOTEL sites and seven gauged catchments. The color coding remains the same for following figures. LTC=Lower Tenderfoot Creek; LSC=Lower Stringer Creek; MSC=Middle Stringer Creek; SPC=Spring Park Creek; UTC=Upper Tenderfoot Creek; SUN=Sun Creek; BUB=Bubbling Creek

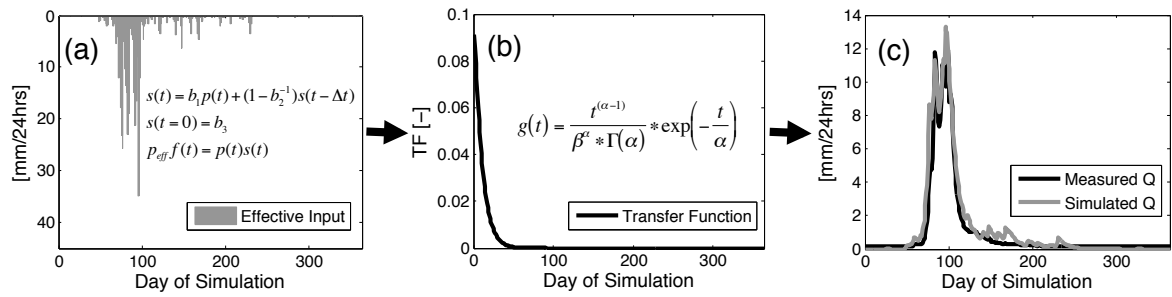


Figure 2.2: Flow chart representation of model implementation: (a) A measured input is transformed into effective precipitation by the loss function; (b) convolution of a calibrated transfer function and the effective precipitation results in (c) discharge. For our study, the parameters of the loss function and the parameters of the transfer function were optimized in 150,000 Monte Carlo simulations.

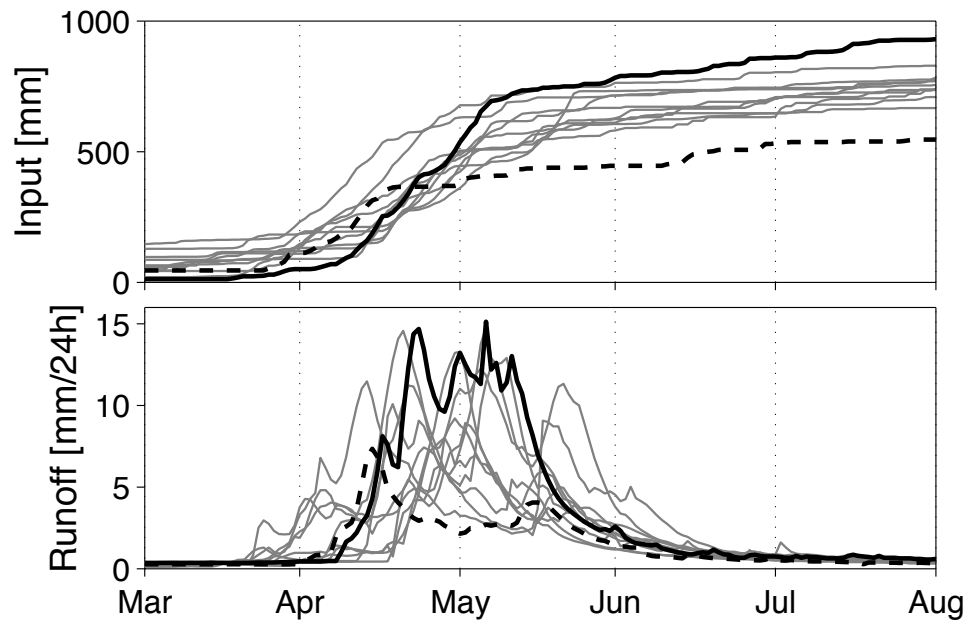


Figure 2.3: (a) Accumulated input (rainfall and snow melt) and (b) variations in discharge from 1996 - 2009. Year with max SWE (1997) denoted by black, bold line, year with minimum SWE (2001) denoted by black, dashed line.

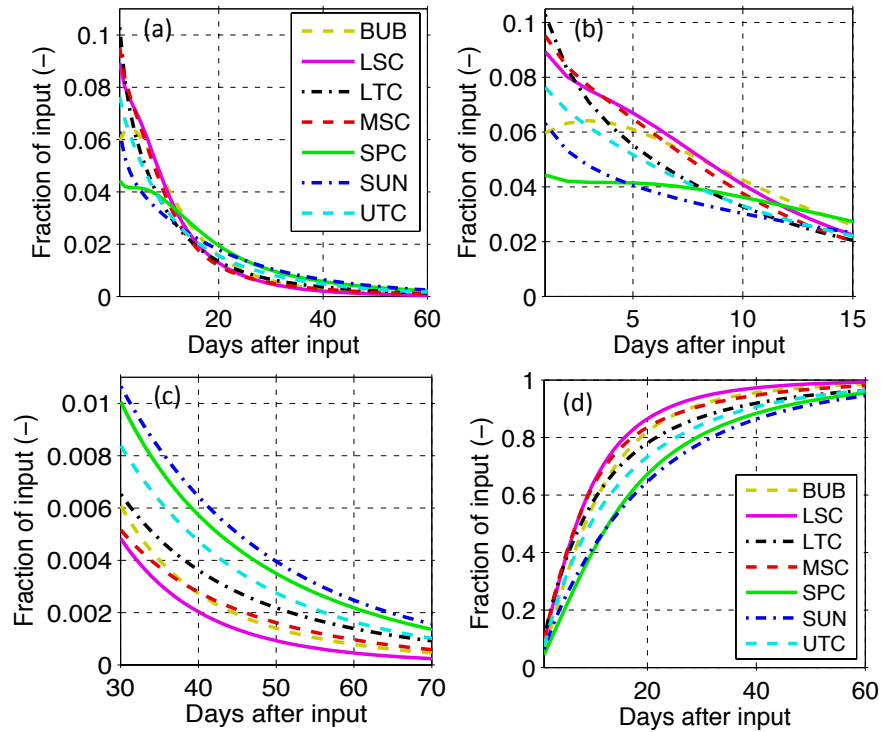


Figure 2.4: (a) Transfer functions for each catchment averaged over 12 simulated years; the y-axis denotes the fraction of the input that left the watershed at a certain day; (b) short term behavior of the transfer functions (first 15 days); (c) long term (baseflow) behavior of the transfer functions (days 30 through 70). Note the order of magnitude difference on the y-axis as compared to the remaining panels; (d) cumulative distribution transfer functions for each catchment averaged over the 12 year simulation period; the y-axis denotes the fraction of an input that left the watershed by a certain time (x-axis); the steeper the curve at short times, the faster the precipitation to runoff response; an elevated curve later on indicates higher baseflow contributions.

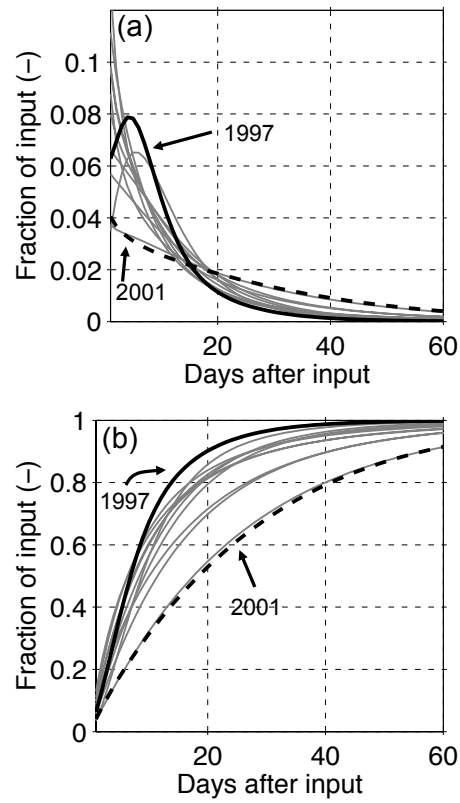


Figure 2.5: (a) Transfer functions and (b) cumulative distribution transfer functions averaged over all seven catchments for each of the 12 years simulated; the black solid lines denote the year with highest SWE (1997), the black dotted lines denote the year with lowest SWE (2001).

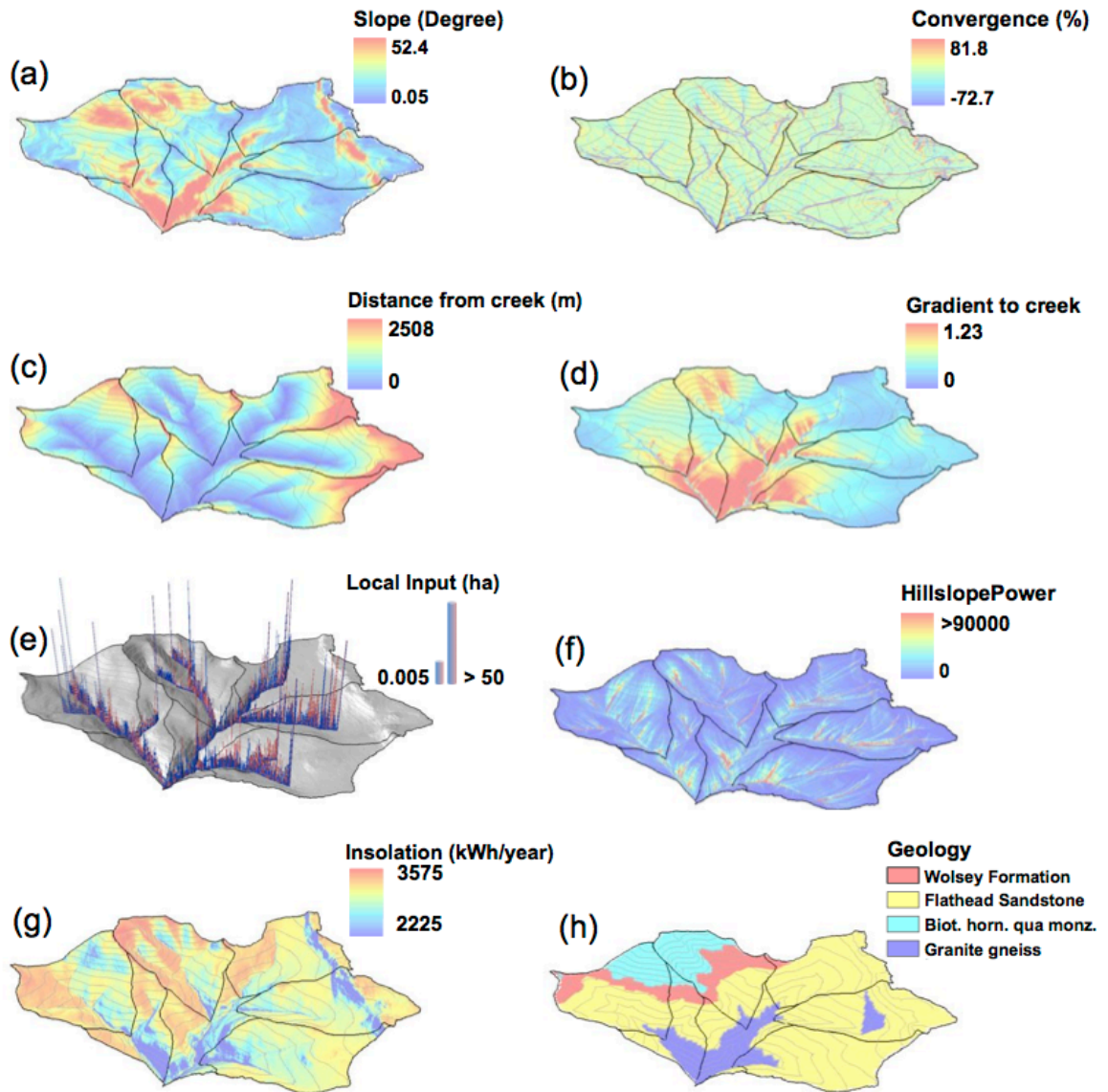


Figure 2.6: Map representation of topographic parameters for seven TCEF catchments, including a) slope, b) convergence, c) distance from creek, d) gradient to creek e) local input, f) hillslope power, g) insolation, and h) geologic strata.

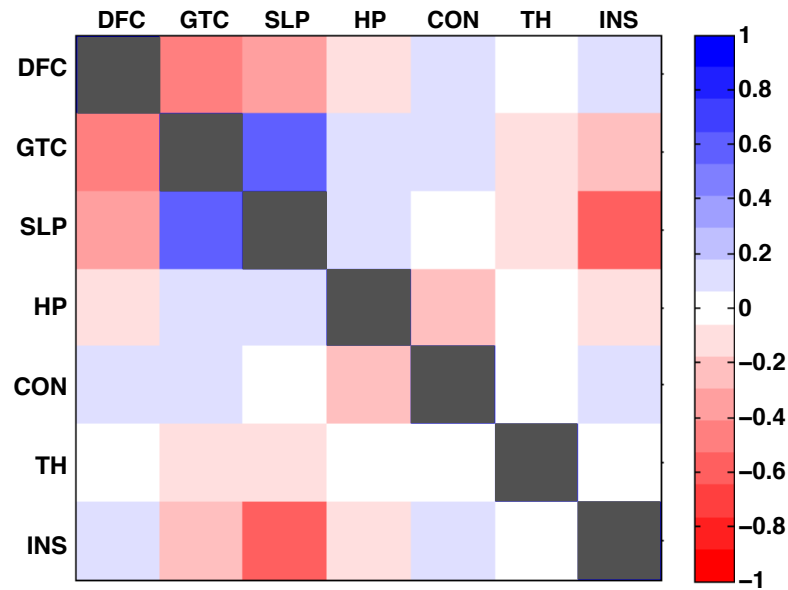


Figure 2.7: Cross-correlation matrix (Pearson's correlation coefficient) between seven landscape metrics. Correlations were calculated on a grid cell-to-grid cell basis using each 1 m grid cell across coverage. Blue colors indicate positive correlations, red colors indicate negative correlations. The gray diagonal is the correlation of a landscape metric with itself.

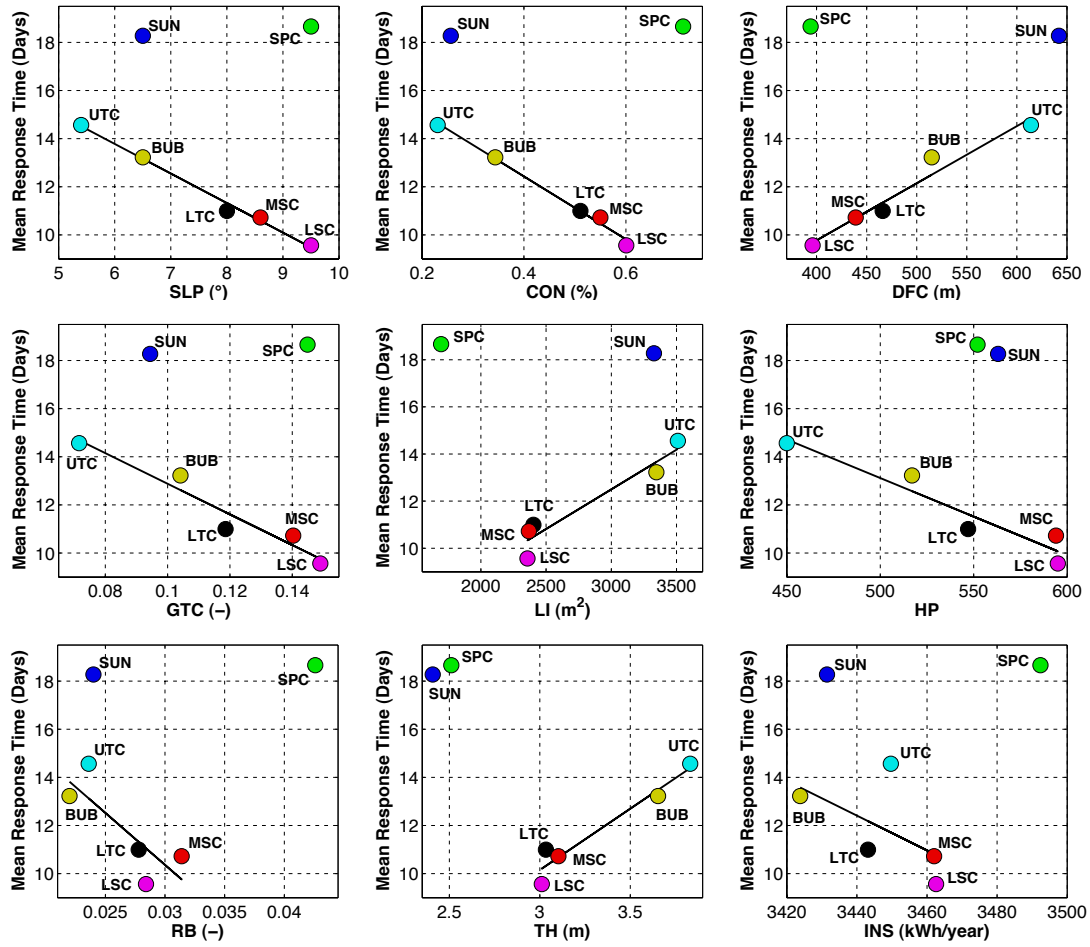


Figure 2.8: Regressions between mean response time (days) and topographic parameters. SUN and SPC have been excluded from the regressions but are shown for completeness; each symbol represents the averaged mean response time for one catchment over 12 years.

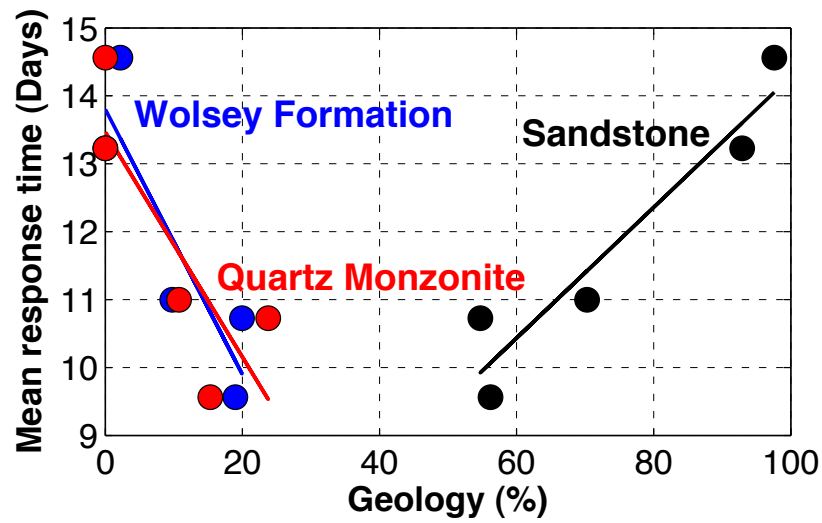


Figure 2.9: Regressions between the mean response time for the five catchments (BUB, LSC, LTC, MSC and UTC) and the percentage of each geologic strata present in each catchment.

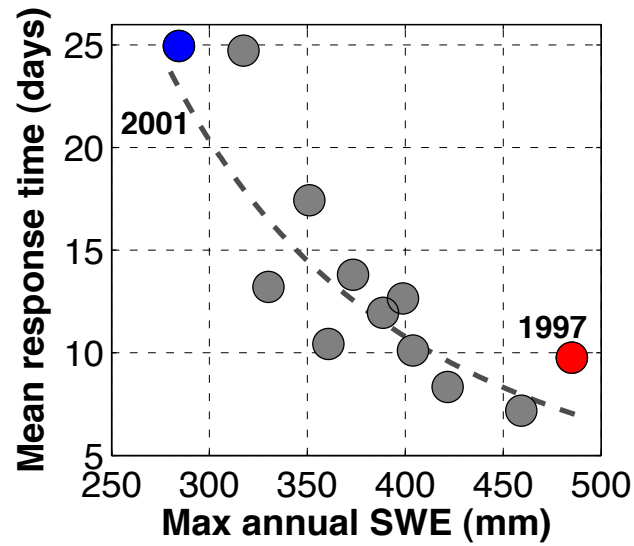


Figure 2.10: Relationship between maximum annual SWE (mm) and mean response time (days) averaged over all seven catchments for each of the 12 years of simulation. The gray dashed line is a power law regression ($R^2=0.75$). The higher the maximum annual SWE, the faster the response time. The blue symbol denotes the year with lowest SWE (2001), the red symbol denotes the year with highest SWE (1997).

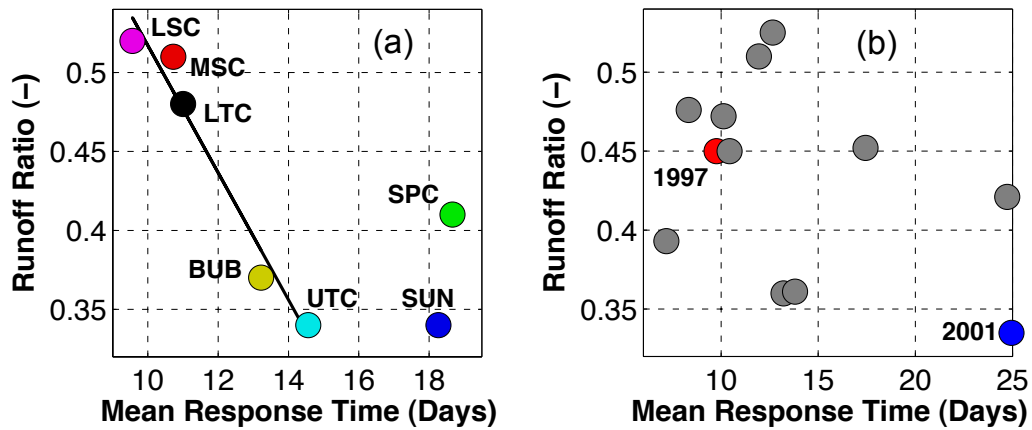


Figure 2.11: (a) Mean response time for catchment averages plotted against catchment runoff ratios (each point represents an average over 12 years). Omitting SUN and SPC there is a strong negative correlation, with smaller runoff ratios associated to longer mean response times; (b) Mean response time for annual averages plotted against annual averaged runoff ratios (each point represents an average of 7 catchments for each year) with no significant correlation (p -value = 0.19). The blue symbol denotes the year with lowest SWE (2001), the red symbol denotes the year with highest SWE (1997).

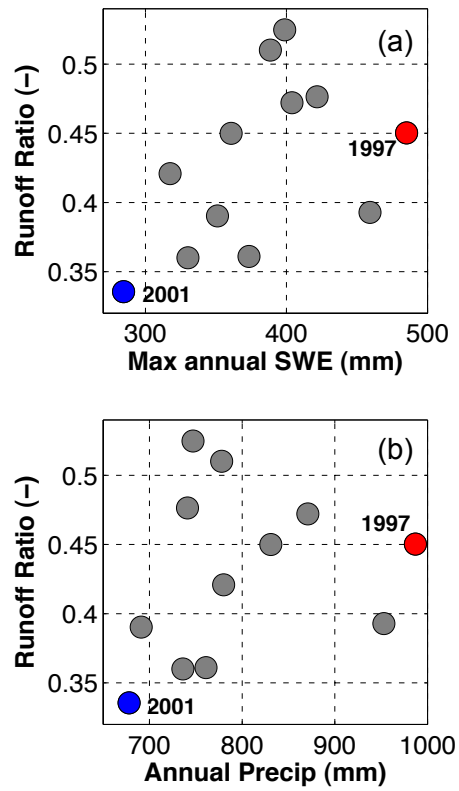


Figure 2.12: (a) Maximum annual SWE versus average annual runoff ratios, $r_p = 0.48$ and $p\text{-value} = 0.11$ (b) Annual precipitation versus average annual runoff ratios, $r_p = 0.23$ and $p\text{-value} = 0.47$. Both correlations are non-significant.

3. WATERSHED MEMORY AT THE COWEETA HYDROLOGIC LABORATORY:
THE EFFECT OF PAST PRECIPITATION AND STORAGE ON
HYDROLOGIC RESPONSE

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Contributions: Facilitated data and partial funding access, contributed critique and ideas at intermediate stage of project progress, commented on final version of the manuscript.

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- ☒ Prepared for submission to a peer-reviewed journal
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- ☐ Accepted by a peer-reviewed journal
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Abstract

The rainfall-runoff response of watersheds is affected by the legacy of past hydroclimatic conditions. We examined how variability in precipitation affected streamflow using 21 years of daily streamflow and precipitation data from five watersheds at the Coweeta Hydrologic Laboratory in southwestern North Carolina, USA. The gauged watersheds contained both coniferous and deciduous vegetation, dominant north and south aspects, and differing precipitation magnitudes. Lag-correlations between precipitation and runoff ratios across a range of temporal resolutions indicated strong influence of past precipitation (i.e. watershed memory). There was a strong runoff ratio dependency on the precipitation of previous time steps at all time scales. At monthly time scales the influence of past precipitation was detectable for up to seven months. At seasonal time scales the previous season had a greater effect on a season's runoff ratio than the same season's precipitation. At annual time steps we determined that the previous year was equally important for a year's runoff ratio than the same year's precipitation. Estimated watershed storage through time and specifically the previous year's storage state was strongly correlated with the residuals of a regression between annual precipitation and annual runoff, partially explaining observed variability in annual runoff in watersheds with deep soils. This effect was less pronounced in the steepest watershed that also contained shallow soils. We suggest that the location of a watershed on a non-linear watershed-scale storage-release curve can explain differences in runoff during growing and dormant season between watersheds with different annual evapotranspiration.

Introduction

Variability in hydrologic response within and among watersheds has long been attributed to watershed physical and biological properties (e.g. *Hewlett and Hibbert* [1967]; *Sidle et al.* [1995]; *Western et al.* [1999]; *Jencso et al.* [2009]; *Jencso and McGlynn* [2011]; *Nippgen et al.* [2011]), and climate (e.g. *Budyko* [1974]; *Arora* [2002]; *Jones et al.* [2012]). While our understanding of the role of spatial variability on runoff has increased, understanding the influence of temporal variability has received less attention, especially the consideration of past precipitation.

Some of the earliest and simplest methods to account for past precipitation are antecedent precipitation or wetness indices (API) that take into consideration the amount of precipitation before some point in time, e.g. a runoff event. APIs have been most often applied on a storm basis (e.g. *Linsley et al.* [1949]; *Sittner et al.* [1969]; *Fedora and Beschta* [1989]; *Sidle et al.* [2000]; *Kim et al.* [2005]) and rarely take into account time periods longer than a few weeks or months prior to the hydrologic event in question. *Fedora and Beschta* [1989] for example found that the effect of antecedent precipitation on runoff events vanished after 72 hours in several watersheds in the north-central Oregon Coast Range. Nevertheless, many studies point out the importance of soil water storage on runoff processes, from individual events to annual water balances. *Pathiraja et al.* [2012] for example showed that incorporating antecedent moisture conditions over longer periods (months) reduced underestimates of design flood peaks for watersheds in Australia. *Jothityangkoon and Sivapalan* [2009] demonstrated that incorporating carry-over storage from previous storms improved the accuracy of model-based annual water

balance predictions in watersheds in Australia and New Zealand. In a modeling study using 30-day lag correlations, *Orth and Seneviratne* [2013] showed that variability in shallow soil moisture was propagated to streamflow in 100 watersheds across Europe. On an even longer time scale, *Istanbulluoglu et al.* [2012] and *Tomasella et al.* [2008] found that precipitation variability led to a carry-over of groundwater storage that affected the water balance of the following year for watersheds in Nebraska and the Amazon, respectively. This carry-over groundwater could possibly sustain baseflow in years with less than average precipitation, acting as a low flow buffer. Understanding the capacity for hydrologic systems to buffer precipitation fluctuations is especially important considering projections of increased drought severity in the coming decades (5th IPCC report, *Hartmann et al.* [2013]). Such knowledge could have far reaching implications for policymakers and water resources managers as they anticipate and manage for increasing water scarcity. Despite the importance of this issue, quantifying the temporal dimension of past precipitation's effect on the current water balance of a watershed has received only modest attention.

The Coweeta Hydrologic Laboratory in the Appalachian Mountains of western North Carolina is one of the longest-running forest hydrology research sites in the United States. While many of the hydrologic experiments conducted at Coweeta investigated the influence of vegetation on water yield, some of the earliest research addressed the effect of soil storage on saturated and unsaturated flow (*Hursh and Brater* [1941], *Hoover and Hursh* [1943]). In what is now an iconic experiment, *Hewlett and Hibbert* [1963] demonstrated that $\sim 1 \text{ m}^3$ of saturated soil in a $\sim 14 \text{ m}$ long, sloping concrete trough (closed at the top to eliminate evaporation from the soil surface) was able to sustain

outflow for more than 140 days. *Hewlett and Hibbert* [1963] estimated that if the trough outflow after 140 days were scaled up to the size of a headwater catchment, it would be sufficient to constitute low flows observed during growing seasons in Coweeta watersheds. In the *Hewlett and Hibbert* [1963] experiment, the soil in the trough was saturated and then simply drained without being recharged again by sprinkling or rainfall. Natural watersheds, on the other hand, experience successive cycles of precipitation events and seasons. Transferring what we learned from the *Hewlett and Hibbert* [1963] experiment to the watershed scale at Coweeta and elsewhere naturally raises the question of how precipitation history, with cycles of wet and dry periods, affects soil water storage and how this precipitation legacy in return influences contemporary and future hydrologic response. Here, based on 21 years of precipitation and runoff data made available from five watersheds of the Coweeta Hydrologic Laboratory in southwestern North Carolina, we address the following questions:

- 1) How do variability in precipitation and system memory modulate hydrologic response across monthly, seasonal, and annual time scales?
- 2) What is the influence of storage state on the annual water balance and how different are storage dynamics under coniferous and deciduous vegetation?

Methods

Site Description

The Coweeta Hydrologic Laboratory is located in southwestern North Carolina in the Nantahala Mountain Range of the southern Appalachian Mountains (lat. 35°03'N, long.

83°26'W, Figure 3.1). Coweeta was established in 1934 and is a site of ongoing watershed experiments that include understanding hydrologic and ecologic changes following vegetation manipulations, such as conversion from deciduous hardwoods to evergreen conifers [Swank and Crossley, 1988]. The Coweeta Basin encompasses 1626 ha with watershed elevations ranging from 675 m near the outlet of Coweeta Creek in the east to 1592 m on the ridge in the west. For this study we were provided with runoff and precipitation data from five watersheds of the 16 currently gauged watersheds in the basin. WS01 and WS02 are south facing paired watersheds, and WS17 and WS18 are north facing paired watersheds. While WS02 and WS18 contain mixed hardwoods generally found in the southern Appalachians, WS01 and WS17 were converted to eastern white pine (*Pinus strobus*) trees (evergreen conifers) in the 1950s. At the beginning of the conversion process, WS01 was burned in 1942 and subsequent regrowth of hardwoods was prevented with chemicals until white pine was planted in 1957. WS17 was clear-felled in 1940 and subsequent regrowth was prevented until white pine was planted in 1956 [Swank and Crossley Jr., 1988]. WS36, the fifth watershed, is a high elevation (1015-1541 m) east facing watershed consisting of mixed hardwoods (Figure 3.1).

Watersheds 01, 02, 17, and 18 are similar in size (12.3 - 15.4 ha) and are located in the lower (elevations ranging from 704-1051 m), eastern part of the Coweeta basin, while WS36 is 49.3 ha and located in the higher, western part of the Coweeta basin. All watersheds are generally steep with mean slopes exceeding 26°. WS36 is the steepest of the five watersheds with a mean slope of >30° (Table 3.1).

Climate at Coweeta is classified as Marine, Humid Temperate (Köppen classification, Cfb) to Humid Subtropical (Cfa) [Swift *et al.*, 1988], especially at lower elevations. Annual precipitation at a weather station in the valley floor near the main basin outlet averages 1791 mm for the period 1937-2011. Precipitation is almost uniformly distributed over the year with slightly more (10-14%) precipitation in the winter months. There is a strong elevation effect on precipitation along the east-west axis with an increase of approximately 5% per 100 m, while this effect is almost negligible for the north-south axis slopes [Swift *et al.*, 1988]. Average annual air temperature near the main basin outlet is 12.6°C with an average monthly low of 3.3°C in January and a high of 21.6°C in July.

The soils at Coweeta are deep sandy loam inceptisols and older more developed ultisols [Swank and Crossley Jr., 1988]. Hoover and Hursh [1943] drew a boundary at approximately 1000 m elevation above which the watersheds are typically steeper and have shallower soils. According to Swank and Douglass [1975] the general depth of the regolith is approximately 7 m. Shallow groundwater well installations by the authors in the low-elevation watersheds 01, 02, 17, and 18 at Coweeta resulted in completion depths of up to 3.5 meters to bedrock, while Hales *et al.* [2009] found soil depths in the high elevation watershed 36 to be only 1.2-1.8 m, based on the excavation of 15 soil pits in various landscape positions.

Hydrometric Measurements

Runoff data were stored as breakpoint data for the water years (WY) 1991-2011. Runoff was measured using 90° V-notch weirs at WSs 01, 02, and 17, and with 120° V-notch weirs at WSs 18 and 36 and recorded using Fisher-Porter Analog to Digital punched recorders and Stevens Type A/E loggers. The breakpoint data were converted to daily runoff totals (mm/day). Runoff data were missing for WY 2011 for WS02 and WYs 1991-1993 for WS36. Missing data for WS17 (2003/08/14-2004/01/27) and WS18 (2003/09/03-2004/06/06) were interpolated using double mass curves with WS01 and WS02, respectively. However, only five out of a total 105 watershed years were missing.

Precipitation data from three rain gauges associated with the watersheds (see Table 3.2 for rain gauge-watershed associations) were available as daily precipitation totals for the same time period (WYs 1991-2011). Precipitation was measured using Belfort Universal Recording Rain Gauges and converted to watershed precipitation with weighting factors established by Swift in 1968 (see *Swift et al.* [1988]) from isohyetal maps based on a network of ~50 recording rain gauges in operation from 1938 to 1958 [*Swift et al.*, 1988].

Landscape Analysis

We calculated the means and medians of three simple topographic metrics: elevation, slope, and gradient to creek (the gradient of flowpaths from a cell to the stream). The metrics were calculated based on 1 m LIDAR (Light Detection And Ranging) data, which were resampled to 5 m resolution to obtain a resolution high enough to adequately

capture the topography in the small watersheds but coarse enough to avoid oversensitivity to microtopography (e.g. fallen trees). To assess differences in available solar radiation among watersheds, we calculated potential solar insolation after *Böhner and Antonic* [2009] for all five watersheds using SAGA open source GIS software (<http://www.saga-gis.org>). Potential solar insolation and thus energy availability can be used as a proxy for differences in potential evapotranspiration between the watersheds, especially between the north- and south-facing pairs. This is of particular interest here because the structure of the Coweeta basin contains both north and south facing watershed pairs, each with different vegetation types. We calculated potential insolation for all watersheds on an annual and seasonal basis.

Empirical Analysis of Runoff and Precipitation Data

We analyzed the runoff and precipitation time series at annual, seasonal, and monthly resolutions. We calculated monthly totals and averages of precipitation (P), runoff (Q), and runoff ratios (RR) and their standard errors to compare the runoff and precipitation regimes of the five watersheds. We calculated the difference in monthly Q totals between the deciduous watersheds and the coniferous watersheds to investigate potential differences in hydrologic response between the two vegetation types.

We calculated Pearson correlation coefficients between P and RR at annual, seasonal, and monthly time scales. In order to quantify the impact of past precipitation –and hence watershed memory– for each watershed, the correlations were calculated as lag-correlations: A lag of 0 means that one time step's RR was correlated with the same time

step's P , at a lag of 1 a time step's RR was correlated with the previous time step's P etc. We calculated lag correlations for up to 12 time steps (i.e. 12 months, 12 seasons, 12 years).

For seasonal statistics, we grouped data into growing season and dormant season, with growing season extending from April 15 to October 14 and dormant season extending from October 15 to April 14 of the following year. This is in agreement with the definitions previously used at Coweeta (e.g. *Vose and Swank* [1994], *Jones and Post* [2004]). While this division is appropriate for the low elevation watersheds, the seasons in WS36 are slightly different. *Hwang et al.* [2011] showed that the timing of leaf-on and leaf-off can differ by several weeks between the low and high elevations at Coweeta. However, for a consistent comparison (i.e. equal length growing and dormant seasons), we assigned all watersheds the same dates for growing and dormant seasons. As a consequence, potential seasonal effects on runoff in WS36 may be weakened as the actual growing season length in WS36 is possibly shorter than the chosen six months period.

In addition to correlations between runoff ratios and precipitation at different temporal scales, we approximated annual watershed storage and the changes in storage in each year from the annual 21-year P and Q time series. We developed a simple method (see equation 2 below) to estimate annual watershed storage using annual P and Q from the 21-year time series, and average ET across the time series. This is in contrast to the more typical water balance approach to calculating ET by assuming no annual changes on storage. Our approach assumes that annual evapotranspiration is relatively constant from year to year in this system where energy rather than water limits evapotranspiration.

Coweeta watersheds are primarily energy limited, as average annual precipitation (~1800 mm) typically exceeds estimates of potential annual evapotranspiration (e.g., 1400-1600 mm for WS17 [Rao *et al.*, 2011]). Relatively constant annual ET values have been reported for many areas (both water and energy limited) in and outside the US (e.g. Wilson *et al.* [2001]; Hanson *et al.* [2004]; Lafleur *et al.* [2005]; Nagler *et al.* [2005]; Kosugi and Katsuyama [2007]; Ohta *et al.* [2008]; Ryu *et al.* [2008]; Jassal *et al.* [2009]). The coefficient of variation in annual ET for the aforementioned studies (where data was published) ranged from 0.02 to 0.14, with the highest coefficient of variation originating from a grasslands site with Mediterranean climate in the lower Sierra Nevada of California and [Ryu *et al.*, 2008].

With typical water balance applications over annual and multi-year time scales ΔS is assumed to be negligible (e.g. Eagleson [1978], Budyko [1974]) and differences between $P-Q$ are attributed to variability in ET . Here we suggest that annual ΔS is much greater than ΔET and therefore we use average ET for each watershed calculated from a 21-year period to estimate annual ΔS and infer the influence of storage on annual Q dynamics. The percolation to deeper groundwater was assumed to be negligible at Coweeta because of impermeable bedrock [Hatcher, 1988]. ET_{mean} for each watershed over the entire study period was calculated as

$$ET_{mean} = \frac{\sum_{i=1}^n (P_i - Q_i)}{n} \quad (1)$$

where n is the number of years in the study period.

Storage, S , was allowed to vary in each year i as

$$\Delta S = P_i - Q_i - ET_{mean} \quad (2)$$

where ΔS_i is the change in annual storage (mm) for each watershed, P_i is annual precipitation (mm), Q_i is annual runoff (mm) and ET_{mean} is the average annual ET for a watershed (mm) over the 21-year study period. ΔS represents the annual deviation from the 21-year mean, whose actual value is undetermined. Cumulative storage, S , for each year and watershed was subsequently calculated as

$$S_i = \sum_{i=1}^n \Delta S_i \quad (3)$$

We used linear regressions between annual P and Q and calculated the residuals of this relationship to examine factors that might explain the variability of this relationship from year to year (e.g. S). For example, a watershed with less storage capacity might exhibit smaller residuals than a watershed with high storage capacity. To test this idea we calculated lag-correlations for the residuals of the regression between annual P and annual Q , and annual watershed storage, i.e. a lag of 0 is the correlation with one year's residual to the same year's storage S , a lag of 1 is the correlation between a year's residual and the previous year's storage term. Based on the storage calculation one year's storage denotes the following year's storage starting point, i.e. initial conditions. All correlations were calculated as Pearson correlation coefficients, r_p .

Results

Landscape Analysis

Based on their location and elevation within the greater Coweeta watershed, WSs 01, 02, 17, and 18 can be characterized as low-elevation watersheds while WS36 is a high-elevation watershed. The four low-elevation watersheds are structurally very similar

(Table 3.1), differing primarily in aspect. WS36, however, is steeper and larger (3-4 times larger in area) than the four other watersheds.

As expected, the south-facing watersheds receive more potential solar insolation over the course of a year than the north-facing ones (Table 3.1). The difference between the aspects is especially pronounced in the dormant season, when the south-facing watersheds (WS01 and WS02) receive on average 79% more direct solar radiation than the north facing ones (WS17 and WS18). This difference decreases to only about 16% more radiation during the growing season.

Climate and Lag-Correlations

The daily data revealed clear seasonality with high runoff in the winter/spring and a period of decreased flows in the summer (Figure 3.2.3.2). The amplitude of this seasonality was variable with pronounced wet (1994-1998) and dry (1999-2002, 2006-2008) periods generally coinciding with high and low precipitation periods. For example, the study period encompassed two of the most extreme dry periods on record at Coweeta (droughts of 1999-2002 and 2006-2008) as well as some of the wettest years [*Laseter et al.*, 2012].

Superimposed on the seasonality in precipitation and runoff were numerous, distinct, short-duration peaks generated by individual storm events. The deciduous (WS02 & WS18) and coniferous (WS01 & WS17) watersheds exhibited similar response to precipitation in terms of timing but with dampened peaks and lower runoff magnitudes evident in the coniferous watersheds. The high elevation watershed (WS36) showed

higher runoff throughout the year with a flashier response to precipitation, e.g. higher peak flows, than the low elevation watersheds.

Distinct differences in runoff behavior between the deciduous and coniferous watersheds were apparent in monthly Q totals (Figure 3.3). Monthly stream flows for coniferous and deciduous watersheds were most similar toward the end of the growing season (denoted with green in Figure 3.3), and most dissimilar near the end of the dormant season (denoted with blue in Figure 3.3), with differences often exceeding 70 mm mo^{-1} , especially in the south-facing watersheds.

Average monthly Q data (over the full record length) highlight the strong seasonality evidenced in the individual daily and monthly totals. Average monthly P was highly variable between months but showed no clear seasonality (Figure 3.4, top panel). However, P was on average greatest in January (175-212 mm) and lowest in July (113-144 mm). Monthly Q averages (Figure 3.4, middle panel) as well as the monthly averages of RR (Figure 3.4, bottom panel) showed a clear seasonal pattern. Q was highest in March, coinciding with high P values. Baseflow typically reached its minimum between August and October, highlighting the strong effect of both precipitation timing and ET on runoff.

The deciduous watersheds had greater runoff and runoff ratios than the coniferous watersheds across almost all months. Over the course of the growing season, Q and RR became more similar, and then diverged again in the fall. While coniferous, evergreen trees can transpire year round [Swank and Miner, 1968], transpiration from deciduous trees effectively shuts down with senescence in the fall until leaf out in the next spring. Because of this, the differences between the vegetation types were greatest during the

dormant season from November through May. Additionally, interception and subsequent evaporation from leaf surfaces is also higher for coniferous trees due to a much higher leaf area index in both dormant and growing season (e.g. *Helvey* [1967]; *Swank* [1968]; *Neary and Gizyn* [1994]; *Ford et al.* [2010]).

Surprisingly there was no strong aspect effect evident in runoff ratios. While WS18 exhibited a slightly higher runoff ratio than WS02 throughout the year, the runoff ratios for WS01 and WS17 were almost identical in the winter months, and in the summer WS01 even exceeded the runoff ratios of WS17.

Mean annual precipitation was 1719 mm for WS01, 1771 mm for WS02, 1968 mm for WS17, 1923mm for WS18, and 2116 mm for WS36. There was an annual difference of approximately 200 mm precipitation between the north- and south-facing watershed pairs despite having similar mean elevations.

Annual Analysis Annual runoff ratios ranged from a minimum of 0.15 in WS01 to a maximum of 0.91 in WS36 (Figure 3.5). Average annual RRs for the deciduous watersheds were 0.33 (WS01) and 0.31 (WS17), while the deciduous watersheds averaged 0.45 (WS02) and 0.50 (WS18) (Figure 3.5). The high elevation watershed (WS36) had consistently higher RRs with an averaged annual RR of 0.77. The deciduous and coniferous watersheds plot as closely matched pairs with no apparent aspect effect.

A lag-correlation analysis between annual P and annual RRs (Figure 3.6, top panel) showed significant correlations for all watersheds without lag (one year's RR with the same year's P , averaged r_p across watersheds = 0.62, $p_{val} < 0.01$) as well as a lag of one

year (one year's RR with the previous year's P , averaged r_p across watersheds = 0.61, $p_{val} < 0.01$), indicating a strong one year memory effect. The lag-correlations became insignificant after the first year. There were no strong differences between the low-elevation coniferous and deciduous watersheds. The fluctuating correlations at later lags are likely not caused by long-term memory effects but rather by periodicity in the P data (Figure 3.6). WS36 exhibited similarly strong correlations for both no lag and a one-year lag ($r_p = 0.68$).

Seasonal Analysis Growing season P was on average 14-20% less (depending on watershed) than dormant season P (Table 3.3); however, these differences were not statistically significant when testing the medians (Kruskal-Wallis test) of growing and dormant season P or differences (Wilcoxon rank sum test) between dormant and growing season pairs (i.e., one dormant season paired with the following growing season). Comparison of growing season and dormant season RR series for each watershed showed different behavior for the low-elevation watersheds and the high-elevation watershed (Figure 3.7). While the RR s of growing and dormant season within the low-elevation watersheds were not significantly different from each other (two-sided Wilcoxon rank sum test at the 5% significance level), the high-elevation watershed showed distinct and significant growing and dormant season RR s (two-sided Wilcoxon rank sum test at the 5% significance level, Table 3, Figure 3.7).

For the low-elevation watersheds, correlations between seasonal P and RR were strongest at a lag of one season (Figure 3.8 and Figure 3.6, middle panel), while WS36

exhibited stronger correlations ($r_p = 0.46$, $p_{val} < 0.001$, rightmost column Figure 3.8 and middle panel Figure 3.4) without a lag. In contrast to the lower elevation watersheds, the correlation at a lag of one season was low, but still significant ($r_p = 0.22$, $p_{val} < 0.001$). Correlations for all watersheds became non-significant for more than one season, e.g. one year's dormant season P did not affect next year's dormant season RR (Figure 3.6, middle panel). The fluctuations at later lags, specifically the higher correlations at a lag of four seasons, can be explained by periodicity of the seasonal data and do not denote correlations caused by memory effects.

Monthly Analysis Lag-correlation analysis between monthly P and RR s for all watersheds showed significant lags for up to 7 months, with the highest correlations at a lag of 1 month (averaged $r_p=0.45$, averaged $p_{val}<<0.01$, Figure 3.6, bottom panel). Due to the large sample size ($N = 252$ months over 21 years) correlations were significant even for small values of r_p . The highest correlation was at a lag of one month, meaning that the previous month's P had a stronger influence on the RR than the same month's P . The lag-correlations gradually leveled off after one month and reached zero after 12 months. The negative but significant correlation at 0-lag (averaged $r_p=-0.29$, averaged $p_{val}<0.01$) is likely caused by the general P regime with alternating low and high P months (see Figure 3.4, top panel). The lag correlations followed a similar pattern for all 5 watersheds irrespective of vegetation type or elevation (Figure 3.6, bottom panel).

Storage

Average annual ET calculated from the water balance over the 21-year study period was lowest for WS36 (462 mm); highest for coniferous WS01 and WS17 (1146 mm and 1351 mm, respectively); and intermediate for deciduous watersheds WS02 and WS18 (964 mm and 947 mm, respectively) (Table 3.4). ΔS and S followed similar patterns in the four low-elevation watersheds (Figure 3.9). The similarities were even more pronounced within the same vegetation type. Interestingly the two coniferous watersheds generally maintained higher S than the deciduous watersheds. It is important to note that the 0 mm storage lines in Figure 3.9 represent the 21-year mean storage for each watershed, whose actual values were undetermined. S fluctuated with P , but the absolute change in storage also depended on Q due to the nature of the water balance calculation. The highest S gain (greatest ΔS) was typically reached in high P years with below average annual Q , e.g. in 2009 where P was high but Q was still low because of the prior dry period. The majority of the annual P then replenished storage. On the other hand, the greatest S decreases occurred during years with normal to low P , but high Q , e.g. 1993.

In general, a net increase in S lead to higher Q in the following year. However, the current year's P also affected Q , so a particularly dry or wet year could reverse or dampen the effect of storage state on Q (Figure 3.9). Changes in S could be great even if P were almost identical in two adjacent years (e.g. 1992 & 1993 or 2006-2008). For example, S can explain why from 1992 to 1993 Q increased despite stationary P : the S built up in 1992 was carried over into 1993 and led to an increase in streamflow. Still low P in 1993 lead to a decrease in S and a subsequent drop in Q in 1994.

Linear regressions between annual P and Q indicated strong, significant correlations, especially for WS36 (Figure 3.10, top row). The stronger correlation for WS36 suggests less storage capacity for the steeper, high-elevation watershed. We evaluated the effect of storage on the linear regressions between annual P and Q by analyzing the relationship between the residuals of the linear regressions and S with zero-lag and one-year lag. In this case zero-lag refers to the storage at the end of the year, while a one-year lag denotes the storage at the end of the previous year and hence represents antecedent conditions for the year in question. At zero lag (one year's residual correlated with the same year's storage), the correlations were weak and insignificant for WSs 01, 02, and 17, but significant for WSs 18 and 36 (Figure 3.10, middle panel). However, at a lag of one year (a year's residual correlated with the previous year's storage state) significant positive correlations emerged for all low-elevation watersheds (Figure 3.10, bottom row). The correlation for WS36 became positive as well but was weaker than without lag, indicating no significant influence of storage on the residuals of the regression between annual P and Q . This corroborates the interpretation that storage has no strong influence on the regression between P and Q . In contrast, the strong positive correlations for the low elevation watersheds suggest a strong influence of antecedent storage state on the relationship between P and Q .

Discussion

The Effect of Watershed Memory on Hydrologic Response Across Different Time Scales

Lag-correlations between monthly, seasonal, and annual precipitation totals and runoff ratios indicate strong influence of past precipitation on present runoff (Figure 3.6). Our storage correlations corroborate the lag correlation results that system memory significantly influences runoff behavior across time scales. The shorter the observed time scale (e.g. monthly vs. annually), the more important was the previous time step's precipitation for the observed runoff ratio of any given time period, e.g. the difference in correlation between zero-lag and a lag of one time step. At the monthly time scale, correlations remained significant for lags of up to 7 months (Figure 3.6, bottom panel). The strong correlation at shorter monthly lags highlights the importance of system memory on shorter time scales, as has been documented by others (e.g. *Kim et al.* [2005]; *Tetzlaff et al.* [2014]). Significant but weaker correlations at longer monthly lags are in agreement with memory effects of more than 6 months found by *Rose* [1998] for the coastal plain of Georgia, USA. The dissipation of water could be expected to be much faster in the mountains relative to coastal areas because of higher gradients; however, deep soils and high soil water storage capacity may offset differences in gradients.

The influence of past precipitation was also evident at seasonal time scales. Similar to the monthly time scale, the lag-correlations were strongest at a lag of one season (Figure 3.6, middle panel). In contrast to the monthly lag-correlations, correlations at 0 lag were positive at the seasonal scale. This indicates that when integrated over a longer period of time, precipitation of one time step gains importance for the runoff ratio of that time step. However, WS36 was the only watershed where the correlation between seasonal precipitation and seasonal runoff was stronger without lag than at a lag of one season. This likely indicates the effects of less storage capacity due to shallower soils, enhanced

drainage because of steeper slopes, higher overall moisture states because of increased precipitation at higher altitudes, or a combination thereof. As a result, precipitation translates to runoff more rapidly, reducing the influence of a memory effect. This is corroborated by *Hewlett and Hibbert* [1967] who found higher quickflow ratios for WS36 than for the low elevation watersheds.

Increasing the time scale from seasonal to annual, both the same as well as the previous year's precipitation have a nearly equal effect on the runoff ratio of a year (Figure 3.6, top panel), highlighting the assumption that when integrated over longer time steps, the current time step gains more importance.

While the lag-correlations showed a significant memory effect across all watersheds and temporal scales, there were no detectable effects of vegetation type or aspect. This suggests that past precipitation influences runoff ratios similarly across watersheds, despite differences in the absolute *RR* values caused by different vegetation types. Furthermore, the lag-correlation calculations demonstrated that it is necessary to consider antecedent precipitation over different lengths of time. For example, to make predictions about a month's water balance it may be sufficient to know the precipitation of several months back, but to assess a full year's water balance it is imperative to know the precipitation history of at least the previous 12 months. *Merz and Blöschl* [2009] showed that for Austrian watersheds even mean annual precipitation can inform prediction of event runoff ratios, suggesting that besides exerting control on the annual water balance, annual precipitation totals also influence event runoff ratios by setting longer-term storage conditions.

The Effect of Storage on the Annual Water
Balance in Coniferous and Deciduous Watersheds

While the lag-correlations suggest that watershed memory is likely responsible for the observed patterns, no actual observations of storage were used in calculating the correlations. Our main assumption was that annual ET at Coweeta was relatively consistent across years (as described in the Methods section). The general climatic conditions at Coweeta are relatively consistent. The standard deviation of annual pan evaporation from the climate station located at the base of the Coweeta watershed over the period 1937 to 2011 was only 57 mm with a mean of 892 mm. Consistent with these long-term observations, *Ford et al.* [2007] estimated ET for Coweeta WS17 using sap flux methodology to be 1290 mm and 1292 mm in two consecutive years (2004-2005). This is within 4% of the 1351 mm averaged annual ET calculated for this study. For a three-year period (2004-2006) *Ford et al.* [2010] found, that growing season ET in the last year was ~29% lower than in the previous two years; however, not all components of ET (e.g., understory ET) were quantified and only a portion of the watershed was measured. It should be noted that *Oishi et al.* [2010] indicated that tree or plot level estimates of ET often suggest more variation in annual ET than is observed at larger spatial scales such as an eddy covariance foot print. In the North Carolina Piedmont *Stoy et al.* [2006] and *Oishi et al.* [2010] found very low inter-annual differences in ET for both a hardwood and a coniferous stand over a four year period, with ranges of 9% and 27% of the annual means as estimated from eddy covariance. At Coweeta the range of $P-Q$ over the 21-year study period was 80% of the annual mean for the deciduous low elevation watersheds, 52% for the coniferous watersheds and 90% for WS36 and thus

well above the values reported by *Stoy et al.* [2006] again indicating strong changes in annual storage. Relatively consistent annual evapotranspiration values are not limited to the southeast. Consistent with our analysis, *Kosugi and Katsuyama* [2007] showed that *ET* values calculated from eddy covariance for three years (mean 735 mm) in a watershed in central Japan were in good agreement with *ET* calculated from a 33-year water balance (mean 749 mm). These studies and data from inside and outside of the Coweeta watersheds suggest that changes in storage over time can be much greater than deviations in *ET* from one year to the next.

Here, the storage calculated from the annual water balance using the 21-year averages of *ET* ($\Delta S = P_i - Q_i - ET_{mean}$) revealed a similar temporal pattern of storage over the study period for all five watersheds (Figure 3.9). This is not surprising since the general rainfall and runoff patterns were similar among the five watersheds. It is important to note, however, that the cumulative storage (dashed black lines in Figure 3.9) should not be treated as an absolute value nor compared across watersheds, as the absolute storage states, i.e. the amount of water stored in the watershed at any given time, were undetermined. While the average change in storage among all watersheds from year to year is just 141 mm, the storage fluctuations within one year can be much greater as shown by *Sayama et al.* [2011] who calculated storage changes for watersheds in California within one wet season as up to 500 mm.

Watershed memory—or the history of storage states—greatly affected annual storage dynamics across both vegetation types. One year’s storage gain or loss was primarily determined by the precipitation of the prior year: in years following high *P* years all watersheds tended to “lose” or discharge extra water (relative to average annual *Q*), often

leading to a negative ΔS , while after low P years all watersheds tended to “gain” water, mostly resulting in a positive ΔS . However, the deciduous low-elevation watersheds underwent greater changes in storage than the coniferous watersheds, resulting in a greater (relative) storage range (Figure 3.9 and Table 3.5). While the smallest storage range for the steeper high-elevation WS36 is in agreement with results from *Troch et al.* [2003] who found faster response and less storage dynamics on the steeper of two idealized hillslopes using a hillslope-storage Boussinesq model, the storage dynamics in the low-elevation watersheds cannot be explained as easily. Furthermore, regardless of aspect, both low-elevation deciduous watersheds and both coniferous watersheds exhibited very similar storage changes (Figure 3.11). In most years even $\Delta S_{conif}/\Delta S_{decid}$ were similar for both aspects, suggesting that at Coweeta vegetation type has a stronger effect on the storage evolution of a watershed than aspect. However, the solar insolation values for each aspect (Table 3.3) indicate, that especially during the growing season the differences in energy input are likely negligible, resulting in nearly identical 21-year average annual ET values in the deciduous watersheds (18 mm difference, 2% of the mean annual ET rate for WS02), while the 21-year average ET difference for the coniferous watersheds was larger (205 mm difference, 15% of the mean annual ET rate of WS17). The difference in annual ET in the coniferous watersheds is likely the result of greater leaf area in WS17 as compared to WS01 (LAI of 7.2 [*Ford et al.*, 2010] and 5.3, respectively [*Vose and Swank*, 1990]). This suggests that at least in the deciduous watersheds the N-S aspect contrast may not have a strong influence on water balances at Coweeta. This is corroborated by comparable runoff ratios (Figures 4 and 5) and watershed memory (Figure 3.6). However, this is in contrast to *Hibbert* [1966], who

found that north and south-facing watersheds showed a very different streamflow recovery after logging, with south facing watersheds generating much less runoff than the north facing ones. However, this might simply imply different water use behavior between young and old stands [*Donovan and Ehleringer*, 1991], where the transpiration rates of young stands can vary much more with insolation but become more similar to each other as they mature. In addition, *Swank and Vose* [1988] reported increased leaf litter evaporation after clear cutting, which could be more susceptible to differences in insolation.

While absolute values of the cumulative storages of the watersheds (dashed black lines in Figure 3.9) cannot be compared, it is possible to make inferences about the general storage state in the coniferous and deciduous watersheds based on different runoff behavior. We hypothesize that because of greater rates of evapotranspiration the actual storage state in the coniferous watersheds is likely lower than the storage state in the deciduous watersheds. This would lead to different magnitudes of runoff between the two vegetation types as storage state increases or decreases over time. The foundation of this concept is a simple storage-release function, as presented for example in *Grayson et al.* [1997]. This assumes that the relationship between storage state and water flux (i.e. runoff) can be described by a power function (Figure 3.12). According to the nature of the relationship between storage and flux, a small change in storage at a high storage state would lead to a greater change in runoff than the same change in storage at a low storage stage (Figure 3.12). This is corroborated by *Ford et al.* [2011], who found increased differences in annual runoff with increasing annual precipitation for the two paired watersheds used in this study as well as other paired watersheds at Coweeta. If the actual

storage state in the deciduous watersheds is generally higher than the storage in the coniferous watersheds, mainly because of higher *ET* rates in the coniferous watersheds, this could have ramifications for runoff magnitudes. During the growing season, both vegetation types shift toward a lower storage state. Due to the low runoff at low storage states, slightly higher storage in the deciduous watersheds would not have a significant effect on runoff and hence both watershed vegetation types would exhibit roughly equal amounts of runoff. The data from Coweeta are consistent with this as demonstrated by small differences in monthly runoff totals during the growing season (Figure 3.3 and Figure 3.12). In contrast, during the dormant season as both watershed types shift toward higher storage states, the runoff from the deciduous watersheds would be greater than the runoff from the coniferous watersheds. This would be due to the position of the watersheds on the storage-flux curve and its high degree of non-linearity. Similar differences in storage between the watersheds would result in greater runoff differences at the higher, wetter end than at the lower, drier end of the storage-flux curve. Those differences could even increase as the coniferous watersheds continue to transpire during the dormant season. The data are consistent with this hypothesis as demonstrated by the monthly runoff differences (Figure 3.3 and Figure 3.12). Continued *ET* in the coniferous watersheds during the dormant season further decreases storage and amplifies this effect. Basically, the longer duration transpiration in the coniferous watersheds leads to a lower absolute storage than in the deciduous watersheds and therefore lower Q . In addition to the annual fluctuations in storage and resulting fluxes, this conceptual model is also transferable to dry and wet time periods. During the two droughts in the study period, overall storage in both vegetation types declined (i.e. the watersheds shifted toward a

drier state on the storage-flux curve in Figure 3.12), resulting not only in less runoff overall, but also smaller differences in runoff between the two vegetation types during dormant seasons (e.g. 1999-2001 and 2006-2008). In the wetter periods (e.g. 1994-1996) storage increased in all watersheds (i.e. the watersheds shifted toward the right on the storage-flux curve in Figure 3.12), thus leading to greater differences in monthly runoff.

In summary, the history of precipitation would determine a watershed's general location on a storage-flux curve, e.g. low or high storage (Figure 3.12). Vegetation could modify this location by water losses through transpiration. The degree of influence would then be a function of vegetation type, with deciduous watersheds being wetter than coniferous watersheds due to lesser *ET*.

Implications

The latest IPCC report suggests drought frequencies and intensities may increase in the later part of the 21st century. At the same time, we will likely see increases in the frequency and intensity of heavy precipitation events (*IPCC* [2013], *Hartmann et al.* [2013]). How watersheds respond to the anticipated changes in climate is largely a function of storage that acts as a buffer between water inputs and streamflow response. Variability in buffering capacity can even be observed over small spatial scales such as within the Coweeta Hydrologic Laboratory. Depending on the location of a watershed within the greater Coweeta basin, storage becomes a more important descriptor of hydrologic response, as demonstrated by the contrast between the low and high elevation watersheds (Figure 3.10). While watershed memory in a system like Coweeta was quantifiable for a period of up to one year (Figure 3.6, top panel), watershed recovery

from severe droughts may actually take much longer. In fact, we have shown that it can take several years after a drought period to refill storages (Figure 3.9, dashed lines, from relative minimum to relative maximum). While we did not observe vegetation influences on watershed memory, both precipitation history and vegetation type can play a role in determining a watershed's storage state. This storage state difference in turn can partially explain the effects of different vegetation (coniferous vs. deciduous) on hydrologic response. This can lead to greater differences not during dry periods, but during wet periods. The location of a watershed on the storage-flux curve (Figure 3.12) thus has implications for a watershed's response to individual precipitation events. As a consequence, drought periods could decrease the effect that different vegetation types have on the water balance. Decreasing precipitation could result in similar water balances for coniferous and deciduous watersheds, especially if the drought periods extended through the dormant seasons, when runoff differences are typically greatest due to differences in storage. Conversely, increases in annual precipitation could increase the differences in runoff between deciduous and coniferous watersheds.

Our observations can also inform interpretations of the effect of land cover change, e.g. clearcutting, on peak flows and storm runoff that is well documented at Coweeta and elsewhere (e.g. *Hewlett and Helvey* [1970]; *Pierce et al.* [1970]; *Harr et al.* [1975]; *Verry et al.* [1983]). Several studies highlight that the differences between cut and uncut watersheds were greatest in late summer and fall, when the moisture differences between the watersheds was greatest (e.g. *Pierce et al.* [1970]; *Harr et al.* [1975]). Here we propose that the differential position on the storage-flux curve of cut and uncut

watersheds or watersheds with different vegetation types serves as a mechanism to explain this observed phenomenon.

Conclusions

Utilizing 21 years of precipitation and runoff data from five watersheds at the Coweeta Hydrologic Laboratory we found strong memory effects (i.e. influence of past precipitation on present runoff) in all watersheds across a range of temporal resolutions. Differences in runoff between the coniferous and deciduous watersheds can be explained largely by different storage dynamics that developed under deciduous and coniferous vegetation. We summarize our findings as follows:

- 1) The Coweeta watersheds with their generally steep slopes and deep soils exhibited considerable memory effects. Past precipitation influenced the runoff behavior of watersheds on monthly, seasonal, and annual time scales. For monthly, seasonal, and annual resolutions, the precipitation of the previous time step had an equal or greater influence on the runoff ratio than the precipitation of the same time step. Neither aspect nor vegetation type appears to have had an influence on lag-correlations between precipitation and runoff ratios at any time scale.
- 2) All watersheds exhibited similar temporal patterns of storage, although the range of relative storage differed in the low-elevation coniferous and deciduous watersheds, and between the deciduous low-elevation and deciduous high-elevation watershed.
- 3) The previous year's storage state explained much of the variability in the relationship between annual P and Q , and explanatory power was greater for the low-

elevation watersheds with deep soils than for the steeper and wetter high-elevation watershed with shallower soils. We demonstrated that watershed storage state set by the previous years' precipitation can be an important component in determining annual runoff.

4) During growing seasons and during multi-year dry periods, runoff differences between deciduous and coniferous watersheds decreased. In dormant seasons and during wetter periods, runoff differences between the vegetation types increased (greater in deciduous watersheds). This suggests that the deciduous low-elevation watersheds at Coweeta had higher storage states on average than the coniferous watersheds. These differences, combined with nonlinearity of the storage–discharge relationship, can partially explain this behavior and further demonstrates the role of vegetation in the watershed water balance.

5) Depending on the geographic region, projected increases or decreases in precipitation as a result of climate change could alter the relative runoff behavior of systems with differing degrees of *ET*, e.g. coniferous and deciduous watersheds. Decreases in precipitation, e.g. increases in drought frequency and intensity, could lead to smaller differences in the observed runoff between different vegetation types; increases in precipitation could increase differences in runoff magnitudes between coniferous and deciduous watersheds.

This study provides insight into how watershed memory can affect runoff in steep headwater catchments. We showed how past precipitation affects storage dynamics under different vegetation and we postulated a conceptual model to explain differences in runoff between coniferous and deciduous watersheds. Given the results from this study,

we hypothesize that climate change may affect watershed memory and the runoff response from watersheds with different vegetation and/or *ET* magnitudes. Future research could include distributed modeling to disentangle the spatial patterns of storage and the degree of hydrologic memory across physiographic and climatic gradients.

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Tables

Table 3.1: Means and medians for landscape structure metrics for the 5 watersheds. Numbers in bold highlight differences between the low-elevation and the high-elevation watersheds

Metric		WS01	WS02	WS17	WS18	WS36
Vegetation		Conif.	Decid.	Conif.	Decid.	Decid.
Size (ha)		15.4	13.0	13.5	12.3	49.1
Elevation (m)	Mean	834	853	894	821	1288
	Median	819	851	894	816	1276
Slope (°)	Mean	27.1	26.9	28.5	27.8	30.6
	Median	27.9	27.5	29.2	29.0	31.7
Gradient to creek (-)	Mean	0.46	0.41	0.48	0.45	0.57
	Median	0.44	0.40	0.48	0.45	0.57
Dormant seas. pot. insolation (kWhrs/m ²)	Mean	927	982	500	568	839
	Median	975	1002	485	552	915
Growing seas. pot. insolation (kWhrs/m ²)	Mean	1384	1409	1187	1218	1342
	Median	1403	1419	1198	1229	1387

Table 3.2: Rain gauge watershed associations and their elevations.

Rain gauge	Watershed(s)	Elevation (m)
RG20	01 & 02	740
RG96	17 & 18	894
RG05	36	1144

Table 3.3: Climate statistics for precipitation (P), runoff (Q), and runoff ratios (RR) for growing (grow) and dormant (dorm) season for all five catchments.

	WS01	WS02	WS17	WS18	WS36
P_{grow}	924	957	1093	1064	1124
	795	814	875	859	972
Q_{dorm}	321	488	382	591	1021
Q_{grow}	255	322	238	387	619
RR_{dorm}	0.34	0.5	0.34	0.54	0.91
RR_{grow}	0.32	0.39	0.26	0.45	0.61

Table 3.4: Mean annual watershed evapotranspiration, calculated from the water balance

	Mean annual ET (mm)
WS01	1146
WS02	964
WS17	1351
WS18	946
WS36	462

Table 3.5: Relative storage ranges over 21-year study period.

	Storage range (mm)
WS01	677
WS02	811
WS17	610
WS18	717
WS36	417

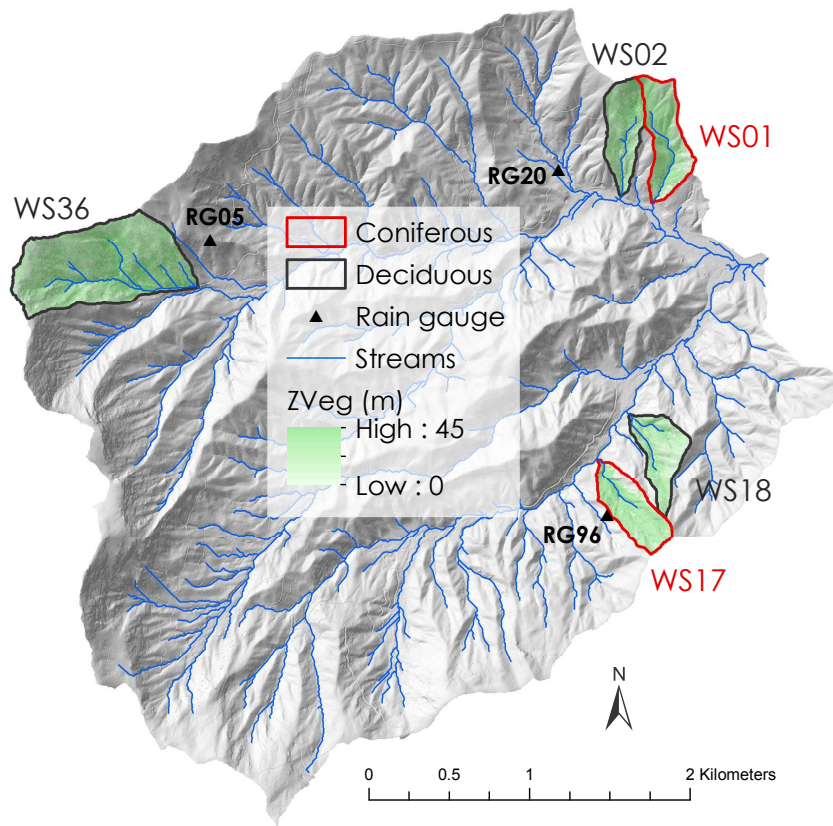


Figure 3.1: Map of the Coweeta Hydrologic Laboratory with the outlines of the five watersheds used in this study. The red color-coding denotes coniferous watersheds and black denotes deciduous watersheds; color scheme remains consistent for all subsequent figures.

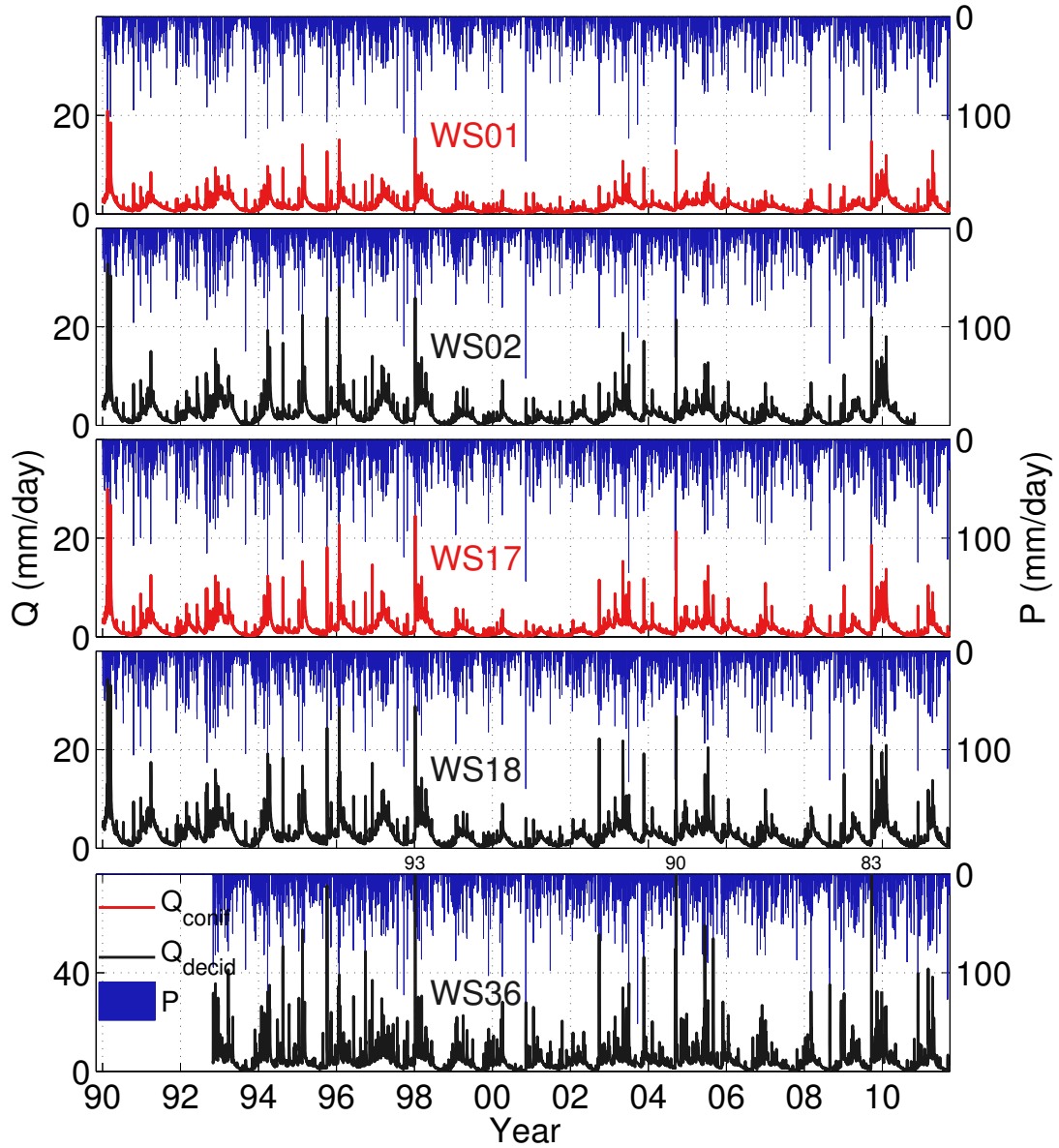


Figure 3.2: Precipitation and Runoff data for the 21-year study period for all five watersheds. Note the greater secondary y-axis range for WS36. Small numbers above WS36 show runoff (mm/day) for days where runoff exceeded y-axis maximum (80 mm/day).

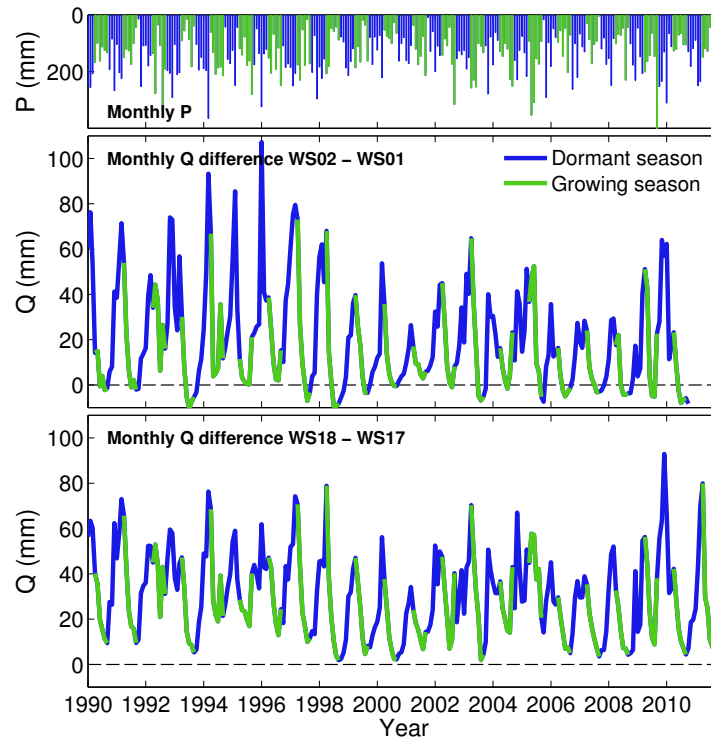


Figure 3.3: Monthly Q differences between the deciduous and coniferous watershed for the south-facing watersheds (middle panel) and the north-facing watersheds (bottom panel). Monthly precipitation for WS01 is shown in the top panel for context. Blue color denotes the dormant season, green color denotes the growing season.

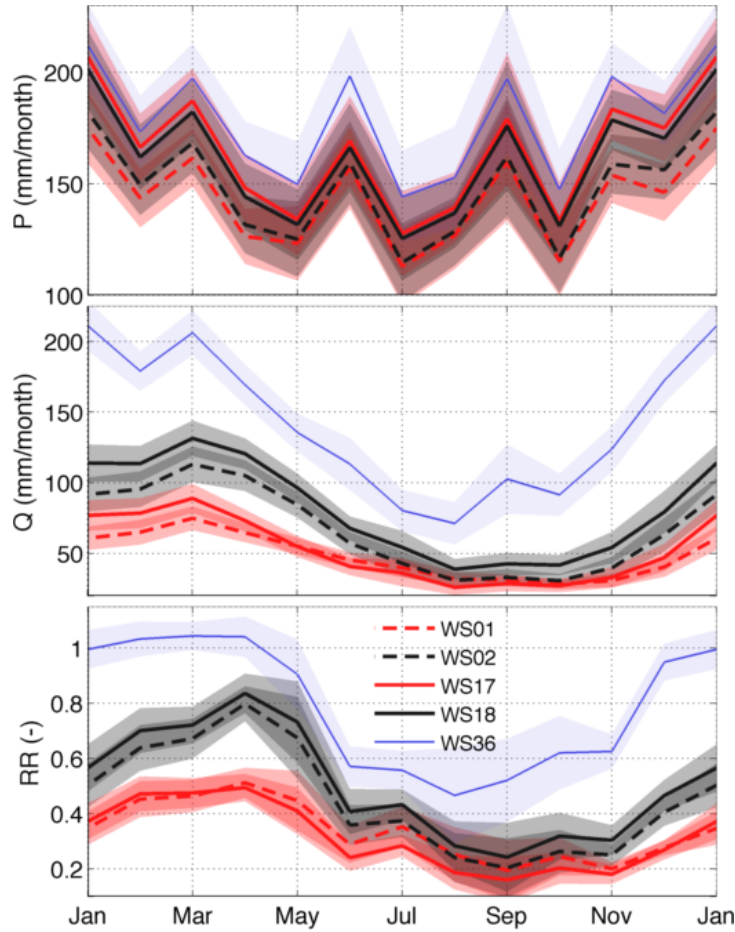


Figure 3.4: Monthly averages of Precipitation (P), Runoff (Q), and Runoff Ratios (RR) over the simulation period. Shaded areas are the standard errors of the respective time series (darker shades indicate overlap). Red denotes coniferous watersheds and black denotes deciduous watersheds. Blue denotes the high elevation watershed.

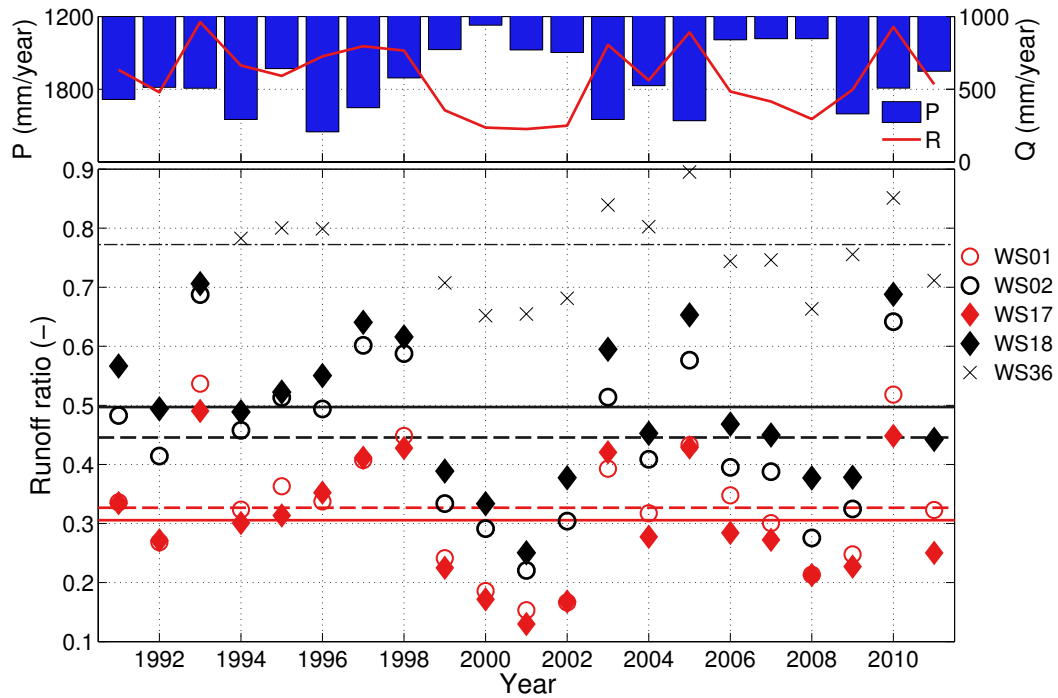


Figure 3.5: Variability in annual runoff ratios for the five watersheds. Diamonds denote the south-facing watersheds (WS01 & WS02) and open circles denote the north-facing watersheds (WS17 & WS18). The gray crosses denote the high elevation watershed (WS36). The red color-coding denotes coniferous watersheds (WS01 & WS17) and black denotes deciduous watersheds (WS02 & WS18). The dotted lines represent the mean runoff ratios over the 21-year study period. For context, the top panel shows P and Q for WS01 representative for all watersheds.

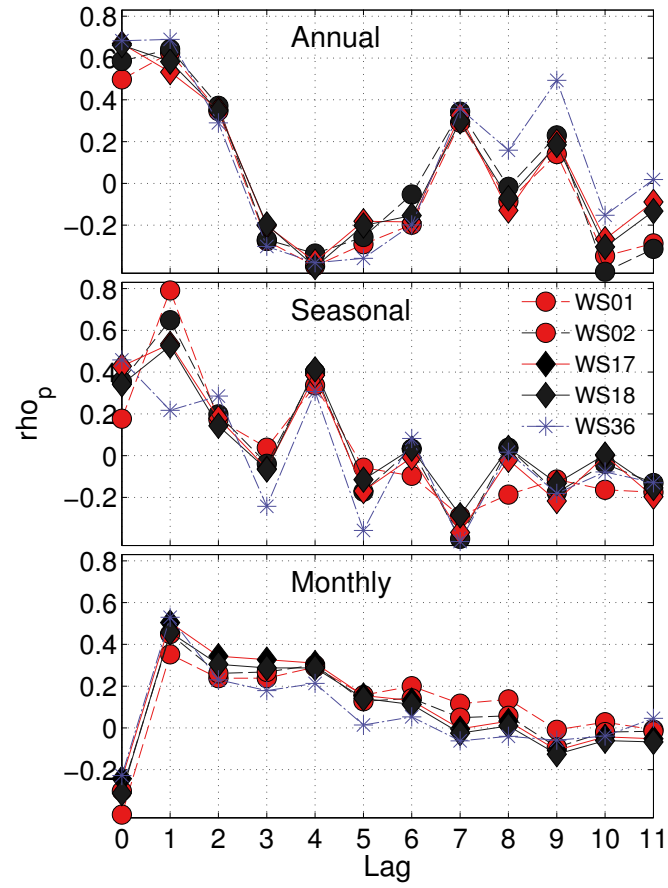


Figure 3.6: Lag-correlations between precipitation and runoff ratios for annual (top panel), seasonal (middle panel), and monthly (bottom panel) time scales. The lag on the x-axis is months (top panel), seasons (middle panel), or years (bottom panel).

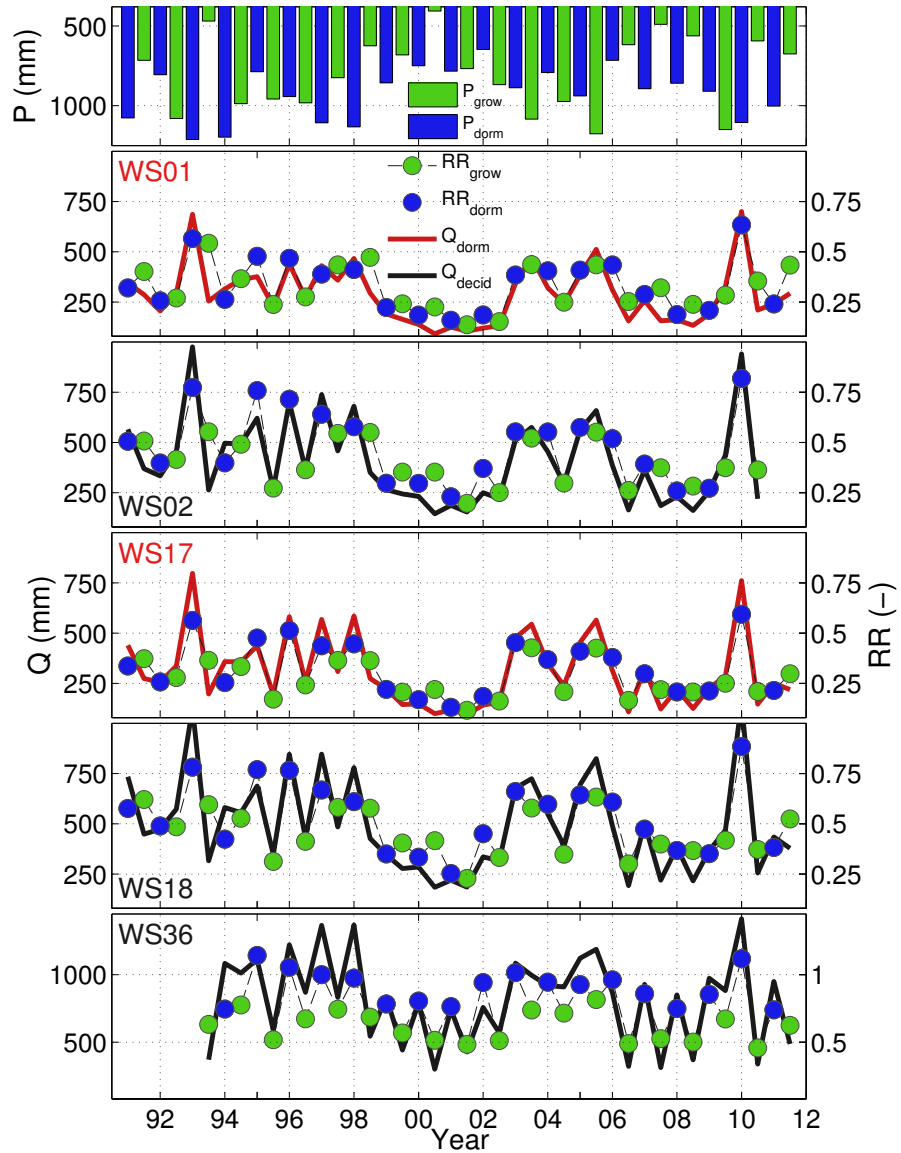


Figure 3.7: Seasonal precipitation (P), runoff (Q), and runoff ratios (RR) for all five watersheds

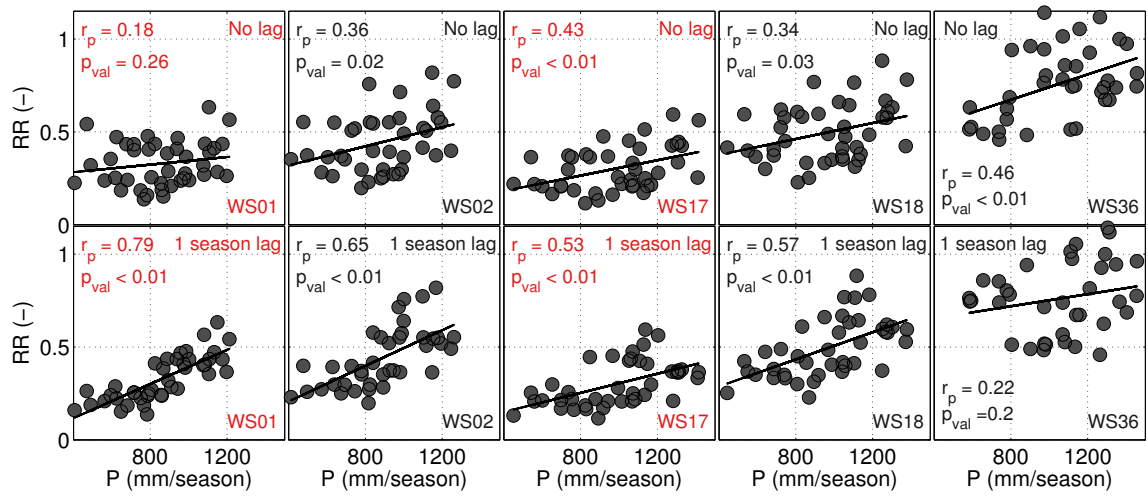


Figure 3.8: Scatterplot of seasonal P versus seasonal RR without lag (top panels) and lagged by one season (bottom panels) for all watersheds.

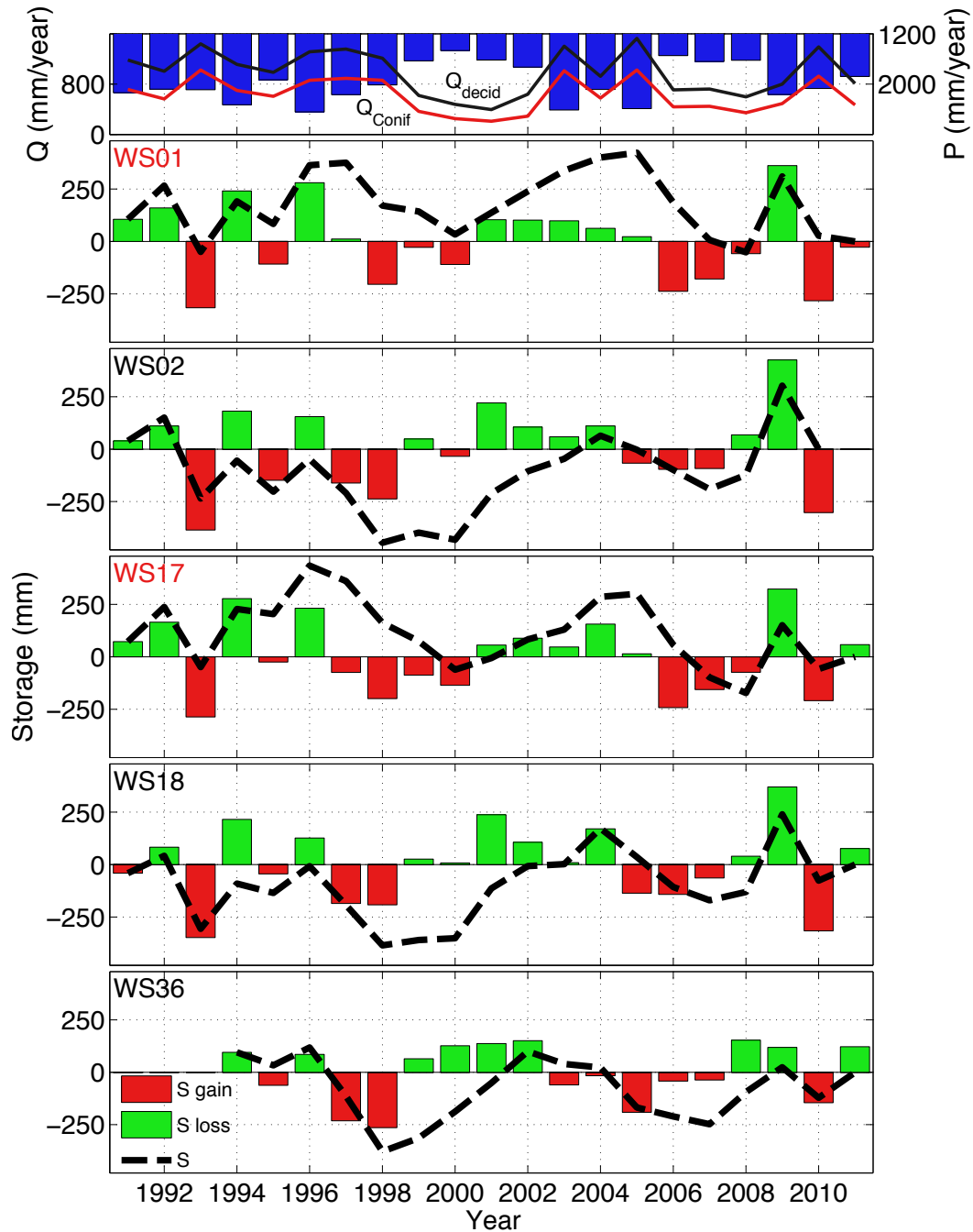


Figure 3.9: Time series of storage and deviation from mean runoff ratio for all watersheds. Storage was calculated as $P - Q - ET_{\text{mean}}$. Bars are changes in S from year to year, black line is cumulative S . Top panel shows P and Q from one deciduous (WS18) and one coniferous (WS17) watershed, representative of the general trend in all watersheds and vegetation types. Please note that the cumulative storage values (dashed black line) are not indicative of absolute storage values because of the arbitrary start value

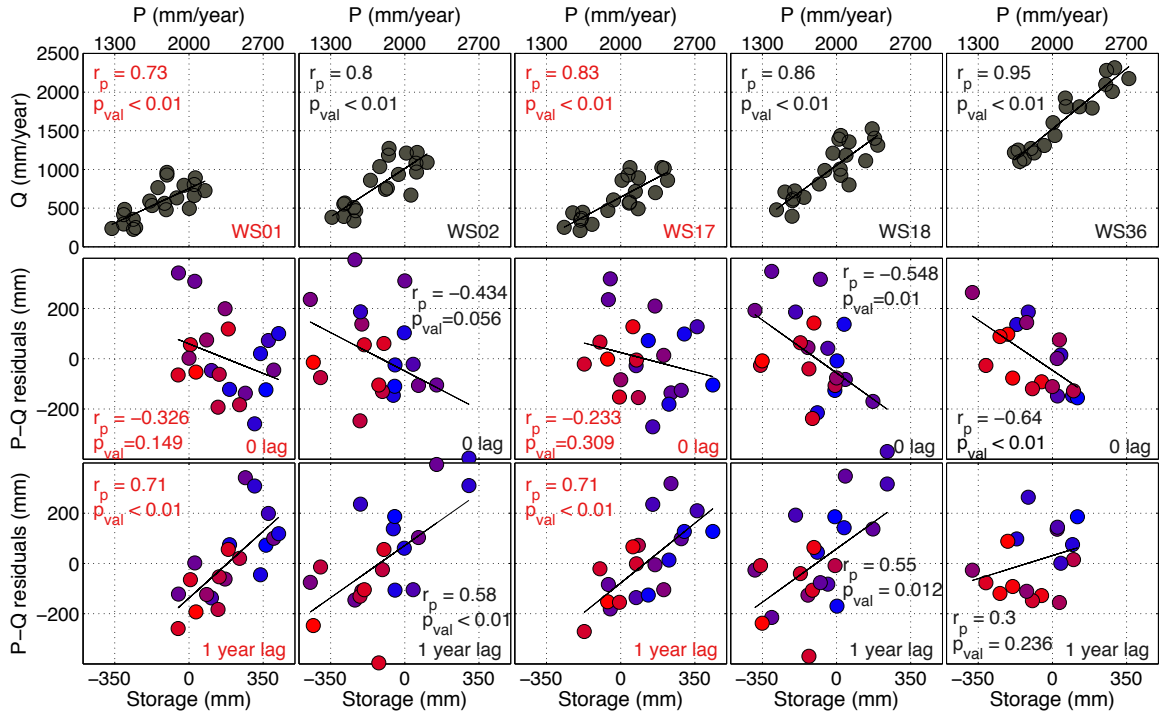


Figure 3.10: Linear regression between annual P and Q (top panel) for each of the five watersheds (each column represents one watershed). Linear regression for the residuals of the regressions between annual P and Q versus cumulative watershed storage without lag (middle panel) and a lag of 1 year (previous year's storage vs current year's residuals, bottom panel). The two bottom rows share the same x-axis. Color-coding blue to red denotes annual P (red = more dry, blue = more wet).

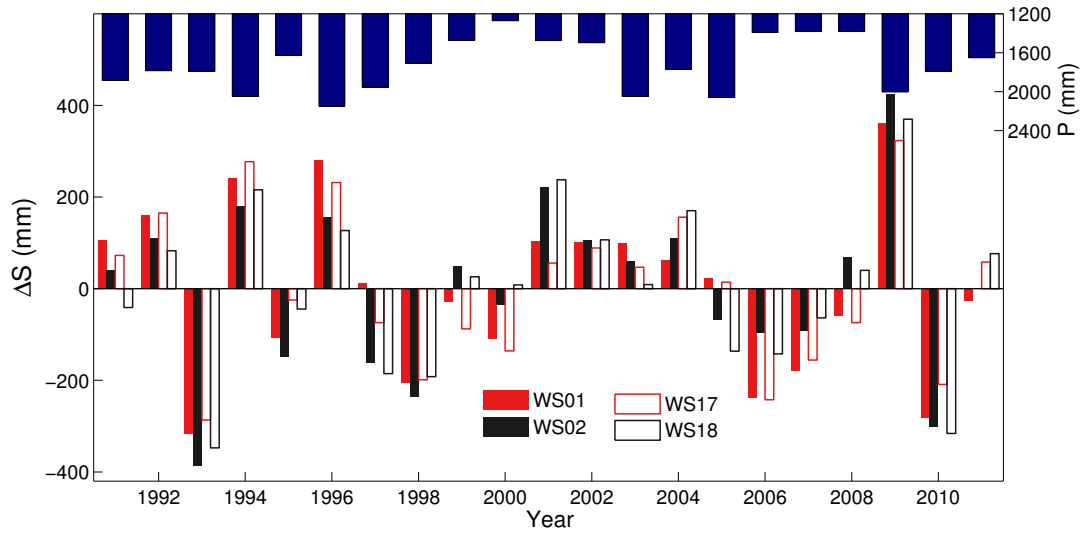


Figure 3.11: Comparison of annual storage changes in the low-elevation watersheds. Red denotes coniferous watersheds, black denotes deciduous watersheds. The filled bars denote the south facing watersheds. The annual precipitation of WS01 is shown and is representative of precipitation dynamics. The patterns of storage changes between the north- and south-facing watersheds are almost identical.

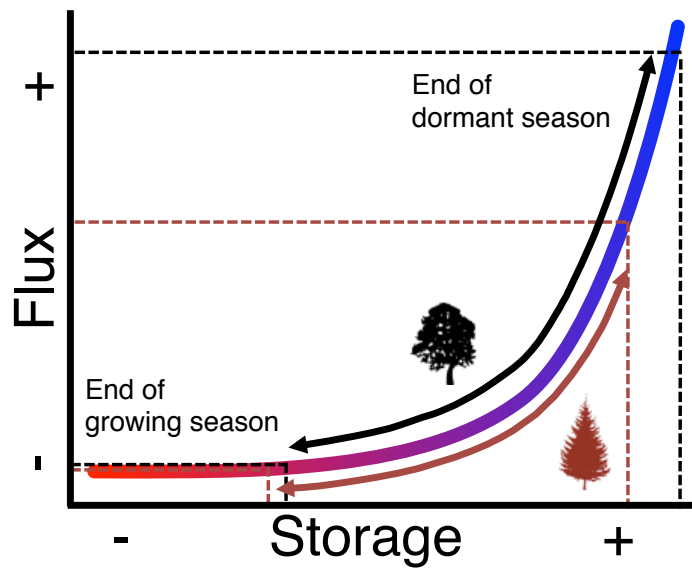


Figure 3.12: Conceptual model of storage states and associated water fluxes in the deciduous and coniferous watersheds. The thick line denotes the general storage-flux relationship, indicating low (red) and high (blue) storage and flux states. The arrows denote the range of storage for the deciduous watersheds (black arrow) and the coniferous watersheds (red arrow). Fluxes are similar at the end of the growing season, due to more similar storage states and minimal differences in resulting fluxes. Storage and flux differences are greatest at the end of the dormant season, when evapotranspiration continues to decrease storage in the coniferous watersheds.

4. THE TEMPORAL EVOLUTION OF VARIABLE CONTRIBUTING AREAS

Contribution of Authors

First Author: Fabian Nippgen

Contributions: Co-developed the project, wrote the model code, analyzed data, created figures, and wrote the manuscript.

Co-Author: Dr. Brian McGlynn

Contributions: Co-developed the project, contributed significant critique and ideas for development of intellectual content within the paper, edited successive versions of the manuscript.

Co-Author: Dr. Ryan Emanuel

Contributions: Co-developed the project, co-secured initial project funding, contributed to development of intellectual content within the paper, commented on final version of the manuscript.

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Abstract

Predicting runoff source areas and how they change through time is one of the grand challenges in hydrology. Topographically induced lateral water redistribution and water removal through evapotranspiration lead to spatially and temporally variable patterns of watershed water storage. These dynamic storage patterns combined with threshold mediation of saturated subsurface throughflow lead to runoff sources areas that are dynamic through time. To investigate these processes and their manifestation in observed watershed runoff, we developed and applied a parsimonious but spatially distributed model (WECOH - Watershed ECOHydrology). Evapotranspiration was measured via an eddy-covariance tower located within the catchment and disaggregated as a function of vegetation structure. This modeling approach reproduced the stream hydrograph well and was internally consistent with observed catchment runoff patterns and behavior. We further examined the spatial patterns of water storage and their evolution through time by building on past research focused on landscape hydrologic connectivity. The percentage of landscape area connected to the stream network ranged from less than 1% during the fall and winter baseflow period to 71% during snowmelt. While most watershed areas were connected at least one day during the two-year simulation period, approximately 10% of watershed area never established a hydrologic connection to the stream network. Runoff source areas during the event shifted from riparian dominated runoff to areas at greater distances from the stream network when hillslopes became connected. Our modeling approach elucidates and enables quantification and prediction of watershed active areas and those active areas connected to the stream network through time.

Introduction

Efforts to trace watershed runoff to its source areas and to determine how much of a watershed contributes to (storm-)runoff have been a central focus of watershed hydrology for many decades. While early studies still emphasized Hortonian, i.e. infiltration excess, surface runoff (e.g. *Betson* [1964]; *Ragan* [1968]; *Dickinson and Whiteley* [1970]), a paradigm shift occurred with recognition that runoff—both during and between storms—can be largely generated in the subsurface (e.g. *Hursh and Brater* [1941]; *Hewlett and Hibbert* [1963], *Tsukamoto*, [1963]). This led to the formulation of the variable source area (VSA) concept, which states that storm runoff is mainly generated by subsurface flow in expanding and contracting near-stream areas [*Hewlett and Hibbert*, 1967]. Variations of the variable source area concept that included saturation excess overland flow from near-stream areas were formulated by *Dunne and Black* [1970]. The variable source area concept also formed the core of one of the more popular research-oriented hydrologic models to date, TOPMODEL [*Beven and Kirkby*, 1979]. A commonality across these studies and concepts was, that even during intense storms only a fraction of the watershed delivered water to the stream network.

Knowledge of runoff generation mechanisms advanced further with studies recognizing that source areas were not limited to lower hillslopes and near-stream areas, but that flow in the subsurface largely via preferential flow pathways extending up the hillslope contributed to watershed runoff (e.g. *Mosley* [1979]; *McDonnell* [1990]; *Woods and Rowe* [1996]). Subsequent studies (e.g. *McGlynn and McDonnell* [2003a]; [2003b]) further documented that the threshold-mediated establishment of shallow water tables on

hillslopes essentially created a hydrologic connection between the watershed uplands and the stream network. *McGlynn and McDonnell* [2003b] for example explained changes in dissolved organic carbon during runoff events with a shift in runoff source areas from riparian (the classic variable source areas) to hillslopes, after the hillslopes exceeded a moisture-based threshold and became hydrologically connected to the stream network. As such, this watershed connectivity can be viewed as an extension of the classic variable source area concept, while including individual landscape elements (e.g. hillslopes) to the source areas without the restriction to specific runoff generation processes. 40 years ago John Hewlett published a comment in *Water Resources Research* defending and further defining the concept of variable source areas [Hewlett, 1974]. This was before research had fully emphasized macropore flow, perched shallow groundwater flow and other preferential runoff generation mechanisms. However, in Hewlett's [1974] definition "subsurface storm flow is any water passing the gaging station that has, however briefly, entered the mineral soil surface and has traveled for some distance, however short, within the soil". This definition allows incorporation of many—if not all—runoff generation mechanisms under the umbrella of the variable (or contributing) source area and helps to define hydrologic connectivity.

Factors leading to the formation of a hydrologic connection and the duration thereof have previously been attributed to (variations in) topography (e.g. *Sidle et al.* [1995]; *McGlynn and McDonnell* [2003b]; *Jencso et al.* [2009]), soil properties (e.g. *Buttle and McDonald* [2002]), and soil type (e.g. *Tetzlaff et al.* [2007]), as well as variability in vegetation cover (e.g. *Hwang et al.* [2012]; *Emanuel et al.* [2014]). Furthermore, the establishment of lateral (subsurface) flow exhibits threshold behavior (e.g. *Sidle et al.*

[2000]; McGlynn and McDonnell [2003a]; Tromp-van Meerveld and McDonnell [2006b]; Lehmann *et al.* [2007]; Penna *et al.* [2011]), where either the exceedance of rainfall depths/intensities or soil moisture states ultimately lead to a response (e.g. activation of a flow pathway) or to a response disproportionately greater than below those thresholds. However, determining thresholds *a priori* can be a challenge because there is no singular precipitation threshold for hillslopes (cf Weiler *et al.* [2005] or Ali *et al.* [2013]).

Despite marked progress in identifying the mechanisms leading to the activation of flowpaths, the detection and prediction of areas contributing water to the stream network remains a challenge. The difference between “active” and “contributing” areas further complicates this assessment. The terms “active” and “contributing” are often used interchangeably especially in reference to the VSA concept. However, as Ambroise [2004] points out, “active” areas are not necessarily “contributing” areas. Ambroise [2004] suggests that “contributing” be used for areas that have an established hydraulic connection to the stream network, while fluxes in “active” areas may be isolated from the stream network and not contribute to watershed runoff when the downhill flowpaths are discontinuous.

Fully distributed watershed models capable of tracking runoff source areas—or contributing areas—are often data intensive and include numerous parameters (e.g. DHSVM [Wigmosta *et al.*, 1994], RHESSys [Band, 1993], PIHM [Qu and Duffy, 2007]), which may increase the degree of equifinality [Beven, 2006] that may lead to various representations of internal system behavior [Kelleher *et al.*, in review]. Here we present a parsimonious but fully distributed modeling framework that incorporates topographically driven lateral water redistribution and eddy covariance derived spatially disaggregated

evapotranspiration measurements to simulate streamflow and the spatial distribution of water stored in the watershed. Utilizing empirical measurements of hillslope connectivity [Jencso *et al.*, 2009; 2010; Jencso and McGlynn, 2011] we approximated watershed connectivity to answer the question: How do runoff source areas change in space and time over the course of two water years in a snowmelt dominated system?

Methods

Site Description

The Tenderfoot Creek Experimental Forest (TCEF) is located in the Little Belt Mountains in the Rocky Mountains of central Montana (lat. 56.55°N, long. 110.52°W). The total area of TCEF encompasses 2300 ha. Tenderfoot Creek is a tributary of the Smith River, which drains into the Missouri. For this study we focused on the 555 ha Lower Stringer Creek (LSC) subwatershed (Figure 4.1). The elevation in LSC ranges from 1992 m to 2426 m, which is also the highest elevation in TCEF. Slope in LSC increases in downstream direction, while the riparian areas (~2% of total area) get narrower and more constrained toward the outlet. Extensive park areas occur mainly in the upper parts of the watershed.

Lodgepole Pine (*Pinus contorta*) is the dominant tree species; Subalpine Fir (*Abies lasiocarpa*), Engelmann Spruce (*Picea engelmannii*) and Whitebark Pine (*Pinus albicaulis*) are also common. Forest composition at TCEF is a mosaic of different aged patches while most substantially large stands are single aged. The majority of LSC areas around the stream and north of the stream regenerated after two fires in the middle of the

18th century. The lower elevations experienced a stand-replacing fire in 1902. Especially stands in the southwestern portion of LSC were more affected by smaller, non-stand replacing fires [Barrett, 1993]. The maximum tree height is just over 30 m. Shrubs and grasses are predominant in the riparian areas.

Soils at TCEF are loamy Typic Cryochrepts along hillslope positions and clayey Aquic Cryoboralfs in riparian zones and parks [Holdorf, 1981]. Based on the installation of >180 shallow groundwater wells, Jencso *et al.* [2009] found an average soil depth of about 1 m, ranging from 0.5 to 2 m.

The geology at TCEF consists of four different strata, all of which are present in LSC. Flathead Sandstone is the most dominant geologic layer in the mid altitude parts while the higher elevations are underlain by shale (Wolsey Formation) and biotite hornblende quartz monzonite. Granite gneiss occurs in the lower part of LSC [Reynolds and Brandt, 2006].

Climate at TCEF can be classified as continental. The average precipitation over the 1961–1990 base period was 880 mm [Farnes *et al.*, 1995]. Approximately 70% of the precipitation falls as snow. The main snowmelt period usually occurs between mid-May and the end of June, runoff normally peaks with snowmelt.

Hydrologic Measurements

Discharge for the study period (WY 2007-2008) was measured every 30 mins with a 3.5-foot H-Flume at the LSC outlet using capacitance rods with ± 1 mm resolution (TruTrack Inc., Christchurch, New Zealand) and later aggregated to an hourly time series.

Precipitation and snowmelt were measured at two Natural Resources Conservation Service (NRCS) SNOTEL sites located within TCEF, one near the headwaters at 2259 m, and the other in proximity to the LSC outlet near the bottom of the Experimental Forest (1996 m)(Figure 4.1a). SNOTEL sites comprise a network of stations across the mountain west and Alaska recording precipitation, snow height, and snow water equivalent (the amount of water in a snowpack at a given time). Actual evapotranspiration (ET) was measured at an eddy covariance tower located in the watershed (see Figure 4.1) and later aggregated to 60 min totals [*Mitchell et al.*, in review].

We used a separate snowmelt module that was not connected to the WECOH rainfall-runoff routine to generate spatially distributed precipitation and snowmelt from observations from two NRCS SNOTEL sites located at TCEF. We simulated distributed snow accumulation and melt, and liquid precipitation for each grid cell across the Stringer Creek watershed using an extended temperature index based model that allowed for cold content approximation following a method outlined by *Schaefli and Huss* [2011]. Details on the method and the calibration of the snowmelt model parameters can be found in Appendix A.

Disaggregation of Eddy Flux Measurements

The evapotranspiration point measurement from the eddy flux tower was disaggregated across the watershed based on a linear relationship between height and leaf area index as suggested for these coniferous forests [*Keane et al.*, 2005], analogous to the

derived distribution approach of *Emanuel et al.* [2011]. Treeless parks and meadows were assigned a 10% baseline of the tower measured ET for each time step. The vegetation height was calculated by subtracting the last returns (bare earth) from the first returns (elevation of canopy) of the 1 m LIDAR coverage. The resulting vegetation heights were subsequently resampled from a 1 m grid resolution to 10 m by averaging the 1 m vegetation heights.

Watershed ECOHydrology Model (WECOH)

WECOH is a parsimonious but fully spatially distributed rainfall-runoff model that explicitly incorporates topographically driven lateral water redistribution and water uptake by vegetation. The only inputs to the model are a digital elevation model (DEM), gridded vegetation height, a precipitation/snowmelt time series, and evapotranspiration from an eddy flux tower (or independently modeled or estimated ET). For computational reasons we ran the model on a 60 min time step, but shorter time steps are possible. The model requires only four parameters, two of which were fixed (Table 4.1). WECOH calculates a water balance for each 10 m grid cell:

$$\Delta S = P + I + O - ET \quad (1)$$

where ΔS is the change in cell storage (mm), P is incoming precipitation or melt (mm) derived from the snowmelt model, I is inflows from other cells (mm), O is outflows out of the cell (mm), and ET is evapotranspiration (mm). Water is routed downhill on a grid cell to grid cell basis using a flow algorithm similar to *Quinn et al.* [1991] that

proportionally routes water down the two steepest gradients. Conductivity is calculated with a simple storage-release function:

$$Cond = Cond_{\max} \times \left(\frac{S}{S_{\max}} \right)^b \quad (2)$$

where $Cond$ is conductivity (mm/min), $Cond_{\max}$ is the maximum conductivity (mm/min), S is actual cell storage (mm), $S_{\max}$ is the maximum possible storage (or effective storage, mm), and b is the shape parameter of the storage-release power function that determines the degree of concavity (Figure 4.2) of the storage-conductivity relationship. Cell outflow is then calculated using the cell cross-sectional area and the gradient between two cells (approximated by local topography). Cell storage was reevaluated every 10 min during the 60 min time step to adjust actual storage and cell outflows. Cells less than 3 m in elevation above the stream cells were designated riparian cells, comparable to *Jencso et al.*, [2009 & 2010], and their gradient was fixed at 0.01 to account for only partial stream penetration of riparian soil profiles and low gradient valley bottoms relative to adjacent uplands. Similar to DHSVM [*Wigmosta et al.*, 1994] the model does not have an infiltration routine and all precipitation immediately becomes storage if the cell is not saturated and can accommodate the incoming precipitation amount. Excess water, the ponded water of oversaturated cells, is routed downhill along a flowpath to the next unsaturated cell, until all overland flow has either become cell storage or reached the stream. Once the water from the hillslopes reaches the stream, it is routed downstream with a constant velocity, Vel_{Stream} .

Model Calibration And Parameter Values

Runoff and precipitation data were available for three consecutive years, 2006, 2007 and 2008. ET data were available for 2007 and 2008. Since no ET data were available for 2006, we duplicated the 2007 WY and used it as warm-up for the two-year model simulation period 2007-2008. Calibrations were performed on averaged daily runoff values to integrate pronounced daily melt and re-freeze dynamics.

$S_{\max}$ was fixed at 500 mm to approximate actual soil properties at TCEF, S_{vel} was set to 10 m/min but was found to be insensitive to model performance in early model simulations, likely because of the overall short stream length in LSC (2.9 km for main channel). The two remaining parameters, $Cond_{\max}$ and b , were calibrated with 5000 Monte Carlo simulations using the observed hydrographs for the 2007 and 2008 WYs. We calculated Nash-Sutcliffe efficiencies (NSE) for both years separately as well as over the full 2-year calibration period.

Active and Contributing Watershed Area

We determined a storage threshold on the storage-release curve below which a cell was considered inactive for our analysis of active and contributing watershed areas. An active cell transmits water to the cells immediately downslope. However, even inactive cells can transmit water downslope, but because of low transmissivities below a certain storage threshold, those amounts are insignificant when compared to the amounts transmitted by the wetter, active cells. A “contributing cell” is an “active cell” whose downslope flowpaths are entirely active and end in the stream [Ambroise, 2004]. Active

cells may become disconnected from the stream network, forming insular patches of active, but non-contributing cells when sections of flowpaths become inactive. The contributing watershed area at any point in time is therefore always equal to or smaller than the active watershed area.

The storage threshold for active and inactive cells was determined based on an empirical connectivity duration curve established for TCEF by *Jencso et al.* [2009]. Based on water table observations from 84 recording shallow groundwater wells, *Jencso et al.* [2009] quantified a linear relationship between the size of the upslope accumulated area (UAA) of a hillslope and the duration of an above bedrock water table across the hillslope-riparian area-stream continuum for the 2007 water year. They extrapolated the relationship to all hillslope cells bordering the riparian area. This resulted in a saturated throughflow connectivity duration curve that informs how much of the hillslope network was hydrologically connected to the stream network. This concept has since been successfully applied at TCEF with the semi-distributed Catchment Connectivity Model (CCM) framework [*Smith et al.*, 2013]. Here, we used the empirical network connectivity curve to calibrate a watershed scale threshold for active/inactive cells. We calculated network connectivity, i.e. the amount of hillslope cells that were connected to the stream network, for 1-500 mm thresholds in one mm increments on a daily basis for the 2007 water year, resulting in 500 modeled network connectivity curves (see results section for details). The modeled network connectivity curves were compared to the empirical network connectivity curve from *Jencso et al.* [2009] using Nash-Sutcliffe efficiencies. The storage threshold that resulted in the best fit to the empirical CDC was used as the threshold to determine active from inactive cells in WECOH.

We calculated active and contributing areas for every hour but here present 12 characteristic days (six in each year) that cover the main melt periods. For each of the 12 days we calculated the hillslope width functions, $W(x)$, of the contributing area. A hillslope width function is essentially a frequency distribution of flowpath lengths for hillslope cells to a stream cell along a flowpath (e.g. *D'odorico and Rigon* [2003]; *Bogaart and Troch* [2006]). The cell count is then rescaled by the total amount of hillslope cells (55240). In the case of the entire DEM, the area under the width function is equal to 1; for the contributing portions of the watershed, the area under the curve equals the fraction that is contributing area.

Results

Snowmelt and ET

Annual P in both 2007 (916 mm) and 2008 (874 mm) was close to the average of the 1961-1990 base period of 880 mm. The snowmelt model calibration resulted in good fits for the 2007 and 2008 water years, with NSEs between observed and simulated SWE time series ranging from 0.85 to 0.99 (see Table A.1 in the Appendix for the calibrated parameters and efficiencies). In 2007, 668 mm of P fell in the form of snow (73% of total annual P) and in 2008 snowfall was 735 mm and comprised 84% of annual P. At the time of greatest SWE accumulation, minimum simulated SWE (lowest elevation) in 2007 and 2008 were 260 mm and 313 mm, respectively, while maximum simulated SWE (highest elevation) in 2007 and 2008 were 533 mm and 592 mm, respectively.

The hourly evapotranspiration measured at the eddy covariance tower was highly variable (Figure 4.3 middle panel). Annual ET was 483 mm for 2007 and 449 mm for 2008. Adjusting the ET time series for sublimation resulted in annual sublimation losses from the snowpack of 105 mm in 2007 (11% of P) and 126 mm in 2008 (14% of P). This reduced the water input through melt to 556 mm in 2007 and 596 mm in 2008. The adjusted annual ET totals (water that will be removed from the soil column during the model simulation) were 378 mm in 2007 and 323 mm in 2008.

Runoff Calibration

The highest Nash-Sutcliffe efficiencies for calibration on the individual water years were 0.8 for 2007 and 0.94 for 2008 (Figure 4.4 left and middle panel). The highest NSE over the entire two-year simulation period was intermediate with 0.86 (Figure 4.4, right panel). However, high Nash-Sutcliffe efficiencies were achieved over a range of parameter combinations, especially in 2008, where efficiencies greater than 0.9 were reached over much of the parameter space (see inner contour line in Figure 4.4, middle panel). Multiple parameter combinations resulted in NSEs that were similar to the third decimal place. We averaged the parameters of the twenty best runs for the combined calibration (all identical efficiencies) to obtain an optimal parameter set. The averaging resulted in a parameter combination that matched the highest efficiency on the combined parameter-efficiency surface grid. However, similar efficiencies were derived for $Cond_{max}$ and b parameters within of 15% and 20%, respectively, of the optimum value. The optimal parameter set for the two-year period calibration resulted in $Cond_{max}$ and b

parameter values of 167 mm/min and 4.82 respectively. While the simulated runoff hydrograph matched the observed runoff well, especially in timing, it generally overestimated peak runoff in 2007 and underestimated peakflows in 2008 (Figure 4.5). The final recession periods were slightly overestimated in 2007 but matched the observed falling limb of the hydrograph well in 2008.

Calibration of Storage Threshold

The empirical connectivity duration curve developed by *Jencso et al.* [2009] (Figure 4.6, black lines in main figure and inset) indicates a maximum connectivity of the stream network to the hillslopes of 72%. Connectivity quickly decreased to approximately 10% at 20% exceedance (i.e. 20% of the time are 10% or more of the network connected). 50% of the year the connectivity is below 5%. Calculating the simulated connectivity duration curves for all possible thresholds between 1 mm (i.e. always connected) and 500 mm (only connected at full saturation) resulted in an ensemble of CDCs (Figure 4.6 inset). The calibration of the connectivity duration curves (CDC) to distinguish between active and inactive cells resulted in a maximum NSE of 0.98 at a storage threshold of 293 mm (Figure 4.6, red line in main figure). However, NSEs were nearly identical over a range of ± 4 mm around the 293 mm storage threshold, only differing in the third decimal. The calibrated peak network connectivity for 2007 slightly overestimated (at 81% versus 72%) the empirical CDC. The largest differences in the CDCs occur between approximately 15% and 30% annual exceedance, when the simulated CDC slightly underestimated network connectivity. However, a linear regression between the observed

and simulated CDCs reveals a good fit, with the regression line matching the 1:1 line closely (Figure 4.7). The 293 mm activation threshold resulted in a conductivity value of 12.7 mm/min on the calibrated storage-release curve (Figure 4.8). Conductivities were generally very low (<2 mm/min) below 200 mm of storage.

Evolution of Watershed Connectivity

The extent of active area was smallest during the baseflow period from summer through spring, before snowmelt started (Figure 4.9 a, f for 2007, g, l for 2008, Table 4.2), with less than 2% of the watershed active (by the definition set forth in this paper). With the onset of snowmelt in the lower elevations, areas in the lower part of the watershed became active (Figure 4.9 b). Over the course of the melt period in 2007, the mid- and upper sections of the watershed became active (Figure 4.9 c, d) until toward the end of snowmelt, runoff was mostly generated in the higher elevations and larger drainages in the lower elevations (Figure 4.9 e). During peak flow approximately 76% of the watershed area was active. Less than 2.5 months after peak runoff, only 2% of the watershed areas were active and only the largest drainages contributed to stream runoff (Figure 4.9 f). During the 2008 melt period the onset of melt was more homogenous with approximately equal areas in upper and lower elevations being active (Figure 4.9 h, i). Active areas during the later part of melt were focused on the mid- and upper sections similar to 2007 (Figure 4.9 j, k).

Contributing areas were highly spatially and temporally variable within years but showed similar patterns in both years of the study period (Figure 4.10, blue shaded

areas). A Pearson correlation coefficient between upslope contributing area and the fraction of time a hillslope cell was connected to the stream network was $r_p=0.69$. For purely topographically controlled water redistribution this correlation could be expected to be closer to 1. However, water uptake by vegetation is an important factor that exerts a control on the storage in a cell and both upslope and downslope flowpath connectivity and water redistribution. Spatially disaggregated ET leads to heterogeneous water withdrawal, further decreasing the correlation between UAA and storage. Other confounding factors influencing storage—and thus connected area—include spatially variable snow accumulation and melt/precipitation dynamics. There was a considerable difference in snow accumulation between the lower and higher elevations in both 2007 and 2008, with the highest elevation DEM cell accumulating 273 mm and 279 mm, respectively, more SWE than the lowest elevation DEM cell. Exceeding the storage threshold for connectivity for cells with similar upslope area is thus more likely for higher elevations that received more snow. The resulting spatial patterns of the active and contributing areas were thus a result of topographically driven lateral water redistribution, water uptake by vegetation, and spatially variable precipitation and snowmelt inputs.

The active and contributing areas at equal runoff were very different for the rising and the falling limb of the snowmelt generated hydrographs in both years (Figure 4.10 and Figure 4.11). While the distinction between rising and falling limb was very distinct, with the low elevations dominating the rising limb and the high elevations producing most of the runoff on the falling limb, leading to hysteretic behavior in runoff source areas, (Figure 4.11, left panel). The differences in 2008 were not as pronounced and the higher

elevations were the dominant source of runoff at those stages of both rising and falling limb of the hydrograph (Figure 4.11, right panel).

The differences in runoff source areas become more apparent in the hillslope width functions (Figure 4.12). The contributing area width functions indicate how far away from the stream network the contributing areas are (following their downslope flowpaths). In both years, approximately 35% of pre-melt runoff was generated mainly in near-stream and riparian areas at a distance of less than 60 m from the stream (black solid lines in Figure 4.12). With beginning snowmelt, those areas expanded to more locations parallel—and close—to the stream network. During the initial rising limb of the hydrograph in 2007, 56% of runoff was generated on areas less than 300 m from the stream network (Figure 4.12, 05/08 in upper panel). Subsequently hillslopes became active and during peak flow. In addition to the at first mostly stream-parallel expansion of contributing areas, After riparian and near-stream areas started contributing, hillslopes along the stream began to contribute to runoff. At peak flow, 50% of contributing areas were at distances of less than 450 m and 500 m in 2007 and 2008, respectively (Figure 4.12, 05/10 in upper panel and 06/03 in lower panel). A shift occurred during the later stages of the snowmelt event, when more areas at distances >500 m became connected, while simultaneously small hillslopes became disconnected. As a result, during the final streamflow recession in both 2007 and 2008, 59% and 67% of runoff, respectively, were generated on areas at distances of at least 500 m from the stream network (Figure 4.12, 06/03 in top panel, 06/27 in bottom panel).

Even though the maximum area connected at any day in the two-year study period was 71%, over the course of the two years 90% of the cells became active for at least one

day (Figure 4.13). The areas that never connected to the stream network were only 10% of the watershed. Those dry areas were located on the (mostly) subtle ridges between the larger hillslope drainages but there were also small individual patches that receive only few inflows, nested within the large drainage networks (Figure 4.13). The average active and contributing areas over the two-year simulation (731 day) period were 7.9% and 5.7%, respectively.

Discussion

Despite long-term effort by the hydrologic community, tracing streamflow back to its source areas remains a grand challenge. To partially address this, we developed and present a parsimonious rainfall-runoff modeling framework that allows for tracking of the spatial distribution of watershed storage and thereby watershed runoff source areas. Utilizing empirical measurements of hillslope-riparian-stream connectivity we approximated active and contributing areas based on a calibrated storage threshold. The discussion focuses on general model performance, the meaningfulness of the calibrated storage threshold, and the implications of temporally variable contributing areas for predicting and interpreting observed streamflow dynamics.

Model assessment

Calibration The runoff simulated over the two-year period matched the observed runoff well with a NSE of 0.86 over the two simulated water years. While fitting the hydrograph is the minimum requirement for any hydrologic model, we were particularly interested in the spatial distribution of storage in order to use the model as an interpretative tool to learn about watershed functioning. The resulting storage patterns for both years reflected a realistic scenario with carry-over storage magnitude and patterns from year to the next. Single-year calibration storage states optimized for individual years could either retain or release too much water in the alternate year, thereby matching an individual year's hydrograph well at the expense of the alternate year(s). This suggests and reinforces the importance of antecedent conditions and model conditioning even if the period of interest is short. In fact, numerous studies have documented the strong memory effect of watershed storage state on future runoff behavior (e.g. Fedora and Beschta [1989]; Istanbulluoglu et al. [2012]; Nippgen et al. [in review]).

Internal functioning and structure Despite the reasonable fit to the hydrograph, the question remains whether the simplified model structure adequately captures the runoff processes at TCEF. The timing and peaks of the simulated runoff time series match the observed runoff well during the melt period and, like the observed runoff, the model does not respond to summer storms. Summer precipitation events at TCEF usually do not lead to an increase in streamflow. The precipitation gets absorbed in the upper centimeters of the soil and subsequently evaporates, similar to what McNamara et al. [2005] reported from a watershed in Idaho with comparable climatic conditions.

Runoff source areas exhibited a strong topographic control during snowmelt and extended far up the hillslopes, with smaller hillslopes becoming connected during the main melt period. During the recession period the contributing areas contracted and only left main drainages and channel heads active. This is on accordance with empirical observations at TCEF that determined that topography drives hydrologic response during the melt period (e.g. *Jencso et al.* [2009]; *Jencso and McGlynn* [2011]; *Payn et al.* [2012]). The good fit between the simulated and empirical connectivity duration curves furthermore exemplifies this.

However, while the general runoff dynamics are captured well, the model slightly underestimates the initial rising limb of the 2007 melt period and overestimates runoff in the fall of 2006. However, it is important to note that those under- and overestimations are likely exacerbated by the modeled precipitation input. While the modeled input is reasonable, it simulates a brief but intense melt period at the Stringer Creek SNOTEL site in early October 2006 with 60 mm of melt over the course of 10 days. This melt period does not occur in the observed melt record at the site. Likewise, the delayed response in

simulated runoff is probably caused by a delayed melt input, i.e. the observed runoff begins to increase 2-3 days prior to the onset of simulated melt. Regardless, these details do not impact the general model behavior but rather indicate model and system sensitivity to precipitation.

In addition to uncertainties in the simulated input, the underestimation of runoff peaks in 2008 may hint that the watershed potentially became too dry over the eight-months baseflow period during the simulation. As a consequence, cell storage states during peak melt may not increase sufficiently to achieve high flux rates. Owing to the nature of the highly non-linear response function, even small differences at high storage states will lead to larger changes in the resulting flux than the same difference at smaller storage states (Figure 4.8), for example the increase in *Cond* from 300 mm storage to 320 mm is 5.3 mm/min, while the increase from 450 mm of storage to 470 mm is 23.5 mm/min.

A possible explanation for overly dry watershed conditions could be the lack of a deeper groundwater component in WECOH as all runoff is generated in the soil horizon from the watershed uplands. While this is a justified assumption for the melt period, this may not be true for the baseflow period. It has previously been documented that hillslope soil water alone can sustain baseflow [Hewlett and Hibbert, 1963], but there is evidence at TCEF that at least during baseflow some of the runoff may originate from deeper groundwater layers. Jencso and McGlynn [2011] and Nippgen *et al.* [2011] determined that elevated baseflow levels and delayed response times were evident in subwatersheds at TCEF underlain by sandstone as opposed to subwatersheds without sandstone surficial geology. This suggests increased recharge into the bedrock groundwater during the melt season and subsequent extended outflow during the growing season, thereby sustaining

more elevated baseflow in some of the TCEF sub-watersheds. *Jencso and McGlynn* [2011] documented and *Payn et al.* [2012] further suggested that the hydrologic controls on runoff at TCEF may shift from mainly topography-dominated flow during the snowmelt period to increasing groundwater contributions toward the end of the growing season and through the winter.

Not including groundwater flow in WECOH may lead to artificially high simulated drainage of the watershed uplands and thus to storage conditions at the end of the baseflow period that are too low to a) be responsive enough at the beginning of the melt period, or b) raise storage at peak flow to storage states that allow for greater water flux. If the cells around the stream within WECOH dry down too much over the course of the summer and through the winter, the initial melt magnitudes may not suffice to raise cell storage enough to achieve higher conductivities for an increase in streamflow. However, it is uncertain if the underestimation of peak runoff in 2008 is a result of storage states that are too low or if it is related to the spatiotemporal estimation of melt and precipitation inputs. Although including a deeper groundwater component in WECOH might slightly improve the response in the early stages of snowmelt or at peak flow, WECOH's simple model structure appears generally adequate to represent a broad range of hydrologic behavior at TCEF.

From storage to contributing areas Our objective was to develop and evaluate a parsimonious distributed water balance model to inform understanding of streamflow source areas through time. In order to translate information on distributed storage state to dominant runoff source areas, we built on widely recognized threshold behavior in hydrologic systems. For example, at Maimai in New Zealand Mosley [1979] documented that exceedance of a rainfall threshold led to activation of preferential flowpathways (i.e. macropores) and rapid water delivery to the stream network. Woods and Rowe [1996] further observed rainfall thresholds that led to variable hillslope trench face runoff magnitudes. McGlynn and McDonnell [2003a] later separated watershed response—also at Maimai—into contributions from riparian areas and hillslopes and concluded that hillslope areas delivered water in addition to the riparian areas as a function of precipitation magnitude and initial soil moisture conditions. These behaviors are similar to what we observed in our study, where at first the runoff source areas were limited to near-stream riparian areas and then extended to the hillslopes as the distributed storage states increased. Precipitation thresholds above which saturated hillslope throughflow response increased were also determined in other geographic regions such as the Panola Mountain Research Watershed [Tromp-van Meerveld and McDonnell, 2006b]. However, in most cases it is likely not just a simple precipitation threshold, but as McGlynn and McDonnell [2003a] suggest a combination of precipitation magnitude and antecedent soil conditions that lead to saturated hillslope throughflow initiation. Therefore, here we did not approximate the connectivity threshold as a precipitation/melt depth but rather as storage state. Since surface runoff at TCEF is restricted to limited areas even during the wettest times, we focus on thresholds for subsurface flow initiation.

We used the connectivity duration relationship established by Jencso et al. [2009] to calibrate the threshold for active/inactive cells. This approach resulted in a storage threshold of 293 mm below which we consider a cell as inactive. Assuming a uniform soil depth of 1000 mm and 50% porosity, this translates to 29% volumetric soil water content (SWC) when integrated over the entire soil profile. This value is similar to SWC thresholds reported at other sites, e.g. 45% at 0-30 cm soil depth in a watershed in the Italian Alps [Penna et al., 2011] and 37% at 70 cm soil depth at the Panola Mountain Research Watershed [Peters et al., 2003]. At TCEF Jencso et al. [2009] determined that hydrologic connectivity was tied to the establishment of a shallow water table on the bedrock-soil interface (saturated throughflow). In the WECOH model, calibration of the storage threshold to differentiate between active and inactive cells was based on Jencso et al.'s [2009] empirical water table observations (i.e. distillation of water table time series to a binary variable of presence/absence). The storage-conductivity relationship used in WECOH itself does not differentiate between saturated and unsaturated conditions. It is therefore a threshold above which the rate of water delivery down-gradient changes more abruptly.

Lehmann et al. [2007] argue that a mean threshold for flux initiation may not be an adequate assumption because of hillslope heterogeneity. For example, the fill-and-spill hypothesis for a hillslope throughflow, as outlined by Tromp-van Meerveld and McDonnell [2006a], suggests that flow on hillslopes initiates when depressions in the bedrock surface fill up and then water spills out of those depressions to fill up other depressions. The maximum outflow originates after all depressions have been filled and are connected. While conceptually compelling at the small scale in certain environments,

fill-and-spill is likely not a widely significant factor at the watershed scale where hundreds of hillslopes of varying length and area contribute to streamflow. At TCEF, soil depth to bedrock has been found to be relatively consistent across the hillslopes (based on the installation on >150 shallow groundwater wells). While variability in bedrock surface will indeed affect the onset of runoff on a small (<1 ha) hillslopes, the larger spatial extent of the entire LSC watershed (555ha) would average out potential heterogeneities in the bedrock.

While a large distribution of hillslope sizes will average out some structural heterogeneity, it may introduce issues in determining a response threshold based on precipitation. Although individual hillslope throughflow development has often been associated with precipitation thresholds (e.g. *Buttle et al.* [2004]; *Tromp-van Meerveld and McDonnell* [2006b]; *McGuire and McDonnell* [2010]), it was likely often due to hillslope storage thresholds rather than event size per se. In this case, different size hillslopes would have different precipitation/input thresholds for hydrologic response because of the storage potential and legacy of past precipitation and redistribution downslope. This means larger hillslopes would require less precipitation to connect hydrologically via shallow throughflow to downslope riparian zones and streams because of their size (cf *Jencso et al.* [2009]). In fact, this is a basic premise of the Catchment Connectivity Model (CCM) recently described by [*Smith et al.*, 2013] that has been shown to represent streamflow and be consistent with internal watershed observations of hydrologic connectivity (albeit at the semi-distributed hillslope scale). There is some similarity to the manner in which stream channel initiation thresholds generally depend on contributing area but where streamflow initiation through time is also sensitive to

storage / precipitation thresholds (e.g. *Montgomery and Dietrich* [1992]). Thresholds based on storage/soil water content conditions encompass both area and precipitation, making them more robust to heterogeneities in watershed structure and hillslope size.

The storage-based subsurface throughflow threshold in WECOH provides defensible representation of general hydrologic response (streamflow, storage patterns) and therefore shows promise as a tool for inference into watershed scale processes including the prediction and documentation of time varying contributing areas.

Variable Contributing Area

Variation in the spatial extent of contributing area ranged from less than 1% in the fall and winter to more than 70% during peak snowmelt (Figure 4.10). This represents an effective watershed area that changed from hour to hour, with effective watershed area defined as the area actively contributing to the runoff generation process at any point in time. This effective watershed area could be located near the stream network or in distal portions of the watershed. It follows then that a second assessment of an area's influence on watershed runoff would include the flowpath distance to the stream network.

At TCEF, in the beginning of snowmelt only a portion of riparian areas and the drainages of the channel heads contributed to runoff (contributing area <2%), those areas expanded with the onset of melt, activating the “traditional” variable source areas parallel—and close—to the stream network. This is also evident (and especially pronounced in 2007) in the hillslope width functions that indicated an increase in contributing area at distances of less than 300 m from the stream network (Figure 4.12). However, as opposed to the *Dunne and Black* [1970] variable source areas, the runoff generated at TCEF was

almost exclusively subsurface flow. After riparian and near-stream areas became contributing areas, some hillslopes along the stream network activated and became contributing areas, extending upslope as a function of upslope area, vegetation mediated ET, and input amount. The increase in contributing area over this period largely occurred up-gradient of the channel heads during this time of maximum connectivity. With snowmelt cessation, many of the hillslope areas fell below the active threshold, leaving only the larger drainages connected. This is demonstrated by hillslope width functions in the transition from 05/10 to 05/28 in 2007 (Figure 4.12, upper panel) and 06/03 to 06/17 in 2008 (Figure 4.12, bottom panel).

In addition to changing connected area spatial extents and distances from the network, specific watershed runoff source areas have major implications for watershed processes driven or influenced by hydrology and can provide a first approximation of solute source areas and sensitive or more influential areas of watersheds.

This spatial and temporal dynamic in contributing area has major implications for watershed processes driven or influenced by hydrology. Knowledge of runoff source areas and how they change through time is critical for deciphering water quantity and composition signals measured at the watershed outlet. For example, the transport of DOC to the stream network (e.g. *McGlynn and McDonnell* [2003b]; *Pacific et al.* [2010]) greatly depends on the spatial extent of contributing areas and the intersection of DOC accumulation and connectivity-driven mobilization of DOC to streams. Temporally dynamic streamflow and composition source areas can change in time, even on individual rain event timescales, and may partially explain observed streamflow solute behavior across a range of systems including but not limited to TCEF.

Variable contributing areas and their extent and associated thresholds are likely different for headwater watersheds across geographic regions (see for example the compilation of hillslope precipitation thresholds in *Weiler et al.* [2005] or *Ali et al.* [2013]). The importance of the smaller scale patterns of hillslope to stream connectivity for hydrograph dynamics, however, can change when moving from headwater systems (zero- to third-order) in a downstream direction to larger river systems ($\geq$ e.g. third or fourth order). There can be a shift from hillslope controlled runoff dynamics in headwaters to network and network geometry controlled runoff in larger river systems (e.g. *McGlynn et al.* [2004]), thereby obfuscating landscape (hillslope-riparian-stream) connectivity with increasing scale. Obviously, however, headwater watersheds and downstream waters are intrinsically linked [*Alexander et al.*, 2007], highlighting the need to better understand how headwaters influence and determine the character of larger river systems (e.g. *Lowe and Likens* [2005]; *Bishop et al.* [2008]). This can become particularly important for assessment of land cover/land use change effects and their mitigation. For example, the effect of harvesting operations and species conversion on the water yield of watersheds has been well documented since the early 20th century (e.g. *Bates and Henry* [1928]; *Hibbert* [1967]). However, if increasing water yield with targeted vegetation removal is desirable, predicting those areas that are larger contributors to streamflow or would become contributors to streamflow with small increases in local storage then becomes a viable model enabled strategy.

Predicting and mapping watershed hydrologic connectivity through time has implications well beyond water quantity. For example, many studies have documented the impact of human induced land use change in headwater watersheds on water quality

(e.g. *Bolstad and Swank* [1997]; *Clinton and Vose* [2006]; *Gardner and McGlynn* [2009]) and have begun to target those changes and landscape locations most influential on downstream water quality [*Gardner et al.*, 2011]. Recognizing that the effective watershed is continuously changing, especially during precipitation events, is a first step toward adequately considering landscape hydrologic and runoff generation processes into streamflow composition source attribution. It should be noted however, that documentation of active and connected contributing areas limit the likely water constituent source areas but that these zones themselves do not equally contribute to streamflow composition. Flowpath distances to the stream within these connected areas and the threshold change from subsurface to overland flow can offer a next approximation of influence within the contributing areas.

Recognition of the dynamic nature of active and connected watershed areas through time is critical to understanding and managing land and related water resources. Interpretation of observed watershed runoff dynamics and magnitudes in the context of the history of precipitation (e.g. *Nippgen et al.* [in review]), its redistribution in the landscape, and how this can influence both baseflow and hydrologic response to the next precipitation event is fundamental for landscape and process attribution of observed behavior. Prediction then requires application of well-constrained spatial models consistent with observed processes to estimate landscape hydrological connectivity to stream networks through time and more refined analyses to address particular challenges or questions.

Conclusions

Recognizing, understanding, and predicting streamflow source areas through time requires fundamental observations of terrestrial hydrology and influences on water redistribution and storage and efficient numerical representation that is consistent with dominant processes and landscape scale hydrologic behavior. Here, we built on a comprehensive history of detailed spatial and temporal observations of landscape hydrologic connectivity, evapotranspiration, and streamflow to synthesize landscape scale process observations and understanding with a newly developed parsimonious watershed model. Our approach (WECOH model) includes the combined effects of topographically driven lateral water redistribution and vegetation mediated evapotranspiration to simulate spatial patterns of watershed water storage through time and streamflow. We calibrated the threshold storage necessary for saturated throughflow and its connectivity across the landscape using empirical observations of water table development (i.e. water table present or not). We calculated active and contributing areas (those active areas hydrologically connected to the stream network) to ascertain and learn from the spatial and temporal evolution of watershed connectivity over the course of two snowmelt seasons. We summarize our findings as follows:

- 1) Runoff source areas shifted over the course of each melt season from near stream areas early in the melt progression to lower hillslopes areas and finally areas further up the hillslopes when runoff was concentrated in the larger drainages. With snowmelt recession, the pattern reversed but hysteresis in the

pattern was evident, leading to different spatial extent and source areas for streamflow on the rising and falling limbs of the snowmelt hydrograph.

- 2) The maximum and minimum areas connected to the stream network at any day over the simulation period were 71% and <1%, respectively. Total active area (combination of active, connected and active, disconnected) was always greater than contributing area during the melt seasons but became similar during late summer and winter baseflow periods.
- 3) Over the course of two years 90% of the watershed area was connected for at least one day. The 10% of area that never became connected (above the threshold) to the stream network were either areas with low upslope accumulated area near ridges or small patches that received little inflows and were interspersed between the more dominant pattern of flow accumulation within the upland drainage network.

The WECOH modeling framework seeks to bridge the gap between knowledge gained from empirical observations and the usefulness of spatially distributed hydrologic models to inform and synthesize understanding at the landscape scale. WECOH is a simple modeling framework to examine how landscape hydrologic connectivity changes over time and how field observations can facilitate the identification of threshold behavior in hydrologic systems. Future research should begin to examine more explicitly the ecohydrologic feedback between vegetation patterns and landscape scale patterns of water redistribution, the utility of this framework across landscape settings and climate

regimes, and coupled hydrological-biogeochemical representation that includes variable contributing areas for water and solutes.

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Model results (snowmelt as well as rainfall-runoff) and streamflow data can be requested from the USFS or the first two authors. SNOTEL and LIDAR data can be downloaded at <http://www.wcc.nrcs.usda.gov/snow/> and <http://www.opentopography.org/>, respectively.

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Tables

Table 4.1: WECOH model parameters

Parameter	Description
$Cond_{\max}$	Maximum conductivity
b	Shape parameter for the storage-release function
$S_{\max}$ (fixed)	Maximum storage capacity
Vel_{stream} (fixed)	Stream velocity

Table 4.2: Active, contributing, and active non-contributing area, as well as runoff for 12 characteristic days of the 2-year study period

Date	Active area (%)	Contrib. area (%)	Active, disconn. area (%)	Q (mm/day)
04/27/07	< 2	< 2	0	0.16
05/08/07	44	28	16	7.26
05/10/07	76	71	5	14.52
05/28/07	73	57	17	11.04
06/03/07	49	35	14	7.20
07/25/07	2	< 2	< 1	0.30
04/13/08	1	1	0	0.06
05/20/08	53	27	26	5.26
06/03/08	68	52	16	10.13
06/17/08	73	57	15	9.83
06/27/08	38	26	12	5.08
07/22/08	3	2	1	0.60

Figures

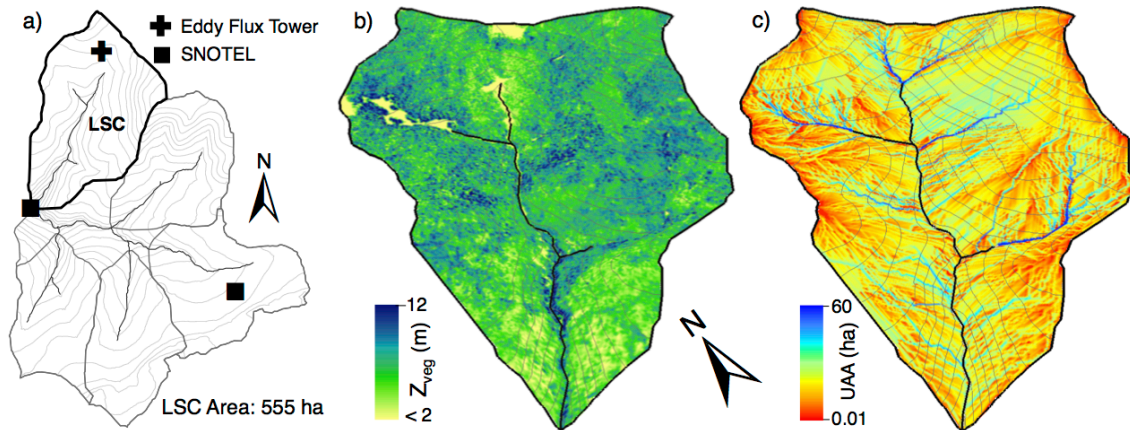


Figure 4.1: Lower Stringer Creek (LSC) subwatershed within the greater Tenderfoot Creek Experimental Forest and locations of SNOTEL stations and eddy covariance tower (a). Vegetation heights (10 m grid resolution) averaged from 1 m resolution LIDAR data (b). Upslope accumulated area (UAA, logged colorscale) calculated using a two-direction flow algorithm (c).

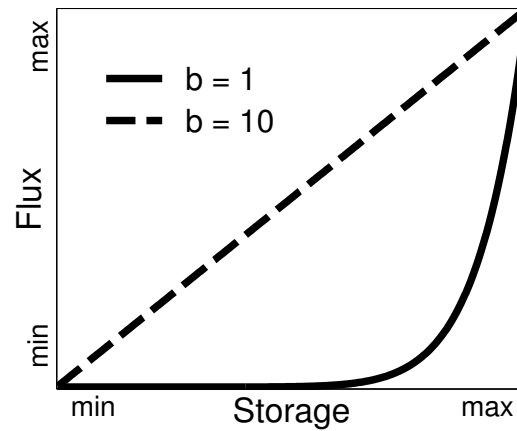


Figure 4.2: Relationship between cell storage and resulting flux. Pictured are two hypothetical relationships for different values of the shape parameter b . $b < 1$ is not shown as it would result in an unrealistic storage-flux relationship.

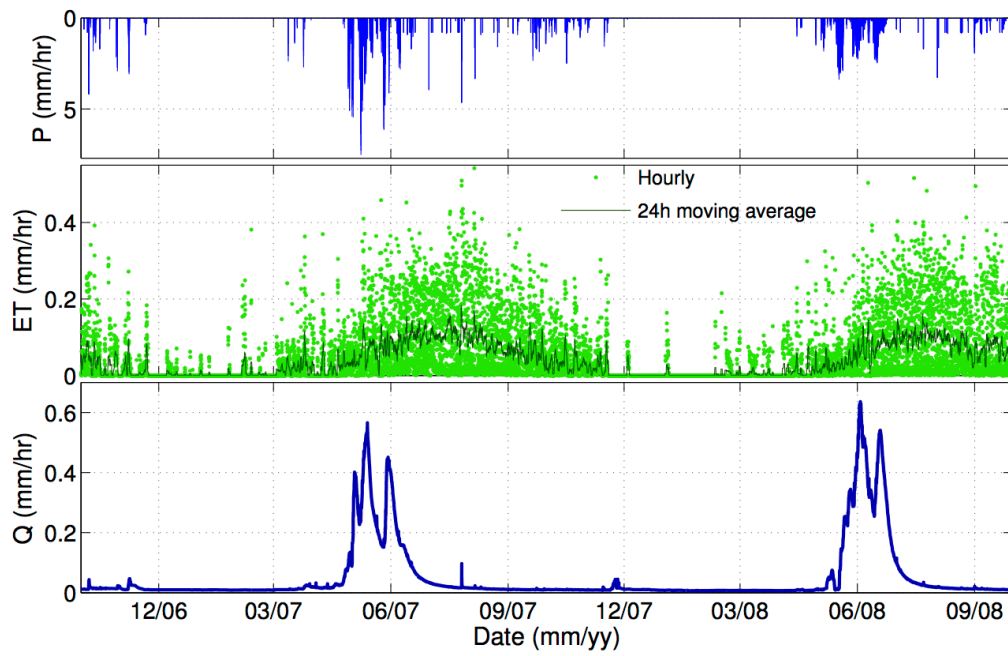


Figure 4.3: Overview of precipitation (P), evapotranspiration (ET), and Runoff (Q) data for the simulation period.

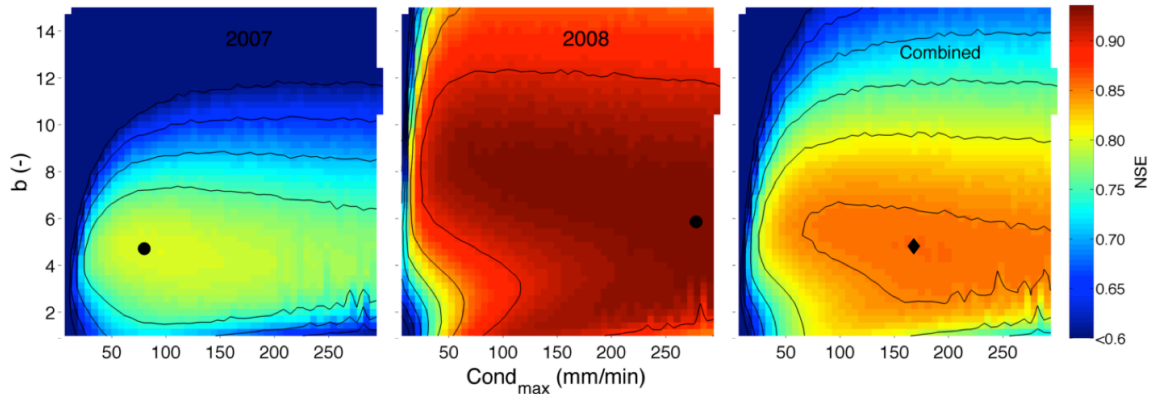


Figure 4.4: Parameter surfaces for runoff calibration for the 2007 (left panel) and 2008 (middle panel) WYs and the combined calibration for both years (right panel). It is important to note that the “combined” surface was not derived by averaging the “2007” and “2008” surfaces, but shows the NSE surface for 2-year runs. The optimal parameter combinations for 2007 and 2008 are denoted by the black dots, the final parameter combination is denoted by a black diamond.

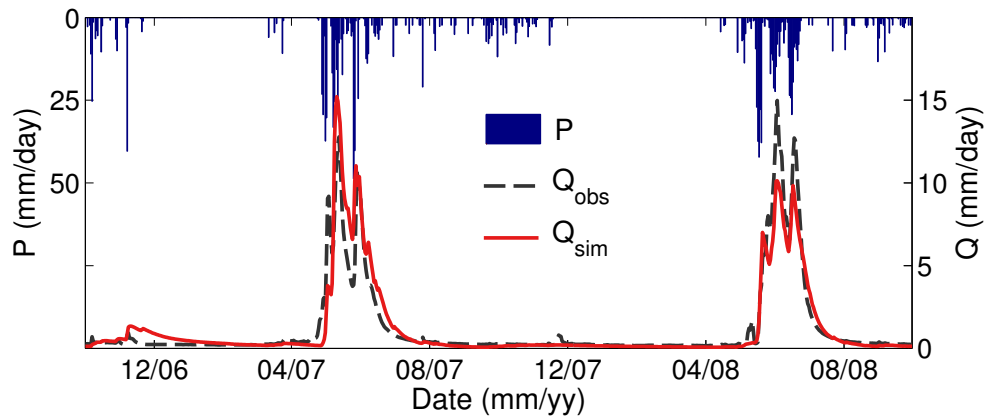


Figure 4.5: Average simulated precipitation (rain and snowmelt) and observed and simulated runoff for the 2-year simulation period.

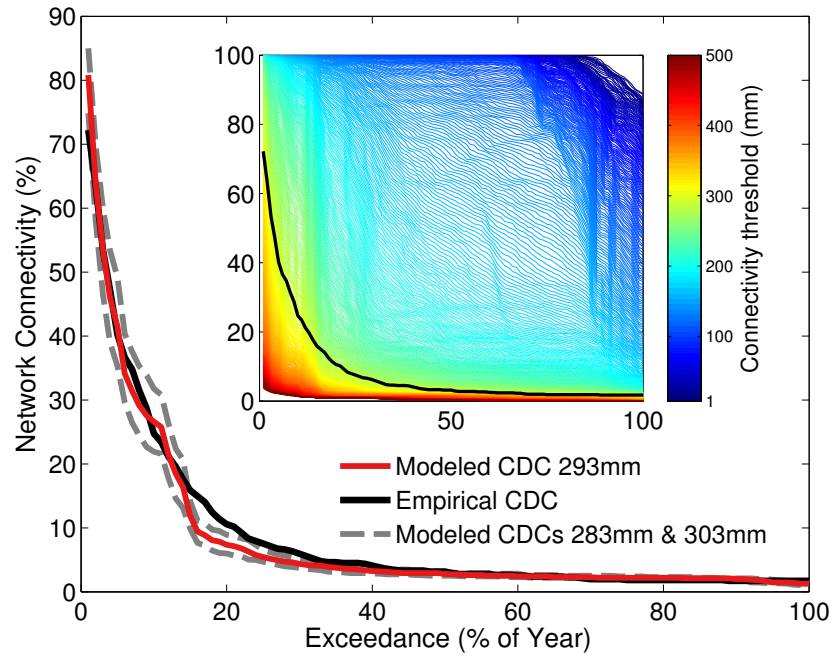


Figure 4.6: Empirical connectivity duration curve (*Jencso et al.*, 2009), denoted by black line, and modeled connectivity duration curve for hillslope-riparian-stream connectivity. A storage threshold between active and inactive cells of 293 mm resulted in a CDC that best fit the empirical CDC (denoted by the red line). Two CDCs 10 mm below and above the best threshold are shown for comparison (denoted by gray dashed lines). The inset shows the suite of CDCs from 1 to 500 mm thresholds in one mm increments as well as the empirical CDC.

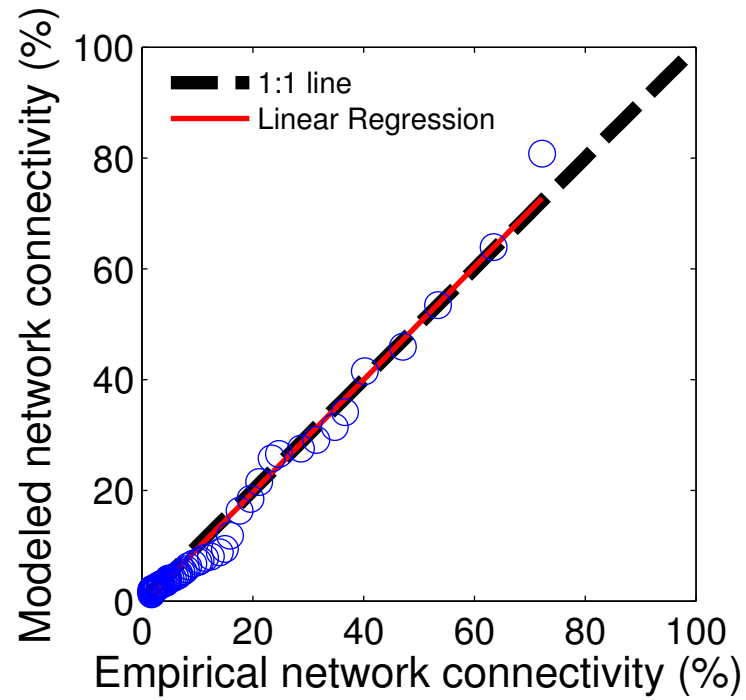


Figure 4.7: Bivariate plot of empirical and modeled CDC (blue circles) as well as the 1:1 line between them and the result from a simple linear regression between the two curves.

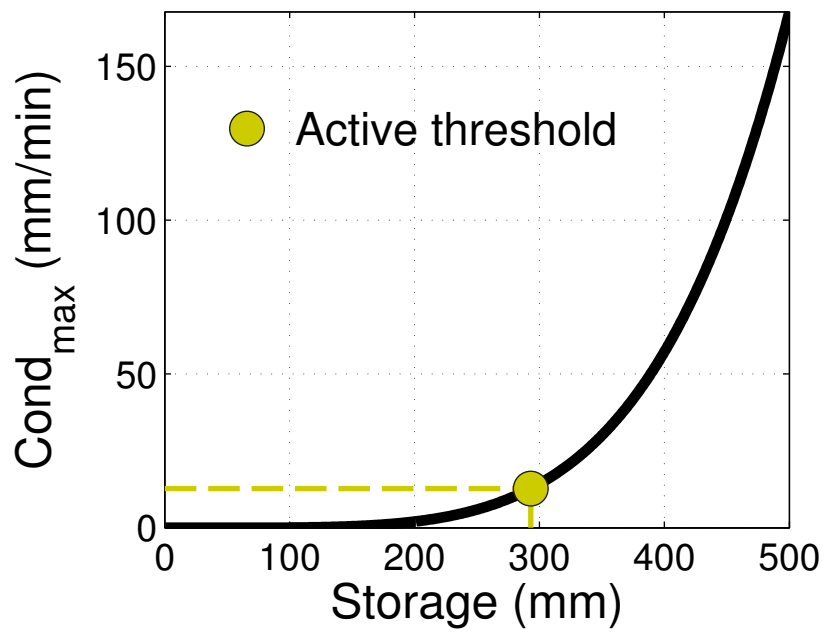


Figure 4.8: Storage-release curve with calibrated model parameters as well as the threshold for active/inactive cells at 293 mm (denoted by the yellow circle).

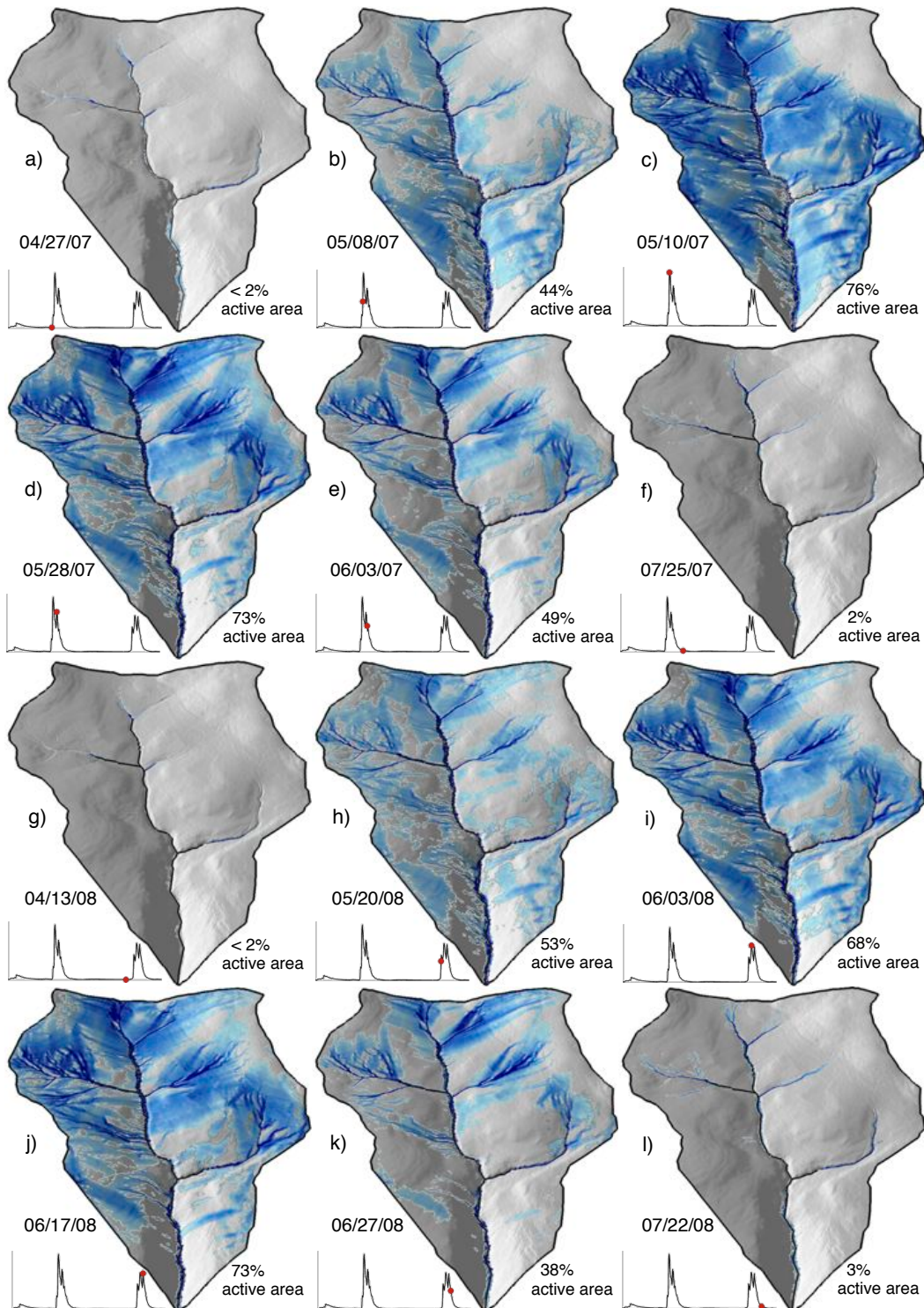


Figure 4.9: Evolution of watershed active area over the course of the 2007-2008 study period. Gray areas were below the 293 mm storage threshold and thus inactive. The blue shading denotes the degree of storage, with light blue areas at 293 mm to dark blue areas at 500 mm. The insets show the location of the snapshot on the hydrograph for reference.

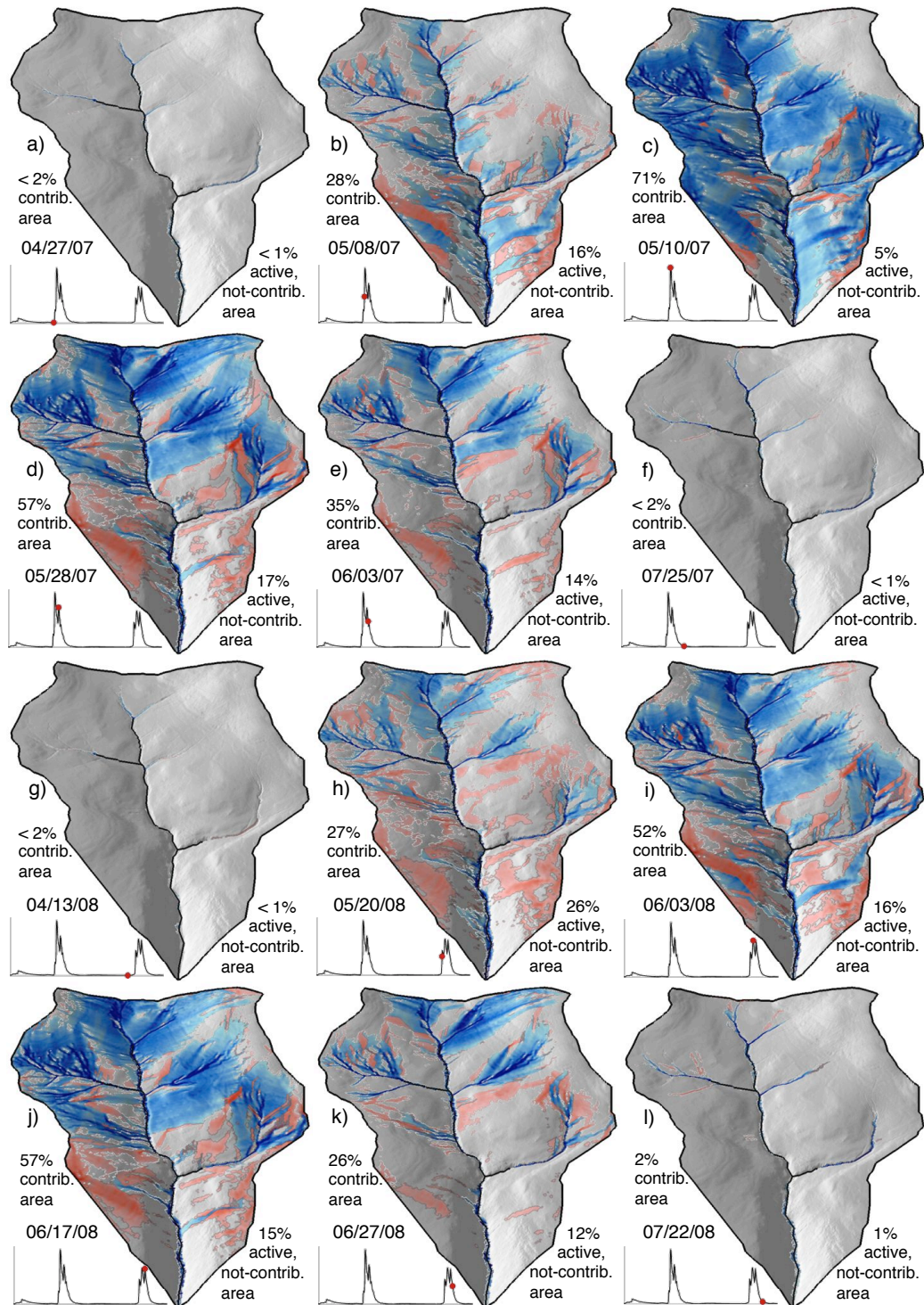


Figure 4.10: Extent of active and contributing areas at 12 spatial snapshots in time. Blue color denotes contributing areas; active, non-contributing areas are shown in red. The shading denotes the degree of storage, with light blue/red areas at 293 mm to dark blue/red areas at 500 mm. The insets show the location of the snapshot on the hydrograph for reference.

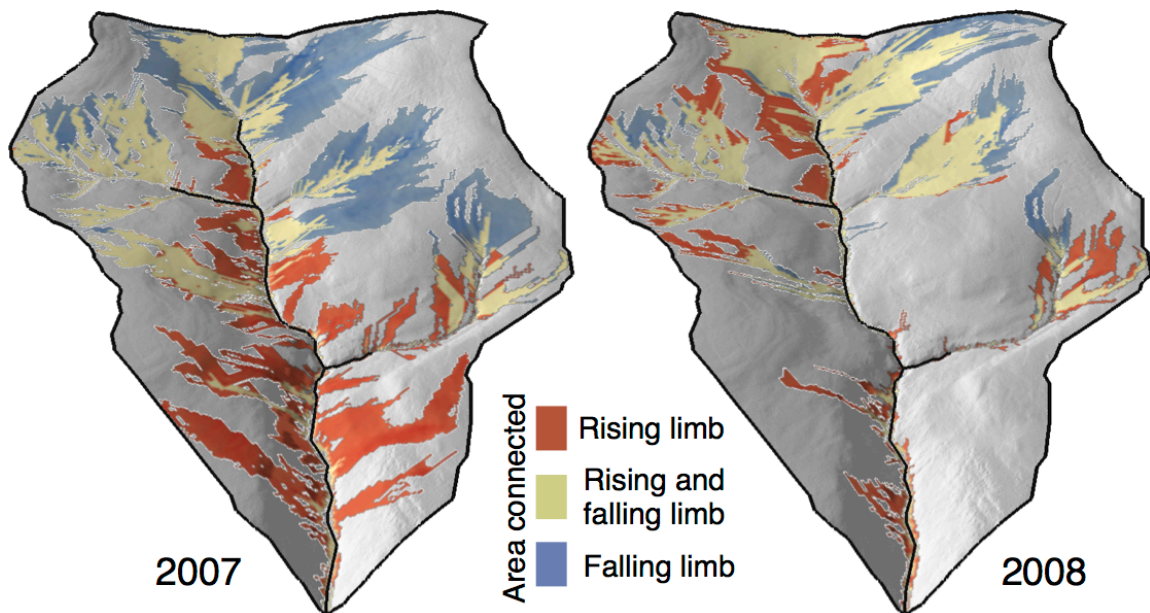


Figure 4.11: Connected areas on one day of both the rising and falling limb for the 2007 (May 8 and June 3) and 2008 (May 20 and June 27) water years. Runoff on select days on the rising and falling limb of each year was approximately equal.

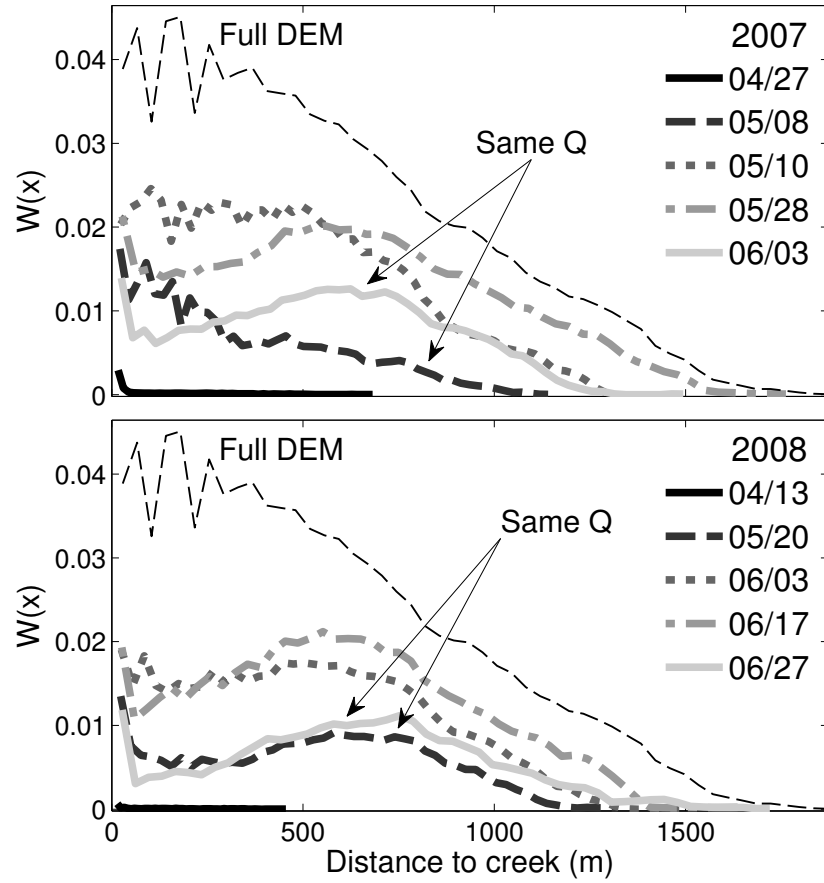


Figure 4.12: Hillslope width functions for contributing area at five days from the 2007 (top panel) and 2008 (bottom panel) water years. The thin dashed line shows the hillslope width function for the entire watershed for reference.

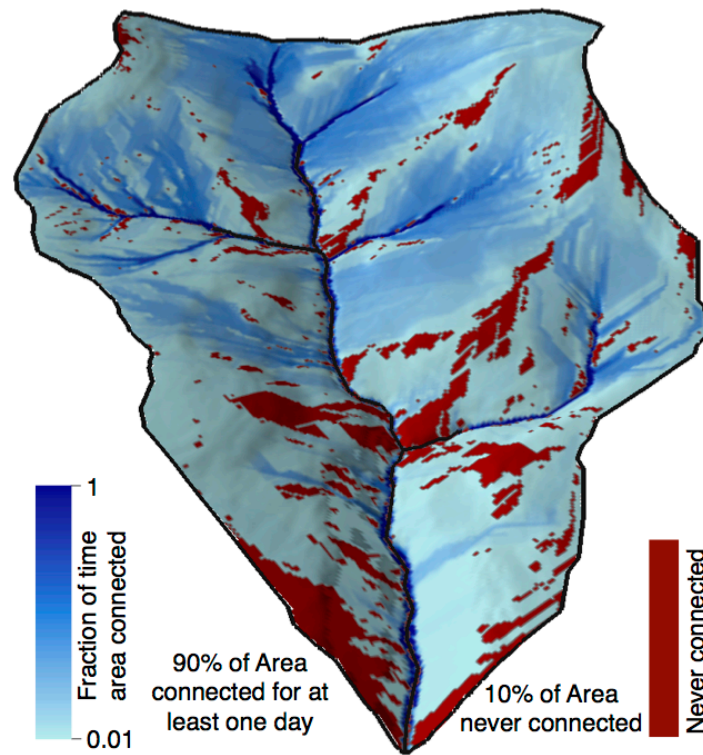


Figure 4.13: Connected areas (shades of blue) and areas that never connected (red) over the course of the two-year study period (colorscale is logged for better visualization).

5. SUMMARY

Much watershed-scale hydrology research has sought to increase understanding of how various feedbacks in the soil-vegetation-atmosphere continuum affect runoff. In fact, this has been a central focus for hydrologists since concentrated research began at the watershed scale. 70 years ago M. D. Hoover and C. R. Hursh [*Hoover and Hursh*, 1943] already correctly identified the drivers of hydrologic response in headwater streams:

“It has been recognized that topography, size, shape, vegetation, and soil-profile differ for individual drainage-areas. These inherent properties of watersheds together with climatic factors are the principal causes for variations in the hydrologic characteristics of different drainage areas.”

While we have gained much insight over the course of the last seven decades on *how* these metrics affect hydrologic response, many challenges remain. With this dissertation I seek to contribute to our understanding of watershed hydrology and have presented research that evaluates the effect of watershed structure (e.g. topography and vegetation) and climatic variability on various aspects of hydrologic response. Specifically, I examined how hydrologic response times are affected by watershed structure and climatic variability, how watershed memory—as mediated by topography, vegetation, and climatic variability—affects watershed runoff, and how the spatial distribution of water stored in a system affects runoff source areas over the course of two years in a snowmelt-dominated system.

In Chapter 2, *Landscape structure and climate influences on hydrologic response*, we demonstrated that hydrologic response times strongly depended on both structural as well as climatic properties. Watershed averaged mean response times showed strong correlations with several topographic metrics and the underlying geology. We demonstrated that even seemingly subtle differences in watershed structure between the watersheds resulted in response time differences on the order of days. Annually averaged mean response times were strongly negatively correlated with maximum annual snow water equivalent, resulting in faster response times with increasing available meltwater. Two watersheds that were outliers in the regressions were likely influenced by a groundwater component that is not present to the same extent in the remaining watersheds. These results suggest that hydrologic response time is dependent on both climatic as well as watershed structural metrics and that even subtle variability in either can lead to significant differences in hydrologic response. Furthermore the research suggested that topography exhibited a stronger control on runoff ratios than climatic variability.

In Chapter 3, *Watershed memory at the Coweeta Hydrologic Laboratory: The effect of past precipitation and storage on hydrologic response*, we examined the effect of precipitation variability on runoff in five watersheds at the Coweeta Hydrologic Laboratory. Cycles of high and low precipitation introduced variability in watershed storage that influenced the hydrologic response. This legacy effect can be described as watershed memory. We found strong memory effects from monthly to annual time scales across all five study watersheds. In general the previous time step's precipitation was as important or more important than the current time step's precipitation in determining the

current time step's runoff ratio. This effect was similar for watersheds with opposing dominant aspect (north versus south) and vegetation types (coniferous versus deciduous). We hypothesize that a watershed's position on its storage-flux curve is strongly influenced by vegetation type. Therefore, differences in storage between coniferous and deciduous watersheds can lead to different hydrologic response. Those differences were greatest during wet times and the dormant season and smallest during dry periods and the growing season.

In Chapter 4, *The temporal evolution of variable contributing areas*, we presented a parsimonious but fully distributed rainfall-runoff model that allowed for spatial quantification of watershed storage through time. Utilizing empirical measurements of hydrologic connectivity, defined as shallow water table connections in the hillslope-riparian-stream zones, we approximated the dynamics of watershed-wide runoff source areas. We showed that the spatiotemporal extent of the connected areas, i.e. the areas that contributed water to the stream network, was highly variable. While the maximum amount of contributing area was 71% for a single day during peak snowmelt, over the course of the two years 90% of the watershed areas were connected for at least one day. Over the course of the spring snowmelt events the contributing areas shifted from near-stream riparian areas and bigger drainages early in the events to lower hillslopes as the contributing areas extended uphill. The areas furthest away from the stream became most active during the hydrograph recession. We have worked to bridge the often-cited gap between hydrologic modeler and experimentalist with the development and application of the WECOH model that incorporates empirically-gained knowledge. This synergy shows

promise for further informing understanding of spatially distributed processes that are difficult to measure in the field.

This dissertation contributed new hydrologic understanding of how watershed properties (topography, geology, vegetation etc.), climatic variability, and the interactions between them affect hydrologic response at the watershed scale. In Chapters 3 and 4 we treated various watershed structural metrics as static parameters to make inferences about the lumped hydrologic response of watersheds in Montana and North Carolina. However, Chapter 4 highlighted that watersheds as a whole are highly dynamic systems. Documentation of changing watershed runoff source areas demonstrated that the effective portions of a given watershed are dynamic in time. This suggests that static metrics or whole watershed metrics might be mismatched with many hydrological and biogeochemical questions. With WECOH, the combination of topographically driven lateral water redistribution, spatially variable melt/precipitation inputs, and water loss through evapotranspiration resulted in temporally dynamic spatial patterns of contributing watershed area. These spatially and temporally variable runoff source areas have implications for water quantity, source composition, and quality. Even though the general notion of the variable source area was introduced more than 50 years ago, the concept has still not been fully embraced in the hydrologic and related sciences.

Clearly, runoff measured at the watershed outlet has a spatial and a temporal component, in terms of *where* it came from and *when* it was generated. This can be particularly relevant for biogeochemical questions and process/source area attribution. Coupling a spatially distributed model that can identify contributing areas with

biogeochemical process models could greatly improve the prediction of nutrient (or contaminant) export from watersheds. Additionally, how watershed memory affects the formation of contributing areas is not yet known. We have highlighted the importance of watershed memory on hydrologic response in Chapter 3. Now, another central question is how the variability of past precipitation affects storage patterns and the formation of active and contributing areas of a watershed. This could be especially crucial for geographic regions with highly variable precipitation and large soil storage potential (such as the southeastern United States). At TCEF, changes in storage may not have an equally strong effect on streamflow from year to year because the soil mantle is much shallower than at Coweeta for example and the watershed storage state may essentially reset with each snowmelt season. In fact, at TCEF we found that the average annual response time was highly correlated with the snow water equivalent of the same year, but not with the snow water equivalent of the previous year (data not published). However, this has yet to be tested rigorously through empirical data analyses and modeling applications.

Based on the results of my research I suggest the following topics for future research:

- 1) We have demonstrated that variability in precipitation lead to changes in watershed storage. In return watershed storage greatly affected runoff in the following year. For systems like Coweeta that exhibit a strong sensitivity to precipitation variability, it will be critical to determine how precipitation induced storage variability affects runoff source areas, for example to make predictions about nutrient or contaminant export.

- 2) Vegetation has a fundamental impact on all components of the water balance and is a central part of ecohydrologic research. Disturbances such as beetle outbreaks, wildfire, or anthropogenically induced land use change make it imperative to quantify the effect of changing vegetation coverage on water resources. It is largely unknown how the spatial configuration of vegetation affects storage, flowpath continuity, and hydrologic connectivity.
- 3) Much of our knowledge on runoff source areas was generated in small headwater catchment or hillslope studies. It is largely unknown how runoff source areas and hydrologic connectivity change when moving from small headwater catchments to larger river systems. Being able to transfer landscape connectivity concepts to large river systems could ultimately assist with the delineation of areas with a “significant nexus” to navigable and other regulated water.
- 4) Urban watersheds exhibit a high degree of variability despite their seemingly organized structure. Response times are usually higher as much runoff is generated on impervious surfaces and then concentrated in storm drain systems. However, determining runoff source areas and hydrologic connectivity in urban areas remains a challenge.

This dissertation presented research on the effects of watershed structure and climate on hydrologic response. The work progressed from lumped representations of watershed response that emphasized the importance of watershed structure and vegetation on streamflow and storage, to detailed spatiotemporal representations of storage and runoff

source areas. We addressed knowledge gaps in the hydrologic sciences regarding the effect of watershed memory on runoff and the evolution of contributing areas in headwater catchments. The empirical analysis of runoff and precipitation data from Coweeta highlighted the importance of long-term data sets to assess the impact of climatic variability on hydrologic response. Furthermore we demonstrated the usefulness of parsimonious models, both lumped and spatially distributed, as learning tools to inform on watershed processes.

Reference cited

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APPENDIX

Snow melt modeling

Cold content was calculated as

$$w_c = \frac{c_i}{c_f} h_s (T_m - T_s) \quad (1)$$

where w_c is cold content in mm, c_i is specific heat capacity of ice (2.06 kJ kg⁻¹°C⁻¹), c_f is latent heat of fusion (334 kJ kg⁻¹), h_s is snow water equivalent (SWE), T_m is a temperature threshold for melt, and T_s is the snowpack temperature deficit (air temperature over snowpack averaged over a calibrated period of time, $Snow_{age}$). In case of $T_s > 0$, T_s was set to 0. Cold content can be considered a heat deficit to overcome before melt begins [ASCE, 1996; Schaefli and Huss, 2011]. For each time step a potential melt is calculated as

$$M_{pot} = ddf(T_i - TT) \quad (2)$$

where ddf is the degree-day factor (mm/deg C), T_i is air temperature at time step i (deg C), and TT is the threshold temperature (deg C) above which melt occurs. If potential melt is smaller than the cold content, the initial melt abstraction is set to potential melt. The actual melt is then calculated as the difference between potential melt and the melt abstraction. If the simulated snowpack contains less water than calculated as melt, the snowpack will fully melt

For our implementation of the snowmelt model at TCEF, the model contains four calibration parameters: a temperature threshold for each SNOTEL site above which melt

occurs and below which precipitation accumulates as snow, the degree-day factor that determines how much melt leaves the snowpack at a certain air temperature (identical for both SNOTEL sites), and a parameter that determines the length of the period to approximate snowpack cold content. Using a 10 m resolution digital elevation model we calculated precipitation, snow accumulation, and melt for each grid cell on an hourly time step based on observed air temperature and precipitation data from the two SNOTEL sites. Air temperature and precipitation at elevations in between and above the SNOTEL sites were calculated with linear temperature and precipitation lapse rates determined by observed data at the two SNOTEL sites.

The model was calibrated separately for the 2007 and 2008 water years with the observed snow water equivalent (SWE) time series of the Onion Park and Stringer Creek SNOTEL stations (see Figure 1 for locations) and the simulated SWE time series from the two cells that encompass the SNOTEL sites. We calculated Nash-Sutcliffe efficiencies (NSE) for the full SWE time series as well as the main melt period from peak SWE to complete meltout since this period represents the main water input to the ground over the course of the year. Furthermore we calculated differences in the timing and magnitudes of observed and simulated peak SWE magnitudes for the two SNOTEL sites.

The optimal parameter sets for both years were determined using a 4-step process. We calculated (1) NSEs for each SNOTEL site for the entire water year, (2) NSEs for each SNOTEL site for the main melt period only (peak SWE till melt out), (3) differences in the magnitudes of peak SWE between simulated and observed SWE time series for each SNOTEL site, and (4) differences in the timing of peak SWE between simulated and observed SWE time series for each SNOTEL site. In a first step we averaged the NSEs of

(1) and took either the 20 best parameter sets or all parameter sets within 2% of the best averaged parameter set (whichever resulted in a smaller subset). In a second step we averaged the melt-only NSEs of the previously derived subset between the two SNOTEL sites of (2) and chose the 10 parameter sets with the highest averaged melt-only NSEs. In step three we averaged the differences in peak SWE magnitudes of (3) between the SNOTEL sites and chose the 5 parameter sets with the minimum combined peak SWE differences. In the final step we calculated the averaged differences in peak SWE timing of (4) between the SNOTEL sites. The parameter set with the minimum differences in peak timing was the final parameter set for a particular water year (see Table A.1).

We also included an adjustment for aspect in the model so that grid cells that receive more solar insolation were assigned slightly increased melt rates than cells with less solar insolation. For this we approximated potential solar insolation during the spring melt phase (March 15 – June 15) for every grid cell using SAGA GIS, following *Böhner and Antonic* [2009], using the default values for solar constant (1367 W/m^2), height of atmosphere (12000 m), and vapor pressure (10 mbar). We then calculated mean insolation and assigned all cells within $\pm 10 \text{ kWh/m}^2$ of the mean the value 1. All grid cells with insolation values above or below this range were assigned values greater than or less than 1 based on a linear regression between the mean and maximum and minimum values, respectively. Since the impact of aspect on melt and runoff at TCEF was found to be much weaker than the elevation effect [*Smith and Marshall*, 2010], the aspect correction was fixed at 5%, i.e. the maximum and minimum values for the insolation correction factor were set to 1.05 and 0.95 respectively. The melt values for each time step were then multiplied by the insolation factor to either increase or decrease snow melt

based on aspect. The calibration was not affected by this factor as both SNOTEL sites fall within the range of $\pm 10 \text{ kWh/m}^2$ around the mean insolation.

Sublimation from the snowpack and canopy can constitute an important part of the annual snow energy balance (e.g. *Dozier and Melack* [1987]; *Molotch et al.* [2007]) but was not automatically accounted for in the snowmelt model. We approximated sublimation from the snow pack with the eddy covariance derived latent heat flux time series. For the adjustment we assumed that, when a snow pack was present, latent heat flux measured at the eddy covariance tower at temperatures below 0°C was sublimation from the snowpack rather than transpiration. The accumulated sublimation over the course of the winter and spring was subsequently subtracted from the SWE value of each cell, i.e. when SWE in a grid cell reached the value of sublimation during the melt phase, SWE of that cell was set to 0. The SNOTEL sites are located in small openings in the forest, while forested areas in LSC comprise more than 90% of the watershed. The sublimation adjustment therefore partially accounts for increased canopy interception in forests as compared to open areas.

Table A.1: Calibrated parameters (top part) and efficiencies (bottom part) of the snowmelt routine for the 2007 and 2008 water years. On is the Onion Park SNOTEL, Str the Stringer Creek SNOTEL, Peak denotes the difference in either peak SWE magnitude or timing between simulated and observed SWE time series)

SWE Calibration	2007	2008
TT_{On} (°C)	2.9	0.38
TT_{Str} (°C)	4.6	0.66
ddf (mm/hour/°C)	0.728	0.208
$Snow_{age}$ (days)	178	176
NSE _{On}	0.98	0.98
NSE _{Str}	0.85	0.99
NSE _{On_melt}	0.87	0.88
NSE _{Str_melt}	0.94	0.93
Peak _{MagnOn} (mm)	2	14
Peak _{MagnDiffStr} (mm)	34	11
Peak _{TimeOn} (hours)	13	16
Peak _{TimeStr} (hours)	66	30