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Publication Date
2011

Peer reviewed|Thesis/dissertation
An Interdisciplinary Approach to Characterize Flash Flood Occurrence Frequency for Mountainous Southern California

A dissertation submitted in partial satisfaction of the requirements for the degree Doctor of Philosophy in

Oceanography

by

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2011
This Dissertation of Theresa Marie Modrick Carpenter is approved, and it is acceptable in quality and form for publication on microfilm and electronically:


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University of California, San Diego

2011
DEDICATION

To my father, Stephen E. Modrick,

for his ever-present love, support, and faith.
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ACKNOWLEDGEMENTS

First and foremost, I would like to acknowledge and thank my family. Without their love, support, understanding, and sacrifice, these years of study and research would not have been possible. I would like to acknowledge Dr. Konstantine Georgakakos for his guidance as the chair of my committee, and for many years of support and mentorship. I would also like to acknowledge Dr. Dan Cayan for his candid comments and discussions throughout the research. I appreciate the time and effort of my entire doctoral committee.

I gratefully acknowledge the Board of Directors of the Hydrologic Research Center (HRC) for their support of my pursuit of this doctoral research and for the computational resources provided for this extensive modeling effort. To my colleagues at HRC, I acknowledge their support and appreciate their time and discussion of ideas.

I also would like to acknowledge the following individuals for providing various sources of data used in my research: Dr. Masao Kanamitsu of Scripps Institution of Oceanography for access to and discussion of the CaRD10 precipitation data; Mr. Tom Haltom, California Water Science Center, USGS, for providing hourly streamflow records; and Mr. Jeff Agajanian and Mr. Al Caldwell of the Poway Field Office, and Mr. Matt Scrudato and Ms. Katie Klock of the Santa Maria Field Office, USGS for allowing access to their stream survey reports.

Chapter 1 has been prepared for submission for publication in the *Journal of Geophysical Research*. The dissertation author was the primary investigator and author of this material.
Chapter 2 has been prepared for submission for publication in *Geomorphology*. The dissertation author was the primary investigator and author of this material.

Chapter 3 is currently being prepared for submission for publication. The dissertation author was the primary investigator and author of this material.

Chapter 4 is currently being prepared for submission for publication in the *Journal of Hydrology*. The dissertation author was the primary investigator and author of this material.
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PUBLICATIONS


ABSTRACT OF THE DISSERTATION

An Interdisciplinary Approach to Characterize Flash Flood Occurrence Frequency for Mountainous Southern California

by

Theresa Marie Modrick Carpenter

Doctor of Philosophy in Oceanography

University of California, San Diego, 2011

Konstantine P. Georgakakos, Chair
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The research presented in this dissertation is a prototype synthesis of hydrologic, meteorological, and geomorphologic principles and modeling toward the characterization of flash flood occurrence frequency in the southern California mountainous region. Globally, flash flooding has the highest fatality rate per person affected among natural disasters and occurs over small spatial (< 500 km$^2$) and short time (<12 hours) scales. This research characterizes regional flash flood occurrence potential with high spatial resolution (O[30 km$^2$]) through an interdisciplinary modeling approach that involves diagnostic analysis to understand the small- and large-scale
forcings and effects of flash flooding in southern California, geomorphologic principles to define the level of flood occurrence in streams draining flash flood prone basins, numerical modeling of orographically-forced precipitation with high resolution, hydrologic modeling of soil moisture evolution, and, as a synthesis of models and data, the simulation of flash flood occurrence potential.

The research begins with an investigation of observed precipitation and streamflow at scales relevant to flash flood occurrence, i.e., hourly to daily timescales (Chapter 1). This analysis, based on a compiled regional database, demonstrates the influence of larger scale climatic forcing on the regional precipitation and streamflow occurrence associated with flash flooding conditions in southern California. Geomorphologic principles defining bankfull streamflow and the hydrologic index for flash flood occurrence are presented in Chapter 2, with region-specific relationships for predicting bankfull conditions with high spatial resolution. High spatial and temporal resolution estimates of precipitation are required in this research due to small scales of flash flood occurrence in nature. Chapter 3 intercompares three alternative methods for estimating orographically driven precipitation in the region based on raingauge data and numerical models, and includes the adaptation and application of a simplified high resolution orographic precipitation model for southern California. The research culminates in Chapter 4 with the characterization of soil moisture climatology (1949-2005), and a multi-model ensemble characterization of the historical regional flash flood occurrence frequency. Frequency is estimated by counting the instances of precipitation exceeding the threshold satisfying bankfull and soil moisture deficit conditions as defined from the earlier geomorphologic and soil moisture modeling.
study. Chapter 4 concludes with an application of the integrative methodology developed to identifying changes in flash flood occurrence frequency under projected climatic change.
CHAPTER 1. Executive Summary: Flash Flood Occurrence in Southern California

1.1 Motivation

Flooding occurs over a wide range of time and space scales – from large river flooding, which may impact areas of several thousands of square kilometers and last several weeks, to local drainage problems in urban areas, impacting very localized areas often within hours of precipitation occurrence. Flash floods are associated with small spatial and temporal scales and this often leaves very short times for warning. This makes flash floods among the deadliest natural disasters. Flash floods caused nearly one-half of the annual flood damage and 80% of the flood-related fatalities in the U.S. over the most recent decade (NWS, 2005). On a global scale, the mortality rate for flash floods, or the number of fatalities relative to the total number of people affected, is disproportionately high when compared to other forms of flooding. Jonkman (2005) reports that this rate is nearly 4% for flash flooding, while for the other types of flooding this ratio is significantly less than 1%, with an average of more than 5000 deaths annually due to flash floods across the globe.

Southern California is not unfamiliar with flash flood occurrence. Examples of recent, widespread flash flooding and other hydrologic hazards occurred in January 2005. A series of storms besieged the region, beginning in late December 2004 and continuing in January and February 2005 (Carpenter et al, 2007). The second storm of this series (7-12 January 2005) produced heavy precipitation from Santa Barbara eastward to the San Bernardino Mountains in San Bernardino County and southward to San Diego County with local rain gauge stations reporting as much as 300% of the
normal (long term average) precipitation for the month of January in this 1-week period. There was widespread flooding, flash flooding, and other hydrologic impacts including the disastrous La Conchita landslide that damaged or destroyed 35 homes and took 10 lives.

Flash floods most frequently occur in small streams and rivers draining areas of a few hundred square kilometers (or less), quickly responding to storm runoff in a few hours (typically 6-12 hours). In mountainous regions, flash flood occurrence is influenced by steep terrain, thin soils, and narrow valleys, which contribute to the rapid production of storm runoff and allow stream channel to rise quickly to significant depths. Orographic enhancement of precipitation in mountain regions also contributes to heavy, localized precipitation. The southern California Mountains have been identified among regions in the U.S. with high topographic relief and an abundant atmospheric moisture source in the Pacific Ocean and yielding high runoff (O’Connor and Costa, 2004). In addition the interaction of low level moist Pacific air flow with the local terrain is an important rainfall generation mechanism in southern California and is an active research area (Small, 1999).

1.2 Drivers of Flash Flood Occurrence

There is a notion that natural flash floods are associated with heavy or extreme precipitation. However, comparing the climatology of heavy precipitation in the U.S. with a prior study on flash flood climatology, Brooks and Stensrud (2000) conclude that flash floods occur 17 times less frequently than heavy precipitation on average. Flash flood occurrence is clearly not driven by heavy or extreme precipitation alone. It has
long been recognized that there are both climatic and physiographic influences on the occurrence of flooding, which include precipitation, topography and soil properties (e.g., Hoyt and Langbein, 1939; Georgakakos, 1986). The land surface plays an important role in modulating the amount of precipitation required to produce flash flooding through soil saturation. The impact is illustrated in Figure 1.1, which presents the relative frequency distributions of watershed-averaged precipitation under estimated flash flood events in southern California over a 55-year historical period for two cases: (a) a single small watershed and (b) a large number of watersheds (833) throughout the region. The single watershed case shows a range of precipitation intensities from 7 to 40 mm/3-hr, while the second case shows a wider range from 5 to 70 mm/3-hr. While very heavy or extreme rainfall (say > 30mm/3-hr for southern California) can lead to flash flooding, the majority of the estimated flash flood events depicted occur under more moderate rainfall. This figure emphasizes that both meteorological and hydrologic conditions must be considered for reliable predictions of flash flood occurrence.

1.3 Scientific Objectives of the Thesis

There is recognition of the importance of flash flood occurrence and associated need for improvement of forecasting capabilities (e.g., AMS, 2000; NRC, 2005). However, there has been relatively little research on flash flood climatology with high spatial detail on national or regional levels. This is largely due to limited long-term historical records of flash flood occurrence within any region with high resolution or with specificity of location. For example, historical records from local weather
Figure 1.1. Relative frequency of basin averaged precipitation during estimated flash flood events in southern California during the period 1950-2005: (a) case for a single small watershed (O[30km²]), and (b) case for a large number of small watersheds.
forecasting offices may indicate a flash flood was reported “in the southeast portion of the county”, rather than including the specific stream name. Often reports are noted only when people are effected and emergency response was required. Although such climatological analyses have been limited, a better understanding of flash flood occurrence as an extreme hydrometeorological event is an important aspect of climate science. The present research contributes to the fields of hydrology and climate science by:

(a) increasing the understanding of the variability in hydrologic processes;

(b) increasing the understanding of hydrologic response to variable climate and climate change; and

(c) following an interdisciplinary approach as deemed necessary in water resource science.

These areas have been noted as important needed developments by the National Research Council (NRC, 1991, 1999, 2001, 2004).

In addition to the lack of long-term records of flash flood occurrence, there is also a lack of spatially distributed records of associated observations, specifically of streamflow and precipitation, at the scales of interest for flash flooding. Such observations can be particularly sparse in mountainous regions where the terrain can make accessibility to higher elevations difficult. This dissertation develops and presents an interdisciplinary approach to characterize flash flood occurrence through detailed atmospheric and land surface modeling. The overarching objective is to examine the spatial character and variability in the occurrence of flash flooding throughout the southern California mountainous region (of the order of tens of thousands of square
kilometers) with high spatial detail (order of a few tens of square kilometers). The modeling approach is presented schematically in Figure 1.2, where the rectangular boxes indicate modeling components for:

- Estimation of orographically driven precipitation with high spatial and temporal resolution (Chapter 4);
- Distributed hydrologic modeling of the land surface (Chapter 5);
- Estimation of small watershed flooding response threshold (Chapter 3).

In this approach, results of the precipitation model are used to drive the distributed hydrologic model (indicated by the arrow). Hydrologic conditions within small watersheds are combined with the geomorphologically defined land-surface response to estimate flash flood response thresholds for each small watershed and at each time step. This flash flood response threshold (FFR) represents the amount of rainfall falling over the catchment with a given duration required to produce conditions of minor flooding in the stream draining the watershed. Flash flood response thresholds are compared with the watershed average precipitation estimates (P) of the same duration, and simulated flash flood occurrence for a given watershed is defined when the precipitation exceeds the FFR.

The approach combines elements from the fields of meteorology, geomorphology, hydrology, and climatology, along with statistical analysis. While large scale atmospheric forcing largely controls smaller scale atmospheric behavior, it is often local or small scale inhomogeneity in the terrestrial properties that substantially
affect the hydrologic response (e.g., Beven, 1989, 2001). Thus, this study links the larger scales of atmospheric forcing to the small scale of hydrologic response for the occurrence of flash flooding. This difference in spatial scales motivates the high-resolution investigation of hydrologic response within the regional focus area. Such a detailed study for flash flood occurrence within any region is lacking in the literature. This dissertation presents the first regional examination of the climatology and spatial variability of flash flood occurrence with high spatial resolution.

1.4 Summary of the Thesis

The chapters of this Thesis are summarized as follows. The dissertation research begins with a diagnostic analysis of observed precipitation and streamflow for
selected stations with long-term records and is presented in Chapter 2. The focus is on
the spatial character of mean precipitation and streamflow, along with the occurrence at
various heavy precipitation and streamflow thresholds. Observational records are
examined for 111 daily precipitation stations, 59 hourly precipitation stations, and 21
streamflow stations for southern California extending from the northern borders of San
Luis Obispo, Kern and San Bernardino Counties to the southern border with Mexico
and spanning the years from 1948 to 2005. The precipitation and streamflow records
show strong seasonal cycles, with highest values generally in January or February and
with low precipitation volumes and flows during the months of June, July and August.
The spatial variation in precipitation suggests orographic influences, with moderate
mean precipitation along the coastal area, heavier mean precipitation in the mountain
regions, and very low precipitation in the eastern region (encompassing the southern
portion of the Mojave Desert). Analysis of streamflow data for the selected stations
indicates higher average annual volume along the Transverse Mountain Range (aligned
east-west from Santa Barbara County to San Bernardino County) and lower average
annual volume along the Peninsular Mountains (align north-south from southwestern
San Bernardino County into Baja Mexico). The occurrence of precipitation and
streamflow is shown to be sensitive to larger scale climatic influences. Mean
precipitation, variability in precipitation and occurrence of precipitation exceeding
various precipitation thresholds increase during wet seasons with higher than normal
total seasonal volume and for climatic indicators of El Nino conditions. Similarly,
mean streamflow and the occurrence of higher streamflow increase during wet seasons
with climatic indicators of El Nino conditions.
Chapter 3 develops the land-surface response for small catchment flooding as indicated in the upper right hand box of Figure 1.2. The chapter presents the development of regional hydraulic geometry relationships that define stream channel cross-sectional geometry characteristics based on the watershed area, and use these relationships to develop the surface response index. The relationships are developed for channel characteristics at the bankfull stage condition. The bankfull stage represents the transition from the active stream channel to the flood plain, and its associated discharge is considered an equilibrium discharge level that maintains the channel dimension in the longer term. Thus, this is a geomorphologically significant level, and has been widely used in hydrologic science and river restoration studies. Regional hydraulic geometry relationships for channel bankfull top width, hydraulic depth, and cross-sectional area were developed using data from stream surveys conducted in 2005 with the author’s participation, together with records available from the U.S. Geological Survey, and are compared with relationships developed from independent surveys and channel geometry estimates. The relationships developed from these different sources are in general agreement.

Use of the developed regional hydraulic geometry relationships is then demonstrated in estimating channel characteristics, along with bankfull discharge under Manning’s steady flow formulation, for small watersheds throughout the southern California mountain region ranging from 13 to 3000 km$^2$ in accumulated drainage area. The small watersheds were delineated based on digital elevation data, with a total of 975 small watersheds covering the mountainous to foothill regions of southern California (excluding major urban areas). Bankfull top width estimates range from 7.2
to 44.5 m; hydraulic depth from 0.35 to 1.15 m; and bankfull discharge from 1.3 to 74 m$^3$/s. Hourly streamflow records for 19 stations are examined to consider the frequency of occurrence of bankfull discharge as estimated using this methodology. The computed average return intervals for these stations ranged from 0.15 to 3.3 years. The final application of the regional hydraulic geometry relationships is to estimate the surface response index. This is a geomorphologically derived estimate of the surface runoff response required to produce bankfull flows at the outlet of the small watersheds. The index is related to the flash flood response threshold by considering precipitation losses (as presented in Chapter 5). The estimates of the surface response index ranges from 2 to 6 mm per 3-hour rainfall duration, with somewhat higher values along the Transverse Mountain Range relative to the Peninsular Mountain Range.

Chapter 4 presents the development of high spatial and temporal resolution precipitation estimates over the historical period from October 1948 through April 2005. Three methods of varying complexity and spatial resolution are used to estimate precipitation with hourly resolution over the southern California region:

(a) an interpolation of hourly observations based on the monthly climatological average spatial distribution on a 1 km$^2$ grid;

(b) a simplified numerical model of topographically-driven precipitation with 3x3 km$^2$ spatial resolution;

(c) output of a mesoscale numerical model of the atmosphere with 10x10 km$^2$ spatial resolution.

The three methods are compared in terms of spatial distribution and magnitude of wet-season and monthly average precipitation, of regional and small watershed (50 to 400
km$^2$) seasonal climatology, and frequency of hourly precipitation at different precipitation rates. The comparison indicates similarities in the spatial distribution and monthly climatology although the magnitude varies among models. The three models each produce higher average precipitation in the Transverse Mountains versus the Peninsular Mountains. The precipitation estimates from the numerical models (b and c) were capable of reproducing the observed sensitivity to larger scale climatic forcing as indicated the climatic indices of El Nino conditions as demonstrated for observations in Chapter 2. The combination of these different methods forms a multi-model ensemble of long-term, high resolution precipitation estimate for the region that may be used in further hydrologic analysis of flash flood occurrence and other hydroclimatological studies.

The results of the research synthesis are presented in Chapter 5, which brings together the modeling approach components of Chapters 3, 4 and 5 to characterize the flash flood occurrence frequency for southern California. This chapter includes distributed modeling the land surface to simulate soil moisture conditions using the precipitation forcing developed in Chapter 4, the integration of the surface response index of Chapter 3 with simulated soil moisture conditions to estimate flash flood response threshold (central oval of Figure 1.2), and the simulation of flash flood occurrence at 950 small watersheds throughout the region. Two hydrological models with different parameterizations are employed to simulated soil moisture evolution over the historical period (October 1948 – April 2005). These models provide insight to seasonal climatology, spatial distribution, and variability of soil moisture throughout the region. While the average soil moisture is relatively low with peak monthly soil
moisture saturation fraction < 0.5 in the upper soil layer and < 0.8 in the lower soil layer, simulated flash flood events are associated with a high degree of soil saturation in both soil layers. The multi-model ensemble approach towards flash flood occurrence frequency is employed using two precipitation estimates from Chapter 4 and two hydrologic models. The wet season flash flood occurrence frequency is computed in terms of the average number of events per year and ranges from < 0.5 events per year to nearly 4 events per year during the historical simulation.

Finally, the utility of the interdisciplinary modeling approach is demonstrated with an application that examines changes in flash flood occurrence frequency under potential climate change. The simplified orographic precipitation model of Chapter 4 is used to produce precipitation estimates based on state-of-the-science climate model output for the control climate (1970-1999) and for a moderate-emissions-scenario future climate (2070-2099). Using these precipitation estimates with the hydrologic models and flash flood occurrence simulation components, the flash flood occurrence frequency between the future and control climates is estimated. Both hydrologic models indicate increases in the average frequency of flash flood events under the projected future climate.

The research synthesis presented in this Thesis opens new areas of fruitful diagnostic and predictive studies:

(1) use the historical simulation results and historical observations to identify atmospheric forcing and other vulnerability factors associated with flash flood occurrence in southern California;
(2) examine the role of uncertainty in quantifying flash flood occurrence frequency under this modeling approach by considering the impact of model parameterization and input errors;

(3) extend the applications of the modeling approach to consider changes in flash flood occurrence frequencies due to (a) land cover changes and (b) climatic changes using other climate models and different greenhouse gas emission scenarios; and

(4) explore the utility of this methodology for real-time, forecasting applications.

1.5 References


CHAPTER 2. Characteristics of Small-Scale Heavy Precipitation and Streamflow in Southern California

Abstract

Characterization of spatial and temporal variability of observed precipitation and streamflow is an important first step toward understanding local and regional hydroclimatology and serves as a foundation for hydrologic modeling studies. This chapter focuses on the hydroclimatology of precipitation and streamflow for southern California, with particular emphasis on scales that are relevant for flash flooding. The paper presents a three-tiered analysis of precipitation and streamflow observations addressing: (a) climatological variability and spatial distribution of high daily streamflow and heavy daily and hourly precipitation; (b) association of the occurrence of such precipitation and streamflow events with larger scale climate oscillations as indicated by extreme seasonal volume and climate indices using the Nino 3.4 index; and (c) association of the synoptic-scale meteorological states with widespread occurrence of such events throughout the region. With widespread and no detailed observational records on the occurrence of flash flooding, thresholds of high precipitation and streamflow are employed to define such events. “Heavy” precipitation is defined by the upper 25-, 10-, and 5% exceedance thresholds at individual observation stations (both hourly and daily), and streamflow thresholds defined by flows of specified return periods are used as surrogate indicators of the initiation of overbank flooding conditions. The recurrence interval flow thresholds considered were the 1.5-year, 2-year, and 5-year return periods. Climatological analysis shows distinct wet versus dry
seasons in southern California precipitation and streamflow, particularly for the cismontane region of the Southern California Bight, with up to 90% of the precipitation/streamflow volume occurring between October and April (inclusive). Distributions of wet season daily precipitation and streamflow at selected long-term stations are shown to be statistically different during the early thirty year period compared to the later thirty year period, with a tendency toward wetter conditions. Meteorologic conditions for anomalously high precipitation reflect differences in composite low and high pressure systems and consequent forcing wind direction and magnitude depending on whether heavy precipitation occurs in the Peninsular versus Transverse Mountain ranges, desert region, or northern coastal regions. Although relatively few observation stations exist at high elevation, the influence of orography on precipitation amount and frequency of occurrence is identified. Both the volume of precipitation and occurrence frequency at various precipitation thresholds are enhanced during wet seasons that are “wetter than average” compared to occurrence during seasons that are “drier than average”. Similarly, precipitation and streamflow volume and frequencies of exceeding specified thresholds are greater during wet seasons when early season climatic indices indicate warming in the equatorial Pacific Ocean associated with El Nino conditions (e.g., high Nino 3.4 index) and conversely during wet season associated with equatorial Pacific Ocean cooling (La Nina conditions).

2.1 Introduction

Flash floods, defined as flood events that occur shortly (generally within 6 hours) after the causative rainfall event, can occur anywhere and with little advanced
warning. Short response times and associated small spatial scales contribute to the
distinction of flash floods as being the deadliest among weather-related events in the
long term average (NWS, 2005). In southern California, the close proximity of the
coastal mountain ranges to the Pacific Ocean promotes the enhancement of precipitation
and the rapid development and concentration of runoff, which may lead to flash
flooding. The flash flood prone areas of southern California are also among the most
heavily populated in the U.S. and have the potential for sustaining significant impacts.

Flash flooding is often tied to intense local precipitation and hydrologic
conditions that promote the generation of significant surface and subsurface flow
exceeding the conveyance of stream channels. Mountainous regions, with steep slopes
and thin soils, allow for the rapid production of runoff and stream flow. Narrow valleys
can further concentrate stream flow, promoting quick rise of water levels to form
significant flow depths. With adequate atmospheric moisture, the high topography of
these regions may also cause orographic enhancement of precipitation leading to
localized heavy rainfall. The Transverse Ranges of southern California have been
identified as having these features among other flash-flood prone regions (O’Connor
and Costa, 2004). Flash flood events in southern California may also be associated with
additional hazards of mud flows or debris flows (USGS, 2005a,b), particularly in areas
devoid of vegetation due to wildfires. A recent example occurred on Christmas Day,
December 25, 2003, along Waterman Canyon in the San Bernardino Mountains, north
of the city of San Bernardino. In a region that typically receives a total of 150-200 mm
of precipitation on average during the month of December, a nearby observation station
at Lytle Creek indicated daily precipitation that exceeded 200 mm. Analysis of the
event suggests sustained precipitation rates of 20 mm/hr with a peak rate of 36 mm/hr (Small and Dandrea, 2004). The resulting flash flood/debris flow killed 14 people in a local campground and destroyed two bridges (San Bernardino County Flood Control District, 2007).

As a first step toward understanding of conditions under which flash floods may occur in southern California, this chapter examines the temporal and spatial character of observed precipitation and streamflow at scales relevant to flash flooding. For this, we utilize available observations of rainfall with both hourly and daily resolution, and daily streamflow records. The analysis of observations focuses on:

(a) The climatological features of precipitation and streamflow in the region, including the occurrence and spatial variability of heavy precipitation and flooding stream flows;

(b) the description of synoptic atmospheric conditions during and prior to the widespread occurrence of heavy precipitation throughout the study region.

(c) the associations of these occurrences with climatic conditions that include wetter- or drier- than normal wet seasons and extreme indices of the El Nino-Southern Oscillation and Pacific Decadal Oscillation;

Generally, widespread and detailed historical records of flash flood occurrence upon which to build a detailed regional climatological analysis of hydrometeorologic
conditions associated with flash flooding are lacking. Surrogate information is necessary, and precipitation has been used (Brooks and Stensrud, 2000; Maddox et al, 1980; Maddox et al, 1979). In this study, focus on scales of flash flooding is achieved by defining high precipitation and streamflow thresholds above which flash flooding is probable. Although typically associated with intense local precipitation, flash flooding can occur under moderate to heavy rainfall depending on hydrologic conditions within the watershed such as the state of soil saturation (e.g., Georgakakos, 1986). For the purposes of this study, several precipitation thresholds are considered with the thresholds defined from the exceedance probability distribution of individual stations records. The upper 25\(^{th}\), 10\(^{th}\), and 5\(^{th}\)-percentile exceedance levels are used. Similarly, streamflow thresholds were estimated based on the observed records at individual stream gauging stations. As a measure of flood occurrence, the initiation of flooding is associated with the times when the stream overflows its banks and flows on to the flood plain. This level is considered to be the bankfull level or bankfull discharge (Leopold, 1994). Based on a variety of data and methods across different hydroclimatic regions of the U.S. and abroad, bankfull discharge has been estimated to have a return period that spans the range from less than 1 year to more than 5 years, with a typical value of 1.5 years (Leopold, 1994; Leopold et al, 1964; Woodyer, 1968; Williams, 1978; Petit and Pauquet, 1997; Castro and Jackson, 2001). This means that streamflow may be expected to exceed bankfull discharge on average once every 1.5 years at a given location. In this study, flow thresholds associated with the 1.5-, 2-, and 5-year return periods are considered.
With these precipitation and streamflow thresholds defined, this chapter is concerned with the description of the spatial variability of these thresholds and their occurrence frequencies relative to larger scale synoptic and climate variability. This differs from other climatological studies (Guirguis and Avissar, 2008; Leung et al, 2003) in that it focuses on the southern California region, and on events pertinent to flash flooding occurrence. It also differs from other studies in southern California (Nezlin and Stein, 2005; Hope et al, 2008) in scale and for longer length of the historical records considered. This study serves as a foundation for hydrometeorologic modeling aimed at characterizing and predicting the spatial and temporal variability of the frequency of flash flood occurrence throughout the region.

The next section offers a description of the study region and the data utilized, together with an overview of relevant prior investigations. Section 2.3 presents the results of the analyses undertaken in four major subsections: (a) climatological description of southern California precipitation and streamflow, including synoptic states during days with anomalously high precipitation; (b) definitions of high streamflow and precipitation thresholds; (c) description of the occurrence frequencies of the selected thresholds; and (d) association of event occurrence with climatic indices. Conclusions and recommendations are presented in Section 2.4.

2.2 Southern California Study Region

2.2.1 General description

The study region is southern California, extending from the San Luis Obispo, Kern and San Bernardino counties south to the U.S.-Mexico border (Figure 2.1). A
notable feature of the region is the proximity of the local mountain ranges to the Pacific Ocean. This includes the Transverse Mountain Range which runs roughly east-west (northwest to southeast) from the western coast of Santa Barbara county, consisting of the Santa Ynez, San Rafael, and San Gabriel Mountains, and extending to the Mojave Desert at the termination of the San Bernardino Mountains in southwestern San Bernardino County. The Tehachapi Mountains extend northward from the Transverse range in Kern County and connect with the southern Sierra Nevada Mountains. The Peninsular Mountain Range extends southward from the San Bernardino Mountains into Baja California, Mexico, and includes the San Jacinto Mountains in Riverside County and Laguna Mountains in San Diego County. The highest mountain reaches 3500m at San Gorgonio peak, followed by Mount San Antonio at 3000m. The interaction of onshore flows of moist Pacific air masses with the topography, particularly when the air is driven by westerly to southwesterly cross-barrier flows, frequently generates significant rainfall events with high potential for triggering flash floods.

The southern California region has a Mediterranean climate, generally typified by relatively dry summer seasons and moderately wet winters. Figure 2.2 shows the monthly climatology of precipitation for select locations (indicated in Figure 2.1). The selected stations are grouped to represent coastal locations (first row), mountainous locations (middle row), and eastern/desert locations (bottom row). Additionally, the plots are arranged from the northernmost to the southernmost location within each row. Generally, there is a clear annual cycle with monthly precipitation volume increasing from October to a peak monthly volume in January-February, followed by decreasing volume through May and with little precipitation in June, July and August for the
coastal and mountain locations. In the eastern locations (bottom row), there may be a secondary peak in monthly precipitation volume in August-September. This secondary peak may derive from the North American monsoon, although this is not investigated herein. The monthly climatologies also hint at the influence of orography on the regional variation of precipitation, with peak average monthly volumes reaching 100 mm along the coastal locations, approaching 250 mm at the mountain locations, and only reaching 10-30mm for the eastern/desert locations in the lee of the mountains.

2.2.2 Recent studies on the variability of California precipitation and streamflow

Numerous studies have focused on western U.S. precipitation and streamflow (Guirguis and Avissar, 2008; Leung et al, 2003; Dettinger et al, 1998; Chen et al, 1996; Hamlet et al, 2007; Lundquist and Cayan, 2002), their variation with climate oscillations (Andrews et al, 2004; Cayan et al, 1999; Jones, 2000; Mo and Higgins, 1998; Cayan and Peterson, 1989), and potential changes with climate change (Dettinger et al 2004; Miller et al, 2003). For California, precipitation tends to be below normal during La Niña events and above normal for El Niño events (Schonher and Nicholson, 1989; Cayan and Redmond, 1994), with further modulation of wet/dry events by the tropical interseasonal oscillation (Mo and Higgins, 1998). With respect to streamflow, Andrews et al (2004) examined the relationship between ENSO and flooding for California coastal streams. They defined flooding in terms of the annual peak flows without regard to occurrence of flood damage, and found strong correlations between this streamflow definition and ENSO phase for 38 coastal streams extending from south of the San Francisco Bay area to southern California. Cayan and Peterson (1989) also
found significant correlations between monthly streamflow and ENSO phase throughout the western United States. Focusing on ‘extremes’ in both precipitation and streamflow again over the western U.S., defined in terms of the median and the 90th percentile of daily values, Cayan et al (1999) found an increased frequency of high precipitation and streamflow for El Nino events in the southwestern U.S. That study suggested an amplification of the high streamflow response to ENSO as high flows are as much as ten times as likely during El Nino versus La Nina years for several basins in the southwestern U.S.

Albeit within the regions of interest for the cited studies, the southern California region was not the explicit focus of these studies. Generally, the focus has been over larger regions, ranging from the State of California to the western United States, and over variables not necessarily associated with flash flooding (e.g., mean daily or monthly variables). Relatively few studies have attempted to characterize variability for the smaller- and shorter-scales that are pertinent to flash flooding conditions, particularly for southern California. Recently, Nezlin and Stein (2005) examined observed daily precipitation from 98 stations within the southern California region. However, their focus was on an intercomparison remotely-sensed precipitation with 1-degree resolution with the observed precipitation. In their study, data from the Global Precipitation Climatology Project (GPCP) during the period 1996-2003 were averaged over southern California watersheds ranging from 300 to 5200 km² in size and compared with commensurate observations. They found low correlation and a low bias in the GCPC mean values, but with higher correlation in the precipitation variability at these scales. Conil and Hall (2006) examined atmospheric variability for the southern
California region through mesoscale numerical modeling. Although their modeling was at high spatial (6 km) and temporal (hourly) resolution, they focus on variability in winds and the relationship of other atmospheric variables (e.g., temperature, precipitation) to the winds. Their analysis is limited to the simulation period of 1995-2003.

The present study expands previous work by focusing on the scales relevant to flash flooding (e.g., hourly to daily rainfall, streamflow exceeding the annual mean), and by focusing in detail and over longer-term records on the southern California region specifically.

2.2.3 Data

Precipitation data used for this study includes both daily and hourly records from the U.S. National Climatic Data Center (NCDC) databases (EarthInfo, Inc., 2005a,b). The criteria for selection of precipitation stations included: (a) at least 50 years of record between 1948 and 2005, beginning not later than 1955 and ending not earlier than 2000; and (b) having no more than 20% missing data coverage. It is not uncommon for stations to be relocated and for two (or more) closely located stations to cover the period of record at a location. However, such stations were not included if the individual stations did not meet the minimum selection criteria stated above. The station records from these locations were not combined to be included in this analysis. A total of 111 precipitation stations with daily resolution, and 59 stations with hourly resolution met these criteria and were used. The station locations are indicated in Figure 2.3 by the plus (+) symbol for daily records and open circle (o) symbol for
hourly records. A concentration of stations is present within the greater Los Angeles region, with fewer stations located to the south and northwest, and sparsely located stations in northeastern San Bernardino, Riverside and Imperial Counties. As commonly noted in other mountainous regions, the station coverage is biased towards lower elevations. For southern California, less than 10% of the stations are located at elevations greater than 1000m with the highest station used located at 1750m. Peak elevation in the terrain data reaches 3500m.

Given the relatively steep terrain in southern California along the oceanic side of the coastal mountains, the streamflow response tends to be rather fast (within a day). To truly examine the hydrologic character of flash flooding, high temporal resolution information at relatively small spatial scales is needed; however, high temporal resolution (e.g., hourly) streamflow data is not readily available for the long historical periods necessary to examine climatological features. The approach taken here is to utilize records available from the U.S. Geological Survey (USGS) National Water Information System (NWIS), with focus toward relatively small and natural watersheds. Natural watersheds are defined as those having minimal impacts from major flow regulation. The available records include annual peak flow series, which are used to derive return-period flows as a surrogate indicator for conditions when small-scale flooding may be initiated, and the mean daily flow series, which are utilized to characterize the occurrence of flows exceeding defined flow thresholds.

The selection criteria for streamflow data were: (a) available mean daily flow records for the period 1950 to 2008; (b) annual peak flow series covering at least 15 years; (c) stations without streamflow regulation or where total regulation was small
and/or impacted a small portion of the watershed. The USGS gauge descriptions were used to assess the presence and extent of regulation, given that small-storage and municipal regulation exists throughout southern California. No regulation was specified in the descriptions for 12 selected locations. An additional 9 gauge locations were selected with no comment regarding regulation in the USGS descriptions. For these locations, without additional information to the contrary, no upstream regulation of the flows at the station was assumed. A total of 21 stations were selected based on the availability of annual peak flow records and are indicated in Figure 2.3 by the filled-triangle (Δ) symbols.

2.3 Characterization of Southern California Precipitation and Streamflow

The analysis begins with an overall description of the annual and monthly climatology of precipitation and streamflow for the region, along with a characterization of the synoptic patterns associated with heavy precipitation and an examination of changes in the distribution of precipitation at stations with long-term records. The analysis then moves into the occurrence of events of various magnitudes relevant to flash flooding and their association with climatic scale forcing.

2.3.1 Climatological, synoptic, and long-term features

2.3.1.a Climatology of southern California precipitation and streamflow

Figure 2.4 presents the average annual precipitation (top panel) and average annual number of days with precipitation (bottom panel). The first striking feature is the distinction between the coastal-mountain region and the desert region. The annual
precipitation for coastal regions is in the range of 200-400mm, increasing along the foothills of the Transverse Mountain Range to 600mm, and reaching a peak average annual precipitation of 1000mm at the Lake Arrowhead station. The desert stations generally show annual amounts in the range of 60-200mm. The annual number of days with precipitation for the desert stations is also distinctly lower than for the coastal and mountain regions. Desert stations average from 9 to 20 days per year with precipitation, while the western stations range from 20 to 50 days per year. Stations along the southern coastal region have somewhat lower total precipitation than stations within the Los Angeles, Ventura, and Santa Barbara coastal region. This north-to-south gradient in precipitation was also noted by Nezlin and Stein (2005). On average, many of the low elevation stations in the Los Angeles region (< ~150m) have fewer days of precipitation (20-25 days) than stations at similar elevation along the southern coast (25-30 days) or northern coast in San Luis Obispo County (30-40 days). The few stations on the leeside of the mountains suggest a rapid decrease in the average number of days, from > 40 days at the mountain stations to < 25 days on the leeside.

Considering two coastal-to-mountain-to-desert transects of the elevation and annual precipitation (a) across San Diego County and (b) from Los Angeles to San Bernardino County, the effects of orography on precipitation are readily discerned. Albeit with variability, there is an increase in average precipitation with elevation suggestive of orographic enhancement. Due to the regional variability in annual precipitation and few stations at high elevation, the variation of precipitation with elevation is obscured when all stations in the study region are plotted together and without consideration of dominant atmospheric flow paths. Likewise, the topographic
Effect on average precipitation is suggested in Figure 2.4 for the region that is an
extension of the California Central Valley, between the mountains along the San Luis
Obispo-Kern County line and the southern Sierra Nevada mountains in Kern County.
The low elevation stations in this region receive low annual precipitation (100-200mm)
and few days of precipitation (20-30 days). Values quickly increase for the bordering
stations in the mountain regions with average precipitation in the range of 200-400mm
and days with precipitation between 30-50 days.

The average annual precipitation (mm), average number of hours with
precipitation, and average precipitation rate (mm/hr) were also computed using the
hourly station data (Figure 2.5). Annual precipitation computed from the hourly
stations is commensurate with that computed from the daily data. The peak annual
amount reaches 750 mm in the San Bernardino Mountains, with values of 200-400 mm
in the coastal regions, and with values as low as 45 mm in the desert. A comparison of
the annual precipitation computed at locations with both hourly and daily records shows
a regional under-representation of the annual amount at hourly locations, largely due to
missing hourly records. Figures 2.4 and 2.5 highlight the importance of the station
distribution in studies of inference pertaining to precipitation variability and in
validation studies that use observed precipitation data as ground truth.

The spatial pattern with lower annual precipitation for eastern/desert stations and
higher values at the coastal and mountain stations is similar to that shown in Figure
2.4a, albeit with fewer stations. The annual average number of hours with precipitation,
shown in Figure 2.5b, ranged from 15 hours at El Centro in Imperial County to nearly
180 hours at Running Springs and Camp Angelus in the San Bernardino Mountains.
The spatial pattern of the number of hours with precipitation is similar to that shown in Figure 2.4b for daily data. Figure 2.5c summarizes average hourly precipitation (mm/hr), conditional on precipitation occurring (e.g., it does not consider hours of zero precipitation). This indicates a more uniform appearance in the spatial variation in average precipitation rate, ranging from 1 to 4 mm/hr.

The annual cycles of precipitation given for several stations in Figure 2.2 exemplify the results throughout the region in terms of seasonal and regional variation. The monthly variations of the number of days and number of hours with precipitation follow that of total precipitation: higher numbers during the late Fall through Spring season (e.g., October through April) and low numbers during the Summer season (June, July, August) for coastal and mountain locations. The number of days with precipitation per month is high from January through March with 5-6 days per month for most coastal stations, and between 5-7 days per month for mountain stations. The maximum average for eastern stations is only 2 days per month, occurring either in January or August.

The diurnal variation of precipitation was also examined at the hourly stations. Figure 2.6 shows the average precipitation during each hour (mm/hour) (left column) and the relative frequency of precipitation occurrence during each hour (right column) computed over the period of record for the stations at Santa Barbara, Camp Angelus, and Hayfield. These stations are located in Santa Barbara, San Bernardino (near Lake Arrowhead), and Riverside Counties, respectively, and were selected represent stations along the coastal, mountainous, and eastern regions. The mean precipitation amount (mm/hr) across all hours of the day is shown in the plots by the horizontal red line.
Although with variability in hour-to-hour occurrence of precipitation, the diurnal variability of precipitation generally did not show strong signal for preferential hours in precipitation amount or for precipitation occurrence. The diurnal plots show relatively uniform amount of precipitation across all hours of the day while the fraction of occurrence during each hour showed more variability from station to station.

Figure 2.7 shows the spatial distribution and monthly cycle of the streamflow climatology. The top panel shows the spatial distribution of annual mean flow volume \((Q_a)\) computed from the mean daily flow records for selected stream gauging stations. The flow volume has been normalized by drainage area and is expressed in mm. The goal of the normalization is to remove the effect of drainage basin size on the computed streamflow volume. The figure shows lower streamflow volumes for the southern stations in Riverside and San Diego counties, with two of these stations located on eastward draining streams (i.e., on the leeside of the Peninsular Range). Highest runoff volumes occur along the Transverse Mountains. Although with far fewer observation points, the general pattern of higher values in the mountain region of San Bernardino and Los Angeles Counties reflects of the precipitation climatologies of Figure 2.4. The bottom panel shows the monthly climatology of mean daily flow during each month for 12 selected stations, identified by their USGS station number. In the figure, the stations are generally arranged by row for northern (top row), Transverse and Peninsular Ranges (middle two rows), and desert (bottom row) locations. A strong seasonal cycle is observed, with peak flow occurring in February or March, and generally very low flows during the summer months (July – September).
Many of the streams in southern California are known to be ephemeral. A few stations, such as 11058500 in Figure 2.7b, indicate low but sustained flows during the summer. This may be indicative of impacts of regulation or diversion, although the record descriptions stated no regulation exists. The two stations on the lee side of the Peninsular Mountain range, 10258500 and 10255810, indicate a small, secondary peak during August, consistent with the precipitation observations. Station 11189500, located in northern Kern County and generally far away from the other stations, has its peak monthly flow occurring in May. The Transverse Range exhibits the highest area-normalized streamflow volumes.

As noted for both the coastal to mountain precipitation and streamflow stations, most precipitation and flow is observed during the wet season (October through April). In fact, this period accounts for 90% or more of the annual flow volume. For the subsequent analysis, focus is only on the wet season, rather than the entire annual period.

2.3.1.b Composite synoptic forcing during periods with heavy precipitation

Large scale synoptic forcings during periods with heavy precipitation are examined next, including variations with the location of heavy precipitation. Focus is on wet season precipitation only given the relative scarcity of stream gauging stations. For this analysis, a total of six index stations were selected for each of four sub-regions within the southern California domain (see Figure 2.8a): (a) northern coastal region, (b) Transverse Mountain Range, (c) Peninsular Mountain Range, and (d) the eastern desert. It is expected a priori that any signal in the desert region will be weakened with respect
to the other regions as the storm system causing heavy precipitation will undergo moisture depletion prior to reaching the desert while it passes over the coastal to mountain regions.

For each of the selected stations, standardized anomalies were computed for daily precipitation exceeding 1mm/day as:

$$a_i = \frac{(p_i - \bar{p}_{mo})}{\sigma_{mo}}$$

where $p_i$ is the precipitation on day $i$, and $\bar{p}_{mo}$ and $\sigma_{mo}$ are the mean precipitation and standard deviation of precipitation for month $mo$, (for $p > 1$ mm/day). Daily anomalies were computed when at least one station recorded precipitation ($p > 1$ mm/day) in the sub-region, and these were then averaged to give a regional anomaly. The regional anomalies were ranked from high to low values, and the dates with the 10 highest anomalies during the wet season were identified for each sub-region. Table 2.1 provides this listing of the dates with the 10 highest ranking precipitation anomalies. If consecutive days with high anomalies occurred, the date with the highest regional anomaly was included. Such multi-day high anomalies occurred for the Peninsular and Transverse Mountain regions and are indicated in Table 2.1. Also, there was some overlap in the dates with highest anomalies among the regions, notably 3 dates in common for the Peninsular and Transverse Mountains, and 2 dates in common for the Transverse Mountains and northern coastal regions. The common dates were included for the analysis of each region.
To elucidate characteristic synoptic scale features, large scale plots of meteorological variables were constructed using the NCAR Reanalysis I dataset (Kalnay et al, 1996). Composite anomalies of select meteorological variables were plotted using the Earth System Research Laboratory website (ESRL, 2010). These meteorological variable anomalies were computed based on deviation from the 1968-1996 climatology, and composite anomalies were plotted based on the dates of heavy precipitation identified in Table 2.1. Composite meteorological anomalies plots were generated for the identified dates and, separately, for 1- and 2-days prior to the peak date. The selected variables examined were: sea level pressure (SLP); air temperature at 925-hPa; geopotential height (GH) at 925-, 850-, and 500-hPa; wind speed and direction at 850- and 700-hPa; vertical velocity ($w$) at 850-hPa; and specific humidity at 925- and 850-hPa. Figures 2.9 through 2.12 present the composite plots of the anomalies of (a) 925-hPa GH, (b) 500-hPa GH, (c) 850-hPa winds, and (d) 850-hPa specific humidity for heavy precipitation dates in each of the four regions respectively.

Common features among the composite maps are low pressure along the Pacific Northwest of the United States, with high pressure west of this and centered over the southern Alaskan coast or northern Pacific Ocean. This pattern is accompanied by anomalously high winds in the Southern California Bight region, directing off-coast high humidity to the southern California region via southwesterly to westerly air flow. Although not presented herein, composite temperature anomaly maps did not show a deviation from climatology in the immediate southern California region, but a positive temperature anomaly to the east over the four corners region of the U.S., and a negative
anomaly to the north over western Canada, for the identified dates of heavy precipitation in the southern California mountain regions.

There are differences among the composite anomaly maps based on the region of heavy precipitation. Considering first the composite anomaly maps for heavy precipitation along the Peninsular Mountains (Figure 2.11), the surface low pressure (e.g., 925 hPa GH) on the day of heavy precipitation is centered near the Oregon coast, with the high negative anomaly (-80m) extending eastward toward the state of Utah. This low pressure anomaly also extends westward over the Pacific Ocean and is of the order of -60m to a longitude of 160W at 30N. Similarly, the high level low pressure (e.g., 500 hPa GH) is centered off the Pacific Northwest coast and extends towards the central Pacific Ocean. The surface high pressure anomaly behind the low is centered over south-central Alaska. By contrast, for heavy precipitation dates along the Transverse Mountains (Figure 2.10), the composite anomaly maps show the surface low pressure is centered over with Pacific Northwest, however without the extension toward the central Pacific Ocean. The surface high pressure anomaly is shifted in a southwest direction relative to this feature for Peninsular Mountain heavy precipitation, and is located more towards the Aleutian Islands. The composite surface high pressure anomaly on the date of heavy precipitation is also deeper (~100m) for the Transverse Mountain case relative to the Peninsular Mountain cases (~65m). The composite high pressure anomaly over eastern North America (ahead of the low pressure) is deeper and more extensive for Transverse Mountain case, particularly at the 500 hPa level. The composite pressure anomaly maps for heavy precipitation over the northern coastal region (Figure 2.9) are similar to the Transverse Mountain case, although with a
southwest-ward shift in the center of the low pressure off the coast of northern California and southward shift in the extend of the high pressure over the Aleutian Islands.

These differences in the surface and upper level pressure composite anomalies drive differences in the forcing winds. Composite wind anomaly maps for heavy precipitation along the Peninsular Mountains (e.g., Figure 2.11c) show westerly-to-southerly winds near the southern California coastline, with wind speed anomaly magnitudes of 12 m/s at 850 hPa and 14 m/s at 700 hPa. For heavy precipitation dates along the Transverse Range (e.g., Figure 2.10c), the composite wind anomaly maps show more southwesterly to southerly winds at the southern California coast, with higher wind speed anomalies (>14 m/s at 850 hPa and 18 m/s at 700 hPa). Again, the composite wind anomaly maps for heavy precipitation along the northern coastal region are similar to those for the Transverse Mountains case; however the extent of high wind speed anomalies is more spatially extensive and, in particular at the 700 hPa level, cover the region from the southern border to northern California near the San Francisco Bay and extends inland into the State of Nevada. For heavy precipitation dates in these three regions, the composite maps show a slow progression of high specific humidity in a northeastward direction towards the southern California coast over the period from 2-days prior to the day of heavy precipitation. The high specific humidity is driven into the southern California region by the westerly to southwesterly wind anomalies. This high specific humidity anomaly appears to draw upon the equatorial Pacific region for its source of moisture (e.g., Figure 2.10d).
As mentioned earlier, the desert region was expected to show a somewhat weaker signal and thus the discussion of this region was reserved to the end. Indeed, the average regional precipitation anomalies for the desert which defined heavy precipitation dates were lower, and nearly one-half the magnitude of the regional precipitation anomalies for the other regions. The composite surface and upper level low pressure anomalies (e.g., Figure 2.12a,b) are similar to those shown for the other regions, with a southward shift in location and centered off the coast of California. The high pressure anomaly behind this low is centered over the western Canadian coast at 2-days prior to the day of heavy precipitation, and weakens substantially until there is only a small region of weak high pressure on the day of heavy precipitation in the composite maps. The location of the maximum composite wind anomaly for the desert case is shifted southward, impinges along the northern Baja California coast at 850 hPa, and extends into the State of Arizona at 700 hPa. The composite anomaly map for specific humidity is similar to other regions, with a high anomaly region over southern California/northern Baja California for the composite dates leading up to heavy precipitation.

2.3.1.c Long-term variation in daily precipitation distribution

Studies on the impact of global climate change have suggested regional changes in the distribution of precipitation towards increases in precipitation, particularly in the occurrence of extreme precipitation (e.g., IPCC, 2007). Groisman et al (2001) present an analysis of 20th century trends in precipitation and streamflow across the U.S. considering station observations over a maximum period of 1900-1998. Their results
indicate an increase in annual precipitation for southern California coastal area and a
decrease in the southern California desert region. The majority of the southern
California stations used in their analysis had records from 1961-1998. It appears that
only one station in Santa Barbara County and two stations in San Luis Obispo County
had data extending into the earlier analysis period in their study. In the present analysis,
changes in the frequency distribution of daily precipitation for the study region are
considered using the available station observations. A total of 13 daily precipitation
stations were identified with long-term continuous records between 1931 and 2005. An
additional 4 stations were identified with missing years either at the beginning or end of
this period, but covering the period 1935-2003 inclusively. The locations of the long-
term stations are shown in Figure 2.8b. For all long term stations, the period of record
was divided into two thirty-year periods, covering wet seasons for the earlier period of
record from 10/1931 to 4/1961 (or from 10/1935 to 4/1965 for the additional stations)
and the latter period of record from 10/1975 to 4/2005 (10/1973 to 4/2003 for the
additional stations). These were compared to test for evidence of changes in the
distributions of daily precipitation.

Figure 2.13 presents the cumulative frequency distributions of daily
precipitation greater than 1 mm/day for the two periods for 3 selected stations: Santa
Barbara, Pomona, and San Diego. The locations of the Santa Barbara and San Diego
stations are indicated in Figure 2.1; the Pomona station is located within the foothill
region of the San Gabriel Mountains near the Los Angeles-San Bernardino County line.
The graphs show the two frequency distribution curves (solid lines) with their
respective 95% confidence bounds (dashed lines) for daily precipitation exceeding
1mm/day. These three examples represent the three possible cases: (a) there is almost no difference between the frequency distributions of the two periods (Santa Barbara); (b) there is some difference in the distributions, but with significant overlap in the confidence bounds (San Diego); and (c) there is an overall shift in the frequency distribution between the two periods (Pomona). A Kolmogorov-Smirnov (K-S) test of the distributions was performed to determine whether the observed frequency distributions from the two periods are drawn from the same parent distribution and are thus statistically identical (e.g., Wilks, 2006). The K-S test is non-parametric and makes no assumption for the underlying distribution of the data. The test statistic is formed based on the maximum distance between the empirical CDFs and number of data points. The K-S tests indicated that only 3 of the 17 total long-term stations showed statistically similar distributions during the two periods at the 95% confidence level (i.e., one could not reject null-hypothesis that the two sample periods were drawn from the same distribution). The stations with similar distributions were Brawley, Blythe (eastern/desert region) and San Bernardino (southwestern San Bernardino County). The remaining 14 long-term stations had statistically different distributions during the early versus later periods. Most locations indicated a shift in the distribution with higher precipitation for a given frequency level, or thus towards a wetter distribution for the later period (1975-2005). However, clear exceptions exist, with drier distributions during the later period for the records at Palmdale and Paso Robles.

These long term changes in precipitation distributions include changes in extreme precipitation. A similar analysis was conducted for daily precipitation exceeding the upper 25%-exceedance threshold and upper 10%-exceedance threshold at
each station. These thresholds were estimated from the observed data over the entire
period of record at each station (period of record equal to 1931-2005 or 1935-2003
depending on the station). For extreme precipitation, the number of daily precipitation
values for these cases is greatly reduced and the confidence bounds about the
cumulative frequency curve are broader. K-S tests on the distributions of extreme
precipitation for the two periods indicate that observed distributions at Brawley,
Palmdale, Riverside, and Tustin are statistically similar for the upper 25%-exceedance
level. At the upper 10%-exceedance level with relatively few data points, the
uncertainty bounds of the two-periods were wide and largely overlapping.

The analysis was repeated for daily streamflow. Five gauging stations had
records dating back to October 1931 and therefore covering the period 1931-2008.
Differences in the distribution of daily wet season streamflow were considered between
the period 10/1931-4/1961 and the period 10/1975-4/2005 (commensurate with the
precipitation analysis). The five stations, shown in Figure 2.8b by the filled triangle
symbol, are all located in the Transverse Mountain Range. An additional seven stream
gauging stations had records dating back to October 1941. These locations are
indicated in Figure 2.8b by the open triangle symbols and span the region. These
stations were used to compare the periods 10/1941-4/1971 versus 10/1978-4/2008.
Because streamflow is temporally correlated (see section 2.3.3 for further discussion
specifically for the Southern California data), a subset of the data was used by selecting
one record every seven days beginning October 1st of each year. Figure 2.14 presents
examples of the cumulative frequency distributions of streamflow exceeding the long-
term average daily flow during the two periods for USGS gauging stations 11058500,
11113500, and 1114500. The threshold is imposed to omit periods of baseflow or dry conditions. A second higher flow threshold was also applied, taken as the upper 10% exceedance flow of the one-in-seven-day daily flows. K-S tests indicate that the distributions are different for all stations for the early-period versus late-period record and for streamflow exceeding both the average daily and the upper 10% exceedance thresholds. As with precipitation, generally there is a shift in the distribution towards wetter conditions during the later period.

2.3.2 Determination of heavy precipitation and flow thresholds

The next focus is on streamflow and rainfall at levels that may be associated with flash flood generation. The reader is reminded that flash flooding can occur for moderate to heavy rainfall depending on hydrologic conditions and degree of soil saturation within a watershed. Thus flash flooding can occur as a result of precipitation amounts that are less than the most extreme events. In the following analysis, “heavy” precipitation is defined by rainfall thresholds representing the upper 25\textsuperscript{th}, 10\textsuperscript{th}, and 5\textsuperscript{th}-exceedance percentiles of the station daily or hourly precipitation distributions. The precipitation thresholds were determined by first fitting typical distributions to the station observed data, and then defining the selected quantiles from the fitted distribution. Only non-zero precipitation data from the wet season were used. Several distributions commonly used for precipitation data were considered in fitting the data: Weibull, Exponential, Gamma, Log-normal, Gumbel, and Generalized Extreme Value (GEV) distributions.
For daily data, a minimum precipitation threshold of 1mm/day was used. Gumbel and GEV distributions did not fit the daily data well and are not shown. The shape parameters of the Weibull and Gamma distributions that were fitted to the observations were found to be close to one, indicating exponential behavior. The fit of the observed non-zero precipitation data for the Weibull, Gamma, and Exponential distributions were quite similar. Figure 2.15 shows examples from stations at Santa Barbara, Lake Arrowhead, and Imperial (locations are shown in Figure 2.1). The figure shows quantile-quantile plots of the observed (y-axis) and fitted (x-axis) values for the Weibull, Exponential and Log-normal distributions. The black dashed line is the 1:1 line representing perfect correspondence between observed and fitted quantiles. The small box in the lower left hand corner of the graph corresponds to the 98th-percentile observed and fitted precipitation. Thus the graphs indicate how well the distributions capture the observed values. The “best” fit distribution was objectively selected as the distribution with the minimum total absolute difference between observed and estimated precipitation quantiles between 2- and 98-percentiles. The selected best fit distributions were found to be one of Weibull, Exponential, or Log-normal. Typically, Weibull or Exponential distributions were selected for stations along the coast to mountain regions and the Log-Normal distribution was selected for the station along the desert side of the Transverse and Peninsular Ranges. However, the Log-normal distribution was selected for a few coastal stations and the Exponential distribution was selected for some eastern stations. As shown from the examples in Figure 2.15, often the fit of different distributions to the observed data was quite similar, and the estimated precipitation quantiles of interest (upper 25-, 10-, and 5-th percentiles) were close in value.
A similar procedure was followed to fit the distribution of the hourly precipitation data, with a threshold of 0.25 mm/hr used to classify non-zero precipitation. The individual station’s best fit distribution was selected using the minimum total absolute difference in precipitation quantiles. Only the Gumbel distribution was not selected as having best fit with the observed precipitation quantiles, while all of the other distributions were selected at different locations. The Log-normal was most frequently selected in the Los Angeles basins, extending from coastal Los Angeles and Orange Counties northeastward to the mountains, and the GEV distribution was the next most widely-selected distribution outside of this region. As with the daily precipitation data, often representations of the observed quantiles by the different distributions were quite similar.

From the selected fitted distributions, the upper 25\textsuperscript{th}, 10\textsuperscript{th}, and 5\textsuperscript{th}-percentile exceedance thresholds were extracted for both the daily stations (in mm/day) and hourly stations (in mm/hour). Figures 2.16 and 2.17 show maps of the precipitation thresholds for the daily- and hourly-resolution data, respectively. The general spatial pattern for the precipitation thresholds is similar to the pattern of annual average precipitation. The lowest values for each threshold level occur for the eastern/desert stations, with moderate values at coastal stations, and the highest values at mountain stations. The precipitation thresholds increase as follows for daily stations: eastern/desert stations have upper 25\%-exceedance (P25) values of 5-10 mm/day, 10\%-exceedance (P10) values of 10-15 mm/day, and 5\%-exceedance (P05) values of 14-20 mm/day. Coastal values are 10-20 mm/day for P25, 20-4 mm/day for P10 and 25-60 mm/day for P05. The highest daily precipitation thresholds occur at stations along the San Gabriel to San
Bernardino Mountains in Los Angeles to San Bernardino Counties. The mountain threshold values are 30-40mm/day for the P25, 60-70 mm/day for P10 and 80-100mm/day for P05. More spatial variability is observed in the precipitation threshold values than in annual average values. In particular, the southern coastal region has lower threshold values than the Los Angeles basin and the Santa Barbara/Ventura coastal region. For example, the P90 threshold values in the southern coastal region are in the range 20-30 mm and in the range 30-50mm for the Los Angeles basin, whereas the annual precipitation appears more uniform with most stations within the regions showing 200-400mm of annual precipitation on average. The spatial variation of hourly precipitation thresholds is similar, notably the variation between eastern/desert stations and western locations. Only a few stations in the mountains show high thresholds of P25 > 4 mm/hour, P10 > 7mm/hour and P05 > 10 mm/hour.

Streamflow thresholds are defined herein by return period flows as possible indicators of bankfull discharge. Following standard methodologies, peak annual flow data for the 21 selected stream gauging stations were used to estimate the return period flow thresholds (e.g., Maidment, 1993). To provide a range of streamflow threshold values, the 1.5-, 2-, and 5-year return period flows were determined for the selected stations by fitting a Log-normal distribution to the annual maxima data. The length of record of the annual maxima series for the selected stations generally ranged from 50 to 75 years, with three stations having longer annual maxima series up to 94 years at Arroyo Seco. The return period flows are computed from, and therefore represent, instantaneous streamflow values. However, long records of instantaneous flow values are not common, while mean daily flow records are readily available from the USGS
and report the average of instantaneous values occurring during the day. For fast-
responding streams (e.g., those with response times less than one day), the daily flow
value may be much less than the peak instantaneous value of the day. This is illustrated
in Figure 2.18 for USGS station 11015000, where the instantaneous flow and daily flow
records are plotted together for the period Jan 1998 to April 1998. There are several
peak events during this period, the largest reaching 70 m$^3$/s in early February in the
instantaneous flow records. The mean daily flow computed on the same date was
11 m$^3$/s.

Table 2.2 presents the estimated average daily and return period flows (in m$^3$/s)
for the selected stream gauging stations. Average daily flow values range from 0.03 to
4.0 m$^3$/s, with the 1.5-year return period flow being an order of magnitude or more
larger than the average daily flow (0.3 to 100 m$^3$/s). To reduce the dependence of flow
magnitude on catchment size, the values are normalized by catchment area, and
computed with units of mm/day. Streamflow reported in such units is sometimes
referred as catchment yield or runoff. The spatial variation of the return period flow
thresholds (in mm/day) follows that shown in Figure 2.7a for the annual flow volume,
with higher flow threshold values along the Transverse Mountain Range and lower flow
threshold values along the southern Peninsular Mountains.

2.3.3 Occurrence of heavy precipitation and flow

To define the occurrence events exceeding the defined precipitation and
streamflow thresholds, it is necessary to consider the relevant time scales and temporal
correlation of precipitation events and streamflow. The autocorrelation was computed
for daily streamflow for lag times out to 30 days, for daily precipitation out to 15 days, and for hourly precipitation out to 36 hours. Temporal correlation in precipitation decreases rapidly, as expected. Considering all non-missing daily precipitation, the individual station lag-1 autocorrelation ranged from 0.1 to 0.38, and fell below 0.2 by the second day for all stations. Considering only days with precipitation (P > 1 mm/day), the spread of the lag-1 autocorrelation ranged from -0.05 to 0.3, again with autocorrelation falling below 0.2 by the second day. For hourly precipitation, autocorrelation may remain high for several hours depending on the station. The autocorrelation for a 1-hour lag ranged from 0.3 to 0.78, and autocorrelation fell below 0.2 between the 3rd and 16th hour.

With these temporal correlation characteristics, the average duration of precipitation events was computed for all daily and hourly stations with the following constraints. For daily data, individual events were defined by the count of consecutive days with precipitation greater than 1 mm/day with a default minimum inter-event period of 1 day. Hourly precipitation events were defined by consecutive hours of precipitation greater than 0.25 mm/hour, with a minimum 18-hour inter-event period. As such, intermittent hourly rainfall or “waves” of precipitation events triggered within a day are not counted independently. The average duration of rainfall events thus defined for daily stations ranged from 1.3 to 2.1 days. The duration of events in the eastern/desert region are generally less than 1.5 days, and the duration for coastal to mountainous stations is between 1.6 and 2.1 days. The average number of events per wet season ranged from 5.1 to 19.3 events. The spatial distribution of the average number of events per wet season generally follows that of Figure 2.4, although with a
smaller range and more uniform distribution of values. Stations on the western side of the mountains have 12-19 events per wet season, on average, and those stations to east of the mountains have 5-12 events per wet season. The longest recorded period of consecutive days with rain was 15 days beginning on January 3, 1995, occurring at stations located at San Luis Obispo and Santa Margarita in the northern coastal region.

The occurrence of heavy precipitation for daily stations was defined by the days for which precipitation exceeded the station-specific upper 25%, 10%, and 5%-exceedance thresholds. As an example, the upper panel of Figure 2.19 shows the precipitation record for Lake Arrowhead (located in the San Bernardino Mountains; see Figure 2.1), along with the three precipitation thresholds shown by the horizontal lines. The lower panel of Figure 2.19 shows the number of heavy precipitation events per wet season defined by the upper 25%-exceedance threshold (P25) for daily precipitation for the same station. The mean number of events per wet season at this station is 5.6 events with a standard deviation of 2.8. The average number of occurrences decreases as the precipitation threshold increases as shown in the figure. The average number of events per wet season for the P10 and P05 thresholds at Lake Arrowhead are 2.6 and 1.5 events, respectively.

Imposing the 18-hour inter-arrival time constraint on hourly stations gave an average duration of events that ranged from 3 to 9.4 hours and an average number of events per wet season that ranged from 5.5 to 26. The number of events so defined based on hourly data is therefore similar to the average number of events determined based on the daily stations, and follows the spatial pattern described above for daily events. Given the temporal correlation in hourly precipitation and duration of events,
the occurrence of heavy precipitation events was defined as those periods of rainfall in which at least one hour exceeded the determined precipitation thresholds for the station. Differences in the wet season occurrence of these heavy events at both daily and hourly temporal resolution are considered further in Section 2.3.4 with respect to large scale climatic indices.

The land surface acts as a low-pass filter of precipitation variability, thus the temporal correlation of streamflow tends to be higher than precipitation. The averaging of instantaneous streamflow data to mean daily values also reduces variability for fast responding streams such as those in southern California. The lag-1 autocorrelation of daily streamflow for the 21 selected stations ranged from 0.4 to 0.8. Generally, autocorrelation decreases below 0.3 within 6-7 days, with the exceptions of stations 10255810 (Borego Palm Creek), 11012500 (Campo Creek), and 11189500 (Kern River). These three stations showed autocorrelation values which fell below 0.6 within 3 days, and then showed either slowly decreasing autocorrelation or sustained correlation values between 0.3 and 0.5 for several days. For purposes of defining the occurrence of streamflow events for subsequent analysis, an inter-event period of 6 days was imposed here. If streamflow exceeded a given threshold over multiple days within a 6-day period, the consecutive days were treated as a single event.

Towards the definition of streamflow events, the difference between mean daily flow and the instantaneous nature of return period flows must be reconciled. Generally, mean daily flow and return period flows are not directly comparable. To utilize the return period flow thresholds with the available long-term mean daily flow records for identification of event occurrence relevant to flash flooding, a method to relate the event
occurrences in the mean daily flow record with the instantaneous flow event occurrence at appropriate scales is necessary. Figure 2.18 presents a comparison between mean daily and instantaneous flows at USGS station 11015000, indicating the reduction of variability and peak magnitude observed with mean daily flow. Of the 21 gauging stations selected in climatological analysis, shorter records of instantaneous flows covering the period October 1995 and September 2005 were obtained for 12 stations. The approach taken here was identify a fraction of the two-year return period flow that would identify the same heavy streamflow events from the mean daily flow records as identified in the instantaneous records based on these shorter records, and apply this fraction to the longer mean daily flow records and remaining stations. The approach is illustrated in the following discussion using Figure 2.18. The figure presents the instantaneous flow (blue line) and mean daily flow (red dashed line) records for the Sweetwater River gauging station for January-March 1998. The horizontal line in the figure indicates the two-year return period flow, Q2. For this example, the instantaneous flow (blue line) crosses the Q2 threshold for four events. In contrast, the mean daily flow record (red line) crosses the Q2 threshold for only two of these events. Using a threshold of 50% of the Q2 with the mean daily flow record identifies all four events in the instantaneous record. Consideration of instantaneous Q2 flow events and the corresponding mean daily flow events for all stations indicated a ratio of 50% of the Q2 return period flow is appropriate to define heavy streamflow events from the mean daily flow records. Flow thresholds based on fractions much higher than 50% of Q2 eliminate some of the peak events that were observed in the instantaneous record, while not over representing the number of peak events.
This fraction of 0.5Q_{thr} is applied to each of the 1.5-, 2-, and 5-year return period flow thresholds, and allows for assessment of the occurrence of heavy flow events during the entire period of record (1950-2008) and for all stations (i.e., for those station without instantaneous flow records). Although the constraint of an inter-event period of 6 days limits multiple peak events, this threshold approach allows for more than one heavy flow event during a given wet season. This is in contrast to the definition of return period flow from annual peak series, which allows only consideration of the single highest event in each year. For the selected stations, the average number of events per wet season exceeding three streamflow thresholds ranged from 0.3 to 2.2 for the low threshold (50% of Q_{1.5}), 0.1 to 2.0 for the middle threshold (50% of Q_{2}) and 0 to 1.3 for the highest flow threshold (50% of Q_{5}). The average number of heavy streamflow events occurring each wet season is significantly less than the average precipitation events and thus highlights the need to consider the land surface processes in addition to precipitation events to understand flash flood occurrence.

2.3.4 Large scale climate influences on the occurrence of heavy precipitation and streamflow

Others have shown variations of California precipitation and streamflow with large scale climate phenomena such as El Nino Southern Oscillation (ENSO) and the Madden-Julian Oscillation (MJO) (e.g., Cayan et al, 1999; Mo and Higgins, 1998; Andrews et al, 2004). In this section, variations of the occurrence of heavy precipitation and streamflow during the wet season are explored, using the precipitation
and streamflow threshold determined in Section 2.3.2. Two general indices of the large scale climate are highlighted in this paper: (a) the overall wet season total precipitation or streamflow volume, and (b) the Nino3.4 index as indicator of the ENSO state. The occurrence of heavy precipitation and high streamflow are compared for seasons when these indices are in the upper and lower tercile of their individual distributions.

2.3.4.a Heavy precipitation occurrence conditioned on wet season volume

In this analysis, the individual station wet season total precipitation volumes over the period of record from 1948-2005 were ranked and divided into thirds for each of the daily (and hourly) stations. Comparison is made for wet seasons that are “wetter than average” (i.e. upper tercile) relative to those wet seasons which are “drier than average” (i.e., lower tercile). As an example indication of these differences, Figure 2.20 shows the cumulative frequency distribution of daily precipitation conditioned on the station wet season volume for 3 stations: Santa Barbara, Lake Arrowhead, and San Diego. For these stations, the frequency distributions of daily precipitation for upper versus lower tercile wet seasons are quite different, with more frequent precipitation across a range of magnitudes for the upper tercile wet season. The vertical lines in each plot indicate the three precipitation thresholds (P25, P10, and P05). For each example case, the frequency of precipitation exceeding these thresholds is higher for the case of upper tercile wet season volume.

Ratios were computed for six characteristics of precipitation, dividing the values for wet seasons in the upper tercile by the values for wet seasons in the lower tercile. The characteristics were: (a) number of days (hourly events) with precipitation (P > 1
mm/day or \( P > 0.25 \) mm/hour), (b) wet season average precipitation (mm), (c) standard deviation of wet season average precipitation, and (d)-(f) number of precipitation events exceeding the P25, P10, and P05 thresholds. Also as an example, Figure 2.21 presents the ratios for numbers of days with precipitation and numbers of days with precipitation exceeding P25 for all stations. This figure, again, shows a distinct difference in ratios between the coastal-to-mountainous regions relative to the eastern/desert stations. The ratios are highest in the eastern/desert region, partially due to the low occurrence frequency. Lower ratios are generally noted for the regions of heavier precipitation in the mountains. The number of days with precipitation for wet years (upper tercile) increases by a factor between 1.6 and 2.4 for the western stations, and generally by a factor between 2.8 and 4.1 for the desert stations relative to dry years (lower tercile).

The increase in average precipitation (not pictured) is more moderate, with ratios for the coastal-to-mountainous region between 1.4 and 2.0, and between 2.0 and 2.3 in the southern desert region (Riverside and Imperial Counties) and generally less than 1.8 for the northern desert region (San Bernardino and Kern Counties). In addition to an increase in the mean precipitation, the ratios of standard deviation indicate that the variability in daily precipitation also increases for upper tercile wet seasons. The ratios of standard deviation generally range between 1.4 and 2.2 in the coastal-to-mountainous region, with values in the southeastern region typically reaching 3.0 and with one station having a ratio of 3.6.

The ratios of the number of occurrences of heavy precipitation for the three precipitation thresholds tend to increase as the precipitation threshold increases. For these ratios, a minimum number of occurrences during the lower tercile wet seasons
(i.e., the denominator) of 5 was imposed. For the P10 and P05 thresholds, many of the eastern/desert stations fail to meet this criterion. For the number of occurrences of precipitation greater than P25, the southern coastal and northern coastal ratios are in the range 2.6 to 4.2. The Los Angeles basin ratios tend to be higher (perhaps due to abundant cloud condensation nuclei), in the range 3.0 to 5.0, and the eastern station ratios are generally greater than 5.0 and reach 13.5.

The hourly data were similarly analyzed, conditioned on whether station wet season total volume of precipitation was in the upper or lower tercile of its distribution. The hourly data also show the increase in average precipitation and standard deviation of hourly precipitation for upper tercile wet seasons relative to lower tercile wet seasons. However, the increases are more moderate than the analysis with daily precipitation. The ratios of average precipitation reach a maximum value of about 1.4 while the ratios of standard deviation reach a maximum value of 2.6. Similar to the upper panel of Figure 2.21, the number of rainfall events, as defined in Section 2.3.3 for hourly events, increases during upper tercile wet seasons by a factor between 1.5 and 5.9. The highest ratios are again found for the eastern/desert regions. The increase in ratio of hourly events reaches a value of 3 along the western side of the mountains. Also similar to the behavior observed with daily data, the ratio of the number of heavy hourly events, or intense rainfall, increases as the hourly precipitation threshold increases. For the coastal to mountainous region, typical values for ratio of the occurrence of heavy precipitation are 2-3 for P25, 3-4 for P10, and up to 7 for P05, with ratio values exceeding 8 for the few eastern locations.
This analysis indicates that wet seasons with high total season volumes experience not only increases in the average precipitation, but also increases in the occurrence of heavy precipitation. This increase in occurrence of heavy precipitation is indicated by the ratio of days with precipitation exceeding the heavy precipitation thresholds. There is also an intense rainfall, as indicated by the ratios for events with high hourly precipitation rates.

2.3.4.b Heavy precipitation and streamflow occurrence conditioned on climate oscillations

In this section, the association of heavy precipitation and high streamflow occurrence with various climate oscillations is explored. Monthly climate indices were obtained to characterize common modes of climate variability, including the Southern Oscillation Index (SOI), Nino 3.4 index (Pacific Ocean sea surface temperature, SST, anomalies), Pacific Decadal Oscillation (PDO) index, North Atlantic Oscillation (NAO) index, and Pacific North American (PNA) index. Monthly time series of the indices were obtained from the following: SOI from University Corporation for Atmospheric Research, UCAR (http://www.cgd.ucar.edu/cas/catalog/climind/soi.html) and follows Trenberth (1984); Nino3.4, NAO, PNA from the Climate Prediction Center (http://www.cpc.noaa.gov/data/); and PDO from the University of Washington (http://jisao.washington.edu/pdo).

The correlations between the monthly climate indices and monthly at-station precipitation or streamflow volume were computed, along with the correlations of wet-season precipitation or flow volume with early wet season (October-December) average
climate indices. The monthly correlation between precipitation and climate indices averaged -0.17 with SOI, 0.15 for Nino3.4, 0.04 for PDO, 0.03 for NAO, and -0.02 for PNA. Similar values were determined for monthly correlations with streamflow volume: -0.18 with SOI, 0.16 for Nino3.4, 0.03 for NAO, and 0.05 for PNA. Higher correlation was found between monthly streamflow and the PDO index than determined for monthly precipitation, with average value among stream gauging stations of 0.14. Seasonal correlations are generally higher. The average correlation values of wet season total precipitation with OND average climate indices are: -0.47 for SOI, 0.42 for Nino3.4, 0.07 for PDO, 0.10 for NAO, and -0.15 for PNA. For wet season total streamflow volume, the seasonal correlations are: -0.46 for SOI, 0.37 for Nino3.4, 0.07 for PDO, 0.03 for NAO, and -0.1 for PNA. Again, the correlation of seasonal streamflow with PDO stands out because the season average value is lower than the monthly average value. The correlations of southern California precipitation and streamflow with the ENSO-related indices of SOI and Nino3.4 are stronger than the other indices. Further, since the SOI and Nino3.4 represent the ENSO phenomenon in reciprocal fashion (i.e., positive SOI/negative Nino3.4 associated with La Nina conditions and negative SOI/positive associated with El Nino conditions), the following discussion of results is presented for the Nino3.4 index only, with the understanding that results for the SOI are consistent with respect to the ENSO phenomenon.

El Nino has been indexed by Nino3.4 SST anomalies greater than 0.4°C (Trenberth and Stepaniak, 2001). Rather than a specific SST anomaly threshold, the analysis considers seasonal Nino3.4 values in the upper tercile of its distribution to be indicative of El Nino conditions, and conversely, the lower tercile of Nino3.4 indices to
be indicative of La Nina conditions. Analysis of observed precipitation and streamflow occurrence followed based on these terciles and allowed the period of record of observed data to be divided into periods of equal length. This analysis differs from the previous section in that the division of wet seasons here is regionally consistent, whereas the seasonal volume division was based on and applied individually at each station. For this analysis, the following set of precipitation and streamflow characteristics are considered:

(a) average daily precipitation greater than 1 mm/day
(b) standard deviation of daily precipitation greater than 1 mm/day
(c) average number of days per wet season with \( P > P_{25} \)
(d) average number of days per wet season with \( P > P_{10} \)
(e) average number of days per wet season with \( P > P_{05} \)

and

(f) average daily streamflow (mm/day)
(g) standard deviation of daily streamflow (mm/day)
(h) average number of days with flow (\( Q > 0.001 \) mm/day)
(i) average number of events (days) per wet season with \( Q > Q_{\text{mean}} \)
(j) average number of events (days) per wet season with \( Q > 0.5*Q_{1.5} \)
(k) average number of events (days) per wet season with \( Q > 0.5*Q_{2} \)
(l) average number of events (days) per wet season with \( Q > 0.5*Q_{5} \)

where the precipitation and streamflow thresholds have been previously defined, and with \( Q_{\text{mean}} \) in line (i) representing the long term averaged mean daily flow. Figures 2.22
and 2.23 illustrate example results for the analysis of daily precipitation and daily streamflow, respectively. The figures present box-plot summaries of the ratios of selected characteristics for all stations, and for wet seasons defined either by the lower or upper tercile of the OND Nino3.4 indices relative to the unconditioned characteristics. Thus, on any plot, a ratio value of one indicates that the value of the selected characteristic for the given Nino 3.4 tercile is equal to the long-term mean value. The lower tercile results are shown by the left box-plot and the upper tercile by the right-hand box-plot.

Figure 2.22 presents the ratios of (a) average precipitation, (b) standard deviation of precipitation, and (c) average number of days per wet season with precipitation greater than the P10 threshold. Ratios greater than 1 are found for all characteristics when the OND Nino3.4 index is in the upper tercile, indicating higher than average values during El Nino conditions. During El Nino condition, there tends to be higher than average mean precipitation (plot (a)), with more variability (plot (b)) and an increase in the number of days with heavy precipitation (plot (c)). The average ratio of the number of days exceeding the three precipitation threshold increases for the upper tercile wet seasons as the precipitation threshold increases, from an average value of 1.35 for the P25 threshold, to 1.42 for P10 threshold, and to 1.5 for P05 threshold. Likewise, the average ratios decline for lower Nino 3.4 tercile wet seasons as the precipitation threshold increases (values of 0.62, 0.58, and 0.51 for P25, P10, and P05, respectively). The ratios for the number of occurrences of heavy precipitation at all thresholds show distinct values for upper versus lower tercile Nino 3.4 wet seasons; that
is, there is no overlap in the individual station ratio values between the two Nino 3.4
teriles.

Figure 2.23 presents corresponding results for streamflow and includes the ratios for (a) the average daily flow, (b) the average number of events with flow greater than the annual mean, (c) the average number of events with flow greater than 50% of Q_{1.5}, and (d) the average number of events with flow greater than 50% of Q_2. The results are similar in character to the precipitation results, indicating higher than average mean flow, greater variability, and an increase in the number of high flow events during wet seasons when the Nino3.4 index is in its upper tercile (i.e., El Nino-like conditions), and opposite behavior when the index is in its lower tercile (i.e., La Nina-like conditions). The average daily flow is, on average, about 45% of the mean (unconditioned) wet season flow value when the Nino3.4 index is in its lower tercile, approximately 80% of the mean seasonal value when Nino3.4 index is in the middle tercile (not shown), and 165% of the mean seasonal value when Nino3.4 is in the upper tercile. The ratios for the number of events with streamflow exceeding the specified thresholds also increase (decrease) as the threshold increases for Nino3.4 indices in the upper (lower) tercile. The average ratio for number of events greater than the annual mean flow for upper tercile wet season is 1.45, approximately 1.8 for flow greater than 50% of Q_{1.5}, 2.0 for flow greater than 50% of Q_2, and 2.2 for flow greater than 50% of Q_5. For lower tercile wet season, the ratio values, in the same order, are: 0.63, 0.35, 0.25 and 0.05. This ratio for events with flow greater than 50% of Q_5 with the seasonal Nino 3.4 in the lower tercile was influenced by several stations where this number of events was zero.
2.4 Conclusions

This chapter has characterized the spatial and temporal variability of precipitation and streamflow in southern California, with emphasis on spatio-temporal scales relevant for flash flooding. It presents the first detailed analysis of hourly- to daily-precipitation and daily-streamflow observations specifically for southern California spanning a historical period of 56 years. The observational records included 111 daily precipitation stations and 59 hourly precipitation stations from the northern San Luis Obispo, Kern and San Bernardino county lines to the southern border of California with Mexico. A total of 21 stream gauging stations with minimal influence of regulation or diversions were also selected for the analyses. These selected locations were primarily along streams flowing to the Pacific Ocean.

The precipitation and streamflow records show clear seasonal cycles, particularly for locations along the coastal to mountainous region. Peak precipitation, both in monthly amount and number of days per month, and peak streamflow occur generally in January or February for the coastal to mountainous regions, and with little precipitation and low streamflow during the months of June, July and August. Precipitation stations located in the southeastern/desert regions generally show and additional peak in precipitation during the late summer/early autumn months of August or September. Despite a limited number of precipitation stations existing at high elevations, there is indication of orographic influences on the precipitation climatology, with a general spatial pattern of high precipitation (annual or wet season volume and number of days) in the mountainous regions with lower precipitation in the foothill and coastal region, and very low precipitation amount and occurrence in the eastern/desert
region. The few numbers of stations at high elevation limit detailed information on the spatial variation of precipitation over the mountainous regions, but typical observed values reach 4, 7 and 9mm/h for the upper 25, 10 and 5% exceedance quantiles, respectively.

Synoptic forcing of heavy precipitation over the region is generally associated with low pressure anomalies in sea level pressure and geopotential height at lower and upper atmospheric levels over the Pacific Northwest coast of the U.S., and high wind and high humidity anomalies in the Southern California Bight region. Variations in these general synoptic features were identified for heavy precipitation occurring over different sub-regions. Heavy precipitation along the Peninsular Mountains was associated with an extension of the low pressure anomalies towards the central Pacific and weaker high pressure anomaly over the southern Alaskan coast, yielding lower 850 hPa wind anomalies with a more westerly flow pattern as wind interact with southern California topography. Heavy precipitation along the Transverse Mountains was associated with a spatially limited low pressure anomaly over the Pacific Northwest and deeper high pressure anomaly over the Aleutian Islands off Alaska. Thus the wind anomalies were stronger and in a more southwesterly to southerly flow direction.

Precipitation and streamflow thresholds are developed to characterize the occurrence of events during the wet season at scales relevant to flash flooding. Often associated with extreme rainfall, flash flooding occurs as streamflow exceeds the capacity of the channel network. The flow rate associated with the capacity has been cited as having a recurrence interval on the order of 1-2 years. Flash flooding can also occur following precipitation that may be less than the most extreme events if
antecedent soil saturation conditions are high. Thus, the occurrence of precipitation and streamflow events over a range of thresholds are considered: precipitation exceeding station-defined thresholds corresponding to 25-, 10- and 5-% exceedance probability, and streamflow exceeding the mean flow and thresholds based on peak flows with return periods ranging from 1.5-years to 5 years.

The occurrence of these precipitation and streamflow events is sensitive to indicators of the climatic state. First, it was shown that station average precipitation and variability increases during wet seasons with higher than average total seasonal volume relative to wet seasons with lower than average total volume. The observational records indicate an increase in the occurrence of heavy precipitation, as evidenced by increases in the number of days with precipitation exceeding heavy precipitation thresholds during the seasons with high total seasonal volume. Finally, an increase in intense rainfall, represented by the number of events exceeding high hourly precipitation rates, was also indicated for these high season volume years.

Secondly, it was shown that the occurrence of southern California precipitation and streamflow events shows an association with variation in the El Nino Southern Oscillation state. The correlation of seasonal precipitation and streamflow volume with early season climate oscillation indices of SOI, Nino3.4, PDO, PNA, and NAO was determined to be highest for SOI and Nino3.4, and on the order of 0.4. Station precipitation and streamflow tend to be higher than average during wet seasons characterized by Nino3.4 indices which are higher than average (or SOI indices lower than average) or El Nino-like conditions. Conversely, precipitation and streamflow tend to be lower than average during wet seasons oppositely characterized by Nino3.4 and
SOI indices, or La Nina-like conditions. Additionally, the occurrence of heavy precipitation and high streamflow events increases during El Nino-favorable conditions and are distinctly different than the occurrence during La Nina-favorable conditions.

This analysis has been limited to observational records and statistically derived indicators of heavy precipitation and high streamflow that may be indicators of conditions favorable for flash flood occurrence. Further insight may be gained with information on the spatial variation of precipitation afforded through numerical modeling of orographically generated precipitation. Such models can simulate precipitation on scales to a few tens of square kilometers and thus at the level of interest for flash flood occurrence. This may provide additional information on the spatial distribution and variability not captured by the sparsely gauged mountainous region. Such detailed, spatially resolved simulated precipitation data may provide input to detail hydrologic models which define flash flood potential on a physical basis. This interdisciplinary modeling approach to define flash flood occurrence is the focus of continuing research for the southern California region in an effort to describe the climatology of flash flood occurrence throughout the region, and to consider links to climatic forcing.

2.5 References


Woodyer, K.D., 1968: Bankfull frequency in rivers, J. Hydrology, 6, 114-142.
Table 2.1. Dates and regional precipitation anomalies for four regions.

<table>
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<tr>
<th>North Coastal</th>
<th>Transverse Mtns</th>
<th>Peninsular Mtns</th>
<th>Eastern/Desert</th>
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* 2 consecutive days in highest ranking anomalies
Table 2.2. Stream gauging station flow properties (flows reported in m³/s).

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<th>Station</th>
<th>Location</th>
<th>A (km²)</th>
<th>Avg Q&lt;sub&gt;daily&lt;/sub&gt;</th>
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<td>4.2</td>
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<td>4.7</td>
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Figure 2.1. Terrain and select locations in southern California region. Cities are shown by circles, and county boundaries in black lines with names in italics.
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Figure 2.11. As in Figure 2.9, but for dates of high precipitation for Peninsular Mountains region.
Figure 2.12. As in Figure 2.9, but for dates of high precipitation for eastern/desert region.
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Figure 2.22. Box-plots of the ratio of precipitation characteristics for wet seasons conditioned on Nino3.4 climate index to unconditional value: (a) average daily precipitation; (b) standard deviation; and (c) number of days with \( P > P_{10} \).
Figure 2.23. Box-plots of the ratio of streamflow characteristics for wet seasons conditioned on Nino3.4 climate index to unconditional value.
CHAPTER 3. Regional Bankfull Geometry Relationships for Southern California Mountain Streams and Hydrologic Applications

Abstract

The estimation of bankfull discharge has become important in many hydrologic, ecosystem, and river restoration studies. Regional bankfull geometry relationships allow for the estimation of bankfull channel dimensions and/or channel discharge at ungauged locations based on detailed surveys at a few locations within a region of interest. This study develops and compares such regional relationships for bankfull channel width, hydraulic depth, and cross-sectional area for southern California mountain streams based on several data sources: surveyed streams, US Geological Survey stream survey reports, and existing literature. Although considerable uncertainty exists in estimating bankfull conditions, the relationships developed from the varying data sources show significant agreement. The utility of the developed bankfull geometry regional curves is then demonstrated for several hydrologic applications. The first is the application of the regional curves to small watersheds to illustrate the spatial variation of estimated bankfull properties. This application demonstrates the estimation of these properties throughout a region with few in-situ observations of channel cross-sections. For small watersheds ranging with drainage area ranging from 15 to 3000 km², the estimates of bankfull top width ranged from 7.2 m to 44.5 m and hydraulic depth estimates ranged from 0.35 to 1.15 m. This analysis was extended in a second application to estimate bankfull discharge based on the bankfull cross-sectional properties using Manning’s steady uniform flow formula. For
selected locations with instantaneous flow records, the occurrence frequency of events exceeding bankfull flow was examined for the available 10-15 year span of observational records. Bankfull discharge estimates for all small watersheds in the region ranged from 1.3 to 74 m$^3$/s, while the range at the selected gauged stream locations was from 2.6 to 16.4 m$^3$/s. The frequency of streamflow events exceeding bankfull occurred on average once in 0.15 to 3.3 years. Stream locations along the Transverse Mountains of southern California showed an average of occurrence frequency of less than 1 year, whereas along the Peninsular Mountains the average occurrence tended to be greater than 1 year. A third application is the use of the regional bankfull geometry relationships to estimate the surface runoff response necessary to produce bankfull conditions at the channel outlets draining small catchments. This surface runoff response may be used as an index for conditions of minor flooding with saturated soils from antecedent rainfall. The surface runoff response index ranges from 2.0 to 5.5 mm for a 3-hour rainfall duration for southern California watersheds greater than 15 km$^2$ in area.

3.1 Introduction

Bankfull discharge has become an important and widely used concept in hydrologic science, ecosystem studies, and river restoration design. Bankfull discharge is the flow in a river channel at the level of transition from the active channel to the flood plain (Leopold, et al. 1964). It is closely associated with the concept of channel forming or dominant flow, which determines and maintains the channel dimension, and with the effective discharge, which is the flow that carries the highest sediment volume.
over time. Leopold (1994, pg. 127-128) showed that computations of bankfull discharge using data from surveyed streams and the computations of effective discharge from sediment surveys were nearly equal for streams of the Yampa River basin in Colorado and Wyoming. The concept of bankfull discharge is most frequently applied to alluvial channels, given the connection with effective sediment transport and erosion. The three terms of bankfull-, effective-, and channel-forming- flow have become strongly interconnected. Bankfull discharge has become a widely-used surrogate for channel forming flow in many studies as it may be recognized based on morphologic field evidence.

The systematic variation of stream channel hydraulic geometric characteristics was first suggested by Leopold and Maddox (1953), who showed the relationship of channel width, mean flow depth, velocity, and suspended sediment load to bankfull discharge in the form:

\[ X = a Q^b, \]  

(1)

where \( X \) represents the dependent variables (width, depth, velocity, total suspended sediment), \( Q \) is the discharge, and \( a \) and \( b \) are derived parameters. Such relationships are discussed in two contexts: (a) “at-station hydraulic geometry” which describes the variation of cross-sectional properties with varying discharge rates at a given location or cross-section, and (b) “downstream hydraulic geometry”, which describes the variation of cross-sectional properties along a length of river with a given flow level, such as the mean annual flow, a selected recurrence interval flow, or bankfull flow. The latter case
reflects the variation of discharge and channel dimension with catchment scale. Thus as
discharge varies with catchment size, this work has been extended to examine the
variation of bankfull discharge and related bankfull channel geometry with catchment
area, A, in a similar form (Dunne and Leopold, 1978):

\[ X = \alpha A^\beta, \]  

with \( \alpha \) and \( \beta \) being the parameters of this relationship. Herein, relationships in the form
of Eq. 2 are referred to as regional hydraulic geometry curves (regional curves), and
they relate the bankfull channel dimensions and discharge to watershed drainage area
and possibly other physical or climatological watershed characteristics such as main
stream length, channel slope or mean precipitation.

Hydraulic geometry analyses have been performed throughout the U.S. and
abroad to empirically derive hydraulic geometry relationships (Eq. 1) or regional
hydraulic geometry curves (Eq. 2) for a single stream or for several streams in a
coherent hydro-physiographical region (e.g., Harvey, 1969; Rhoads, 1991; Dudley,
2004; Messinger and Wiley, 2004; Lawlor, 2004; Chaplin, 2005). As regional
hydraulic geometry curves allow for the estimation of bankfull channel dimensions at
unsurveyed locations, regional curves have had practical application in a variety of
studies of fluvial or channel processes (Merigliano, 1997; Buhman, et al. 2002;
Stewardson, 2005), river restoration or natural channel design (Rosgen, 1994, 1996;
Metcalf et al, 2009), habitat assessment and ecosystem studies (Singh and McConkey,
Development of regional hydraulic geometry curves relies on estimates of channel geometric characteristics and flow from detailed field surveys and field identification of bankfull conditions. Leopold (1994) gives the following as principal field indicators of bankfull stage (in order of usefulness):

- Height of point bars
- Changes in stream bed and vegetation (e.g., from channel bed sediment to grasses or change in vegetation species)
- Topographic break or change in channel side slope
- Changes in the size distribution of bed materials (e.g., sand to gravel or cobble stones)
- Changes in debris deposited along banks (although caution is warranted as this may be associated with a single significant recent flooding event).

Uncertainty or differences in estimates of bankfull level has been noted (Williams, 1978; Johnson and Heil, 1996; Radecki-Pawlik, 2002; Navratil et al, 2006; Harman et al, 2008; Xia et al, 2010). Williams (1978) cites 11 variations on the definition of “bankfull”, based on morphologic evidence including various sedimentary surfaces, vegetation and geomorphic boundary features. Such variations may lead to different estimates of the bankfull level at a single cross-section. Xia et al (2010) discuss difficulties in identifying and estimating bankfull level based on field evidence in the
complex and braided channel network of the Lower Yellow River in China. Indeed, there may be variation in channel morphology over relatively short stream reach distances which influence the selection of bankfull elevation. Further recommendations to improve reliability of field-based bankfull stage estimates include the use of multiple indicators at a single cross-section, and evaluation of multiple locations within a short stream reach (up to 20 times the stream width; Leopold, 1994).

After identifying the bankfull elevation, varying methods for computing the associated discharge exist and may involve observed streamflow records and at-station rating curves (i.e., stage-discharge relationships), existing at-station hydraulic geometry relationships, flood frequency information or empirical flow relationships (Williams, 1978). Williams (1978) and Johnson and Heil (1996) give 16 estimates of bankfull discharge at selected stream locations to examine variation in the estimates. Such studies often examined the definition of bankfull in context of estimating the recurrence interval of bankfull flow (Williams, 1978; Petit and Pauquet, 1997; Castro and Jackson, 2001; Navratil et al, 2006). Leopold’s definition of the bankfull discharge as having a 1.5-year return period is frequently cited (Leopold, 1994). Some studies show good agreement with this period; for example, Castro and Jackson (2001) report recurrence intervals of 1-3.1 years with an average of 1.4 years for streams in the Pacific Northwest United States. Other studies report larger variability in the recurrence interval. Williams (1978) reported a range of bankfull recurrence intervals from less the 1 year to more than 30 years. Petit and Pauquet (1997) report a range of values from 0.7 year to 5.3 years for streams in Belgium, with variation in the recurrence interval further classified by catchment size and by permeability of stratum.
The concept of bankfull discharge and the development of hydraulic geometry relationships have historically been applied for alluvial, and often low-gradient, streams. Until recently, such relationships for streams with more erosionally-resistant channel bed, including bedrock substrate, and colluvial processes have received little attention (Wohl and Wilcox, 2005). Wohl et al (2004) found relatively poor correlation in downstream hydraulic geometry relationships for a high-gradient stream in Colorado, with improved correlation with the addition of reach-scale controls such as channel gradient. In contrast, Wohl and Wilcox (2005) found well-developed downstream hydraulic geometry relationships for two high-gradient streams in New Zealand, and with exponent values in the relationship commensurate with those originally suggested by Leopold and Maddox (1953). Montgomery and Gran (2001) explore variation in hydraulic geometry relationships of the form of Eq. (2) for five mountain streams that included both alluvial and bedrock reach sections. They found good agreement among the exponents of the channel width to drainage area relationships between bedrock reaches and alluvial reaches, with the exception of a small drainage basin (A < 1 km$^2$) in Oregon. The exponent values found were also in agreement with reported exponent values for alluvial streams. Finally, Wohl and Merritt (2008) examined an extensive collection of data from 335 mountain stream reaches throughout the western continental United States, Alaska, Panama, New Zealand, and Nepal. There was agreement among the exponents of the hydraulic geometry relationships for channel width and depth (as functions of bankfull discharge) with reference values for alluvial streams, but with variation in the relationships considered for different channel form types (e.g., pool-riffle, step-pool). These recent studies suggest that regional hydraulic geometry curves
may be produced for mountain streams, with relationship exponent values similar to those found in the wealth of literature on alluvial streams.

This chapter develops regional bankfull geometry relationships for streams in southern California utilizing stream survey data from different sources. This region includes mountainous and foothill streams of the Transverse and Peninsular Mountains of southern California. The first data source is from a set of streams for which field surveyed were conducted following significant hydrometeorologic events in January 2005. This dataset contains channel cross-sectional survey data with estimates of channel hydraulic depth and width measurements based on field survey indicators to identify bankfull conditions. The second data source is derived from two types of stream survey reports from the U.S. Geological Survey (USGS). These reports include field notes from regular discharge measurements or post-event reports of significant flood events. Both USGS reports provide detailed channel cross-section survey data from which bankfull conditions were estimated based on cross-section shape only. From each survey data source, regional relationships for the channel width, hydraulic depth, and channel cross-sectional area at bankfull conditions were developed as a function of drainage area.

These relationships are compared to similar relationships based on data on bankfull channel width and depth available in the literature. This includes estimates from an independent survey in the San Bernardino Mountains of southern California (Azary, 1999) and data extracted from the Environmental Protection Agency’s Wadeable Streams Assessment (EPA WSA; EPA, 2006; Faustini et al., 2009a), which covered streams throughout the southern California region. Estimates of bankfull width
and depth from these sources were used to develop separate regional relationships, and were compared with the relationships developed from survey data.

Three applications of the developed regional curves are shown. The first application examines the spatial variability of estimated bankfull geometry parameters and additional derived properties for small watersheds throughout the region. The derived properties include the discharge at bankfull conditions, as estimated using Manning’s steady uniform flow relationship (Maidment, 1993, pg. 8.22). To apply this relationship, an estimate of channel roughness is required. Manning’s roughness factor has been recognized as varying with the depth of flow, as the flow interacts with the channel bed surface. Jarrett’s (1984) relationship for Manning’s roughness factor as a function of flow depth and channel slope for high-gradient streams is employed herein.

Estimates of the recurrence interval of bankfull discharge have not been reported for southern California streams. Using bankfull discharge estimated via Manning’s steady uniform flow equation and the regional bankfull geometry relationships, the occurrence of flow exceeding bankfull discharge is examined for specific stream gauging locations as the second application of the regional relationship. The average occurrence frequency of bankfull flows is computed for these locations over the available record of instantaneous flows.

The third application of the regional curves is to estimate the watershed surface runoff response necessary to produce bankfull conditions. This is a threshold indicator following the threshold runoff theory as developed by Carpenter et al (1999) and which has been utilized in operational flash flood warning systems (Georgakakos, 2006; Villarini et al, 2010).
This chapter presents a description of the study region and data sources in the following section. In Section 3.3, the developed regional bankfull geometry curves are presented and compared. Section 3.4 presents the results of the three applications of the regional curves to hydrologic analyses. Finally, conclusions and recommendations are presented in Section 3.5.

3.2 Data Description

The study region is the cismontane region of the Southern California Bight (Figure 3.1). The striking geomorphologic feature of the region is the close proximity of the southern California mountains to the Pacific Ocean. These include the Transverse Mountains, running approximately east-west from north of Santa Barbara to east of San Bernardino, and the Peninsular Range Mountains, running north-south from San Bernardino County into the Baja Peninsula of Mexico. This topographic setting leads to strong precipitation gradients from the coastal region to the mountains, with heavy and often localized precipitation occurring in the mountains. The narrow mountain valleys and relatively short stream lengths lead to significant potential for hydrologic hazards including flash flood occurrence, landslides and debris flows when the region receives heavy precipitation. Recent examples of events with fatal impacts have been widespread flash flooding and La Conchita landslide in January 2005 and the Waterman Canyon debris flow of December 25, 2003 (Carpenter et al, 2007; Jibson, 2005, San Bernardino County Flood Control District, 2007). Southern California has a Mediterranean climate with precipitation that occurs predominantly during the cool season between October to April and with dry moderate summers. Average annual
precipitation reaches 1000 mm in the mountains, is in the range of 200-400 mm along
the coastal region, and is between 80 and 200 mm in the desert regions to the northeast
and east of the Transverse and Peninsular Range Mountains. Of particular interest in
this study, as it is motivated by flash flood occurrence, are the mountain-to-foot hill
regions within Santa Barbara, Ventura, Los Angeles, San Bernardino, Riverside,
Orange and San Diego counties. The region is marked by large metropolitan and
populous urban areas around the cities of Los Angeles, San Diego, San Bernardino, and
Riverside, which are primarily downstream of the main focus areas.

3.2.1 2005 southern California field surveys

Starting in late 2004 and continuing in early 2005, southern California
experienced a series of heavy rainfall events that caused significant impacts throughout
the region. The most significant precipitation event occurred between January 5 and 12,
2005 with the 7-day event total precipitation exceeding 700mm of precipitation at
observation stations in the San Gabriel Mountains region of Los Angeles to San
Bernardino Counties. The climatological mean for the month of January in this region
is on the order of 250mm. These storms caused widespread flooding and other damages
including a massive landslide at La Conchita, CA which destroyed or damaged 35
homes and caused 10 fatalities (Carpenter et al, 2007; Jibson, 2005). As part of an
NSF-funded (SGER) grant, the author participated in a fast-response survey to collect
evidence of the hydrologic impacts of these storms. Field surveys were conducted in
January 2005, with one of the survey objectives to identify the channel width and depth
at bankfull and high water mark locations on relatively small streams throughout the
region affected by the storm. Follow-up surveys were conducted in early March to add coverage of streams in Riverside and San Diego Counties. Typically, a limited channel cross-section survey was made at each site with depth and distance measurements across the channel at high water and bankfull indicators, and if possible at other significant changes in channel side slope, current water level, and/or the channel thalweg. The surveys were done in relatively low flows, when personnel could safely wade through the flow. Field identification of bankfull conditions at each survey site considered changes in channel side slope, vegetation, and/or bed material size. Often survey personnel identified multiple indicators of potential bankfull conditions, and the bankfull elevation was selected in subsequent evaluation of the channel cross-section data. Post-survey evaluation included the computation of the bankfull channel top width, $B_b$, cross-sectional area, $Ax_{sb}$, and hydraulic depth, $D_b$ (defined as the cross-sectional area divided by top width, $D_b=Ax_{sb}/B_b$). A total of 31 sites were surveyed and are indicated in Figure 3.2 by cross (+) symbols. At several locations, two cross-sections were measured in relatively close proximity to provide an indication of the uncertainty in assessing bankfull conditions. The practice of locating multiple indicators of bankfull stage along a stretch of river is recommended to improve reliability of estimating bankfull elevation (Leopold, 1994). Table 3.1 summarizes the field survey locations and estimated bankfull geometry properties. Location information includes the stream name, county, approximate drainage area, and number of cross-sections surveyed. Average bankfull geometry values are included for those locations with two cross-sections. The drainage area was estimated based on GIS-
processing of digital elevation data to demarcate watershed boundaries. The drainage area of the field survey locations ranged from approximately 0.5 to 50 km$^2$.

### 3.2.2 USGS surveyed sites

Additional channel cross-sectional data was collected from the Poway, CA and Santa Maria, CA Field Offices of the US Geological Survey. A database was developed from (a) hardcopy notes of regular discharge measurements or (b) post-flood survey reports conducted by USGS personnel for USGS stream gauge locations. Both types of reports include channel cross-sectional survey information detailing the distance and depth at points across the channel section. The dataset included three surveys which were done for critical depth measurements, at locations with controlled weirs. For these reports, the surveyed approach cross-section was used. While focus was toward smaller, natural (non-channelized) streams, collection of the USGS data attempted to gather as much information as possible. Often, multiple reports (post-flood reports and discharge measurement notes) were collected for a given stream location. Discharge measurement surveys are conducted quite frequently by the USGS over a range of flows. For the purposes of this work, the discharge at the time of measurement was compared with the mean annual flow to eliminate low flow measurements that would likely not include the bankfull condition. Post-flood surveys, as the name implies, are typically made in association with relatively high discharges, which are often larger than bankfull flow.

The USGS survey data were digitized to produce plots of the channel cross-sections at each location and for each survey date. For each, the “bankfull condition”
was identified based only on the cross-section, using changes in the channel side slope as the primary indicator (e.g., a break in the slope indicating a transition to the flood plain or perhaps from a previously incised channel). Typically, the field notes did not include additional detailed information, such as changes in vegetation, which could have been used in conjunction with the cross-section data to assess the bankfull stage. Multiple surveys at each site were used to aid in identifying consistent estimates of bankfull. Figure 3.3 illustrates examples of the cross-section data, along with the selected bankfull stage for Matilija Creek in Ventura County (USGS station 11114495), and Las Flores Creek in San Diego County (USGS station 11046100).

Data from a total of 29 stream gauge locations were analyzed and are listed in Table 3.2. The list includes the location, USGS-specified drainage area, number of cross-sections available, and average bankfull geometry estimates. The drainage areas for these stream surveys are generally for larger basins than the 2005 field survey streams (see Table 3.1), and range in area from 14 to 2294 km². The locations of the USGS cross-sectional survey sites are shown in Figure 3.2 by the open circle (o) symbols.

3.2.3 Independent reports of bankfull geometry for southern California streams

Independent sources of bankfull channel geometry for streams within the southern California region are identified in this section. Leopold (1994, Figure 8.5) presented bankfull top width, hydraulic depth, and cross-sectional regional curve estimates for the San Francisco Bay region of California. Carpenter et al (1999) derived the regional relationships for California based on this figure, and this is also used in
later sections as a baseline comparison. However, since the basis of those relationships was northern California, additional information for the southern California region was sought. Two sources were discovered that provide estimates of bankfull channel geometry for streams with the region:

3.2.3.1 Azary (1999)

Azary (1999) conducted field surveys of seven small streams in the San Bernardino Mountains of southern California, generally located to the northwest to northeast of the city of San Bernardino. The stream locations, drainage areas, and estimated bankfull properties are included in Table 3.3. The drainage areas of the seven sites range from 5.7 to 99 km². Three of these locations were near USGS gauging stations, from which the author estimated the bankfull discharge from flood-frequency analysis of USGS peak flow data as the flow having a 1.5-year return interval. The bankfull elevation was determined during the surveys from field evidence following guidelines of Harrelson et al. (1994) and Rosgen (1996), and with reference to the stream gauge station data for the three locations. The bankfull properties estimated include bankfull top width, mean depth, and cross-sectional area. It is noted that several of these streams are also covered by the data collected from the USGS (Plunge and City Creeks) and the 2005 field surveys (Mill Creek), although the surveys may have been conducted at different locations along the stream.

3.2.3.2 EPA Wadeable Streams Assessment
In 2000, the U.S. Environmental Protection Agency (EPA) began a pilot program of surveying the ecological condition of streams throughout the U.S. (EPA, 2006; Faustini et al, 1999a). The surveys cover physical, biological, and chemical attributes of wadeable, perennial streams and rivers within the continental U.S., with specific and detailed measurements taken at each survey site, and with the goal of providing a monitoring framework and dataset for the assessment of the ecological condition of streams. The term “wadeable” indicates the streams could be sampled without the need or use of a boat. According to their estimates, nearly 90% of the total stream length within the conterminous U.S. is classified as wadeable; only 10% of the stream length belongs to major rivers. The dataset is extensive, with sampling at 1392 stream locations nationwide, and is freely available (Faustini et al, 1999a).

Of interest for this analysis was the collection of physical geometry properties of channel cross-sections at southern California locations. From the database, a set of 62 locations were identified within the southern California region and the channel data extracted with their locations shown in Figure 3.4. Several locations fall along the same streams as surveyed by the authors, the USGS and Azary (1999). The channel data included bankfull width and mean depth measurements along a series of cross-sections taken at each site. For the EPA assessment, the bankfull channel was defined as “the channel that is filled by moderate-sized flood events that typically occur every one to two years” (EPA, 2004). Standardized procedures were employed in the assessment to identify bankfull geometry. Field evidence used to identify the height of bankfull flow included:
(a) obvious slope break differentiating the channel from a relatively flat floodplain terrace;
(b) transition from exposed stream sediment to terrestrial vegetation;
(c) moss growth on rocks along the banks;
(d) presence of drift material in overhanging vegetation
(e) transition from flood- and scour-tolerant vegetation to vegetation with is relatively flood- and scour-intolerant.

At each site, eleven cross-sections were measured along a channel reach length of approximately 40 times the bankfull width. At each cross-section, the following measurements were recorded: bankfull channel width, height of the bankfull elevation above the water surface at the time of survey, and mean depth of water. Bankfull mean depth was computed as the sum of the reported mean depth and height of bankfull elevation above the water. For each survey site, the mean bankfull width and bankfull depth from the eleven surveyed cross-sections were computed and used in the regional curve analysis. These mean values are included as independent estimates of bankfull properties.

3.3 Regional Bankfull Geometry Curves

As presented by Dunn and Leopold (1978), regional hydraulic geometry curves relate the bankfull channel dimensions and discharge to the watershed drainage area. Drainage area is often used as the primary predictor in such curves, although some researchers have included other reach-scale controls such as channel slope (e.g., Wohl
et al, 2004). Considering drainage area only, the power-law form of the relationship as given in Eq. 2 is used:

\[ B_b = \alpha A^\beta, \]
\[ D_b = \varepsilon A^\delta, \]
\[ Axs_b = \lambda A^\gamma, \]

where \( B_b, D_b, \) and \( Axs_b \) are the bankfull width, depth, or cross-sectional area, respectively; \( A \) is the watershed drainage area to the survey site; and \( \alpha, \beta, \varepsilon, \delta, \lambda, \) and \( \gamma \) are the estimated parameters of the regional relationships. This section presents the developed regional curves for southern California mountain streams based on the different data sources. The focus is on small mountain streams with drainage area less than 2000 km\(^2\). The regional curves are developed for channel cross-sectional geometry properties only; bankfull discharge was not included as this was not estimated at all survey sites. The regional curves developed based on southern California stream surveys are presented first, as the author estimated the bankfull geometry properties. These relationships are then compared with the independently surveyed and estimated data.

3.3.1 Southern California 2005 survey-based curves

Figure 3.5 presents the bankfull geometry properties as a function of survey site drainage area from the 2005 southern California field surveys. For all figures of this
section, the bankfull top width estimate is shown in the top panel, bankfull depth is shown in the center panel and cross-sectional area is shown in the bottom panel. For the 2005 field surveys, bankfull top width was determined directly from the survey data. The cross-sectional area was computed from depth-distance measurements, and bankfull hydraulic depth computed as the cross-sectional area divided by top width at the bankfull elevation \(D_b = Ax_s / B_b\). At the seven locations where multiple cross-sections were surveyed in a relatively short distance along the channel, the bankfull channel properties estimates were examined to provide an indication of the at-site variability in the estimates relative to the differences between different sites. Although the number of sites with multiple estimates of bankfull geometry is limited, the estimates at individual sites were similar, and the average bankfull properties at these sites were used in the determination of regional curves. Figure 3.5 presents the average values for these locations.

The parameters of the regional curve were fit to the data by regression analysis. The regional curves (solid lines), along with the 95% confidence bounds of the regression curve (red, dashed lines), are shown with the data in Figure 3.5. The regional curve relationship for each bankfull property is presented with the regression correlation coefficient, \(R\). The linear correlation varies from 0.46 for hydraulic depth, 0.61 for cross-section area, and 0.66 for top width. Also included in the plots of Figure 3.5 are the relationships presented by Leopold (1994) for northern California (San Francisco Bay region) by the dashed gray line. Although the data from southern California lie near the regional curves for northern California, the curves for southern California fall slightly below the Leopold’s relationships for all properties. This
suggests that smaller stream cross-section dimensions characterize bankfull conditions for the southern California streams. It is noted that the 2005 field survey was limited to relatively small drainage basins. The size of the drainage basins on which Leopold’s relationship for northern California is not known, but it is likely to include larger basin streams.

Other watershed characteristics may be included in the development of regional geometry curves. Typically, these can include stream length, L, or channel slope, S, and the regression relationship may take the form:

\[ X = \alpha A^{\gamma^*} L^{\delta^*} S^{\beta^*}, \]  

(6)

where \( X \) may be any of the cross-sectional properties, \( B_b, D_b, \) or \( Axs_b \), of Eq. 3-5, and the regression parameters, \( \alpha, \gamma^*, \delta^*, \) and \( \beta^* \) are different from those of Eq. 3-5. The addition of these watershed variables as bankfull geometry predictors was considered for the 2005 field survey analysis using a piece-wise linear regression model. The additional watershed properties were derived from GIS digital elevation processing for basins with outlet near the survey site. However, the additional predictors did not improve the estimates at the 95% confidence level and those analyses are not included herein.

Figure 3.6 presents the bankfull geometries estimated from the field surveys conducted by the USGS. Different symbols are used to indicate survey data collected from the different USGS field offices: the Poway, CA Field Office (+ symbols),
representing data from San Diego, Riverside, Orange, San Bernardino and Los Angeles counties, and the Santa Maria Field Office (x symbols), covering stations in the northwestern portion of the study region including Santa Barbara and Ventura counties. Given the similarity in values, the data from the two regional offices was treated as a single source.

It is noted that the USGS surveys covered streams with larger drainage basins, with to an upper limit of 2000km$^2$ imposed for this analysis (thus the largest drainage area, for the location on the Cuyama River, was excluded). The USGS surveys were conducted on streams with drainage area larger than 10km$^2$, whereas nearly one-half of the 2005 surveyed streams were below this size. As with the 2005 survey data, if the multiple estimates from a given site were available, these estimates were averaged prior to determining the parameters of the regional curve relationship. The regional curve is shown in Figure 3.6 by the solid line with 95% confidence bounds shown by the dashed lines and the relationship for each bankfull geometry property given on the graphs. The reference relationship of Leopold is also shown by the thick gray line. The correlation coefficients are slightly higher based on this data: 0.49 for hydraulic depth, 0.715 for bankfull width, and 0.725 for bankfull cross-section area. Relative to the 2005 survey data, the exponent for cross-sectional area is approximately the same, while the exponent for top width is lower and for hydraulic depth is higher. These differences in exponents suggest a steeper change in hydraulic depth with area and milder change in bankfull width with area relative to the 2005 survey-based relationship.

Figure 3.7 plots the bankfull geometry estimates from both the 2005 field survey (square symbols) and USGS surveys (x symbols). This figure includes only one
estimate for each site (i.e., the average is used for sites with multiple estimates). There is a continuum of data across the range of drainage areas from 1 to ~2000km² with a fair amount of overlap in the middle range of drainage areas. The regional relationships for bankfull geometry determined from this combined dataset are:

\[ B_b = 2.961 \ A^{0.338} \]  \hspace{2cm} (7)
\[ D_b = 0.1956 \ A^{0.219} \]  \hspace{2cm} (8)
\[ A_{x_{sb}} = 0.5939 \ A^{0.551} \]  \hspace{2cm} (9)

Given that the cross-sectional area is defined by the product of the top width and hydraulic depth, one expects the exponents of the top width and hydraulic depth relationships to sum to the exponent of the cross-sectional area relationship, apart from numerical inaccuracies, as observed in the above equations. The thick lines in Figure 3.7 show the regional curves fit to this set of combined data, with the 95% confidence bounds of the regression shown by the dashed lines. The correlation values of 0.77, 0.505, and 0.735 for bankfull width, depth, and cross-sectional area respectively, are improved over the relationships for the individual datasets (with exception of cross-sectional area from the USGS survey data, with its correlation of 0.75). Table 3.4 summarizes the regional curve relationships determined from this analysis.

### 3.3.2 Comparison with independent surveys

In this section, the regional curves and data from the combined results as presented in Figure 3.7 are compared with the bankfull geometry estimates from
independent surveys. Figure 3.8 plots the data from Azary (1999, open circles) with the combined survey data (gray cross symbol, +). The Azary data falls well within the survey data and is within the 95% confidence bounds of the regional curve derived from the combined survey data (shown by the dashed red lines). A regional curve is fit using only the Azary data, shown by the solid black line and equation. It is noted that the Azary data included the mean cross-sectional depth at bankfull elevation, whereas the hydraulic depth was computed for the 2005 survey and USGS data. Although similar, these values would only be equal, and thus directly comparable, for a rectangular cross-section. The Avary regression yields lower values for bankfull width and cross-section area over the range of drainage areas of the Azary data. For each bankfull property, the exponent of regional curve from the Azary data is higher than that from the combined survey regional curves. This would result in potentially large deviations if the Azary regional curves were extrapolated over the range of drainage areas covered by the survey data, with lower estimates at smaller drainage areas and larger estimates at larger areas than estimated by the survey data. The regional curve regression relationships based on the Azary data are also summarized in Table 3.4.

Figure 3.9 shows the average bankfull geometry estimates from the EPA WSA data for southern California. To be commensurate with the surveyed data, sites extracted from the WSA database were limited to locations with drainage areas up to 2000km². The largest drainage area meeting this criterion is much smaller. The range of drainage area for the selected 54 EPA WAS sites was from 2 to 300 km². The average bankfull width and mean depth computed from the multiple cross-sections at each site are shown in the figure by the circle symbols. The cross-sectional area is
approximated as the product of the width and mean depth \((Ax_{b0} = B_{b0} * D_{m0})\), which would be applicable for rectangular channel cross-sections. The WSA provides more data and thus there appears to be higher variability than the previous results. The regression equation fit to the WSA data is given for each parameter as shown by the solid black line (and included in Table 3.4). The regression correlation coefficients are lower than the combined survey results (Figure 3.7) with values of 0.67, 0.27, and 0.54 for bankfull width, mean depth and cross-sectional area respectively. The regression exponent values are all smaller than those found for the combined survey data of Figure 3.7. This implies a flatter slope of the regional curve in the log-log plots, and less change with larger drainage area differences than predicted by the combined survey data. The gray dashed lines show the 95% confidence bounds of the combined survey regional curve of Figure 3.7. The bankfull depth regressions show the most significant agreement in values, even though the survey data uses hydraulic depth and WSA data uses mean depth. The WSA shows a flatter slope for depth and weaker correlation. Estimates of the bankfull width (and consequently, cross-sectional area) from the WSA are lower than the combined survey regional curve. However, with the exception of a few locations, the estimates of bankfull width and cross-sectional area fall within the 95% confidence bounds of the combined survey regional curves.

Faustini et al (2009b) also present a relationship for bankfull top width as a function of drainage area for southern California mountains. The regression curve equations from this work and Faustini et al are:

\[
B_{b0} \text{(herein)} = 2.06*A^{0.27}
\]  

(10)
\[ B_b \text{ (Faustini)} = 2.17 * A^{0.24} \]  \hspace{1cm} (11)

The parameters of the regression relationships are in fairly good agreement, confirming the lower value of the exponent for bankfull width regression determined for the EPA WSA data as compared to the combined survey regional regression. Both relationships for bankfull width (Eq. 10 and 11) report correlation coefficients of approximately 0.67 \((r^2 = 0.45)\). Faustini et al (2009b) note that 37 stations in southern California are used, in comparison with the 54 stations used herein. Faustini et al also report a relationship for bankfull width using data from 171 locations in California (see Table 3.4). This relationship is similar to the southern California regression for WSA data presented in Figure 3.9, but with a higher coefficient compared to Eq. 10 which would yield larger bankfull width values from the California relationship.

Table 3.4 also includes regional hydraulic geometry curve relationships for other locations. Included are relationships reported by Montgomery and Gran (2001) for the Yuba River in the Sierra-Nevada mountains of northern California, the Pacific Northwest and Pacific Northwest maritime streams reported by Castro and Jackson (2001), and the semi-arid region of central Arizona and New Mexico as reported by Moody et al (2003). The exponent values for bankfull width reported for these locations are higher than determined and reported for the EPA-WSA data and generally closer in value to the relationship determined herein for southern California streams (Eq. 7). The exponent values for bankfull width are higher for the Yuba River in northern California and for the Pacific Northwest, where exponents for bankfull mean depth and cross-sectional area were also higher than determined herein. This implies a
greater change in cross-section as drainage area increases. However, it is noted that the
data of Castro and Jackson included larger streams than those considered in the
southern California analysis. The regional relationships determined here are most
similar to those reported by Moody et al for Central Arizona.

3.4 Hydrologic Applications of Southern California Regional Curves

Bankfull geometry regional curves may be used for the estimation of bankfull
properties at un-surveyed locations, assuming the sites fall within the same hydro-
physiographical region and the developed relationships apply throughout. In this
section, the regional relationships developed for southern California based on the
combined survey results (i.e., Eq. 7-9) are applied in three analyses:
(a) examining the spatial variation in bankfull geometry estimates and other derived
properties;
(b) estimating the frequency of occurrence of bankfull discharge for selected locations
with gauged streamflow records;
and (c) determining a surface runoff response index for reaching bankfull discharge.

3.4.1 Spatial variation of bankfull geometry and derived properties

The interest in this section is the spatial variation of channel bankfull geometry
and other derived properties for small drainage basins with natural stream channels in
the mountainous region of southern California. As a first step, small drainage basins
were delineated in the region through GIS processing of digital elevation data. The GIS
processing was performed using the watershed delineation routine of GRASS (Neteler et
al, 2008), with a target drainage basin size of approximately 30km$^2$. Input to the delineation processing was the SRTM digital elevation data with 90m resolution (Jarvis et al, 2008). Although other higher resolution digital terrain elevation data is available from the USGS at 30m and 10m resolution (Gesch et al, 2009), small scale comparisons showed that the 90m resolution data resulted in basin delineations comparable to those using higher resolution data at the targeted basin scale of interest while requiring significantly less processing time. Additionally, the 10m resolution data did not extend beyond the southern border of the United States, which would have impacted the southern reach of the study area.

Southern California is home to several large urban areas, including the greater Los Angeles, San Bernardino/Riverside, and San Diego areas. Many streams have been channelized in these urban areas and thus are not controlled by the geomorphologic processes that define the bankfull channel dimension. Thus, these major areas were excluded from analysis as much as possible while maintaining the hydrologic connectivity of the stream networks. The TIGER urban area GIS layer (US Census Bureau, 2010) was plotted over the delineation results in order to identify and exclude basins with significant urban areas. Figure 3.10 depicts the resulting 975 drainage basins (multi-color shaping). A total of 975 subbasins are depicted. A few subbasins have accumulated drainage area greater than the 2000km$^2$ limit used in the regional curve analysis for small streams. These subbasins fall along the downstream reaches of the Santa Clara River in Ventura County, and Santa Ynez River in Santa Barbara County. The average subbasin size was 26 km$^2$, with an average accumulated drainage area of 207 km$^2$. The lower panel of Figure 3.10 shows drainage basin characteristics...
computed from the GIS delineation output, divided into properties of individual subbasins (left column) and accumulated properties along the drainage networks (right column), limiting the accumulated drainage area to 3000km$^2$ or less.

The bankfull geometry regional curves of Equations 7-9 were applied to this set of basins. The estimates of bankfull top width and hydraulic depth are shown in the top panels of Figure 3.11. Since the bankfull geometry properties are only functions of drainage area, these properties show similar spatial patterns with increasing property values (darker colors) downstream as drainage area increases. Bankfull width estimates range from 7.2 to 44.5 m, while hydraulic depth estimates range from 0.35 to 1.15 m.

In addition to the bankfull geometry properties, it was desired to estimate the bankfull discharge for each of these small basins. Discharge can be estimated using Manning’s steady uniform flow equation (e.g., Maidment, 1993; Bras, 1990):

$$Q = \frac{1.0}{n} A_{xs} R^{2/3} S^{1/2} \tag{12}$$

where $A_{xs}$ is cross-sectional area in m$^2$, $R$ is the hydraulic radius (and equal to the cross-sectional area divided by wetted perimeter, $A_{xs}/P$) in m, $S$ is the channel slope (dimensionless), and $n$ is Manning’s roughness factor. Assuming a nearly rectangular cross-sectional shape with the width much larger than depth, the hydraulic radius may be approximated by the hydraulic depth of the channel. Applying this assumption for bankfull conditions, Eq. 12 can be written:
and the bankfull discharge may be estimated based on the channel geometry estimates of width \( B_b \) and hydraulic depth \( D_b \), the channel slope determined from the basin delineation results, and an estimate of the channel roughness.

Manning’s roughness factor is generally considered a function of channel bed material and size, and varying with flow depth. Cited roughness values for mountain streams are 0.04 to 0.07 (e.g., Maidment, 1993) or up to 0.1 for natural streams with heavy debris (Chow et al, 1988). Chow (1959) presents roughness values up to 0.15. Carpenter et al (1999) presented an adaptation of a relationship derived by Jarrett (1985) for the variation of roughness with local stream slope and flow depth for steep sloping streams:

\[
Q_{bf} = \frac{10}{n} B_b D_b^{5/3} S_b^{1/2}
\]  

where \( S_c \) is the local slope (m/m) and \( D \) is the hydraulic depth (m). Jarrett’s original relationship was related roughness to local slope and hydraulic radius and was developed with data from steep gradient streams in Colorado. The relationship given by Carpenter et al (1999) utilized the hydraulic depth instead of hydraulic radius. The Colorado data were from streams with slopes that ranged from 0.002 to 0.034, hydraulic radii between 0.15 and 1.68 m, and hydraulic depth between 0.15 and 2.0 m. The GIS properties (Figure 3.10) show that the mean slope for this study region (0.04 for local

\[
n = 0.43 S_c^{0.37} D^{-0.15}
\]  

where \( S_c \) is the local slope (m/m) and \( D \) is the hydraulic depth (m). Jarrett’s original relationship was related roughness to local slope and hydraulic radius and was developed with data from steep gradient streams in Colorado. The relationship given by Carpenter et al (1999) utilized the hydraulic depth instead of hydraulic radius. The Colorado data were from streams with slopes that ranged from 0.002 to 0.034, hydraulic radii between 0.15 and 1.68 m, and hydraulic depth between 0.15 and 2.0 m. The GIS properties (Figure 3.10) show that the mean slope for this study region (0.04 for local
slope) is slightly higher than observed in Jarrett’s study, while the estimated hydraulic
depth values are within the range of Jarrett’s data.

Equation 14 was used to estimate roughness for the basins of southern
California, and the spatial variation is shown in Figure 3.11 (lower left panel).
Roughness values range from 0.05 to 0.29, with higher values along the Transverse
Mountains than the southern portion of the Peninsular Mountains in Riverside and San
Diego counties, and the western subbasins in Santa Barbara County. The figure
indicates the influence of steepness on the estimates of roughness, as the highest
roughness values are fall along the high mountain slopes. However, the computed
values are higher than the generally cited range (0.04-0.1), and are higher than Jarrett’s
peak roughness value of 0.159. This is likely the result of the high slope values
determined by the GIS for the small subbasins. Other properties being equal, higher
values of roughness will produce more conservative estimates of bankfull discharge.
These roughness values were used with Manning’s steady flow equation (Eq. 13) to
produce bankfull discharge estimates for each subbasin. The bankfull discharge
estimates are shown in Figure 3.11d. Eliminating basins with accumulated drainage
area greater than 3000km², the bankfull discharge values range from 1.3 m³/s to nearly
45 m³/s. The influence of drainage area on the channel capacity is again apparent in the
spatial distribution of values, as the bankfull discharge values increase as one continues
downstream and the basins along the largest rivers stand out in darker shading,
including the Santa Clara River in Ventura County and lower reaches of the Santa Ynez
River in Santa Barbara County.
If one considers the influence of roughness on the computation of discharge, it is clear from Eq. 13 that an increase in roughness yields a decrease in the estimated discharge. If the estimation of roughness using the Carpenter et al (1999) formula of Eq. 14 indeed results in an over-estimation of roughness, one could reduce the roughness estimates to produce values which fall within the cited range for mountainous streams. The average of roughness values computed for southern California is 0.15. Applying a 50% reduction in the estimated roughness values would maintain the spatial variability of roughness values while yielding a range of 0.025 to 0.15 and mean value of 0.075. The resulting change in estimated bankfull discharge would be doubling of values as shown in Figure 3.12. This plot shows an exponential relationship between bankfull flow and drainage area as this plots as a straight line with log-log axes. The variability in estimated bankfull discharge at a given catchment scale results from variation in stream slope. The impact of this change of roughness on the frequency of bankfull flow occurrence is explored further in Section 3.4.2.

To examine regional variation in the bankfull flow, the estimates for basins along the Transverse and Peninsular Mountain ranges were compared. The basins were divided at the headwaters of the Santa Ana River in San Bernardino County, with these headwater basins and those westward making up the Transverse Range basins and those basins to the south making up the Peninsular Range basins. Figure 3.13 presents the cumulative frequency distributions of bankfull discharge estimated for basins of the two regions. The distributions show differences primarily in the tails, e.g., for low frequency (< 0.2) and high frequency (>0.9). The Kolmogorov-Smirnov (K-S) test considers the maximum distance between the two distributions and number of data
points used to assess whether the two sample distributions are likely to be drawn from the same underlying distribution (e.g., Wilks, 2006). The K-S test for bankfull discharge among the two mountain regions indicates that they are, in fact, statistically different at the 90% confidence level.

3.4.2 Frequency of bankfull flow exceedence

Bankfull discharge is often referred to as having a returned period of 1.5 (Leopold, 1994). The return frequency is traditionally determined by (a) conducting a field survey at a gauging station to estimate the bankfull stage, (b) use the stage-discharge rating relationship at the gauge site to determine the discharge associated with bankfull stage, and (c) then conducting an annual peak flood frequency analysis, interpolating the return period of the bankfull discharge from the flood-frequency curve (Maidment, 1993). Here, an alternative approach is undertaken to examine the frequency of exceeding bankfull discharge, which was estimated using the bankfull geometry regional curves to compute the channel cross-section properties and Manning’s steady flow formula as presented in Section 3.4.1. As opposed to annual flow maxima frequency analysis, this approach follows a flow-over-threshold frequency analysis. A set of instantaneous flow records were obtained for 20 USGS gauging stations for the period October 1990 - September 2005, with one station eliminated due excessive missing data. For these sites, either cross-section surveys were not obtained or reference to gauge height was not available in the survey notes to link the survey measurement to the rating curve. The stations, USGS station numbers, drainage areas, and estimated bankfull discharges are listed in Table 3.5. The drainage area upstream
of the 19 stream gauging stations ranged from 23 to 347 km². The channel slope required by Manning’s formula was derived from the GIS delineation results for the basin with outlet nearest the gauging station, and varied from 0.022 to 0.1 m/m. Bankfull discharge estimates range from 2.6 m³/s for the East Twin Creek basin to 16 m³/s for the largest basin, Temecula Creek.

Using these estimates of bankfull discharge, a count of events exceeding bankfull flow was made through comparison with the instantaneous record. Although mean daily flow records are available for an extended period of record, the comparison to bankfull flow is appropriately made with instantaneous flow data. For small, fast responding basins such as those in southern California, peak flow in a day may be much higher than the daily average, and thus mean daily flows may not capture the occurrence of bankfull flow within a day. For the purposes of this analysis, the initiation of a bankfull flow event was marked when the flow exceeded bankfull discharge at the site, and ended then the flow receded below bankfull discharge. Such an event was considered one occurrence, regardless of the duration of the event (whether a few hours or much longer). If multiple peaks occurred without the flow falling below bankfull, this was considered one event. Additionally, if several peaks occurred in short succession, these were treated as a single event as they are likely driven by “waves” of precipitation falling within the same synoptic scale system. For some streams, the instantaneous data showed that multiple peaks can occur and return below bankfull flow, even within a day or two. To eliminate such multiple peaks, a minimum inter-event arrival period of 5 days was enforced as representative of synoptic scale atmospheric forcing. This approach corresponds to the partial duration series
frequency analysis and, as such, allows for multiple occurrence of bankfull discharge within a year (in contrast to annual series frequency analysis, which considers only the largest flow occurrence each year).

With this definition of and constraint on bankfull event occurrence, the behavior of stream flow in southern California was such that it was typical to observe multiple occurrences of bankfull flow during a particular year(s), and no bankfull occurrences during other years or extended periods. The count of events during the period of record is also given in Table 3.5. The average return interval was computed by dividing the number of years of non-missing data by the total number of occurrences. Interestingly, at about ½ of the locations, the average return interval is less than 1 year. Borrego Canyon Creek, on the leeward side of the Peninsular Range with respect to the Pacific Ocean, did not yield any occurrences of bankfull discharge within 9 years of record. Thus, the estimated bankfull discharge at Borrego Canyon Creek is suspected of being too high for this station. The other two basins on the leeward side (Deep Creek and Big Rock Creek) are the basins with average return intervals of 3.3 and 1.1 years.

Arroyo Trabuco, located in coastal Orange County, yielded a very large number of occurrences. It was typical of this station’s record to exceed the estimated bankfull discharge more than once a month and in successive months during high flow periods yielding the lowest return period estimate of 0.15 year. This lower return interval is in contrast with the adjacent San Juan Creek watershed, which had a return period nearly three times greater at 0.4 year. The high number of occurrences at Arroyo Trabuco suggests that the bankfull discharge may be under-estimated.
Excluding these two basins, the range in return interval was 0.33 to 3.3 years, with an average value of 1 year. With the exception of East Twin Creek, the stations located in the Transverse Mountain Range (USGS station numbers 11058500 through 1112850) have average return periods less than one year, while stations along the Peninsular Range have a wider range of values. The Peninsular Range locations with average return periods less than 1 year are located at lower elevations in the foothills of the mountains and near the coast (USGS station numbers 11022200 and 11044350-11047300). The remaining four locations along the Peninsular Range yield average return periods greater than 1 year.

Magdych and Moore (2005) considered the recurrence frequency of ordinary high water marks (OHWM) along five streams in southern California. They state that OHWM may be similar to the bankfull stage, although the given definition of OHWM as the level of normal fluctuations of streamflow within the main channel, implies this level may be lower than bankfull. OHWM may be identified by similar physical/geomorphologic characteristics as bankfull stage including sedimentary and vegetation changes. They estimate the recurrence intervals of these streams to be in the range between 0.4 and 0.8 year. This is consistent with the findings of this study, particularly if the OHWM is lower than bankfull stage.

To explore the sensitivity of the bankfull flow occurrence frequencies estimated here, especially under the uncertainty in the estimation of Manning’s roughness factor, the computations of the average return intervals were repeated under two cases: (a) under a 30% reduction in roughness, or a 42% increase in discharge; and (b) under a 50% reduction in roughness, or doubling of bankfull discharge. Table 3.6 presents the
average return interval computed under the three cases. Excluding the noted extreme cases of Borrego Canyon Creek and Arroyo Trabuco, the case of a 30% decrease in roughness produces average return period values between 0.38 and 3.3 years, and the case of a 50% decrease in roughness yields average return period values between 0.42 and 4.3 years. The average return interval at Deep Creek (USGS station number 10259200) is invariant with the change in Manning’s roughness, while the remaining stations show an increase in the average return interval as the roughness value is reduced. With the 50% decrease in roughness, several of the streams along the Transverse Range show average return values greater or equal to 1 year. These results suggest that the computation of bankfull flow occurrence frequency is sensitive to the uncertainty in estimation of bankfull discharge.

3.4.3 Surface response index

Under the assumption that the runoff response of watersheds is linear with respect to input precipitation forcing when the watershed is at or near saturation conditions, it is possible to define a characteristic surface response index that represents the surface runoff volume per unit catchment area required to bring the channel at the catchment outlet to the bankfull discharge level. This follows the threshold runoff theory presented by Carpenter et al (1999) and which has been used in operational flash flood warning (e.g., Georgakakos, 2006; Ntelekos et al, 2006; Villarini et al, 2010). Under the given assumption, the catchment unit hydrograph may be defined to relate the runoff response of the catchment to a unit input of effective rainfall (e.g., after infiltration and evapotranspiration abstractions have been applied) of a given duration
and which is uniformly distributed over the catchment. If catchment response is
approximately linear with respective to effective rainfall input, then the peak response
of the catchment is given as:

\[ Q_p = q_{pR} A I \]  \tag{15}  

where \( Q_p \) represents the flow at the catchment outlet ([m\(^3\)/s]), \( q_{pR} \) is the unit hydrograph
peak response ([m\(^3\)/s/mm/km\(^2\)]), and \( I \) is the effective rainfall ([mm]) of a given
duration, \( t_R \), applied over the catchment area, \( A \) ([km\(^2\)]. Effective rainfall is defined as
the amount of rainfall that contributes directly to catchment surface runoff, and does not
include the rainfall that is intercepted, infiltrated into the soil, or evaporated from the
soil or through the plants.

By setting \( Q_p \) equal to the bankfull discharge, then \( I \) represents the desired
surface runoff response that is just sufficient to yield bankfull conditions at the
catchment outlet, and is a characteristic of the watershed for a given duration of
effective rainfall. Bankfull discharge is defined using the Manning’s steady uniform
flow formulation of Eq. 13, using the bankfull geometry regional curves (i.e., as shown
in the lower right panel of Figure 3.11).

The geomorphologic unit hydrograph (Rodriguez-Iturbe and Valdez, 1979) is
used to define the catchment unit hydrograph response. This is preferred as the
geomorphologic formulation uses catchment and channel characteristics, and eliminates
empirical coefficients inherent to traditional unit hydrograph approaches. The original
formulation (Rodrigues-Iturbe and Valdes, 1979) expressed the peak magnitude and time to peak of the instantaneous unit hydrograph in terms of the channel length, catchment velocity, and geomorphologic structure. Rodriguez-Iturbe et al (1982) converted the expressions to give the peak response \((Q_p)\) as:

\[
Q_p = \left( \frac{2.42_i A t_R}{\Pi^{0.4}} \right) \left[ 1 - \frac{0.218 t_R}{\Pi^{0.4}} q_{pR} \right]
\]  (15)

where \(i\) is the effective rainfall intensity (cm/hr), \(A\) is the catchment area, \(t_R\) is the duration of effective rainfall, and the parameter \(\Pi\) is a function of catchment properties:

\[
\Pi = \frac{L^{2.5}}{i A R L \alpha^{1.5}}
\]  (16)

with \(L\) equal to the main channel stream length, \(R_l\) representing Horton’s length ratio, and \(\alpha\) defining a channel shape parameter:

\[
\alpha = \frac{1}{n B^{2/3}}
\]  (17)

The channel shape parameter is evaluated for bankfull conditions (i.e., \(B=B_b\)). Recognizing that the effective rainfall intensity multiplied, \(i\), by the effective rainfall duration, \(t_R\), equals total effective rainfall \((I)\), this may be substituted into Eq. 15-16 to express the unit hydrograph peak response as a function of the surface runoff response
The solution of the surface runoff response index is found through the non-linear solution of Eq. 15 when bankfull discharge as estimated by Manning’s formula is substituted as $Q_p$.

Figure 3.14 shows surface runoff response indices computed for a 3-hour rainfall duration for the southern California region. The values of this index indicate that effective rainfall in the range of 2 to 6 mm in a 3-hour period is necessary to produce bankfull flows in the small streams of the region, with a mean value of 3.6 mm and standard deviation of 0.64 mm. These values are lower than estimated in Northern California by Carpenter et al (1999), although also showing a similar narrow distribution. The figure suggests higher values along the Transverse Mountains streams than the Peninsular Mountains streams. This is confirmed in Figure 3.15, which shows the cumulative frequency distributions of the 3-hour surface runoff response index for subbasins within the Transverse and Peninsular Ranges. Indeed a shift in the distributions of surface runoff response is observed for the two sub-regions with higher values for the Transverse Range at a range of frequency levels. A K-S test was also formulated for this index and indicates that the two sample distributions are statistically different at the 90% confidence level.

The underlying theory for the development of the surface runoff response index inherently imposes limitations on the application of this index. The assumption that runoff responds linearly to rainfall excess under saturated conditions, i.e., that unit hydrograph theory is applicable, limits the size of catchments. Very small catchments are more non-linear than larger ones (e.g., Wang et al, 1981), and unit hydrograph theory is generally more applicable at higher flows than lower flow in small catchments.
(Mesa and Gupta, 1987). Also, the assumption of uniform distribution of effective rainfall also limits the size of catchments for which the unit hydrograph approach is reasonable, as spatial non-uniformity of rainfall grows as area increases. Thus, in this application, the focus has been on catchments with accumulated drainage area less than 2000km$^2$ and with the smallest catchment size limited to 13km$^2$. The use of the bankfull discharge as the target flow of Eq. 15 assumes that the channels in the region of application are natural or non-channelized so the geomorphologic processes that allow for the development of regional hydraulic geometry curves governs channel size. This application also assumes that the regional curves developed in Section 3.3.1 from a relatively few surveyed sites are applicable to all catchments.

Ntelekos et al (2006) identify two major sources of uncertainty in the surface response index calculation: the first is GIS-derived catchment scale properties and the second is the regional hydraulic geometry relationships. They conclude that the index is particularly sensitive to the regional hydraulic geometry relationships for bankfull top width and hydraulic depth, drainage area, and stream length, with greater effect at smaller spatial scales. An increase in uncertainty with smaller spatial scale has also been noted in hydrologic simulations by the authors (Carpenter and Georgakakos, 2004). Research on such threshold indices continues to improve their use in operational flash flood forecasting for ungauged basins (e.g., Reed et al, 2007; Norbiato et al, 2009), and considering additional uncertainties including hydrologic model parameters (Ntelekos et al, 2006) and rainfall estimates (Villarini et al. 2010).

3.5 Conclusions
This chapter described the development of regional relationships for bankfull channel geometry characteristics for streams in the southern California mountainous region. Data is drawn from field surveys, covering 31 locations, and from USGS survey reports, covering 29 stream gauge locations. The regional bankfull geometry curves define the variation of bankfull width, hydraulic depth, and cross-sectional area with catchment drainage area. The correlation coefficients for the developed regressions were lowest for hydraulic depth (0.5) and highest for channel width (0.77). The assessment of these regression relationships suggests agreement among the relationships developed herein and those from independent data sources and reported in available literature.

Using the developed regional relationships, estimates of bankfull channel properties were made for small watersheds throughout the southern California region ranging in accumulated drainage area from 13 to 3000km². Bankfull channel width estimates were in the range from 7.2 to 44.5 m, while hydraulic depth estimates ranged from 0.35 to 1.15 m. The spatial distribution of the properties reflected the dependence on drainage area, with the largest values occurring along the largest drainage basins of the Santa Ynez River in Santa Barbara County and Santa Clara River in Ventura County. The estimated cross-sectional properties were then used to compute bankfull discharge at each basin utilizing Manning’s steady uniform flow formula. Manning’s roughness coefficient was estimated as a function of hydraulic depth and local channel slope, yielding a range in values of 0.05 to 0.29 which are considered high relative to commonly cited roughness values for mountainous regions. The spatial distribution of these values reveals that the highest roughness values were computed for basins along
the steep mountain slopes, and lower values along the flatter downstream reaches. The computed bankfull discharge values ranged from 1.3 to 74 m$^3$/s. The spatial distribution of computed bankfull discharge largely reflects the dependence on drainage area.

Using available instantaneous streamflow records for 19 stream gauging stations, the frequency of bankfull flow occurrence was examined. The streamflow records were available for different periods between 1990 and 2005. The occurrence of bankfull discharge events were identified by recorded flow exceeding the estimated bankfull discharge and subsequent recession below this level, with an imposed inter-event arrival period of 5 days to distinguish synoptic scale influences. The magnitude of the peak flow and the possible existence of multiple peaks within the event periods were not evaluated. This approach is consistent with partial duration series or flow-over-threshold frequency analysis and allows for more than one occurrence per year (unlike annual maxima frequency analysis). The average occurrence frequency was represented by the average return interval, computed by dividing the number of years of non-missing data by the number of occurrences.

The average return interval ranged from 0.15 to 3.3 years, with an average value across locations of 1 year. This suggests that on average bankfull flows may occur more frequently for southern California streams than the often cited recurrence period of 1.5 years. Uncertainty in the estimation of Manning’s roughness coefficient in deriving bankfull discharge was accounted for by considering reductions in the Manning’s values. A reduction in roughness of 50% produce a small increase in the average return interval, with the range in return interval values of 0.2 to 3.3 years. Given the relatively short period of record and often infrequent occurrence of bankfull
flow, the change in average return interval may be due to just a few events that are not captured as the Manning’s roughness coefficient is reduced.

The regional bankfull geometry curves developed for southern California were also employed to estimate an index of the surface runoff response required to produce bankfull flows at the outlet of small watershed throughout the region. The surface runoff response index represents the amount of effective rainfall necessary to produce bankfull flows. Surface runoff response index values ranged from 2 to 6 mm per 3-hour rainfall duration, with somewhat higher values determined for catchments along the Transverse Mountain Range relative to the Peninsular Range.

An extension of this work would be to utilize the derived surface runoff response together with high resolution precipitation estimates and hydrologic modeling of the land surface that accounts for soil moisture and evaporative losses to examine the historical occurrence of precipitation with exceeds the surface runoff response for catchments throughout the region. The objective of such research would be to quantify the historical occurrence frequency of exceeded bankfull conditions with spatial detail not available through observations alone.

Another possible extension of this work relates to the regionalization of bankfull geometry curves. There were relatively few surveyed sites with which the regional bankfull geometry curves were developed and thus limited data available to examine with-in region variation of the regional curves. A potential improvement to the regional curves would be to expand the available cross-sectional data and consider spatial variation in the curves based on additional features such as channel form, terrain characteristics (aspect or slope), geology, and land cover.
3.6 References


Harvey, A.M., 1969: Channel capacity and the adjustment of streams to hydrologic regime. *J. Hydrology*, 8, 82-98.


Table 3.1. Location and cross-sectional geometry estimates for 2005 field survey sites.

<table>
<thead>
<tr>
<th>Location</th>
<th>County</th>
<th>Area (km²)</th>
<th>NXS</th>
<th>B (m)</th>
<th>D (m)</th>
<th>A XS (m²)</th>
</tr>
</thead>
<tbody>
<tr>
<td>La Jolla Ind Res Creek</td>
<td>San Diego</td>
<td>0.5</td>
<td>1</td>
<td>2.7</td>
<td>0.17</td>
<td>0.5</td>
</tr>
<tr>
<td>Plaisted Creek</td>
<td>San Diego</td>
<td>7.1</td>
<td>1</td>
<td>2.2</td>
<td>0.08</td>
<td>0.2</td>
</tr>
<tr>
<td>Pauma Creek</td>
<td>San Diego</td>
<td>30.1</td>
<td>1</td>
<td>12.1</td>
<td>0.52</td>
<td>6.5</td>
</tr>
<tr>
<td>NF San Jacinto River</td>
<td>Riverside</td>
<td>14.2</td>
<td>2</td>
<td>6.5</td>
<td>0.43</td>
<td>4.0</td>
</tr>
<tr>
<td>Strawberry Creek</td>
<td>Riverside</td>
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<td>1</td>
<td>7.3</td>
<td>0.23</td>
<td>1.7</td>
</tr>
<tr>
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<td>San Bernardino</td>
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<td>1</td>
<td>18.0</td>
<td>0.35</td>
<td>6.3</td>
</tr>
<tr>
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<td>1</td>
<td>8.7</td>
<td>0.35</td>
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</tr>
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<td>1.1</td>
</tr>
<tr>
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<td>Los Angeles</td>
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<td>9.7</td>
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<td>26.7</td>
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<td>0.53</td>
<td>4.5</td>
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<td>Silverado Canyon</td>
<td>Orange</td>
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<td>13.8</td>
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<td>Ventura</td>
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<td>Ventura</td>
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<td>2</td>
<td>6.6</td>
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<td>Ventura</td>
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<td>0.37</td>
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<td>Ventura</td>
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<td>1</td>
<td>3.8</td>
<td>0.27</td>
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<td>Ventura</td>
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<td>1</td>
<td>7.9</td>
<td>0.44</td>
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<td>Ventura</td>
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<td>5.7</td>
<td>0.57</td>
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<td>Santa Barbara</td>
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<td>9.4</td>
<td>0.30</td>
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<td>Sulphur Creek</td>
<td>Santa Barbara</td>
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<td>1</td>
<td>2.6</td>
<td>0.18</td>
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</table>

A = drainage area; N XS = number of cross-sections used; B = bankfull top width; D = bankfull hydraulic depth; A XS = bankfull cross-sectional area.
Table 3.2. Location and cross-sectional geometry estimates for USGS sites.

<table>
<thead>
<tr>
<th>USGS ID</th>
<th>Stream Name</th>
<th>Area (km$^2$)</th>
<th>NXS</th>
<th>B (m)</th>
<th>D (m)</th>
<th>A$_{XS}$ (m$^2$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>11023340</td>
<td>Los Penasquitos Crk</td>
<td>109</td>
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<td>22.1</td>
<td>0.43</td>
<td>9.4</td>
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<tr>
<td>11044000</td>
<td>Santa Margarita R</td>
<td>1522</td>
<td>6</td>
<td>37.3</td>
<td>1.74</td>
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<td>Rainbow Creek</td>
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<td>14.8</td>
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<td>1.3</td>
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<td>13.8</td>
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<td>46.0</td>
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<td>6.0</td>
<td>0.66</td>
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<td>Santa Clara River</td>
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<td>11.2</td>
<td>0.86</td>
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<tr>
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<td>1.04</td>
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<td>20.5</td>
<td>0.99</td>
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<td>0.62</td>
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<td>3</td>
<td>4.6</td>
<td>0.22</td>
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<td>0.86</td>
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<td>0.23</td>
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<td>27.8</td>
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<td>Santa Cruz Creek</td>
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<td>5.0</td>
<td>0.36</td>
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<td>Salsipuedes Creek</td>
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<td>0.39</td>
<td>3.8</td>
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<td>San Antonio Creek</td>
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<td>5</td>
<td>14.4</td>
<td>0.97</td>
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<td>Sisquoc River</td>
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Table 3.3. Location and cross-sectional data from Azary, 1999.

<table>
<thead>
<tr>
<th>Location</th>
<th>Area, $A_{bf}$ (km$^2$)$^+$</th>
<th>Width, $W_{bf}$ (m)$^+$</th>
<th>Depth, $d_{bf}$ (m)$^+$</th>
<th>XS Area, $X_{A_{bf}}$ (m$^2$)$^+$</th>
</tr>
</thead>
<tbody>
<tr>
<td>EF Devils Creek</td>
<td>5.7</td>
<td>3.29</td>
<td>0.26</td>
<td>0.85</td>
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<td>0.14</td>
<td>0.38</td>
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<td>4.82</td>
<td>0.27</td>
<td>1.32</td>
</tr>
<tr>
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<td>7.32</td>
<td>0.36</td>
<td>2.61</td>
</tr>
<tr>
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<td>43.8</td>
<td>9.88</td>
<td>0.28</td>
<td>2.74</td>
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<tr>
<td>City Creek</td>
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<td>5.03</td>
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<td>2.19</td>
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<tr>
<td>Mill Creek</td>
<td>98.6</td>
<td>9.05</td>
<td>0.65</td>
<td>5.88</td>
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</table>

$^+$ original units have been converted to km$^2$, m, and m$^2$. 
Table 3.4. Summary of regional bankfull geometry regressions.

<table>
<thead>
<tr>
<th>Location/Source</th>
<th>Number of Sites</th>
<th>Regression*</th>
<th>R²</th>
</tr>
</thead>
<tbody>
<tr>
<td>Southern California, surveyed this paper</td>
<td>60</td>
<td>$B_b=2.961 , A^{0.538}$, $D_b=0.1956 , A^{0.219}$, $A_{xb}=0.5939 , A^{0.551}$</td>
<td>0.59</td>
</tr>
<tr>
<td>Southern California, Azary this paper</td>
<td>7</td>
<td>$B_b=1.794 , A^{0.366}$, $D_m=0.099 , A^{0.371}$, $A_{xb}=0.169 , A^{0.749}$</td>
<td>0.74</td>
</tr>
<tr>
<td>Southern California, EPA WSA this paper</td>
<td>54</td>
<td>$B_b=2.058 , A^{0.277}$, $D_m=0.255 , A^{0.117}$, $A_{xb}=0.525 , A^{0.387}$</td>
<td>0.45</td>
</tr>
<tr>
<td>Southern California Faustini et al (2009b)</td>
<td>37</td>
<td>$B_b=2.17 , A^{0.24}$</td>
<td>0.45</td>
</tr>
<tr>
<td>California Faustini et al (2009b)</td>
<td>171</td>
<td>$B_b=2.49 , A^{0.277}$</td>
<td>0.47</td>
</tr>
<tr>
<td>Yuba River, California Montgomery and Gran (2001)</td>
<td>12 – alluvial</td>
<td>$B_b=0.405 , A^{0.377}$, $B_b=0.458 , A^{0.45}$</td>
<td>0.93</td>
</tr>
<tr>
<td>Yuba River, California Montgomery and Gran (2001)</td>
<td>12 - bedrock</td>
<td></td>
<td>0.60</td>
</tr>
<tr>
<td>Pacific Northwest Castro and Jackson (2001)</td>
<td>76</td>
<td>$B_b=3.00 , A^{0.38}$, $D_m=0.307 , A^{0.24}$, $A_{xb}=0.751 , A^{0.643}$</td>
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</tr>
<tr>
<td>Pacific Maritime Mountain streams Castro and Jackson (2001)</td>
<td>22</td>
<td>$B_b=3.08 , A^{0.3}$, $D_m=0.167 , A^{0.39}$, $A_{xb}=0.932 , A^{0.739}$</td>
<td>0.59</td>
</tr>
<tr>
<td>Central Arizona Moody et al. (2003)</td>
<td>53</td>
<td>$B_b=4.13 , A^{0.317}$, $D_m=0.215 , A^{0.219}$, $A_{xb}=0.860 , A^{0.540}$</td>
<td>0.82</td>
</tr>
<tr>
<td>Eastern Arizona/New Mexico Moody et al. (2003)</td>
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<td>$B_b=2.64 , A^{0.28}$, $D_m=0.127 , A^{0.237}$, $A_{xb}=0.348 , A^{0.512}$</td>
<td>0.80</td>
</tr>
</tbody>
</table>

+ reported relationships converted if necessary to have following units: $B_b$ and $D_b$ in meters, $A_{xb}$ in m$^2$, and $A$ in km$^2$. 


### Table 3.5. Estimated bankfull flow and occurrence frequency for selected USGS gauging stations.

<table>
<thead>
<tr>
<th>USGS ID</th>
<th>Stream Name</th>
<th>Area (km²)</th>
<th>$Q_{bf}$ (m³/s)</th>
<th>Period of Record</th>
<th>$N^+$</th>
<th>Average Return Period *</th>
</tr>
</thead>
<tbody>
<tr>
<td>10255810</td>
<td>Borrego Canyon</td>
<td>56.4</td>
<td>4.9</td>
<td>1995-2003</td>
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<tr>
<td>10259200</td>
<td>Deep Creek</td>
<td>79.2</td>
<td>6.6</td>
<td>1995-2005</td>
<td>3</td>
<td>3.3</td>
</tr>
<tr>
<td>10263500</td>
<td>Big Rock Creek</td>
<td>59.3</td>
<td>5.3</td>
<td>1995-2005</td>
<td>9</td>
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</tr>
<tr>
<td>11012500</td>
<td>Campo Creek</td>
<td>220.1</td>
<td>11.4</td>
<td>1996-2005</td>
<td>6</td>
<td>2.2</td>
</tr>
<tr>
<td>11015000</td>
<td>Sweetwater River</td>
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<td>7.5</td>
<td>1995-2005</td>
<td>9</td>
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</tr>
<tr>
<td>11022200</td>
<td>Los Coches Creek</td>
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<tr>
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<td>Santa Maria Creek</td>
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<td>8.6</td>
<td>1995-2005</td>
<td>8</td>
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<td>Temecula Creek</td>
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<td>14</td>
<td>1.1</td>
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<td>1991-2005</td>
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<td>Alamo Pintado Crk</td>
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<td>5.6</td>
<td>1991-2005</td>
<td>14</td>
<td>0.86</td>
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</table>

$^+$ $N$ = Number of Occurrences  
*Average return period is computed as number of years with non-missing records divided by the number of bankfull discharge events.
Table 3.6. Sensitivity of average return period calculation to uncertainty in Manning’s N.

<table>
<thead>
<tr>
<th>USGS ID</th>
<th>Stream Name</th>
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<th>Avg Return Period, N’=0.7N</th>
<th>Avg Return Period, N’=0.5N</th>
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4.1 Introduction

Orographically-driven precipitation is of interest over a wide range of scales. On long time scales, this interaction between the land surface and the atmosphere sets the climatology of mountainous regions and has influenced natural ecosystem development, and in many cases, human water resources development. On shorter time scales, this interaction may lead to intense, localized precipitation causing other natural hazards including flash floods, landslides, and debris flows. The mountainous regions of southern California are certainly no exception to the occurrence of such natural hazards. In January 2005, a series of heavy rainfall events preceded a landslide at La Conchita, CA, which damaged or destroyed 36 houses and killed 10 people (Jibson, 2005). Similarly, heavy precipitation in the San Bernardino Mountains of San Bernardino County, CA, occurred on December 25, 2003 and lead to several mudslides and debris flows in a region that had recently burned in October 2003. The resulting mudslide in the Waterman Canyon region killed 13 people (Douglas, 2004).

The objective of this chapter is develop methods for estimating orographic precipitation at fine spatial and temporal scales over the southern California coastal-to-mountainous region in order to understand the dominant spatio-temporal characteristics of the regional precipitation processes. The focus at finer spatio-temporal scales reflects interest in the associated potential for flash flood occurrence and to facilitate modeling studies of this occurrence. In this chapter, three methods for estimating
orographically-enhanced precipitation in the cismontane region of the Southern California Bight are compared over the historical period from October 1948 to April 2005. The methods are: (a) an interpolation of station observations; (b) a mesoscale numerical model of the atmosphere, and (c) a simplified numerical model for estimating topographically-driven precipitation adapted for the southern California region. This chapter discusses the three methods of precipitation estimation, with some emphasis on the formulation and sensitivities of the simplified orographic precipitation model for southern California.

The various precipitation estimation methods have different advantages and limits, including the level of physical processes represented in the models (low for the interpolation method, but high for the numerical modeling), computational demand (low for the interpolation, intermediate for the simplified model, and high for the numerical model), and sensitivity to the spatio-temporal characteristics of available observations (high for the interpolation). However, the combination of different methods offers a multiple model ensemble of long-term precipitation estimates over southern California with high spatial and temporal resolution. This multi-model ensemble represents a valuable tool for climate and hydrologic research for the region.

Orographic precipitation has received significant research attention (e.g., recent reviews by Smith, 2006; Roe, 2005; Barros and Lettenmaier, 1994). Some of these studies have aimed at understanding processes and characteristics (Minder et al, 2008; Roe and Baker, 2006; Lin et al, 2001; Alpert and Shafir, 1991), or developing predictive models (e.g., Smith and Barstad, 2004; Kuligowski and Barros, 1999; Andrieu et al, 1996; Sinclair, 1994), while other focus on hydrologic applications of orographic
precipitation (e.g., Georgakakos, 1987; Leung et al. 1996). In California, much of the research on orographic precipitation has focused on the Sierra Nevada mountain range of northern and central California. In addition to including the highest mountains in the State (exceeding 4400m at Mt. Whitney), the Sierra Nevada mountains intercept most of the precipitation that falls in the state and thus is residence for most of the state’s developed water resources. An understanding of orographically-enhanced precipitation in this region has been important for water resources management. Studies of the regional precipitation process in Northern California include those of Pandey et al., 2000; Dettinger et al. 2004; Reeves et al., 2008; Galewsky and Sobel, 2005; Wang and Georgakakos, 2005; and Georgakakos et al., 2006.

Whereas most precipitation and water resources storage occurs in northern California, a very significant portion of the state’s population lives in southern California and can be severely impacted by heavy, orographically driven precipitation and related hazards. The examples given earlier are some of the more deadly recent events. Other smaller scale damages occur without the loss of life. Fewer studies have examined the occurrence of precipitation/heavy precipitation events in southern California. Neiman et al. (2004) present a detailed study of the February 2-3, 1998, storm event in the Southern California Bight coastal region as part of the CALJET experiment. This study represents a comprehensive observational investigation of the influence of orography on the wintertime storms in this region. With a hydrologic focus, Nezlin and Stein (2005) describe the spatial and temporal characteristics of precipitation over 12 major watersheds in southern California based on the Global Precipitation Climatology Project (GPCP) 1-degree resolution dataset and for the period
1996-2003, despite finding a fairly low correlation between the GPCP-based precipitation estimates and in-situ observations. Chapter 1 of this thesis examined the long term temporal and spatial variability of observed station precipitation in the region based on National Climatic Data Center (NCDC) dataset. This covered a total of 111 daily and 59 hourly precipitation stations. Spatial variation and climatological characteristics of both average and heavy precipitation are described. The analysis identified orographic and larger scale climate variability influences on regional precipitation.

In contrast to the observation-based studies, other recent research has focused on models of the occurrence of precipitation in the region, as is the focus of the present chapter. Several recent studies examined the simulation of heavy rainfall events, each utilizing the MM5 numerical weather simulation model. Carpenter et al (2007) examined the January 6-12, 2005 storm event over the southern California region. They show agreement both in magnitude and location of precipitation between the simulation and observations for this strong, orographic precipitation event. Similarly, Cerezo-Mota and Cavazos (2006) evaluated the MM5 model for simulating heavy precipitation for events occurring in January 1993 over northern Baja California, Mexico and southern California, U.S.A. Hughes et al (2009) also utilized the MM5 model to investigate the role of coastal blocking on the spatial distribution of precipitation in the southern California region. Their study considered the climatological distribution and monthly precipitation for the 11-year period between May 1995 and April 2006. The unique aspects of the present study include the long duration of the historical period considered (55+ years), and the focus toward small spatial scales.
The chapter is arranged as follows. After a brief narrative of the southern California study region is given in Section 4.2, the three models are presented in Section 4.3. Section 4.4 discusses formulation and application of the simplified orographic precipitation model to southern California along with results from sensitivity analyses undertaken. Section 4.5 presents the comparison of precipitation estimates from three different methods at varying spatial and temporal scales, ranging from monthly climatologies to hourly frequencies and from regional (O[10,000s km\(^2\)]) to small watershed (O[100 km\(^2\)]) to grid (O[10s km\(^2\)]) characteristics. The comparison shows significant resolution effects and differences in magnitude of precipitation, but with similarity in basic spatial characteristics. Conclusions and future extensions are discussed in Section 4.6.

4.2 Study Region

The study region is generally the cismontane region of the Southern California Bight (see Fig. 3.1). Southern California has a Mediterranean climate, typified by dry summer seasons and moderately wet winters. This distinction between seasons in the region is apparent in the precipitation record, with the majority of the average annual precipitation (as much as ~90%) occurring during the “cool” or “wet” season between October and April. Another distinctive feature of this region is the close proximity of the Transverse and Peninsular mountain ranges to the Pacific Ocean. The Peninsular Ranges are aligned north-south, and the Transverse Ranges run approximately east-west from Santa Barbara County through San Bernardino and eastward toward Palm Springs. The mountain ridgeline occurs approximately 60-80 km from the coast. The Peninsular
Ranges include the Laguna Mountains of San Diego County and San Jacinto Mountains of Riverside County, and the Transverse Ranges include the Santa Ynez Mountains in Santa Barbara County and San Gabriel Mountains of Los Angeles and San Bernardino Counties. The two ranges intersect at the San Bernardino Mountains in San Bernardino County. The San Gabriel and San Bernardino mountains are separated by Cajon Pass, while the San Bernardino and San Jacinto Mountains are separated by Gorgonio Pass. The peak elevation is 3507 m at San Gorgonio peak in the San Bernardino Mountains, followed closely by Mount San Antonio (a.k.a. Mount Baldy) in the San Gabriel Mountains at 3000 m. When moisture-laden Pacific air masses are directed into these mountains via southwesterly low-level air flow, significant orographically-enhanced rainfall can occur and lead to damage from flash flooding and other related natural hazards. Such interaction of low-level atmospheric flow with the terrain is important for rainfall generation in southern California and is an area of research and a forecasting challenge (e.g., Small, 1999). Operational research focus is in part on identifying atmospheric forcing characteristics for flash flood events (Small, 2005). Additional recent research by the U.S. Geological Survey has involved an effort to define precipitation thresholds and susceptibility factors associated with post-wildfire debris flow occurrence, at small spatial and temporal scales (Cannon et al. 2008; Rupert et al. 2008; Cannon et al. 2007).

4.3 Models of Precipitation

The objective of this study is capturing the spatial variability of surface precipitation in southern California at high spatio-temporal resolution with an interest in
small-scale hydrologic hazard potential. Due to the rapid hydrologic response of watersheds in the region and the interest in flash floods, necessarily the focus is at sub-daily time scales. Figure 4.2 indicates the locations of hourly precipitation observation stations in southern California from the National Climatic Data Center as distributed by EarthInfo (EarthInfo, 2005). These stations were selected because they have at least 30 years of data over the historical period from 1948 to 2005. There is a higher density of stations at lower elevations and in more heavily populated regions, with a relative scarcity of stations in high elevations. In fact, only 15% of the stations are located above 1000 m. Modeling of precipitation is used as the variability of precipitation may not be well represented in observations alone due to the relative scarcity of precipitation stations and non-uniformity in spatial locations, as well as the high variability inherent in the short-duration precipitation process.

4.3.1 Gridded observation model using PRISM climatology

The first model of orographic precipitation considered here is based on an interpolation of hourly observations to a regular grid covering southern California using the Precipitation-elevation Regression on Independent Slopes Model (PRISM) dataset (Daly et al, 1994). PRISM distributes climatological and monthly precipitation from observed stations using linear regression developed on a monthly basis considering elevation, scale, slope orientation, and other environmental factors. Monthly and longer timescales rather than individual storm characteristics are used to estimate the distribution of precipitation from many storms over a large region. The model was initially developed in the Western U.S. and has expanded to include the conterminous
U.S. PRISM datasets have been used various climate and hydrologic applications, often as a validation source (Widmann and Bretherton, 2000; Hunter and Meentemeyer, 2005; Shamir and Georgakakos, 2006; Kanamitsu and Kanamura, 2007). Two datasets are available from the PRISM project website (http://prism.oregonstate.edu): (a) monthly precipitation at 2.5 arc-minute (4km) spatial resolution, and (b) climatological precipitation for each month at 30 arc-second (800m) resolution based on the period 1971-2000. The second product was used to develop the model for gridded observations given its higher resolution. Precipitation at a grid node for any given time step, \(t\), is given by:

\[
P_{gi,t} = \frac{\sum (w_{li} P_{oj})}{\sum w_{li}} \frac{P_{mon,gi}}{P_{mon,oj}}
\]

where the subscript \(gi\) indicates grid location \(i\), and subscript \(oj\) indicates a specific observation station. \(P_{gi}\) is the estimated precipitation at grid \(i\), given by the weighted sum of precipitation at nearby observations, \(P_{oj}\). The weighted sum is scaled by the relative magnitude of the climatological precipitation (PRISM value) at the grid node versus the station location for the month, represented in Eq. 1 by \(\overline{P_{mon,gi}}\) and \(\overline{P_{mon,oj}}\), respectively. The number of observed precipitation sites used to estimate precipitation at a grid is limited by a distance of 17 km, which is the average distance among all observations, and at least 3 non-missing observations were required to define precipitation at a grid node.
The weighting factor, $w_{i,j}$, is adopted from recommended operational procedures in mountainous terrain, and considers both distance from the grid to station and difference in elevation. This form of the weighting factor has been used for interpolation of mean areal temperature estimates (e.g., Smith et al, 2003), and is determined as:

$$w_{i,j} = \frac{1}{d_{ij}^2 + F_e(\Delta \text{elev})}$$

(2)

The coefficient $F_e$ determines the influence of elevation differences between locations on the weighting factor. For strong influence of elevation as in temperature interpolation, recommended values may reach $100 \text{ km}^2/\text{m}$. Given this application for precipitation interpolation and considering topography and average distance between stations, a $F_e$ value of 2 was used for this study. This gridded observation model allows for interpolation of the hourly observations using the climatological relationship of precipitation with topography, and is expected to have a more realistic variation than would be obtained from a simple interpolation of the station data. Issues that limit this method’s ability to reproduce hourly observed precipitation are the induced smoothing of the observations by the interpolation approach, the significant effects of the lack of observations at highest elevations, the use of climatological spatial distribution weights that may misrepresent the spatial distribution of individual storms, and the association of climatological parameters with a set historical period that may not be representative of the prevailing precipitation climate in other periods.
### 4.3.2 California Regional Downscaling (CaRD10) model

The California Regional Downscaling at 10 km (CaRD10) model was developed at Scripps Institution of Oceanography to provide a relatively high resolution regional dataset for climate research and applications (Kanamitsu and Kanamaru, 2007). It represents a dynamic downscaling of the NCEP-NCAR global reanalysis dataset using the Regional Spectral Model (RSM) over a domain covering all of California and Nevada, and portions of neighboring states. RSM (Juang and Kanamitsu, 1994; Kanamitsu et al, 2005) is a numerical model of the atmosphere, solving the hydrostatic primitive system of equations including mass conservation, momentum, thermodynamics, and atmospheric moisture. These equations are appropriate under the assumption that horizontal motions are much larger than the vertical scales of motion. For California, a 10-km horizontal resolution was selected by the developers to capture dominant atmospheric processes and to resolve the complex terrain throughout the State.

The terrain in the model was defined from USGS GTOPO30 digital elevation dataset, which has a native horizontal resolution of 30 arc-second (or ~1km). Mean elevation over each 10-km grid cell defined the topography of the model. Boundary conditions are defined from the NCEP-NCAR global reanalysis (Kalnay et al, 1996). The NCEP-NCAR reanalysis dataset has 200-km spatial resolution and 6-hour temporal resolution. Lateral forcing from the reanalysis was downscaled directly to the 10-km CaRD10 grid, with the model vertical levels selected to match the reanalysis sigma levels so that vertical interpolation is not performed at each input time step. Physical
processes parameterized in RSM include convection, boundary layer processes, long wave and short wave radiation, clouds, vertical diffusion, and the land surface. These parameterizations have been tested as part of the global model counterpart of RSM with simulations comparable to other global models (Robertson et al, 2004). The parameterizations used in the global counterpart are the same for the CaRD10 runs. Roads (2004) compared the RSM model with its global counterpart (GSM) over the United States, with the spatial characteristic of the RSM more resolved, and found overall similar error statistics.

Kanamitsu and Kanamaru (2007) evaluated the CaRD10 model output of surface temperature, precipitation, and winds against station observations and gridded climate products at various temporal scales. For precipitation, they considered a monthly comparison against the PRISM database and comparison against the Higgins et al (2000) gridded precipitation dataset at daily timescales. They conclude that CaRD10 precipitation has a strong correlation with the other datasets, but with a positive bias in heavy precipitation (heavy precipitation defined as > 10mm/day). For southern California, their presented results indicate a negative bias in monthly precipitation relative to PRISM along the southern California mountains in January. For a specific observation location near the Los Angeles county coast (location C3 in their Figure 9), there is a positive bias in monthly mean precipitation reaching approximately 50% during the months of January, February, and March. However, the overall season cycle is fairly well represented.

For the present study, hourly precipitation rates were extracted from the CaRD10 database and used for comparison with the other models. The CaRD10 model
represents results from a full numerical model, although at a courser resolution than the other models.

### 4.3.3 Simplified Orographic Precipitation (SIMOROP) model

Relative to full mesoscale prediction models, including RSM, the orographic precipitation (SIMOROP) model considered here is a simplification based on a decoupling of the momentum from the moisture conservation equations in the atmosphere. An analytical solution of potential theory wind flow over complex terrain under steady forcing is used to estimate equilibrium (or steady state) three-dimensional air velocities. The velocities then provide input to a three-dimensional moisture conservation model with bulk microphysics to produce estimates of precipitation rates over the terrain. This simplification makes the model computationally efficient and appropriate for long term simulation of high resolution orographic precipitation, while it maintains microphysical process parameterizations for cloud and precipitation generation and the deterministic signal of topographically-forced precipitation. The computational efficiency allowed for the generation of a relatively long (57-year) historical simulation of hourly precipitation and sensitivity analyses for this research.

The model was developed by Georgakakos et al (1999, 2004). The model was adapted for application to southern California at high resolution (3-5km), and included a series of sensitivity runs and model configuration changes. The SIMOROP model has been incorporated as part of an operational hydrometeorologic forecasting system for the Panama Canal watershed (Georgakakos et al, 1999) and also as a precipitation downscaling component of a large-scale decision-support project for water resource
management at several large reservoirs in the Sierra Nevada Mountains in northern California (Georgakakos et al., 2004). Initial model tests of the Panamanian system showed the model provided good short term forecasts of precipitation, which then fed a hydrologic modeling component of sub-watershed streamflows with hourly resolution (Georgakakos et al., 1999). For northern California, initial validation of the precipitation model over the American River watershed (with sub-catchment drainage area of O[1000km²]) showed correlations near 0.5 for hourly mean area precipitation (MAP) comparisons and near 0.65 for comparison with daily data, although with an over estimation bias for most sub-catchments (Georgakakos et al., 2004, 2006). Details of the formulation and its application to southern California are provided in the following sections.

4.3.3.a Potential theory wind

A three-dimensional estimate of the velocity field over mountainous terrain is obtained by application of potential theory flow. Potential theory flow is an idealization in which the flow is assumed to be irrotational, incompressible, and without momentum sources or sinks, over the meso-scale domain of interest. Under these assumptions, one can define the velocity potential, \( \phi \):

\[
\phi(x,y,z) = U(x) + Z(z) \times Y(y) \times X(x)
\]  

(3)
which satisfies the Laplace equation:

\[ \nabla^2 \phi = 0. \]  \hfill (4)

The functions \( U, X, Y, \) and \( Z \) are determined through application of the boundary conditions. Solution of the velocity potential provides the three dimensional velocity field as, by definition, \( \mathbf{U} = \nabla \phi \), or:

\[ u = \frac{\partial \phi}{\partial x} \] \hfill (5a)
\[ v = \frac{\partial \phi}{\partial y} \] \hfill (5b)
\[ w = \frac{\partial \phi}{\partial z} \] \hfill (5c)

The boundary conditions are defined for a rectangular domain as depicted in Figure 4.3, where the horizontal domain represents the region of interest of length \( L \) in the \( x \)-direction and \( K \) in the \( y \)-direction. The upper boundary (at \( z=0 \)) is the top of the atmospheric boundary under consideration and is located in the upper troposphere. The lower boundary (at \( z=-H \)) is the terrain surface, represented by the function \( S \). The horizontal axes are defined such that the \( x \)-axis is aligned with the forcing or free air stream velocity, \( U_0 \). Under this arrangement, the boundary conditions state that the directional derivatives vanish along the \( y \)- and upper boundaries, and are constant and equal to the free stream velocity along the \( x \)-boundaries:
\[ \frac{\partial \phi}{\partial x} = U_o \quad \text{at} \quad x=0, x=L \] (6a)

\[ \frac{\partial \phi}{\partial y} = 0 \quad \text{at} \quad y=0, y=K \] (6b)

\[ \frac{\partial \phi}{\partial z} = 0 \quad \text{at} \quad z=0 \] (6c)

Along the lower boundary, the vertical derivative is forced by the topographic gradient along the x-direction:

\[ \frac{\partial \phi}{\partial z} = U_o \ast \frac{\partial S}{\partial x} \quad \text{at} \quad z=-H \] (6d)

Under these boundary conditions, Georgakakos et al (1999) derive an analytic steady-state solution for the velocity potential in the form of a series of harmonic functions given by:

\[ \varphi = U - U_o \sum_{m=1}^{\infty} \sum_{n=1}^{\infty} c_{mn} \cos \left( \frac{m \pi x}{L} \right) \cos \left( \frac{n \pi y}{K} \right) \frac{\cosh(\beta_{mn}z)}{\beta_{mn} \sinh(\beta_{mn}h)} \] (7)

where the coefficients \( c_{mn} \) are defined from the boundary conditions and with

\[ \beta_{mn} = \pi^2 \left( \frac{m^2}{L^2} + \frac{n^2}{K^2} \right). \] (8)

Given the analytical solution, the velocity components are determined from the derivative of the velocity potential (Eq. 5). The solution provides the three-dimensional velocity vectors at each grid in a three-dimensional rectangular domain encompassing...
the region, and supplies the forcing for the atmospheric moisture model and orographic precipitation computations. The solution is computed at each of a time sequence of boundary conditions leading to a quasi-steady state evolution of the wind patterns. The steady state solution of the wind component is a reasonable assumption if flow fluctuations traverse the domain within the time interval of interest.

4.3.3.b Atmospheric moisture model

The orographic precipitation computations are based on the atmospheric moisture model for cloud and precipitation first proposed by Kessler (1969), which describes the response of the water content of air to air motions and microphysical process (see also Georgakakos and Krajewski, 1996; Tsintikidis and Georgakakos, 1999). In the present formulation, we incorporate a term for environmental mixing with the ambient atmosphere. Following an air parcel along its path (Lagrangian framework), the conversation laws for precipitation and cloud may be written as:

\[
\frac{\partial M}{\partial t} = -v \frac{\partial M}{\partial x} - u \frac{\partial M}{\partial y} - (V+w) \frac{\partial M}{\partial z} - M \frac{\partial V}{\partial z} + Mw \frac{\partial \ln \rho}{\partial z} + AC + CC - EP - k_5 |w|M
\]

(10)

\[
\frac{\partial m}{\partial t} = -v \frac{\partial m}{\partial x} - u \frac{\partial m}{\partial y} - w \frac{\partial m}{\partial z} + wG + mw \frac{\partial \ln \rho}{\partial z} - AC - CC + EP - k_5 |w|(m-m_e)
\]

(11)

where, \(M\) and \(m\) represent respectively the model states of (a) precipitation content
(always \( \geq 0 \)) and (b) either cloud content (when \( m > 0 \)) or the saturation deficit (when \( m < 0 \)). Both \( M \) and \( m \) have units of \( \text{gm/m}^3 \). The horizontal and vertical velocities \((u, v, \text{ and } w \text{ in m/s})\) are provided by the potential theory wind component; \( \rho \) is the air density; \( V \) is the terminal velocity of precipitation relative to the air and parameterized by Eq. 12 below; and \( G \) is the condensation function. The first terms of Eq. 10 and 11 represent the horizontal and vertical advection of precipitation and cloud (e.g., \(-w \partial M / \partial z\)). Compressibility effects are incorporated via the \( \partial \ln \rho / \partial z \) term. The condensation function describes genesis of cloud and is approximated by a linear function of the vertical gradient of the saturation mixing ratio. The last terms are bulk parameterizations of the microphysical processes of autoconversion of precipitation from cloud (AC), cloud collection (CC), evaporation of falling precipitation (EP), and environmental mixing (\( k_5 \) term). These terms are determined as follows:

\[
V = -38.3 \, N_o^{1/8} M^{9/8} \exp\{kz/2\} \\
AC = k_1(m-\alpha) \\
CC = k_2E N_o^{1/8} m M^{7/8} \exp\{kz/2\} \\
EP = k_3 N_o^{7/20} m M^{13/20}
\]

These processes are functions of altitude \((z)\), precipitation \((M)\) and/or cloud content \((m)\), the efficiency of cloud conversion \((E)\), and parameter \( N_o \). Initial parameters values were based on recommended values as given by Kessler (1969), and on prior applications (Georgakakos, et al, 1999, 2005). The initial parameter values are given in
Table 4.1. Several limitations of the microphysical processes are noted: (a) autoconversion occurs only when the cloud content exceeds the autoconversion threshold (i.e., \( m > \alpha \)); (b) cloud collection occurs only when cloud exists (i.e., \( m > 0 \)); (c) evaporation of falling precipitation occurs only when there is a saturation deficit (i.e., \( m < 0 \)), and (d) environmental mixing can occur only when cloud or precipitation exist (i.e., \( M, m > 0 \)).

4.3.3.c Limits of Applicability

The fundamental assumptions of the potential theory wind model are:

(a) flow is steady and uniform over the time interval of interest;
(b) the atmosphere is near saturation;
(c) the scale of flow fluctuations is longer than the topographic fluctuations; and
(d) the Coriolis effect is negligible over the spatial domain of interest.

Assumptions for the atmospheric moisture model are stated in Kessler (1969):
(e) cloud is condensed water which shares fully the motion of the air;
(f) cloud is formed in rising saturated air and evaporates in saturated descending air, cloud exists only in saturated air and unsaturated air does not contain cloud;
(g) precipitation particles, once formed, are distributed in size according to an inverse exponential distribution; precipitation particles collect cloud particles or evaporate in sub-saturated air according to approximations of the natural accretion and evaporation processes;
(h) precipitation shares the horizontal motion of the air, but vertical mass transport is based on the fall speed of the median-diameter precipitation particle.

This formulation of the simplified orographic precipitation model differs from earlier simplified approaches (e.g., Rhea, 1978) in that produces consistent, three-dimensional velocity fields over complex terrain while it includes explicit microphysical parameterizations for cloud and precipitation generation. Similarly, relative to linear upslope models (e.g., Smith, 2003; Smith and Barstad, 2004), the model includes microphysical processes. It also differs from similar models applied for tropical regions (i.e., Georgakakos et al, 1999; Tsintikidis and Georgakakos, 1999; Lee and Georgakakos, 1990), in that the present formulation does not explicitly incorporate convective updrafts present in regions of strong convection. In such regions, the total updraft, $w_t$, may be expressed as:

$$w_t = w_c + w_o$$

(10)

where $w_c$ represents portion of the updraft due to convection and $w_o$ represents that portion that is orographically forced. Lee and Georgakakos (1990) express this convective updraft as a function of the convective available potential energy, $E_c$:

$$w_c = a \sqrt{2E_c}$$

(11)
and a is a parameter with values which may be on the order of 0.01.

It is of interest to know whether convection is an important component of the precipitation regime of southern California. The convective available potential energy, or CAPE, is a measure of the buoyant energy an air parcel may have to accelerate the parcel vertically through its surroundings if the parcel is able to reach the level of free convection. CAPE is typically taken as a measure of atmospheric instability or potential convective storminess, with CAPE values equal to 0 indicating stable conditions; values less than 1000 J/kg signifying very weak instability; and values greater than 2500 J/kg indicating strong instability for extratropical regions. CAPE values estimated from radiosonde observations (available from the University of Wyoming, Department of Atmospheric Sciences, http://weather.uwyo.edu), were obtained for the San Diego/Miramar, CA station (NKX) with twice daily atmospheric soundings for the period September 1989 to September 2008. Of the more than 2600 soundings, fewer than 10 indicated CAPE values greater than 800 J/kg, and only 3 of which occurred during the wet season months of October through April. Although localized convection is known to occur in the study region, the low values of CAPE suggest this is not the primary driving force for precipitation for this study, particularly during the wet season. No attempt was made to model convection in the simplified orographic model formulated.

4.3.3.d Application over simple mountain geometry
To illustrate the model response, this section presents the application of the SIMOROP model to a hypothetical, simple geometry of a single mountain, elongated in the north-south direction with a steep rise to a peak elevation of 3500 m. The domain of the hypothetical model covers 200 km x 200 km. The mountain geometry and updraft velocities for a unit-magnitude westerly forcing wind are shown in Figures 4.4a and 4.4b. Peak vertical wind updraft velocity is about 0.06 m/s. An idealized atmospheric sounding was defined (based on radiosonde data for NKX on a random selection of days with rainfall). This baseline case incorporated a moisture lower atmosphere with relative humidity set to 95% to the 750hPa level and decreasing to 65% at 400hPa. This represented a single, moist lower atmospheric layer. The atmospheric model component was run with a forcing wind ($U_o$) of 20 m/s, model parameters defined as in Table 4.1, and with a moisture model integration time step of 225 seconds for a period of 6 hours. The total volume of precipitation generated is shown in Figure 4.4c, with a peak of 64 mm/6-hour.

The purpose of this simplified geometry case was to examine impact in changes in the atmospheric input, resolution, and parameters. Several variations of this idealized case were run and evaluated in terms of changes to the peak precipitation and the areal extent of precipitation generated. Table 4.2 presents this comparison of precipitation and areal extent of precipitation exceeding several thresholds for a number of sensitivity cases. First, variation in the amount of moisture input from the sounding was considered. The first sensitivity run introduced a secondary, upper layer of moist area with relative humidity of 98% between 600-500hPa, and another run reduced the lower moisture to 93% relative humidity between 1000-850hPa. This sensitivity showed the
expected results, with only small changes in precipitation for the secondary upper moist layer case and reduced precipitation for drier conditions in the lower atmosphere. Next, several cases were run which perturbed the model resolution and the atmospheric model parameters of efficiency \((E)\) and autoconversion moisture threshold \((\alpha)\) from the initial values listed in Table 4.1. The efficiency was reduced from 1.0 to 0.2, producing a large reduction in the peak precipitation, but with little impact on the spatial extent of precipitation. The autoconversion threshold was increased from the initial value of 0.3 to a value of 0.6 and reduced to a value of 0.1. The higher value would require more moisture in the atmosphere before autoconversion to hydrometeors would occur. In contrast to the efficiency, this parameter had a smaller impact on the peak precipitation and changed the areal extent of precipitation with lower thresholds increasing the extent of precipitation for the selected thresholds of 0, 2 and 10 mm. These sensitivity runs confirmed expected model behavior and gave insight to the sensitivity of the atmospheric model parameters. This sensitivity is confirmed in larger scale sensitivity runs performed for southern California as described in Section 4.4.2.

4.4 SIMOROP Model Application for Southern California

This section describes the adaptation of the SIMOROP model for southern California and presents a series of sensitivity studies that examined model configuration and parameterizations.

4.4.1 Southern California model configuration
For the southern California model, the National Elevation Dataset (NED, USGS, 2010) was used to define the terrain. The chosen dataset has a native resolution of 30m, which was sampled at a resolution of 1 km. The southern California domain totaled 391 km by 581 km. The wind forcing and moisture boundary conditions were derived from the NCEP/NCAR global reanalysis dataset (Kalnay et al, 1996). This dataset was selected as it contains 6-hourly atmospheric data extending historically to 1948, albeit at coarse resolution, that allowed for the long term simulation of precipitation for the region. The spatial resolution is 2.5 degrees, or ~250km. The locations of the nearest reanalysis grid centroids are indicated in Figure 4.2 by the cross symbols. For this study, a single grid node was selected to provide the forcing and boundary condition depending on wind direction. The selected node was either off the coast of San Diego County (centered at [117.5W, 32.5N]) for free atmosphere winds (taken as 700 hPa level) coming from the southeast to northwest directions to represent conditions for generally oceanic, onshore flow. In all other cases, the node to the north/northeast of the region (centered at [115.0W, 35.0N]) was used.

The reanalysis input included the atmospheric variables of temperature, pressure, humidity, and winds extracted at standard atmospheric levels for each 6-hour period for 1948-2005 record. The free air stream velocity forcing, $U_o$ in Figure 4.3 and Eq. 3 above, is taken as the 700 hPa wind from the reanalysis. Table 4.3 presents a comparison of the 700hPa wind characteristics for the off-coast reanalysis node with data from the nearby upper air radiosonde station at NKX (San Diego/Miramar). The table shows the fraction of input periods with wind from the different directions and the average wind speed magnitude in each directional bin. The comparison is done for wet
season (October – April) data periods only. The reanalysis results are shown for two
time periods: (a) the entire historical record available, 10/1948-4/2005, and (b) the
period of overlap with radiosonde data, 1/1994-4/2005. There is agreement between the
different periods and data sources, showing that highest frequency of winds come from
southeast to northern direction (~255 – 360°), with average wind magnitude exceeding
11 m/s for southwesterly (SSW, 240°) to northwesterly (NNW, 345°) winds.

Figure 4.5 presents the lower level (925 hPa) specific humidity annual cycle for
the two (off-coast and inland) reanalysis nodes for all periods in the record (solid lines)
and when the wind is from a southwest-to-west direction (210-285°; shown by dashed
lines). This general wind direction is of interest as this drives the moist Pacific air into
the southern California mountains. Both reanalysis nodes reflect this increase in
specific humidity during the wet season with southwesterly air flow.

For enhanced computational efficiency, the terrain data is rotated to align the
axis of the computational grid with the forcing wind direction. The forcing wind
direction was discretized into twenty-four 15° orientations, and the potential theory
3-dimensional wind field was solved a priori for a unit magnitude wind forcing. The
1-km terrain data was rotated, and the slope was computed along the direction of the
wind (i.e., ∂S/∂x). For the initial model configuration, the slope was then averaged on a
3-km grid before solving for the 3-dimensional velocity field. The slope is computed at
the 1-km resolution to maintain detail in the topographic variations, and the averaging at
3-km provided reasonably fine resolution with a fairly short computational time and
without numerical sensitivities. A total of 15 vertical layers were used with a resolution
of 500m, thus extending to a height of 7500m, and incorporating essentially all the
moisture in the atmosphere. At each time step in the simulation, the free air stream velocity (700 hPa) was used to linearly scale the pre-computed velocity components to drive the atmospheric moisture model. The reanalysis atmospheric variables were used to derive an atmospheric sounding from the pressure, temperature, and humidity at twelve standard levels (1000, 925, 850, 700, 600, 500, 400, 300, 250, 200, 150, and 100 hPa). However, the higher levels (< 300 hPa) are typically above the vertical limits of the computational grid. Other meteorological variables, such as air density and saturation vapor pressure, were derived from this input and used as necessary to define the initial conditions and parameters of the atmospheric moisture model. The initial moisture conditions were uniformly applied across the computational domain. For each 6-hourly NCEP file, the forcing wind was assumed constant, and model was run for 6 hours with an integration time step of 225 seconds. Surface precipitation rate was computed for each hour during the integration and associated back to the original, non-rotated, rectilinear grid in the domain. The results and comparisons presented in Section 4.5 are based on the hourly precipitation on the original 1-km resolution grid.

4.4.2 Sensitivity analysis

Once the SIMOROP model was configured for the southern California domain, a series of sensitivity runs were performed to explore the impact of variations in the configuration on the simulated precipitation. The analyses were confined to a shorter record focusing on the winter months of December, January, and February for the five year period from December 2000 to February 2005. The configuration variations considered: (a) the vertical resolution of the model; (b) the selection of atmospheric
level to define the forcing wind; (c) the strength of mixing of moisture with the
surrounding environment; (d) the spatial distribution of input moisture conditions; and
(f) the forcing input data source.

The baseline configuration case, as described in Section 4.4.1 above, has 15
vertical layers with a resolution of 500m and forcing from the NCEP-NCAR reanalysis,
free stream velocity defined at 700 hPa, and with no environmental mixing (k_5 term of
Eq. 11 set equal to zero). Table 4.4 summarizes the sensitivity run cases which
included: a change in the vertical resolution to 350 and to 1000m; free stream wind
forcing defined by the 850 and 500 hPa atmospheric levels; distribution of initial
moisture conditions defined by horizontal interpolation based on inverse square distance
weight of the input from 4 reanalysis grid nodes; variation of environmental mixing
term through a change in the value of k_5 of Eq. 10-11; and finally, use of an alternative
input forcing data source, specifically radiosonde observations for the NKX (San
Diego/Miramar) location. The radiosonde data was obtained from the Earth Systems
Research Laboratory of NOAA (ESRL, 2008), and used to develop the input
atmospheric sounding as expected for the model. The NKX was used due to location
and the relatively completeness of observations (at 0 and 12 UTC for the period 1994 –
2007). The radiosonde data contains information at additional atmospheric levels used
to develop a more vertically-detailed initial moisture condition description. However,
the radiosonde data was available only 2 times per day, as compared to 4 times daily
records for the reanalysis data, the model was run for a period of 12 hours with constant
forcing wind.
Table 4.5 presents a comparison of summary validation measures among the various sensitivity runs. For this table, the hourly simulated precipitation was aggregated to daily values over 25 small watersheds across the region, compared with mean areal precipitation estimates based on daily observations over those watersheds. The watersheds were selected based on a minimum of three observations station located within or nearby the watershed from which observed mean areal precipitation (MAP) could be computed. The measures presented are based on average values across the selected watersheds. The measures include:

(a) fraction of hours without precipitation (Frac0);
(b) fraction of observed hours without precipitation also predicted by the model (Hit0);
(c) maximum distance in the cumulative frequency distributions (CDFs) of non-zero precipitation between observations and simulations (maxD);
(d) correlation (CCOR) and bias of daily precipitation; and
(e) correlation (CCOR) and bias when observed precipitation exceeded 10mm/day.

There are inherent difficulties in the comparison of model output to station observations. Notably, different spatial scales are represented by each, a grid average relative to a point observation. Averaging over multiple locations aims to reduce the impact of the spatial discrepancy. Precipitation observations are also subject to uncertainties and bias, particularly gage catch deficiencies (Larson and Peck, 1974; Sieck et al 2005, Ciach, 2003). The baseline case for the 5-year, 3-month sensitivity run indicates the model has a low positive bias of 0.19 over all locations, and average correlation of 0.46. The model over-simulates days with no precipitation (Frac0 is
lower), and captures about 75% of the non-raining days. The various configuration runs generally produced similar correlations and maximum CDF difference statistics (with exception of NKX radiosonde forcing, which had lower correlation). In some sensitivity cases, improvement in representing the no-precipitation days came at the cost producing a large negative bias.

The model configuration associated with the baseline case was then used to simulate the long term, wet season precipitation. Table 4.6 presents validation measures as considered in the sensitivity runs for the long-term simulation. This indicates an average correlation of 0.43 against daily observations, with an overall positive bias (all precipitation), and a negative average bias for precipitation exceeding 10mm/day. The overall bias is partially due to the lower frequency of simulating days with zero precipitation, where this fraction is 0.77 in the simulations and 0.84 for the observations. The combination of these measures suggests that the orographic model is over-simulating the frequency of low precipitation. Although this is a simplified model, a tendency of other numerical models to over-simulate low precipitation (i.e., drizzle) has been noted by others (e.g., O’Connor et al 2007). An examination of the long term average monthly wet season total volumes indicated very high, but localized, bias when compared to the gridded observations. The simulated long term average wet season total value maximum precipitation in the region exceeded 2.5 m. While the observations have limitations, especially in high elevations, this high average wet season total is roughly twice the magnitude of the maximum average annual precipitation at observation stations, indicating a model positive bias.
A secondary, limited set of sensitivity runs examined the parameters of the atmospheric moisture model and its sensitivity to spatial resolution. These sensitivity runs examined the precipitation only over the months of January and February 2005 (which included the heavy and damaging precipitation events described in the Introduction), and considered only the peak monthly volume and extent of precipitation under variations in the model spatial resolution and moisture model parameters. Observation stations recorded over 800mm of precipitation in January and 700mm in February of 2005. The moisture model parameters considered were the efficiency (E), and autoconversion moisture threshold (α). The initial values of these parameters, as shown in Table 4.1, were adopted from the prior applications for Panama and Northern California. These models had different spatial resolutions of 5 and 10km, respectively. The value of the efficiency, given a value of 1.0, is quite high for the present region of application. Several studies have looked at estimates of estimating precipitation efficiency, with values varying with specific storms, storm system type, updraft (e.g., Market et al, 2003; Ferrier, et al, 1996; Fankhauser, 1988) Market et al (2003) tabulate efficiencies listed in the literature ranging from 5% in mid-latitudes to 130% in tropical convection. Jiang and Smith (2003) cited a study of estimated precipitation efficiency in the Santa Ynez and San Gabriel Mountains of southern California with values between 20 and 30%. The model efficiency value of 1.0, which perhaps more typical of a tropical region as Panama, may be significantly too high for southern California. Table 4.7 presents a listing of the secondary sensitivity runs conducted and evaluation measures for the two simulation months. Considering three different model resolutions of 3, 5, and 10km, the efficiency was varied with values of 1.0, 0.4, and 0.1, and the
autoconversion moisture threshold varied with values of 0.6, 0.3, and 0.1. Also included was a change in the vertical resolution from 1000m to 600m for the 5km-horizontal resolution model. The table is intended to give representative response of the large scale model sensitivity runs; not all parameter configurations were run for each spatial resolution.

The table does show responses similar to the simple geometry sensitivity runs. The efficiency largely controlled the magnitude of the peak precipitation at a given resolution, and the autoconversion threshold affected both peak precipitation magnitude and the areal extent of precipitation. Lower autoconversion thresholds represent conditions more conducive to the formation of precipitation droplets, such as the existence of more abundant cloud condensation nuclei (CCN). A higher concentration of CCNs downwind of urban environments has been noted (Kocmond and Mack, 1972; Ochs and Semonin, 1979), and with recent evidence of suppression of precipitation generation due to urban air pollution (Rosenfeld and Givati, 2006; Collier, 2006; Givati and Rosenfeld, 2004). CCN source and influence of urban centers for the Los Angeles-Riverside area have been studied as part of the Study of Organic Aerosols at Riverside (SOAR-1) campaign (e.g., Cubison et al, 2008). The potential for urban aerosols exists near these regions along with the metropolitan areas of Orange County and San Diego.

The interplay with peak precipitation and model horizontal resolutions is largely related to the updraft velocities. Coarser resolution results in greater smoothing of the terrain and lower slopes, thus yielding lower updraft velocities. For other parameters being equal, the higher horizontal resolution configurations produced higher peak precipitation volume. The sensitivity results reflect a substantial decrease in peak
precipitation from the parameterization of the long term runs (3 km horizontal resolution, $E=1.0, \alpha=0.3$). The final configuration of the SIMOROP model was based on these sensitivity runs, considering peak magnitude and full extent of precipitation generated by the model. The final configuration incorporated a horizontal resolution of 5km, with 15 vertical layers of 600m, and an efficiency coefficient of 0.1. A variable autoconversion threshold was used to represent potential effects of increased CCN for the urban areas. A value of 0.01 was applied to low elevations and a value of 0.5 to elevations > 750m. This configuration was used to produce the long term simulations presented in the next section. It is noted that fine-tuning of the model parameters involves trade-offs of being able to produce long term precipitation statistics well versus individual storm extremes, and a choice must be made between these two when selecting parameter values. In this work we opted for preserving the long term statistics of precipitation mainly as they are akin to being able to estimate antecedent moisture conditions better for flash flood applications.

4.5 Model Inter-comparison at Various Temporal and Spatial Scales.

4.5.1 Long-term climatological features

Figure 4.6 shows the average wet season total precipitation from the 3 models computed over the simulations period 10/1948 – 4/2005. The depiction of precipitation is limited by the coast lines, which effects the display of precipitation over the ocean as simulated by the CaRD10 model. The figure highlights the differences in resolution of the models. The spatial interpolation GRIDOBS model shows a relatively smooth variation across the region and a peak value of 1.1 m in the San Gabriel Mountains near
the San Bernardino County line. The SIMOROP model has a narrower extent of precipitation, with higher volume of precipitation produced only near the mountain ranges. The coastal and lower elevation regions tend to have a small to zero average wet season total simulated by the model. Although the regions receive rainfall as apparent in observations, this low or absence of precipitation in the SIMOROP model results is not unexpected given the lack of topographic forcing. In the mountainous area, SIMOROP produces higher total wet season volume than the GRIDOBS model, with a peak value of 2.2 m. Although the SIMOROP model may over simulate peak precipitation, the magnitude of the precipitation in the mountain is probably not well observed due to the scarcity of stations at high elevation. The coarser resolution of the CaRD10 model precipitation is apparent relative to the other models. The CaRD10 resolution impacts the magnitude of wet season precipitation yielding a lower average value of 850 mm. The CaRD10 precipitation occurs somewhat upstream (southwest) of the high elevations, depicted by the contour lines in the Figure. This may result from the difference in the topographic data and possible difference in geo-registration of the GTOPO30 data used for the CaRD10 relative to the NED data used for the SIMOROP model.

The location of precipitation occurrence is consistent among the models and generally across months during the wet season, as shown by example for the months of October (drier month) and February (wettest month) in Figure 4.7. Peak precipitation occurs along the Santa Ynez, San Gabriel, and San Bernardino mountains of the Transverse Range (Santa Barbara, Los Angeles, and San Bernardino Counties, respectively). The color scale for each month is consistent for all three models. Thus,
although the location of precipitation is similar in all models, the lower monthly average precipitation for the GRIDOBS and CaRD10 models in October is apparent. The SIMOROP model produces higher values for Santa Ynez, San Gabriel and San Bernardino Mountains, as also shown in the GRIDOBS model, although the magnitude is higher along the Santa Ynez Mountains for the SIMOROP results. The SIMOROP model also produces a higher magnitude of precipitation along the lower Peninsular Range in San Diego County than the GRIDOBS model. The higher precipitation along the southern Peninsular Range is also indicated by the CaRD10 model, and the precipitation is generally less there than along the Transverse Range. Considering the location of observation stations as shown in Figure 4.2, there appears to be few observations at the higher elevations in this region which may influence the estimates from the GRIDOBS model.

From the simulations and gridded results, an average precipitation was computed over the Transverse and Peninsular Ranges each month to illustrate the wet season cycle and differences between the Ranges. The grid locations with elevation greater than 950 meters were selected in each Range, with the Peninsular Range defined from the San Diego County border to just north of San Gorgonio Pass (at latitude 34.1°N). The Transverse Range latitudinal extent was limited to 34.1° – 34.9°N, and extended from Santa Barbara through San Bernardino Counties. This definition eliminates the high precipitation simulated in the SIMOROP model along the Tehachapi Mountains near -119W, 35N.

The monthly climatology of precipitation defined for these two mountain regions is shown in Figure 4.8. The seasonal cycle in precipitation is similar amongst
the models, with variation in the magnitude. The models peak precipitation occurs in January-February, and with the Transverse Range having higher average precipitation than the Peninsular Range. The wet season average precipitation from the GRIDOBS model is 464 mm for the Transverse range and 410 mm for the Peninsular Range. The average values are similar as computed from the SIMOROP model (450 and 439 mm, respectively), and reduced for the CaRD10 model (350 and 240 mm, respectively). The lower average precipitation from the CaRD10 model is also impacted by the shift in location relative to the terrain. While the wet season average values are similar, the amplitude of the seasonal cycle of the SIMOROP model is reduced relative to the GRIDOBS model. The SIMOROP model yields higher precipitation during the months of October and April by approximately 20mm on average while the peak in February is reduced, particularly for the Transverse Range. The increase in simulated precipitation for the month of October is apparent also in Figure 4.7. The GRIDOBS model shows low, but wider extent of precipitation reaching about 55 mm/month in October, whereas the SIMOROP model shows narrow extent but high precipitation reaching 200mm/month.

The comparison of long term average precipitation among the three models indicates similarity in spatial features and overall seasonal behavior on a large scale representing the two mountain ranges. However, significant differences exist particularly in the magnitude of precipitation.

4.5.2 Frequency of hourly precipitation
Although the wet season total precipitation may be significant, the nature of precipitation is episodic with no precipitation occurring most of the time. The initial sensitivity studies of Section 4.4.2 indicated the fraction of daily periods with no precipitation averaged about 0.77 for observation stations during the 2000-2005 period analyzed (Table 4.5) and a value of 0.84 for the longer term period (Table 4.6). For the shorter period, this fraction was lower for the SIMOROP model, indicating more days with precipitation were simulated by the model. This section considers the frequency of precipitation occurrence at different thresholds, rather than non-occurrence, for all three models. Figure 4.9 presents the frequency of precipitation greater than precipitation thresholds of 0.25mm/hr and 5 mm/hr at all grids. The average frequency is computed over all the wet season and as a fraction of all hours simulated. For the low (trace) precipitation threshold, the GRIDOBS shows the region between the coast and mountains with a range of values near 0.03-0.035. The highest frequency of trace precipitation occurs in the San Bernardino Mountains and reaches 0.08. For the 5mm/hour threshold, the frequency for GRIDOBS model generally confined along the Transverse Range, with localized peak values reaching 0.01. The CaRD10 model shows higher frequency values for trace precipitation, but slightly lower values at the 5 mm/hour threshold. The pattern along the Transverse Range is also apparent at the higher threshold, but shifted to the south. The SIMOROP model frequency appears intermittent in the coastal region relative to the other models, due to the fact that very low frequency values are not depicted. The SIMOROP model produces the highest frequency values of 0.28 for trace precipitation, along the northwest portion of the domain. The SIMOROP simulated frequency along the southern Peninsular Range is
higher than the GRIDOBS model, and higher than simulated in the San Gabriel and San Bernardino Mountains. This is also reflected in the CaRD10 model, although with a lower magnitude. At the 5mm/hour threshold, the SIMOROP frequency range is generally lower in magnitude and more spatially limited than the other models.

This analysis was expanded to look at the frequency of precipitation exceeding various thresholds, ranging from 0.5 mm/hour to 15mm/hour. The focus was narrowed to the mountainous region by selecting a few higher elevation locations along each mountain range, and comparing the average frequency of precipitation from the observational record with the average frequency computed from a corresponding model node. The locations of the selected observation stations are indicated in Figure 4.2 by the filled symbols. A total of 6 stations were selected in each mountain range with station elevation ranging from about 900m to 2000m. The frequencies of precipitation exceeding a range of thresholds were computed on a monthly basis as the number of hours with precipitation exceeding the threshold divided by the number of hours in the record for that month. The observed frequencies for the selected stations are shown in Figure 4.10 for the month of February. Model frequencies were computed for the corresponding grid locations, and the average observed and simulated frequencies over the locations were compared as in Figure 4.11 (for month of February). Due to the point estimates of observed precipitation, it is important to note that this comparison is sensitive to the location of the observing station and provides a rough measure of performance.

As expected, there is significant agreement across all precipitation thresholds for the GRIDOBS model, with a slight under representation of the frequency of higher
precipitation rate (> 10mm/hour), resulting from the smoothing of high precipitation
rates in the interpolation. Little difference is seen between the agreement of model and
observed frequencies between the Peninsular and Transverse Ranges for the GRIDOBS
model.

The CaRD10 model shows a reduction in simulated frequencies across the range
of precipitation thresholds, with the Peninsular Range average simulated frequency
closer to the average observed frequency than the Transverse Range. The SIMOROP
model produces a higher frequency at low precipitation thresholds, and under-simulates
the frequency of heavier precipitation significantly, on average, not exceeding an
average precipitation rate of 7.5mm/hour over the selected grid locations. The results of
Chapter 1 suggest that this upper limit corresponds to the 90th percentile of observed
precipitation. In light of the mean seasonal volume results of the previous section, this
suggests that the SIMOROP model tends to over-simulate light precipitation (or drizzle)
and under-simulate heavy precipitation while maintaining the overall seasonal volume.
This to a large measure is due to the choice to emphasize long term average statistics in
the model parameter estimation as discussed earlier. While the SIMOROP model is
based on simplifications relative to full mesoscale numerical models, the occurrence of
over-simulating light precipitation has been noted with such models (e.g., O’Connor,
2008; O’Connor et al, 2007).

An analysis was performed to attempt to adjust the simulated precipitation in a
way that would maintain the spatial detail provided by the model but better represent
the frequencies of precipitation at the different thresholds. A quadratic adjustment of
the form:
was considered for application to the precipitation of the selected grid, with monthly adjustment coefficients estimated to reproduce the average observed frequencies across the selected locations. Several iterations were undertaken, considering variation in the number of locations selected, the elevation range over which the adjustment applied, and different adjustments for the Transverse versus Peninsular Range. However, the methodology proved sensitive to the adjustment configuration and was not pursued further in this work.

4.5.3 Small watershed climatology

One of the driving motivations for this analysis is to consider precipitation occurrence at the smaller spatial scales associated with flash flooding. Therefore, the modeled precipitation is also compared on a small watershed basis. A total of eight small watersheds were selected across the region based on size (drainage area < 500 km$^2$), location, and distance to hourly observations stations. Watershed characteristics and basic statistics are presented in Table 4.8. The statistics are computed for the SIMOROP and CaRD10 models relative to the GRIDOBS model, and include the long-term bias and correlation both on daily and hourly timescales. Figure 4.12 shows the wet season cycle in average and standard deviation of monthly precipitation for the three models for 4 of the watersheds. The size of watersheds cover a range from 65 to 400 km$^2$ in drainage area, and are distributed throughout the mountainous region with average elevation from 520 to 1800m. The seasonal cycle in
precipitation is the same as shown for the large scale region (Fig. 8) with a broad peak in January-February. The variability of precipitation follows the mean pattern for all watersheds. However, the comparison of the magnitude in seasonal cycle between the three models for any given watershed is quite variable. For particular watersheds, e.g., Alamo Pintado Creek (11128250) in Santa Barbara County, the magnitude of the seasonal cycle is similar between all three models. For other watersheds, the SIMOROP and CaRD10 under represent the mean monthly precipitation for all months (e.g., Sespe Creek, 11111500).

The statistics shown in Table 4.8 also indicate this variability in average precipitation relative to the GRIDOBS model. While all three models have limitation and do not truly represent the true occurrence of rainfall at hourly resolution, for this calculation, the GRIDOBS model is used as the baseline. The table shows the mean bias, and linear correlation at both daily and hourly time steps. For each watershed, the first line in the table presents the statistics for the SIMOROP model and the second line is for the CaRD10 model. The bias ranges from -0.7 to 0.6, with the highest positive bias resulting from the SIMOROP model along a Peninsular Range watershed. The largest negative bias is also from the SIMOROP in the San Gabriel Mountains, followed closely by the CaRD10 model at the Sweetwater River watershed where the SIMOROP model has positive bias. Correlation of MAP at hourly timescales ranges from 0.15 to 0.38 for the CaRD10 model and 0.3 to 0.53 for the SIMOROP model, and at daily timescales, the correlation values are 0.27 to 0.76 for the SIMOROP model and 0.3 to 0.65 for CaRD10. These correlation values are similar to values reported for other comparisons of numerical modeling of precipitation (e.g., Kanamitsu and Kanamaru,
2007; Wang & Georgakakos, 2005). Overall, this comparison for small watershed MAP climatology indicates reasonable correlations, but that the magnitude of precipitation from a particular model can vary substantially with respect to the rest of the models.

4.5.4 Large scale climate influences

In Chapter 2, analysis of observed station precipitation data showed a fairly strong seasonal correlation with climate oscillation indices, particular SOI and Nino 3.4 (seasonal correlations 0.45 and greater). This also indicated a clear distinction in the occurrence of precipitation conditioned on the climate index Nino3.4. When the early wet season (October-December or OND) Nino3.4 index fell in the upper tercile of its distribution, wet season precipitation was found to be higher than normal, with higher variability and a more frequent occurrence of heavy precipitation. A similar methodology was explored to investigate whether this feature was captured in the simulations. Conditioned on the OND Nino 3.4 index occurring in the upper and lower terciles of its distribution, the average wet season total precipitation for those seasons was compared to the unconditional mean at each grid. Figure 4.13 presents the spatial distribution of the ratios for the SIMOROP and CaRD10 models. The GRIDOBS model was not included as this would necessarily follow the observations results. The upper row figures show the ratio when the Nino3.4 is in the upper tercile, and the lower row shows the ratio when Nino 3.4 index is in the lower tercile. The characteristic feature of an increase in the mean wet season total precipitation relative to the unconditional mean with high Nino 3.4 indices is captured by both models, along with
the lower than normal mean wet season precipitation with low Nino 3.4 values. For high Nino3.4 indices, the increase is on the order of 1.2 to 1.6, and conversely for low indices, the reduction is near 0.8. The models suggest some within region variability. However, in particular for the SIMOROP model, the region outside of the mountainous area, the model does not simulate significant precipitation.

4.6 Conclusions

The paper has inter-compared the spatial and temporal characteristics of long-term simulations of orographically-driven precipitation over southern California under three different models of varying complexity. The objective is to capture the spatial variability of precipitation at high resolution (O[10s of km²]) and at hourly timescales to support analysis of flash flood occurrence in the region. The first model is based on precipitation observations and uses a spatial interpolation method that utilizes climatological relationships of precipitation and terrain features. The second model is a full mesoscale numerical model of the atmosphere while the third model is a simplified numerical precipitation model aiming to reproduce orographic precipitation in the region.

The main conclusions are:

1) Models exhibit similarities in the distribution of precipitation over the southern California region, reflecting their consideration of the relationship of precipitation with the mountainous terrain.
2) Differences in the model representation of precipitation stem from limitations in the spatial distribution of observation stations, model input and spatial discretization scales, and model consideration of the physics and dynamics involved.

3) Sensitivity analyses of the simplified orographic model demonstrate variation in both precipitation magnitude and spatial extent under model configuration and model parameterization changes. The sensitivity analysis notes the trade-off between representing long-term mean versus individual storm statistics.

4) Comparison of the numerical model estimates to estimates of mean areal precipitation based on observations for specific watersheds indicate good reproduction of climatologies and their spatial variability. Results show cross-correlations that reach 0.5(0.7) for hourly (daily) precipitation.

5) Significant spatial variability of small scale precipitation rates is estimated for southern California, which suggests that existing observation stations are not adequate to reproduce it.

6) The numerical models produce the association of wet season precipitation volume with ENSO indices similar to that found for observations of seasonal precipitation from raingauge stations.

A significant outcome of this research is the development of 3 long-term, high resolution, hourly datasets of precipitation over the southern California region. Using all three estimates represents capability of a multi-model ensemble approach for further analysis of regional precipitation at high resolution.
A significant and fruitful next research effort to complement this work is to utilize the high resolution precipitation datasets to examine flash flood occurrence in the southern California region. Even though the numerical models may overestimate or underestimate precipitation on certain regions or times, because of the interaction with the land surface that acts as a buffer to precipitation peaks and because flash flood occurrence may happen with modest precipitation rates for soils near saturation, it is likely that all models could be useful for use in flash flood applications.

With respect to the sensitivity of the simplified orographic model, additional future research could pursue analysis of the model simulations to other observed sources (e.g., radar or satellite precipitation) or to other numerical model simulations (e.g., Weather Research and Forecasting model, WRF) at commensurate spatial resolution. The nature of data available and numerical modeling would limit such comparison to shorter analysis periods (decadal period or event-based).

4.7 References


Table 4.1. Atmospheric moisture model constant initial values.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>(\alpha)</td>
<td>0.3</td>
<td>(k_1)</td>
<td>(1 \times 10^{-4})</td>
</tr>
<tr>
<td>(K)</td>
<td>0</td>
<td>(k_2)</td>
<td>(6.96 \times 10^{-4})</td>
</tr>
<tr>
<td>(E)</td>
<td>1.0</td>
<td>(k_3)</td>
<td>(-1.93 \times 10^{-6})</td>
</tr>
<tr>
<td>(N_0)</td>
<td>(10^7)</td>
<td>(k_5)</td>
<td>(5 \times 10^{-5})</td>
</tr>
</tbody>
</table>

Table 4.2. Sensitivity runs statistics for simple geometry case.

<table>
<thead>
<tr>
<th>Sensitivity Case</th>
<th>Maximum Precip (mm)</th>
<th>Area (frac)</th>
<th>Area P &gt; 2mm</th>
<th>Area P &gt; 10mm</th>
<th>Area P &gt; 30mm</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Idealized</strong></td>
<td>64.0</td>
<td>0.098</td>
<td>0.021</td>
<td>0.011</td>
<td>0.003</td>
</tr>
<tr>
<td><strong>2nd Moist Layer</strong></td>
<td>63.9</td>
<td>0.103</td>
<td>0.024</td>
<td>0.011</td>
<td>0.003</td>
</tr>
<tr>
<td><strong>Drier Sounding</strong></td>
<td>59.4</td>
<td>0.071</td>
<td>0.014</td>
<td>0.007</td>
<td>0.002</td>
</tr>
<tr>
<td><strong>Efficiency ((\alpha=0.3))</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>E=1.0</td>
<td>64.0</td>
<td>0.098</td>
<td>0.021</td>
<td>0.011</td>
<td>0.003</td>
</tr>
<tr>
<td>E=0.4</td>
<td>36.6</td>
<td>0.101</td>
<td>0.021</td>
<td>0.010</td>
<td>0.001</td>
</tr>
<tr>
<td>E=0.2</td>
<td>29.7</td>
<td>0.100</td>
<td>0.021</td>
<td>0.010</td>
<td>0.000</td>
</tr>
<tr>
<td><strong>Autoconversion Threshold (E=0.4)</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>(\alpha=0.6)</td>
<td>29.6</td>
<td>0.06</td>
<td>0.016</td>
<td>0.006</td>
<td>0.000</td>
</tr>
<tr>
<td>(\alpha=0.3)</td>
<td>36.6</td>
<td>0.101</td>
<td>0.021</td>
<td>0.010</td>
<td>0.001</td>
</tr>
<tr>
<td>(\alpha=0.1)</td>
<td>39.9</td>
<td>0.124</td>
<td>0.026</td>
<td>0.014</td>
<td>0.001</td>
</tr>
</tbody>
</table>
Table 4.3 Comparison of 700mb wind characteristics from NCEP-NCAR reanalysis (NNR) node and radiosonde station NKX.

<table>
<thead>
<tr>
<th>Wind direction (° from N)</th>
<th>NNR FREQ (48-05)</th>
<th>NNR Avg Vel (48-05)</th>
<th>NNR FREQ (94-05)</th>
<th>NNR Avg Vel (94-05)</th>
<th>NKX FREQ (94-05)</th>
<th>NKX Avg Vel (94-05)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0-15</td>
<td>0.049</td>
<td>9.49</td>
<td>0.047</td>
<td>9.72</td>
<td>0.041</td>
<td>8.82</td>
</tr>
<tr>
<td>15-30</td>
<td>0.035</td>
<td>8.26</td>
<td>0.035</td>
<td>8.15</td>
<td>0.036</td>
<td>8.08</td>
</tr>
<tr>
<td>30-45</td>
<td>0.024</td>
<td>7.13</td>
<td>0.023</td>
<td>7.36</td>
<td>0.028</td>
<td>8.06</td>
</tr>
<tr>
<td>45-60</td>
<td>0.019</td>
<td>6.52</td>
<td>0.019</td>
<td>6.93</td>
<td>0.020</td>
<td>8.80</td>
</tr>
<tr>
<td>60-75</td>
<td>0.017</td>
<td>6.13</td>
<td>0.013</td>
<td>6.04</td>
<td>0.017</td>
<td>7.43</td>
</tr>
<tr>
<td>75-90</td>
<td>0.014</td>
<td>5.81</td>
<td>0.009</td>
<td>5.43</td>
<td>0.016</td>
<td>6.02</td>
</tr>
<tr>
<td>90-105</td>
<td>0.015</td>
<td>5.46</td>
<td>0.011</td>
<td>5.27</td>
<td>0.014</td>
<td>5.25</td>
</tr>
<tr>
<td>105-120</td>
<td>0.014</td>
<td>5.51</td>
<td>0.012</td>
<td>6.11</td>
<td>0.013</td>
<td>5.56</td>
</tr>
<tr>
<td>120-135</td>
<td>0.013</td>
<td>5.40</td>
<td>0.012</td>
<td>4.95</td>
<td>0.011</td>
<td>6.85</td>
</tr>
<tr>
<td>135-150</td>
<td>0.014</td>
<td>5.31</td>
<td>0.011</td>
<td>4.68</td>
<td>0.014</td>
<td>6.23</td>
</tr>
<tr>
<td>150-165</td>
<td>0.015</td>
<td>5.92</td>
<td>0.014</td>
<td>5.71</td>
<td>0.015</td>
<td>6.63</td>
</tr>
<tr>
<td>165-180</td>
<td>0.016</td>
<td>7.09</td>
<td>0.015</td>
<td>6.43</td>
<td>0.019</td>
<td>7.60</td>
</tr>
<tr>
<td>180-195</td>
<td>0.021</td>
<td>7.58</td>
<td>0.018</td>
<td>7.96</td>
<td>0.019</td>
<td>8.97</td>
</tr>
<tr>
<td>195-210</td>
<td>0.026</td>
<td>8.54</td>
<td>0.026</td>
<td>9.60</td>
<td>0.030</td>
<td>8.91</td>
</tr>
<tr>
<td>210-225</td>
<td>0.034</td>
<td>10.36</td>
<td>0.032</td>
<td>10.05</td>
<td>0.036</td>
<td>10.73</td>
</tr>
<tr>
<td>225-240</td>
<td>0.050</td>
<td>10.92</td>
<td>0.048</td>
<td>11.23</td>
<td>0.052</td>
<td>10.80</td>
</tr>
<tr>
<td>240-255</td>
<td>0.059</td>
<td>11.27</td>
<td>0.059</td>
<td>11.55</td>
<td>0.071</td>
<td>12.18</td>
</tr>
<tr>
<td>255-270</td>
<td>0.067</td>
<td>11.66</td>
<td>0.071</td>
<td>11.76</td>
<td>0.073</td>
<td>11.51</td>
</tr>
<tr>
<td>270-285</td>
<td>0.077</td>
<td>11.13</td>
<td>0.082</td>
<td>11.45</td>
<td>0.095</td>
<td>12.60</td>
</tr>
<tr>
<td>285-300</td>
<td>0.085</td>
<td>11.59</td>
<td>0.092</td>
<td>11.35</td>
<td>0.098</td>
<td>11.55</td>
</tr>
<tr>
<td>300-315</td>
<td>0.090</td>
<td>11.81</td>
<td>0.098</td>
<td>11.58</td>
<td>0.091</td>
<td>11.69</td>
</tr>
<tr>
<td>315-330</td>
<td>0.094</td>
<td>11.85</td>
<td>0.102</td>
<td>11.84</td>
<td>0.083</td>
<td>11.42</td>
</tr>
<tr>
<td>330-345</td>
<td>0.085</td>
<td>11.21</td>
<td>0.083</td>
<td>11.10</td>
<td>0.060</td>
<td>9.52</td>
</tr>
<tr>
<td>345-360</td>
<td>0.071</td>
<td>10.37</td>
<td>0.070</td>
<td>10.15</td>
<td>0.050</td>
<td>9.33</td>
</tr>
</tbody>
</table>

Table 4.4. Southern California SIMOROP model sensitivity analyses.

<table>
<thead>
<tr>
<th>Sensitivity Case</th>
<th>Run Name</th>
<th>Vertical Resolution</th>
<th>Wind Level (hPa)</th>
<th>K5 value</th>
<th>Input Moisture</th>
<th>Data Source</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Baseline</strong></td>
<td>Baseline</td>
<td>500m</td>
<td>700</td>
<td>0.</td>
<td>Single</td>
<td>Reanalysis</td>
</tr>
<tr>
<td><strong>Vertical Res</strong></td>
<td>Vres350</td>
<td>350m</td>
<td>700</td>
<td>0.</td>
<td>Single</td>
<td>Reanalysis</td>
</tr>
<tr>
<td></td>
<td>Vres1k</td>
<td>1000m</td>
<td>700</td>
<td>0.</td>
<td>Single</td>
<td>Reanalysis</td>
</tr>
<tr>
<td><strong>Forcing Wind</strong></td>
<td>Wlvl850</td>
<td>500m</td>
<td>850</td>
<td>0.</td>
<td>Single</td>
<td>Reanalysis</td>
</tr>
<tr>
<td></td>
<td>Wlvl500</td>
<td>500m</td>
<td>500</td>
<td>0.</td>
<td>Single</td>
<td>Reanalysis</td>
</tr>
<tr>
<td><strong>Distrib Moist</strong></td>
<td>Distrb4</td>
<td>500m</td>
<td>700</td>
<td>0.</td>
<td>4 nearest NCEP nodes</td>
<td>Reanalysis</td>
</tr>
<tr>
<td></td>
<td>DistrbC</td>
<td>500m</td>
<td>700</td>
<td>0.</td>
<td>4 central NCEP nodes</td>
<td>Reanalysis</td>
</tr>
<tr>
<td><strong>Enviro. Mix</strong></td>
<td>EnvMix 1</td>
<td>500m</td>
<td>700</td>
<td>1.0e-4</td>
<td>Single</td>
<td>Reanalysis</td>
</tr>
<tr>
<td></td>
<td>EnvMix 2</td>
<td>1000m</td>
<td>700</td>
<td>1.0e-4</td>
<td>Single</td>
<td>Reanalysis</td>
</tr>
<tr>
<td></td>
<td>EnvMix 3</td>
<td>1000m</td>
<td>700</td>
<td>5.0e-5</td>
<td>Single</td>
<td>Reanalysis</td>
</tr>
<tr>
<td><strong>Data Source</strong></td>
<td>NKXforce</td>
<td>1000m</td>
<td>700</td>
<td>5.0e-5</td>
<td>Single</td>
<td>Radiosonde (NKX)</td>
</tr>
</tbody>
</table>
Table 4.5. Comparison of validation measures for sensitivity runs.

<table>
<thead>
<tr>
<th>Sensitivity Case</th>
<th>Frac0* (obs)</th>
<th>Frac0* (sim)</th>
<th>Hit0*</th>
<th>Avg Max D(^1)</th>
<th>Avg CCOR (all)</th>
<th>Avg Bias (all)</th>
<th>Avg CCOR (p_{obs}&gt;10)</th>
<th>Avg aBias (p_{obs}&gt;10)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Baseline</td>
<td>0.775</td>
<td>0.635</td>
<td>0.739</td>
<td>0.177</td>
<td>0.458</td>
<td>0.194</td>
<td>0.273</td>
<td>-0.510</td>
</tr>
<tr>
<td>Vres350</td>
<td>0.713</td>
<td>0.814</td>
<td>0.208</td>
<td>0.445</td>
<td>-0.221</td>
<td>0.296</td>
<td>-0.639</td>
<td></td>
</tr>
<tr>
<td>Vres1k</td>
<td>0.649</td>
<td>0.758</td>
<td>0.155</td>
<td>0.484</td>
<td>0.783</td>
<td>0.340</td>
<td>-0.154</td>
<td></td>
</tr>
<tr>
<td>Wlvl850</td>
<td>0.722</td>
<td>0.828</td>
<td>0.260</td>
<td>0.503</td>
<td>-0.241</td>
<td>0.309</td>
<td>-0.642</td>
<td></td>
</tr>
<tr>
<td>Wlvl500</td>
<td>0.573</td>
<td>0.675</td>
<td>0.177</td>
<td>0.432</td>
<td>0.751</td>
<td>0.251</td>
<td>-0.371</td>
<td></td>
</tr>
<tr>
<td>Distrib4</td>
<td>0.448</td>
<td>0.526</td>
<td>0.187</td>
<td>0.511</td>
<td>0.702</td>
<td>0.324</td>
<td>-0.351</td>
<td></td>
</tr>
<tr>
<td>DistribC</td>
<td>0.444</td>
<td>0.522</td>
<td>0.187</td>
<td>0.515</td>
<td>0.745</td>
<td>0.327</td>
<td>-0.328</td>
<td></td>
</tr>
<tr>
<td>EnvMix,1</td>
<td>0.814</td>
<td>0.892</td>
<td>0.294</td>
<td>0.430</td>
<td>-0.596</td>
<td>0.220</td>
<td>-0.820</td>
<td></td>
</tr>
<tr>
<td>EnvMix,2</td>
<td>0.738</td>
<td>0.880</td>
<td>0.223</td>
<td>0.473</td>
<td>-0.330</td>
<td>0.331</td>
<td>-0.636</td>
<td></td>
</tr>
<tr>
<td>EnvMix,3</td>
<td>0.758</td>
<td>0.858</td>
<td>0.202</td>
<td>0.472</td>
<td>-0.206</td>
<td>0.329</td>
<td>-0.584</td>
<td></td>
</tr>
<tr>
<td>NNX Forcing</td>
<td>0.746</td>
<td>0.829</td>
<td>0.199</td>
<td>0.229</td>
<td>0.257</td>
<td>0.038</td>
<td>-0.599</td>
<td></td>
</tr>
</tbody>
</table>

\(+ \text{ Frac0} = \text{the fraction of days with no precipitation}\)

\(* \text{Hit0} = \text{the fraction of coincident days with no precipitation both in observations and simulation}\)

\(^1\) \text{MaxD} = \text{maximum difference between cumulative frequency distributions of non-zero precipitation.}\)
Table 4.6. Average validation measures for initial 57-year historical run.

<table>
<thead>
<tr>
<th></th>
<th>Frac0* (obs)</th>
<th>Frac0* (sim)</th>
<th>Hit0*</th>
<th>Avg Max D1</th>
<th>Avg CCOR (all)</th>
<th>Avg Bias (all)</th>
<th>Avg CCOR (p_{obs}&gt;10)</th>
<th>Avg aBias (p_{obs}&gt;10)</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Historical</strong></td>
<td>0.838</td>
<td>0.772</td>
<td>0.885</td>
<td>0.129</td>
<td>0.433</td>
<td>0.270</td>
<td>0.246</td>
<td>-0.466</td>
</tr>
</tbody>
</table>
Table 4.7. Comparison of monthly simulated precipitation for model parameterization variations.

<table>
<thead>
<tr>
<th>Sensitivity Case</th>
<th>Jan 2005 Peak Precip</th>
<th>Jan 2005 Area, P &gt; 10 mm</th>
<th>Feb 2005 Peak Precip</th>
<th>Feb 2005 Area, P &gt; 10 mm</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Model Resolution 3km</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>E=1.0, a=0.3</td>
<td>4000 mm</td>
<td>0.48 (0.24)</td>
<td>2800 mm</td>
<td>0.52 (0.25)</td>
</tr>
<tr>
<td>E=0.4, a=0.3</td>
<td>2800</td>
<td>0.46 (0.23)</td>
<td>2350</td>
<td>0.50 (0.24)</td>
</tr>
<tr>
<td>E=0.1, a=0.3</td>
<td>1900</td>
<td>0.46 (0.22)</td>
<td>1800</td>
<td>0.50 (0.23)</td>
</tr>
<tr>
<td><strong>Model Resolution 5km</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>E=0.4, a=0.3</td>
<td>2350</td>
<td>0.47 (0.23)</td>
<td>2200</td>
<td>0.54 (0.25)</td>
</tr>
<tr>
<td>E=0.1, a=0.3</td>
<td>1700</td>
<td>0.46 (0.22)</td>
<td>1750</td>
<td>0.53 (0.24)</td>
</tr>
<tr>
<td>E=0.1, a=0.6</td>
<td>1500</td>
<td>0.38 (0.17)</td>
<td>1550</td>
<td>0.45 (0.20)</td>
</tr>
<tr>
<td>E=0.1, a=0.1</td>
<td>1900</td>
<td>0.55 (0.26)</td>
<td>1900</td>
<td>0.60 (0.28)</td>
</tr>
<tr>
<td><strong>Model Resolution 10km</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>E=0.4, a=0.3</td>
<td>1600</td>
<td>0.47 (0.20)</td>
<td>1250</td>
<td>0.55 (0.22)</td>
</tr>
<tr>
<td>E=0.1, a=0.3</td>
<td>1450</td>
<td>0.46 (0.19)</td>
<td>1200</td>
<td>0.54 (0.21)</td>
</tr>
<tr>
<td><strong>Model Resolution 5km, 600m vertical</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>E=0.4, a=0.3</td>
<td>1500</td>
<td>0.52 (0.26)</td>
<td>1350</td>
<td>0.62 (0.30)</td>
</tr>
<tr>
<td>E=0.1, a=0.3</td>
<td>1250</td>
<td>0.52 (0.25)</td>
<td>1200</td>
<td>0.62 (0.30)</td>
</tr>
</tbody>
</table>
Table 4.8. Small watershed characteristics and MAP statistics.

<table>
<thead>
<tr>
<th>Basin (USGS ID)</th>
<th>A (km$^2$)</th>
<th>Elev (m)</th>
<th>Bias</th>
<th>$\rho$ (day/ hour)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Pintado 11128250</td>
<td>106</td>
<td>507</td>
<td>-0.08</td>
<td>0.65 / 0.42</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>0.15</td>
<td>0.64 / 0.38</td>
</tr>
<tr>
<td>Atascadero 11120000</td>
<td>48.9</td>
<td>285</td>
<td>-0.70</td>
<td>0.58 / 0.35</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>0.33</td>
<td>0.77 / 0.50</td>
</tr>
<tr>
<td>Sespe 11111500</td>
<td>165</td>
<td>1485</td>
<td>-0.29</td>
<td>0.76 / 0.52</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>-0.52</td>
<td>0.54 / 0.30</td>
</tr>
<tr>
<td>Lone Pine 11063500</td>
<td>65</td>
<td>1452</td>
<td>-0.70</td>
<td>0.53 / 0.27</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>-0.47</td>
<td>0.46 / 0.23</td>
</tr>
<tr>
<td>Deep Creek 10260500</td>
<td>406</td>
<td>1810</td>
<td>0.23</td>
<td>0.59 / 0.33</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>-0.58</td>
<td>0.30 / 0.15</td>
</tr>
<tr>
<td>Arroyo Trabuco 11047300</td>
<td>140</td>
<td>437</td>
<td>-0.30</td>
<td>0.52 / 0.27</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>0.20</td>
<td>0.67 / 0.36</td>
</tr>
<tr>
<td>Temecula Creek 11042400</td>
<td>339</td>
<td>1105</td>
<td>0.56</td>
<td>0.54 / 0.31</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>-0.09</td>
<td>0.71 / 0.41</td>
</tr>
<tr>
<td>Sweetwater R 11015000</td>
<td>270</td>
<td>1070</td>
<td>0.59</td>
<td>0.51 / 0.30</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>-0.68</td>
<td>0.41 / 0.19</td>
</tr>
</tbody>
</table>

Bias and correlation for SIMOROP (top number in each cell) and CaRD10 (lower number) models relative to the GRIDOBS model.
Figure 4.1. Southern California study region. Major mountain ranges are identified in boldface type; County boundaries and names are in gray.
Figure 4.2. Location of hourly precipitation station used in development of GRID OBS model
Figure 4.3. Illustration of SIMOROP wind model configuration.
Figure 4.4. Simple mountain geometry example for SIMOROP model: (a) elevation, (b) updraft velocities, and (c) simulated precipitation (mm/6hour).
Figure 4.5. NCEP-NCAR Reanalysis monthly average specific humidity ($q_s$) for two nodes used to derive input to SIMOROP model. Dashed lines indicate average over times wind free air stream velocity, taken as 700hPa height) is from southwest quadrangle.
Figure 4.6. Average wet season total precipitation (10/1948 – 4/2005): (a) GRIDOBS model, (b) SIMOROP model, and (c) CaRD10 model.
Figure 4.6. Average wet season total precipitation (10/1948 – 4/2005), Continued.
Figure 4.7. Average October (left column) and February (right column) precipitation for 3 models.
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Figure 4.9. Average frequency of precipitation exceeding 0.25 mm/hr (left column) and 5.0 mm/hr (right column).
Figure 4.10. Observed frequency of exceeding range of precipitation thresholds at 12 select stations during month of February.
Figure 4.11. Comparison of observed and simulated average frequency of exceeding select precipitation thresholds of precipitation for month of February.
Figure 4.12. Wet season cycle in precipitation (average in solid lines, standard deviation in dashed lines) for selected small watersheds.
Figure 4.13. Ratio of average wet season total precipitation to unconditional mean for years with OND Nino 3.4 index in its upper tercile (upper row) and in its lower tercile (lower row).
Chapter 5. Soil Moisture Modeling and Flash Flood Occurrence Simulation

5.1 Introduction

Soil moisture is often a key state and of fundamental interest in hydrologic modeling. It has a strong influence on a range of hydrologic processes including flooding, erosion, ecologic impacts, and land surface-atmosphere interactions. Although soil moisture has high spatial variability and knowledge of such variability is important to understanding the hydrologic processes mentioned (Western and Blosch, 1999), there remains a lack of large scale observations of soil moisture through various soil depths (e.g., Georgakakos and Baumer, 1996). Hydrologic models representing soil moisture conditions are often calibrated using observed streamflow and are used to simulate soil moisture conditions in a watershed (e.g., Reed et al, 2004; Winchell et al, 1998; Boyle et al; 2000) or are used to infer soil moisture characteristics with limited soil moisture observations (e.g., Georgakakos and Carpenter, 2006, Koren et al, 2008).

Of interest herein is the role of soil moisture in runoff generation that leads to flash flooding. Soil moisture content and evolution greatly influences the division between slower responding subsurface runoff and base flow, and the faster responding surface runoff generation. While soil moisture supplies deep soil and ground water, generally precipitation on saturated soils produces higher surface runoff volume. Significant production of surface runoff may lead to increased potential for flash flooding if rapid confluence of runoff volume exceeds the capacity of the receiving stream channel to carry that volume.

Flash floods are defined by a rapid rise and fall of flow following heavy or
intense rain, usually within a few hours of the causative rainfall events, and are characterized by small spatial scales (AMS, 2000; NWS, 2005a). Flash floods have the dubious distinction of being among the deadliest weather related hazards. In the U.S., nearly 80% of flood-related fatalities are associated with flash floods (NWS, 2005b), and on a global scale, the ratio of flash flood fatalities to number of people affected is more than 4 times higher than any other type of flooding (Jonkman, 2005).

The occurrence of heavy precipitation is not the only causal ingredient for flash floods. Land surface factors contribute as well, including topography, soil properties and soil moisture conditions (e.g., Hoyt and Langbein, 1939; Georgakakos, 1986). O’Connor and Costa (2004) examined observations of streamflow across the U.S. and identified regions with high unit discharge (discharge divided by drainage area, a measure of runoff generated over a watershed area). They note that among the characteristics of regions with high unit discharges are abundant atmospheric moisture sources, topography, and significant relief that lead to rapid concentration of streamflow. The Transverse Mountain Range in southern California was identified as having high unit discharge and these about characteristics in the study.

Analysis of historical flash flood occurrence has largely been limited to event-specific studies (e.g., Ogden et al, 2000; Li et al, 2003; Gaume et al, 2004), forecasting capabilities (e.g., Collier, 2007; Javier et al, 2007; NRC, 2005; Doswell et al, 1996), and relatively recently, warning effectiveness (e.g., Drobot and Parker, 2007; Gruntfest and Ripps, 2000). Relatively few studies examine climatological aspects of flash flooding, and those studies have been focused on meteorological characteristics (e.g., Maddox et al, 1979; Laing, 2004; Ntelekos et al, 2007). Considering climatologies on a national
basis published studies indicate that flash flood occurrence is 20 times less frequent than the occurrence of heavy precipitation (Brooks and Stensrud, 2000; Maddox et al, 1979), which highlights the importance of the land surface in flash flood production.

Flash floods occur over small spatial scales, typically for drainage areas less than $500\text{km}^2$ given the rapid development of flow volume (e.g., within 6-12 hours of rainfall). At such scales, hydrologic response is often affected by small scale inhomogeneity in the terrestrial properties including soil characteristics (e.g., Beven, 1989, 2001). Characterization of hydrologic model input that accounts for the spatial variation in both precipitation and land surface properties, and the development of distributed hydrologic models and associated a priori parameterizations have been an active research areas (e.g., Beven and Hornberger, 1982; Finnerty et al, 1997; Smith et al, 2004; Beven, 2001; Koren et al, 2004; Reed et al, 2004; Duan et al, 2000). The move towards the use of distributed hydrologic models for small watersheds introduces issues of scale. Carpenter and Georgakakos (2004b, 2006) demonstrate that hydrologic model response to uncertain parameters and input forcing is scale dependent, with the uncertainty in response increasing substantially at smaller scales. This may be particularly relevant to modeling hydrologic response of flash flood occurrence at small spatial scales.

This chapter presents a final step in the development of a multi-disciplinary modeling approach to characterize flash flood occurrence frequency in the mountainous region of southern California. The aim of this chapter is two-fold: (a) to simulate soil moisture conditions throughout the region, providing insight to spatial variability of long term soil moisture in the region; and (b) to develop estimates of flash flood
occurrence frequencies under a multi-model ensemble approach to consider uncertainty in both precipitation input and model parameterization. The effort is among the first attempts to characterize regional flash flood occurrence potential with high spatial detail through modeling. The utility of the modeling approach is then demonstrated through an application to characterize potential changes in the frequency of flash flood occurrence potential under a changing climate.

The chapter is organized as follows. The southern California study region and terrestrial characteristics are introduced in Section 5.2. Section 5.3 describes two hydrologic models and their parameterizations used in the research as members of a multi-model ensemble. Section 5.4 illustrates the application of the two hydrologic models for selected focus watersheds in terms of reproduction of observed streamflow under regional adjustments to a priori parameterizations. The regional application of the models to simulate soil moisture characteristics for small watersheds is presented in Section 5.5. Section 5.6 presents the synthesis of the modeling components described throughout this thesis towards estimation of flash flood occurrence frequency. This includes discussion on the uncertainty in historical frequency estimation under the multi-model ensemble and the change in simulated flash flood occurrence frequency under potential climate change. Conclusions and future extensions are presented in Section 5.7.

5.2 Study Region

The focus area for this work is the cismontane region of the Southern California Bight, bounded by the Transverse Mountain Range to the north and the Peninsular
Mountain Range to the east (see Figure 5.1). This region is generally described as having a Mediterranean climate with two distinct seasons: a moderately cool, wet season (October-April) and a dry warm season (May-September). Annual rainfall is 200 mm along the coastal regions and reaches 1 m in the mountains. Beyond the mountains and to the east/northeast, the climate becomes arid, and includes the southern portion of the Mojave Desert.

The southern California coastal region is residence to nearly two-thirds of the State’s population, including the large metropolitan cites of Los Angeles, San Diego, San Bernardino, Ventura, and Santa Barbara. Indeed, the presence of large population centers means that exposure to hydrologic hazards such as flash flooding remains high, even though such hazards may be relatively infrequent, especially where the population has concentrated along the mountain foothills. Large urban areas cover the coastal region from Santa Barbara to Los Angeles and southward through Orange County to San Diego and eastward to Riverside and San Bernardino. The natural drainage network has been altered in these highly urbanized areas through channelization and other technological features. However, this study is concerned with small natural watersheds within the mountain-to-foothill regions that may be the source of flash floods that exposure the urban corridor to risk.

These small watersheds were delineated for the study region basin from digital elevation data (90-m SRTM data, Jarvis et al, 2008), and are shown in Figure 5.2. The small watersheds are sub-basin units of larger watersheds which drain into the Pacific Ocean. A few headwater basins which flow into the Mojave River in the north and the Whitewater River to the east are included in the San Bernardino Mountains. A total of
975 individual basins are delineated, with an average local drainage area of 30km$^2$. The cumulative drainage area for any watershed is limited to ~4,500 km$^2$. However, this limit was reached only for subbasins along the Santa Clara River in Ventura County.

Figure 5.2 shows several physical features of the watersheds. The channel slopes indicate fairly steep slopes reaching 0.25 m/m. Surface soil texture and depth to bedrock are derived from the STATSGO database (NRCS, 1994). Soil texture, depth, and other soil properties are given for various STATSGO ‘map units’ and map unit components in the database. The presented values were computed as (a) the area-dominant soil texture in the upper 20cm of soil depth, and (b) average depth to bedrock over the map units which were identified within each small watershed. The surface soil texture is classified primarily as sandy-loam and loam. A few regions exhibit clay-loam surface texture, particularly in the western Transverse Range. The depth to bedrock ranges from about 40 mm to over 1.5 m. The lower depths are primarily found along the Transverse Mountains and in the southern portion of the Peninsular Range.

Figure 5.3 presents land cover and geologic features of southern California. The land cover data (Figure 5.3b) is taken from the 2000 Southern Coastal California Land Cover/Land Use dataset from the National Ocean and Atmospheric Administration’s (NOAA) Coastal Services Center (NOAA, 2010; see also Dobson et al, 1995). Outside of the urban areas near the coast, the land cover is primarily chaparral along the foothill to mountainous regions, with a transition to deciduous and evergreen forest at higher elevation. Grassland cover is interspersed in the chaparral regions along with cropland and orchards, particularly toward coastal Ventura County. Forest cover is more extensive in Santa Barbara and northern Ventura Counties, while it is more spatially
limited in the eastern Transverse and Peninsular Ranges. Many of these forested regions are part of the Los Padres, Angeles, and Cleveland National Forests. The geologic map (Figure 5.3a; USGS, 2010) shows similarities between in the eastern Transverse and Peninsular Ranges with both consisting of granitic rock of Mesozoic age, with some pockets of older metamorphic and sedimentary rock. The western Transverse Range consists mostly of tertiary sedimentary rock with regions of Mesozoic sedimentary rock. These features are similar to the coastal region of Orange and San Diego Counties and suggest availability of groundwater storage and potential strong links with the surface water processes.

5.3 Hydrologic Model Description

This research employs two hydrologic models for the simulation of soil moisture and for the estimation of flash flood occurrence for small watersheds throughout the southern California mountain-to-foothill region: (a) a detailed conceptual model used in routine forecast operations by the U.S. National Weather Service; and (b) a two-layer physical process-based model with simplified parameterizations. This section describes the models, their parameterizations, and input forcing data. The ability of both models to reproduce observed streamflow for select stations in southern California is presented in Section 5.4.

5.3.1 Sacramento Soil Moisture Accounting Model

5.3.1.1 SAC-SMA Description
The Sacramento soil moisture accounting model (SAC-SMA; Burnash et al, 1973; Burnash, 1995), is a conceptual hydrologic model describing the movement of moisture through a soil column and production of runoff over a watershed of interest which is input to the stream channel draining the watershed. It is frequently used operationally by the U.S. National Weather Service for the forecasting of streamflow for watersheds that range from 100s to several 1000s km$^2$ in area. A time-continuous adaptation of the operational model is employed herein (Georgakakos, 1986).

The SAC-SMA model conceptualizes the soil column as a two layer system depicted in Figure 5.4, and treats soil moisture as the depth integrated water volume in each layer. Each layer contains of tension and free water components, and allows different paths for runoff generation. The upper layer is relatively thin and fast responding, while the lower layer is thicker and represents the bulk of the soil moisture storage with slower response. Tension water components are bound closely to the soil particles and their contents are removed only through evapotranspiration. Free water moves freely in the soil column under gravitational forces. The lower zone free water is divided into two storage components to allows for complex of hydrologic response by representing both fast responding baseflow (supplemental) and slow, long lasting baseflow (primary component). Thus, there are five model storage components or model states indicated by $X_1$ to $X_5$ in Figure 5.4. A sixth state is defined that represents water content in the areas around the streams and other surface water bodies that becomes impervious as the tension water components approach saturation. The storage capacities of each storage component are model parameters and represented by $X_1^0$ to
The arrows in Figure 5.4 represent the soil moisture flow paths within the model and model parameters are indicated in parenthesis.

Inputs to the model are watershed averaged precipitation ($P$) and evapotranspiration demand ($etd$). Precipitation enters the model at the upper zone tension water component ($X_1^0$). A fraction of the watershed may be specified as impervious (e.g., densely constructed urban areas) and precipitation on the impervious area produces direct surface runoff. Actual evapotranspiration is computed based on the demand and moisture availability in the tension water components. The evapotranspiration from the upper zone tension water component is given by:

\[
ET_1 = etd \times \left( \frac{X_1}{X_1^0} \right)
\]  

(1)

where $X_1$ represents the model state of the upper tension water component at a given time step and $X_1^0$ is the capacity of the upper zone tension water, both having units of depth of water. Once the upper zone tension water component reaches capacity, soil moisture moves into the upper zone free water storage component. Thus, the equation for the evolution of the upper zone tension water state is:

\[
\frac{dX_1}{dt} = \left[ 1 - \left( \frac{X_1}{X_1^0} \right)^{m1} \right] P - etd \left( \frac{X_1}{X_1^0} \right)
\]  

(2)

where the terms $X_1$, $P$, and $etd$ are time-varying. The second term on the right hand side represents the overflow of soil moisture from upper zone tension to the free
component, and the parameter $m_1$ is given a large value to activate this flux when the upper zone tension reaches saturation. The equation of state for the upper zone free water component, $X_2$, expressed also in units of water depth, is:

$$\frac{dX_2}{dt} = P \left( \frac{X_1}{X_1^0} \right)^{m_1} \left[ 1 - \left( \frac{X_2}{X_2^0} \right)^{m_2} \right] - d_u X_2 - PERC$$  \hspace{1cm} (3)

Here, the first term on the right hand side represents the influx of moisture coming from the upper zone tension component. The second term, or the influx multiplied by $\left( \frac{X_2}{X_2^0} \right)^{m_2}$, represents the flow of excess moisture in the upper zone free component (when this component approaches saturation) simulating surface runoff. Like $m_1$, the exponent $m_2$ is specified as a large value to activate this flux under saturation. The third component, $d_u X_2$, is the interflow rate and the final term is the percolation to the lower soil layer.

Percolation of moisture from the upper layer to deeper soils is a function of both the availability of moisture in the upper zone free component and the deficiency of moisture in the lower zone. Percolation, based on early empirical field studies, is defined by:

$$PERC = C_1 \left( 1 + \varepsilon y^\theta \right) \left( \frac{X_2}{X_2^0} \right)$$  \hspace{1cm} (4)

where $y$ is the lower zone deficiency ratio:
\[ y = \left( \frac{x_3 + x_4 + x_5}{x_3^0 + x_4^0 + x_5^0} \right) \]  \hspace{1cm} (5)

and \( C_1 \) is the maximum lower zone baseflow flux:

\[ C_1 = d_{l1}x_4^0 + d_{l2}x_5^0 . \]  \hspace{1cm} (6)

The parameters \( \varepsilon \) and \( \theta \) control the maximum percolation and shape of the percolation curve. The constant \( \varepsilon \) represents the maximum percolation rate with higher values allowing more moisture to transfer to the lower zone over a given time step. The exponent \( \theta \) controls the shape of the percolation curve based on the lower zone deficiency ratio. Higher \( \theta \) values require higher deficiency values (or lower saturation) to increase transfer of moisture to the lower zone.

The lower zone state (in units of water depth) equations are given in Eq. 7-10:

\[
\frac{dx_3}{dt} = (1 - p_{free}) \left[ PERC \left[ 1 - \left( \frac{x_3}{x_3^0} \right)^{m_3} \right] \right] - etd \left( 1 - \frac{x_1}{x_1^0} \right) \left( \frac{x_3}{x_1^0 + x_3^0} \right) \]  \hspace{1cm} (7)

\[
\frac{dx_4}{dt} = PERC \left[ 1 - (1 - p_{free}) \left[ 1 - \left( \frac{x_3}{x_3^0} \right)^{m_3} \right] \right] \left[ \left( C_2 \frac{x_5}{x_5^0} - 1 \right) \left( \frac{x_5}{x_5^0} \right) + 1 \right] - d_{l1}x_4  \hspace{1cm} (8)

\[
\frac{dx_5}{dt} = PERC \left[ 1 - (1 - p_{free}) \left[ 1 - \left( \frac{x_3}{x_3^0} \right)^{m_3} \right] \right] \left[ \left( 1 - C_2 \frac{x_5}{x_5^0} \right) \left( \frac{x_5}{x_5^0} \right) \right] - d_{l2}x_5  \hspace{1cm} (9)
The parameter \( p_{free} \) represents a portion of the influx to the lower zone which goes directly into the free water components, simulating faster movement of soil moisture into the active lower zone. The terms on the right hand side of the Equation 7 for the lower zone tension water component are (a) the percolation influx from the upper zone (PERC term) and (b) the excess overflow from the lower tension to lower free components when tension approaches saturation (\( X_3 \) term), multiplied by the fraction entering the tension water, and (c) the evapotranspiration loss from the lower zone tension component (etd term).

The lower zone free water components appear more complex. The first terms represent the lower zone free water component influx, multiplied by a factor to apportion the influx into the primary and supplemental storages based on the relative saturation of the storages and the maximum outflow rates. The last terms of Eq. 8 and 9 represent the primary and supplemental baseflow withdrawal from the lower zone free storages. The complex first terms portion the influx between the primary and supplemental storages. Finally, the final state equation for the water content (in units of water depth) in the area that becomes impervious around streams and surface water bodies as the tension water storages become saturated is:

\[
C_2 = \left( \frac{d_{ip}x_A^0}{d_{ip}x_A^0 + d_{ls}x_S^0} \right)
\]
A summary of the model states and parameters is given in Table 5.1. A total of five runoff components are represented in the model: (a) direct runoff from impervious area, (b) surface runoff from the upper layer exceeding saturation, (c) interflow, and (d-e) baseflow from primary and supplemental storage components. The total channel input may be expressed as:

\[
\frac{dX_6}{dt} = \left[ 1 - \left( \frac{X_6}{X_3^0} \right)^2 \right] \left[ 1 - \left( \frac{X_2}{X_1^0} \right)^m \right] \left( \frac{X_1}{X_1^0} \right)^m \frac{P}{1 - \frac{X_1}{X_1^0}} \left( \frac{X_6}{X_3^0 + X_3} \right).
\]  

(11)

where the last two terms represent direct and surface runoff generated from the additional impervious area. The parameter \( \mu \) in Eq. 12 adjusts the lower zone baseflow component for losses to deep groundwater recharge that do not appear at the outlet of the basin under consideration. The flow of recharge is equal to:

\[
gwr = \left( \frac{\mu}{1+\mu} \right) \left( d_{lp}X_4 + d_{ls}X_5 \right) \left( 1 - \beta_1 - \beta_2 \right)
\]  

(13)
5.3.1.b SAC-SMA Parameterization

Prior to model application, the parameters of the SAC-SMA model given in Table 5.1 must be estimated. The parameters are not directly observable quantities within a watershed of interest and are typically calibrated based on observed streamflow and estimates of mean areal precipitation and evapotranspiration demand. Mean areal precipitation may be estimated on the basis of point precipitation observations, while mean areal evapotranspiration demand may be estimated from surface meteorological data with adjustment factors to account for the transpiration of the plant cover of the region of interest.

At the scales of interest for this study, streamflow observations are not available at all the watershed locations defined. In fact, streamflow observations are relatively sparse, particularly in the upper or headwater regions of the watersheds in southern California. The approach taken herein is to utilize a priori parameter values based on available spatial datasets to allow for application of the model throughout southern California. Analysis of simulations at select watersheds (see Section 5.5) suggests that regional adjustment to a priori parameters yields better reproduction of observed streamflow. Thus, the approach employed for the small watersheds uses regionally-adjusted a priori parameters. This approach has implications on the simulation of flash flood frequency as discussed later in Section 5.5.

Koren et al (2000) developed a method for regional estimation of the SAC-SMA model parameters based on soil texture. Information regarding soil texture may be available based on local or regional soil surveys or in national soil survey databases including STATSGO (NRCS, 1994). Physical soil properties, including porosity ($\theta_s$),
field capacity ($\theta_f$), wilting point ($\theta_w$), and saturated hydraulic conductivity ($K_s$) have been associated with various soil texture classes (e.g., Cosby et al, 1984; Rawls and Brakensiek, 1985). Such associations are used to define the SAC-SMA model parameters in this method. The following relationships are adapted from Anderson et al (2006) and Duan et al (2001) for the model parameters.

First, the depths of the soil layers are defined. The maximum depth, $Z_{\text{max}}$, may be defined through geologic information as the depth to bedrock, or as the maximum depth of survey information. The maximum depth may be limited to represent the active soil column in terms of soil moisture fluxes, and will not include the saturated soil column or groundwater. For the upper layer, the concept of an initial abstraction and the soil curve number (CN) were employed to give:

$$Z_{up} = 5.08 \frac{10^{\frac{1000}{CN}}}{\theta_s - \theta_f}$$  \hspace{1cm} (14)

$$Z_{lo} = Z_{\text{max}} - Z_{up}$$  \hspace{1cm} (15)

Using the soil parameters as defined in Table 5.1, the capacities of the five storage zones are given by:

$$X_1^0 = (\theta_f - \theta_w)Z_{up}$$  \hspace{1cm} (16)

$$X_2^0 = (\theta_s - \theta_f)Z_{up}$$  \hspace{1cm} (17)

$$X_3^0 = (\theta_f - \theta_w)Z_{lo}$$  \hspace{1cm} (18)
Given the determined layer depths, the capacities typically are expressed in units of mm. The exponent $n$ in Equations 20 and 21 controls the proportioning of lower zone free water into the supplemental and primary storages. Duan et al (2001) present a value of 1.6 for the exponent, which yields a ratio of 3:1 for the capacity of the lower zone free primary storage to that of the supplemental storage. Next, the storage component withdrawal rates were based on empirical relationships and are given by:

\[ d_u = 1 - \left( \frac{\theta_f}{\theta_s} \right)^n \]  \hspace{1cm} (22)

\[ d_{ls} = \frac{d_u}{1 + 2(1 - \theta_w)} \]  \hspace{1cm} (23)

\[ d_{lp} = 1 - \exp \left( \frac{-\pi^2 K_s D_s^2 Z_{lo} \Delta t}{3.5 (\theta_s - \theta_f)^{1.66}} \right) \]  \hspace{1cm} (24)

These are given as daily rates (1/day). The lower zone primary baseflow withdrawal rate introduces properties of the saturated hydraulic conductivity ($K_s$), a stream channel density ($D_s$), and the time interval ($\Delta t$). The percolation parameters are estimated from the given capacities and withdrawal rates to give the maximum percolation rate for
constant \( \varepsilon \), and relative to the wilting point capacity of a sandy soil for the percolation exponent:

\[
\varepsilon = \frac{x_0^d + x_0^d (1 - d_{ls})}{d_{lp} x_0^d + d_{ls} x_0^d} + \frac{x_0^d (1 - d_{lp})}{d_{lp} x_0^d + d_{ls} x_0^d} 
\]  

(25)

\[
\theta = \left( \frac{\theta_w}{\theta_{w,SAND}} - 0.001 \right)^{0.5} 
\]  

(26)

Finally, the portion of lower zone inflow which passes directly to the lower zone free storages is estimated by:

\[
p_{free} = \left( \frac{\theta_w}{\theta_s} \right)^n 
\]  

(27)

Eq. 16-27 allow for the estimation of 11 SAC-SMA model parameters based on soil properties of porosity, field capacity, wilting point capacity, and saturated hydraulic conductivity, along with the geomorphologic property of stream channel drainage density of the watershed or region, and an estimate of the curve number to define the upper soil zone depth. Typical values of the soil properties for different soil texture classes are presented in Table 5.2. It is noted by Duan et al (2001) that the objective of the a priori parameter estimation is to provide reasonable initial estimates of the parameters, and that subsequent calibration using observed data should be performed to represent specific local watershed conditions. Reed et al. (2004) describe an international model intercomparison effort that tested a variety of model types including
the SAC-SMA model and several process based models. They report good performance of the SAC-SMA even with a priori parameter values.

5.3.1.c Development of a priori SAC-SMA model parameters for southern California

The Hydrology Laboratory of the NWS developed a national gridded dataset of a priori parameters for their River Forecast Centers operational development of forecast models (S. Reed, NWS, personal comm., 01/2002). This dataset utilized the national STATSGO soils database and employed the above formulation to derive the 11 SAC-SMA model parameters over a 1-km spatial grid. The dataset was obtained from NWS and used to estimate an a priori parameter set for all small watershed in the study region. Both average and area-dominant parameters within each watershed area were computed, typically with little numerical difference as the size of the watersheds is small compared to the STATSGO map unit discretization.

5.3.2 Flash Flood Soil Model

5.3.2.a FFSM Description

This section presents an alternative hydrologic model of soil moisture, the Flash Flood Soil Model (FFSM). FFSM is a physical process-based, conceptual model where the soil column is represented by various soil layers. In this application, two soil layers are used, similar to the SAC-SMA model, although more layers may be implemented if information on physical characteristics of watershed warrant. FFSM differs from the SAC-SMA model in that model parameterization is based on physical land surface characteristics and soil properties directly.
Figure 5.5 depicts the two-layer FFSM model and soil moisture flux paths are given by the arrows. Each layer is comprised of tension and free water storage elements with contents expressed in units of water depth. The capacities of each element are:

\[ X_{ti}^s = (\theta_f)Z_i \] (28)
\[ X_{fi}^s = (\theta_s - \theta_f)Z_i \] (29)

where \( i \) represents the layer number, \( Z_i \) is the layer depth, and \( \theta_s \) and \( \theta_f \) are the porosity and field capacity. For this study, a subscript \( u \) is used to denote the upper layer and a subscript \( l \) denotes the lower layer.

Precipitation enters the model through the upper layer tension water element. Moisture is removed by evaporation from the tension elements as:

\[ E_{tu} = \text{etd} \left( \frac{X_{tu}}{X_{tu}^s} \right) \] (30)
\[ E_{tl} = (\text{etd} - E_{tu}) \left( \frac{X_{tl}}{X_{tu}^s} \right) \] (31)

Once the tension elements become saturated, soil moisture moves into the free storage elements. This represents the filling of soil moisture to field capacity, beyond which gravity-driven soil moisture movement dominates. In the upper layer, the model allows for a portion of the incoming precipitation to move directly into the free water elements. This is controlled by the parameter \( \alpha \), and may be useful when physical information
suggests significant soil cracking or joints may exist which permits free or faster movement into the soil. The upper layer flux from tension to free water is given by:

\[ f_{tu} = (\alpha P - E_{tu}) \left( \frac{x_{tu}}{x_{tu}^s} \right)^{m \rightarrow \infty} \]  

(32)

where \((\alpha P - E_{tu})\) represents the net influx into the upper tension element. The exponent \((m \rightarrow \infty)\) indicates that this flux is activated as the tension element reaches saturation. The analogous flux for the lower layer tension element is:

\[ f_{tl} = \text{perc} \left( \frac{x_{tl}}{x_{tl}^s} \right)^{m \rightarrow \infty} \]  

(33)

where \(\text{perc}\) represents the percolation or infiltration of soil moisture to the lower layer and thus the influx into the lower tension element. Percolation is controlled both by the supply of moisture in the upper layer and the deficit of moisture in the lower layers. This is expressed as:

\[ \text{perc} = \left( \frac{x_{fu}}{x_{fu}^s} \right) \left[ 1 + c \left( 1 - \frac{x_{tl} + x_{fl}^s}{x_{tl}^s + x_{fl}^s} \right)^{mp} \right] \]  

(34)

where \(c\) and \(mp\) are parameters of the percolation function and are defined by soil texture. If the upper free water element becomes saturated and influx to the element exceeds depletion, the excess leads to the generation of surface runoff and is expressed:
By definition, excess of the lower zone free capacity is not allowed within the model and removal is through subsurface outflow (i.e., baseflow). Subsurface flow from the upper and lower free water elements are defined by:

\[
R = [f_{tu} + (1 - \alpha)P] \left( \frac{X_{fu}}{X_{fu}^s} \right)^{m \rightarrow \infty}
\]

(35)

where \( S_0 \) is the land surface slope to the channel, \( D \) is the drainage density, and \( K \) is the hydraulic conductivity. The latter is given as a function of saturated hydraulic conductivity (\( K_s \)), a property which has been related to soil texture, as:

\[
K = K_s \left( \frac{X}{X_s} \right)^{c1}
\]

(38)

where \( c1 \) is the pore disconnectedness, and typical values have been defined for different soils types. Total output from the watershed is the sum of surface runoff generated and the subsurface outflow from the free water elements:

\[
CH_{in} = R + b_u + b_l
\]

(39)
5.3.2.b FFSM Parameterization for southern California

Parameterization of the FFSM model requires estimation of the soil properties of porosity, field capacity, saturated hydraulic conductivity, and pore disconnectedness, along with the percolation parameters. Empirical studies have linked such properties values for different soil types (e.g., Cosby, et al, 1984; Rawls and Brakensiek, 1985). The values used in this study are listed in Table 5.2 by soil texture class and have been derived as noted from Anderson et al (2006), Anderson (2002), and Bras (1990). For some properties, the values given are interpolated from fewer soil texture classes.

To derive properties for the southern California watersheds, the STATSGO database was used to define the dominant soil texture within each of two soil layers: the first layer covered the top 20 cm of soil depth; the lower layer extend to a maximum depth of 1.5 m or until the occurrence of bedrock. This dominant soil texture classification was defined by the texture class with highest areal fraction when processing the various component fractions within a given STATSGO map unit and different depth layers within each component and the areal fraction of different STATSGO map units within a given small watershed. The dominant soil texture had the highest fraction out of 12 soil texture classifications. Alternatively, a definition of dominant soil texture based on the capacity of the soil to transfer soil moisture could be made. The justification for such a method is that although it may not be spatially dominant, the transfer of moisture within this soil texture is much higher than through the adjacent different soil textures. This approach was not pursued in this work.
Additionally, two land surface properties are needed in the parameterization of the FFSM model: overland slope to the channel and drainage density. These may be estimated from analysis of digital elevation models (DEM). However, such estimation is sensitive to the spatial resolution of the DEM, particularly in the case of drainage density (Stanislawski et al, 2005). Drainage density is defined as the total length of streams divided by the total drainage area. Thus, a change in spatial resolution of the DEM or mapping source or a change in the threshold to define the beginning of a stream can change the determination of the length of streams for a given drainage area.

An initial calculation of drainage density was performed for the study region using a 90-m resolution DEM and with a minimum stream area threshold of 4 km$^2$. This yielded drainage density values on the order of 0.2 to 0.8 km/km$^2$. Other work in the region suggests higher drainage density values. Azor et al (2002) reported drainage densities in the range of 4 – 6 km/km$^2$ along a specific ridge mountain ridge near Ventura. The USGS has analyzed several hundred small (< 50 km$^2$) basins in steeply sloping regions of southern California as part of their debris flow monitoring and response program (S. Cannon, USGS, personal comm., 10/2010). This analysis was based on 10-m DEM, and gave a range in drainage density values from 0.18 to 5.5 km/km$^2$. The average value was 2.8 with a standard deviation of 0.7 km/km$^2$.

Stanislawski et al (2005) presented a determination of drainage density for the conterminous U.S. based on the National Hydrography Dataset (NHD, http://nhd.usgs.gov), and also considered the effect of both minimum stream area and map resolution. The national map of drainage density based on medium resolution (1:100,000-scale) NHD indicated values ranging from 0.2 to 2.8 km/km$^2$, with the
southern California region characterized by a value of approximately 1.0 km/km². They considered the effect of resolution by the higher resolution (1:24,000-scale) NHD. Increases in drainage density on the order of 1.5 to 1.7 at specific locations are shown for the higher resolution dataset. They also indicate the dependence of computed drainage density on minimum stream area. Drainage densities increased an order of magnitude from 0.1 to 1.8 as minimum stream area decreased from 90km² to 0.1 km².

Based on the presented drainage density value for southern California and the relative increase with higher resolution datasets, a constant drainage density of 1.5 km/km² was employed throughout the study region in application of the FFSM model.

5.3.3 Model Forcing

For both hydrologic models, input forcing includes mean areal estimates of precipitation and potential evapotranspiration (PET). The precipitation estimates are derived from the models of orographic precipitation described in Chapter 4, with the following exception. As noted in Chapter 4, precipitation estimates from the CaRD10 model exhibit a consistent low bias in the mountainous regions of southern California relative to the other models. This would lead to soil water underestimation with severe implications for the surface runoff and flash flood development. Thus, a choice was made to not utilize this model’s estimates to force the hydrologic models. As part of the multi-model approach, precipitation forcing includes the two member ensemble using gridded observations and the simplified orographic precipitation model.

Potential evapotranspiration may be estimated using a variety of methods and with different temporal scales. PET may be computed using the Penman-Monteith
formula, which requires observations of atmospheric states including temperature, wind, and radiation (e.g., see Bras, 1990 or Brutsaert, 2005). When detailed observations of the atmospheric states are not available, simpler methods may be used including temperature-based rates or pan evaporation data adjusted for vegetation (Sumner and Jacobs, 2005).

Due to limited availability of atmospheric observations both spatially and temporally, for this study monthly PET values were developed. The monthly values were established using two sources of information: (a) computed reference evapotranspiration from the California Irrigation Management Information System (CIMIS), a division within the California Department of Water Resources; and (b) monthly PET values as estimated by the California-Nevada River Forecast Center (CNRFC) of the U.S. National Weather Service. These sources differ in that the former source is based on point observations of atmospheric state variables whereas the latter represent average or characteristic over watersheds of varying size.

CIMIS (CIMIS, 2010) infrastructure includes an array of automated weather observation stations measuring such variables as precipitation, air and soil temperature, winds, humidity, and solar radiation. The records include approximately 130 active stations, and archival records for inactive stations, with varying record lengths dating to the early 1980s. In 2000, CIMIS computed monthly reference ET values based on the Penman-Monteith formulation for stations having at least 5 years of record at the time. For this assessment, a total of 25 currently active CIMIS locations were selected and the CIMIS computed monthly average reference evapotranspiration were extracted. Of the selected CIMIS stations, nine were outside of the primary region of interest, located to
the east of the Peninsular or north of the Transverse Range, and most stations are located at lower elevations.

The monthly PET rates from CNRFC are derived from calibrated operational hydrologic forecast models throughout the region. CNRFC employs the SAC-SMA model for operational streamflow forecasts for numerous locations through southern California with drainage areas ranging from 80 to greater than 2000km². Calibration of the model parameters, including the establishment of PET demand rates, has the objective of reproducing historical streamflow. The CNRFC PET values represent watershed integrated rates including aggregate effects of vegetation and elevation.

Figure 5.6 shows the locations of the selected CIMIS stations along with the watershed outlet locations of the CNRFC forecast basins. The figure also presents a comparison of the annual cycle of evapotranspiration for selected locations on a regional basis. Figure 5.6b represents locations in the Santa Barbara region, or in the northwest portion of the study region, while Figure 5.6c is for locations in the southern, San Diego region. Between selected stations and the CNRFC estimates, there is general agreement particularly in the seasonal cycle. However, the amplitude in the season cycle from the CNRFC locations is larger, with higher PET rates particularly in the summer months (June – August). The magnitude of the estimates may differ by as much as 100% between the two sources during the summer. To minimize site-specific variations, the season cycle was normalized by the maximum monthly values, as shown in Figure 5.7a for the CIMIS stations. The normalized curves are quite similar in shape, with the CIMIS stations exhibiting less variability than the normalized curves based on CNRFC estimates (not shown).
The average monthly ET values for July (peak month) were also plotted against station elevation (see Figure 5.7b, limited to cismontane stations from the southern California coast). Although a direct relationship between monthly ET and station elevation is not apparent, there is a general tendency towards a regional variation. To illustrate this tendency, the stations are grouped by relative position. Specifically, CIMIS stations 97, 107, and 99 are near-coastal locations by Santa Barbara, Ventura and Los Angeles. Peak monthly ET values range from 4.2 to 4.5 mm/day. Stations along the San Diego and Orange Counties coast (102, 49 and 66) have slightly higher peak monthly values of 4.6 to 5 mm/day although these stations have similar elevation range as the northern locations. Next, CIMIS stations along the mountain foothills show higher values, with the stations along the Transverse Range foothills (64, 73, and 78) having peak monthly ET of ~5.4 mm/day while those stations along the Peninsular Range foothills (44, 62, 74) are again slightly higher with values around 5.6-6mm/day. CIMIS stations are limited to the foothills, with the possible exception of station 101, located within the Transverse Range near the Ventura-Los Angeles County line. The peak monthly value is highest at the location at nearly 6.5 mm/day. The CNRFC forecast basins extend to higher elevations than the CIMIS station locations. Peak monthly ET values reach 8.1 mm/day the watershed with outlets at CCHC1 (Santa Ynez mountains north of Santa Barbara), and at KNBC1 (San Bernardino Mountains), and 8.9 at HAWC1 (Peninsular Range in San Diego County). The peak monthly values at two locations in the CNRFC records were much higher than other values. This included a value of 11.7 mm/day in July for the watershed NRWC1 (Santa Ynez River at Narrows) and a value of 15.7 mm/day for the YDRC1 watershed (Santa Margarita River...
at Ysidora). These were uncharacteristically high relative to the remaining basins and were considered outliers for the remaining analysis.

The seasonal variation with location grouping was used to establish the input ET demand for the southern California flash flood basins, as depicted in Figure 5.8. This presents the July (peak month) ET demand, and shows increasing demand from coastal areas inland. The coastal Santa Barbara region has slightly lower values around 4.5 mm/day relative to the coastal Orange County values near 4.8 mm/day. The mountainous July ET demand values along the Transverse range are also generally lower, reaching 7 mm/day, until the San Bernardino Mountains and Peninsular Range with reach a value of 8.5 mm/day. Individual month average ET demand values for the remaining months and for each small watershed were derived from this mapping and an average seasonal cycle based on the normalized curves shown in Figure 5.7b.

5.4 Application of Hydrologic Models for Focus Watersheds

To illustrate the application of the hydrologic models to small watersheds, this section presents the application of the two hydrologic models to a few focus watersheds throughout the region. Although the desired assessment of the models is the ability to represent soil water conditions, long term records of in-situ soil water measurements are generally not available within the southern California. In recent years, the USGS has deployed short-term soil moisture monitoring sensors at locations recently burned by wildfires as part of the landslide susceptibility research (http://landslides.usgs.gov/monitoring). However, these sites typically have limited historical records which are not concurrent with the historical period over which the
simulations are made and have limited spatial extent which is not necessarily representative of the entire watershed. Other methods, specifically remotely sensed estimates of soil moisture, detect moisture over a shallow depth of soil and are thus not necessarily commensurate with the hydrologic model simulations.

Streamflow is an integrative measure of aggregate soil moisture and hydrologic response within a watershed. When soil moisture conditions are high, less precipitation may be infiltrated by the soil column and quick-response runoff to the stream channel is also high. The hydrologic models represent the various pathways for runoff production of stream channel flow, so comparison can be and is made with streamflow records as a surrogate indicator of how the model represents soil moisture conditions. For this assessment the two models are run for select watersheds using the gridded observation precipitation model and monthly ET demand and starting with a priori parameters determined for each watershed, and the resulting streamflow simulations compared with observed discharge records. The selection of focus watersheds considered several criteria:

(a) Availability of streamflow records with hourly or daily resolution;

(b) Small drainage areas (A < 500km$^2$);

(c) Record length of at least 10 years;

(d) Limited to no influence of regulation or diversions, as noted the gauge descriptions;

and

(e) Coverage of the watershed by at least two precipitation observation stations with historical records;
A limited number of gauging stations were identified. Five stations were selected which had hourly streamflow records provided by the USGS with a maximum record period of 1990-2005. The sites of these stations are shown in Figure 5.9 by the darker-shaded basins. Two of the sites are located within the western part of the Transverse Range, one is located in the eastern part near San Bernardino, and the remaining two sites are along coastal Orange County. Due to the lack of coverage in the Peninsular Range, five additional basins were selected which had daily streamflow observations (and are shown by the lighter shaded basins in Figure 5.9). These included three watersheds along the Peninsular Range in San Diego and Riverside Counties, one in central Los Angeles County and an additional basin near the hourly record stations in the western Transverse Range. The size of these focus watersheds ranged from 41 to 392 km$^2$.

As presented in Section 5.3, the hydrologic models produce estimates of total input to the stream channel or total basin runoff. For comparison with observed streamflow records, one must account for the travel delay from the production of runoff to reach the stream channel and the stream outlet of the watershed. This is accomplished with a simple, linear reservoir routing component which converts the total runoff generated by the model (i.e., total channel inflow) to streamflow at the basin output through a cascade of two linear reservoirs (Carpenter and Georgakakos, 2004a; Georgakakos and Bras, 1982):

$$\frac{dS_1}{dt} = CH_{in} - \alpha_r S_1(t)$$  \hspace{1cm} (40)
where $CH_{in}$ is the channel inflow as represent by Eq. 12 or 39 (with units of mm/hour), $\alpha_r$ is the reservoir routing parameter (units of 1/hour), and $S$ is the storage in the $j^{th}$ reservoir (in mm). The inverse of parameter $a$ represents the time delay associated with each reservoir.

The two hydrologic models were run with a priori parameters on an hourly basis. For comparison with daily flow records, the hourly simulation results were averaged to daily rates. Tables 5.3 and 5.4 present basic statistics of the a priori parameter simulations for the select watersheds at hourly and daily timescales, respectively, and covering the available observations. The tables indicate the length of the comparison period. It is also noted that the statistics presented are for periods where the observed flows exceeded a low flow threshold. With an interest toward flash flood occurrence, the focus of this evaluation was not specifically on the reproduction of base flow. This threshold, set as the median observed flow over period of observations for the selected stations, is intended to focus the assessment for the medium and higher flows which may lead to flash flooding.

In the tables, the bias is computed as:

$$B = \frac{q_x - q_o}{q_o}$$

(42)
where \( q_s \) represents the simulated mean flow, and \( q_o \) is the observed mean streamflow. Generally, a large positive bias in streamflow is found for all basins and for both models. The bias generally reaches a maximum value of 3.7, with the exception of the daily simulations for Alamo Pintado Creek (ID=11128250) where the bias exceeds a value of 9. This value appears high given that surrounding basins do not exhibit such a large bias. This bias may result from:

(a) precipitation overestimation,
(b) deficiencies in the a priori parameterizations,
(c) additional losses not captured by the model, such as higher evapotranspiration demand rates or recharge to deep groundwater aquifers.

Groundwater recharge was not included in the a priori simulations. Relatively few studies have been conducted specifically on groundwater recharge in the southern California region. Groundwater basins tend to be along the downstream reaches of major rivers, with the largest basin in this study region being the Santa Clara River groundwater basin. The USGS has published two studies of ground water in: on the Santa Clara groundwater basin and the Mojave River (Hanson et al, 2003; Stamos et al, 2001). These reports identify natural recharge to the groundwater basins coming from infiltration of streamflow, mountain front recharge and artificial recharge. Stamos et al (2001) state that 80% of the total recharge in the Mojave River groundwater basin is from streamflow leakage. Hanson et al (2003) breakdown simulated natural recharge into mountain front, valley-floor, and streamflow recharge, with the latter two combining to 78% of the simulated natural recharge. Stamos et al (2001) includes the
average annual discharge of the Mojave River downstream of the recharge area and an estimate of total mountain front recharge. Comparing these estimates, the proportion of mountain front recharge relative to observed discharge is on the order of 14%. These studies support the occurrence of groundwater recharge in the region. In larger watersheds, it is possible that the groundwater recharge occurs in the headwater regions and may appear downstream within the basin as part of the baseflow. For the size and location of watersheds of interest in this study, the recharge is likely of the mountain front recharge type and represents a loss to the total basin channel input.

Primarily to reduce the large bias and improve the model representations of observed streamflows, general adjustments to the a priori parameter runs were made starting with the inclusion of a groundwater recharge component to account for necessary additional loss. Additional adjustments were considered in a parsimonious fashion to improve simulations over the focus watershed. This was intended not as a full calibration of the hydrologic models, but rather to establish regional adjustments which could be applied to the a priori parameters for the regional small watershed runs. For the SAC-SMA model, adjustments were made to specific model parameters at each of the focus watersheds and included:

1. groundwater recharge by specifying a positive value for parameter $\mu$;
2. Increase in the lower zone free and tension water capacities ($X_3^0, X_4^0, X_5^0$);
3. Increase in the $p_{free}$ parameter to move moisture into lower zone free elements, making it available for recharge;
4. Centering of the percolation exponent near a value of 1.7;
(5) Possible adjustments to upper zone storage capacities.

For the FFSM model, a mechanism for groundwater recharge was added to the model. This followed the structure of groundwater recharge as represented in the SAC-SMA model. A parameter, $\mu$, was introduced to represent the amount of subsurface flow which would be lost to groundwater recharge:

$$gwr = \left( \frac{\mu}{1+\mu} \right) (b_u + b_l).$$

The channel inflow (represented by Eq. 39) was reduced by this amount. Additionally, the percolation constant defined based on Table 5.2 yield fairly low values (values of 25 or less for much of the higher elevations of the Peninsular Range and the Eastern Transverse Range). An increase in the percolation constant values for the focus watersheds were also considered to increase the flux of water to the lower zone. The remaining parameters were defined by soil texture determined for each watershed. This relative character was maintained, but an increase in the depth of soils was considered. The depth of the lower soil zone was limited in the a priori run by the average depth to bedrock, and increasing the depth of the soil layers effectively increases the capacities of the storage elements. Finally, for six of the ten focus watersheds, it was noted that when the soil texture classification of the lower zone was of a clay type, the simulated response of the watershed depleted the streamflow during dry periods. This resulted possibly due to the high pore disconnectedness values. When more moderate values of
saturated hydraulic conductivity and pore disconnectedness were used, the flow response during low flow periods better represented the observed flows. This substitution of the soil texture classification for clay soils was applied, using the soil properties of the sandy-loam texture which was more prevalent across the region. A total of four possible adjustments for the FFSM model a priori parameters are: (1) inclusion of groundwater recharge; (2) increase in the percolation constant; (3) increase in the depth of soils; and (4) a change in dominant soil texture classification for the lower layer if clay soils were initially identified as dominant based on areal extent.

Table 5.5 presents the flow statistics for the focus watersheds after the above adjustments were implemented. These adjustments produced a reduction in the large bias of the a priori parameter simulations, while maintaining or improving correlation between simulated and observed flows. For the hourly-record locations, the SAC-SMA model simulations given bias values in the range -0.08 to 0.05 and correlation of 0.55 to 0.92. The daily flow comparisons for these locations show similar values in the statistics, with the exception of Alamo Pintado Creek which had a larger bias (although reduced) and a significant improvement in the correlation. The statistics for the FFSM model are similar for a given basin as the SAC-SMA model. The locations on Campo Creek and Sweetwater River in the southern Peninsular Range have the lowest correlation with observed daily flows. This may suggest more uncertainty in the model parameterizations and adjustments along this area.

As an example, Figure 5.10 presents the cumulative distribution of observed and simulated flows for two of the focus watersheds: Alamo Pintado Creek (hourly flows), and Sweetwater River (daily flows). These locations show low bias and moderate
correlation statistics. Again, flows greater than the minimum threshold defined for each station are compared. The flows are transformed using a Box-Cox transformation (Box and Cox, 1964):

\[ q_t = \frac{q^{\lambda - 1}}{\lambda} \] (44)

where \( q \) is the observed or simulated flow (mm/hour or mm/day), \( q_t \) is the transformed flow, and \( \lambda \) is a parameter which a value set to 0.3 for this work. This transformation is often used in hydrology to represent data which is highly skewed. The figure presents results for the SAC-SMA model, but the comparison is similar with the FFSM model. The examples indicate reasonable reproduction of observed flows over a range for flow values. Overall, specific statistics at the focus watersheds may considered less than acceptable for a typical hydrologic model calibration, the statistics are considered reasonable for objective of parsimonious adjustment of a priori parameters to allow regionalization.

5.5 Regional Soil Water Modeling for Small Watersheds

This section presents the regional hydrologic modeling results under the small multi-model ensemble approach. The multi-model ensemble consists of 4 members, for the two hydrologic models (SAC-SMA and FFSM) each driven with precipitation input from the gridded observations model (GRIDOBS) and the simplified orographic model (SIMOROP). Note, the model abbreviation for SAC-SMA is shortened to SAC in the
results section and associated figures. Each model was run with hourly resolution and for the period 10/1948 to 4/2005. Because the orographic precipitation model was run only for wet seasons (October through April), the gridded observation model was used to fill the spring and summer months (May through September) for the orographic model. Although little precipitation falls during these months, this allowed the model to begin each wet season with an appropriate distribution of soil moisture conditions, rather than climatological values or other approximation. Results presented highlight the wet season and are confined to periods after 10/1949, allowing the soil moisture conditions to develop over the first year of simulation.

It is noted that the hydrologic simulations for the small watersheds are limited to local basin runoff production. The modeling did not include the linear reservoir routing component described in Section 5.4 for within basin flows, nor did it include a component for the stream network and streamflow routing. The flash flood occurrence simulation is based on a precipitation threshold index and thus the production of streamflow hydrographs throughout the region is not the objective of this modeling effort. Rather, it is to produce estimates of within basin soil moisture conditions which are used in the flash flood occurrence simulation.

Figure 5.11 presents the climatological values of basin average precipitation (MAP), runoff, upper and lower layer soil saturation, and standard deviation of upper and lower soil saturation for the SAC model run forced by the GRIDOBS precipitation model and for the month of February. The month of February is selected as this is where the peak in the annual cycle of precipitation occurs. MAP and basin runoff average values were computed based on monthly totals. The hourly estimates of soil
saturation, computed as the soil moisture content divided by soil capacity in the given layers, were averaged over each daily, and then statistics of the daily values computed. The MAP shows a fairly large area of high MAP across the Transverse Mountains, reaching 200mm/mo, with lower values along the Peninsular Range, reaching about 130 mm/mo. This variation in precipitation across the Transverse and Peninsular ranges was noted in the precipitation simulation results discussed in Chapter 4. In comparison, the runoff produced by the SAC model is more spatially limited with only a few locations with high average runoff across the Transverse range, and low average runoff of less than 20mm/mo throughout most of the region, including all of the Peninsular Range watersheds. This reduction in runoff relative to precipitation reflects the losses to deep recharge.

The spatial distribution of the average soil moisture saturation, computed as the ratio of soil moisture content to the capacity, is fairly uniform with values near 0.35. The lower values of average soil saturation occur along the higher elevation values in the Peninsular Range and in the San Bernardino Mountains. The average lower soil saturation fraction shows a wider range of values, from near 0.05 to 0.60. The variability in soil moisture is also presented as the standard deviation of daily soil moisture saturation fraction. As in the average values, the upper zone soil saturation is more spatially uniform than the lower zone.

Figure 5.12 presents analogous results for the SAC model forced by the SIMOROP precipitation model. As noted in Chapter 4, the SIMOROP model forcing produces a narrower extent of watersheds with significant precipitation. Many watersheds show average February MAP values less than ~20 mm/month with the
SIMOROP model where the GRIDOBS model shows mean values from ~70mm/month to 100mm/month in these regions. This spatial pattern in precipitation is then reflected in the average monthly averages for runoff generation and soil moisture. Average runoff exceeding ~40mm/month for the month of February is limited to a relatively few basins across the Transverse Range with a few in the Peninsular Range. In the regions with higher MAP, the average and standard deviation of upper layer soil saturation is similar to the values shown for the GRIDOBS forcing, and with slightly higher values along the Peninsular Range. For the lower layer, the average soil saturation layer shows a few watersheds reaching values of 0.7-0.8, higher than the average determined with the GRIDOBS forcing.

Figure 5.13 and 5.14 present the analogous results for the FFSM model forced by the GRIDOBS and SIMOROP models, respectively. The spatial patterns of the hydrologic model output are consistent with respective results from the SAC model presented in Figures 5.11 and 5.12 for the two precipitation forcing models. There is general similarity in the hydrologic response of the two models. The model structures are similar in the representation of the two-layer soil column, but parameterization and parameter adjustments varied. The FFSM model produces more runoff, particularly for watersheds in the western Transverse Range. Also with the SIMOROP forcing, watersheds in the southern Peninsular Range indicate higher runoff. The average upper soil saturation tends to be lower in for FFSM model than the SAC model with values typically less than 0.3, except for those noted watersheds along the western Transverse and southern Peninsular Ranges. In contrast, the lower layer saturation fraction in the
FFSM model is higher across the region, with values typically from 0.4 to 0.7 for the GRIDOBS forcing.

While the average soil saturation gives an indication of the mean hydrologic behavior, conditions under which flash floods may occur are not typically during average conditions. Higher quick-responding surface runoff occurs as the soil is saturated and precipitation cannot infiltrate to deeper soils, thus increasing the potential for flash flood occurrence. It is anticipated that soil moisture would be higher than average for conditions when the threat of flash flood occurrence is high. Therefore, in addition to the average soil saturation statistics, quantiles of daily soil saturation were also computed at the 75th- and 90th-percentiles. At the 75th percentile, upper layer soil saturation reaches 0.65 along the Transverse Range watersheds for the GRIDOBS-FFSM model and 0.8 for the SIMOROP-FFSM model. At the 90th percentile, these locations with high soil saturation exceed 0.9 for the FFSM model. Given the higher average soil saturation of the lower layer, these values reach near saturation condition for the 75th percentile in many of the small watersheds. The soil saturation of the upper layer for the GRIDOBS-SAC model remains more spatially uniform than the FFSM model, with values near 0.45 for the 75th percentile and 0.55-0.6 for the 90th percentile.

5.6 Flash Flood Occurrence Climatology: Historical and Potential Future Changes

This final section of the chapter represents the synthesis of the multi-disciplinary modeling described in this thesis towards the estimation of flash flood occurrence frequency for the southern California region. This modeling has included components of geomorphology (Ch. 3), precipitation modeling (Ch. 4), and hydrology and soil
moisture modeling (current chapter). The premise for this modeling approach stems from the lack of records of flash flood occurrence with regular or uniform spatial and temporal coverage at relatively high spatial resolution. It is shown in this section how the modeling approach opens other research possibilities regarding flash flood hazards in southern California.

5.6.1 Defining flash flood occurrence potential

The simulation of flash flood occurrence is built on the operational flash flood warning concept of flash flood guidance (FFG). FFG is defined for a particular watershed as the amount of precipitation of a given duration and applied uniformly over the watershed necessary to generate runoff that will cause the stream outlet of the watershed to reach minor flooding conditions, typically taken as the bankfull condition. This bankfull condition is chosen as it is a physically-meaningful and geomorphologically significant level, although it is a conservative measure of flooding as it is not generally associated with significant flood damage. Given the short forecast lead times associated with flash flood occurrence, (generally < 6-12 hours), the approach is valuable as it provides a relatively fast determination of a precipitation threshold based on conditions within the watershed being favorable of producing runoff to yield flooding conditions, and does not require the generation of full basin response and streamflow hydrograph of the peak response. Norbiato et al (2008, 2009) suggests the usefulness of the FFG concept, particularly for ungauged watersheds and basins with limited observations. FFG is used by the U.S. National Weather Service and other locations to produce flash flood warnings (e.g., Georgakakos, 2006; Mogil et al, 1978)
The surface runoff index (SRI) was developed in Chapter 3 for the southern California region. The SRI is the amount of runoff resulting from effective rainfall of a given duration and uniformly applied over a watershed which is necessary to cause bankfull flow at the watershed outlet, as an indicator of the initiation of flooding. Under stable geomorphologic, land surface, and climatic conditions, SRI is a time-invariant characteristic of the watershed. Flash flood guidance is the time-varying quantity which is computed based on SRI by accounting for precipitation losses to the land surface and evapotranspiration. Figure 5.15 illustrates the relationship of FFG with SRI and soil moisture at a given computational time and given duration of rainfall (here, duration time, \( t_d = 3 \) hours). At this time, the computation of FFG considers the current soil moisture state and moisture fluxes, determines the soil moisture deficit, and applies increasingly larger precipitation within the hydrologic model to determine if all soil moisture and evapotranspiration demand is met. The precipitation at which the deficit is satisfied is the FFG. This may be represented by the black arrows in Figure 5.15 for soil moisture deficit 1 (at 90% saturation). At another computation time, the soil moisture deficit may be higher, represented by the line labeled soil moisture deficit 2 (at 75% saturation), the computed FFG has a higher value.

Based on the SAC-SMA model, Georgakakos (2006) presented an analytical solution for FFG. Through substitution, one may arrive at an expression for FFG as (using notation of Table 5.1):

\[
FFG_{td} = \left[ etd + \frac{SRI}{t_r} + (1 - \beta_2 - \beta_1)(p_o - d_uX_2^0) \right] t_d 
\] (43)
where \( t_d \) denotes the time of rainfall for FFG and \( t_r \) denotes the time associated with the SRI. \( p_o \) represents the percolation to the lower soil layer and other terms are defined in Table 5.1. This formulation describes the capacity of the soil and land surface to generate precipitation losses through evapotranspiration, filling of upper soil capacity, percolation to deeper soil depths, and interflow. The focus is on the upper soil layer only as this is the fast responding soil layer that is most relevant at the time scales of flash flooding, and the loss via percolation to deeper layers is incorporated \((p_o)\). Given the similarity in structure of the two-layer FFSM models as applied for southern California, Equation 43 was adapted for application with the FFSM model.

From the hourly simulations of soil moisture at the small watersheds of southern California, all information needed for the estimation of FFG is available, and is used to define the occurrence of flash flooding. In the context of this study, a flash flood occurrence is signaled when precipitation of a given duration exceeds the FFG value (of the same duration) estimated at the start of the precipitation:

\[
FFO_{ti} = \begin{cases} 
1 & \text{if } PRC_{ti+dur} > FFG_{dur,ti} \\
0 & \text{if } PRC_{ti+dur} \leq FFG_{dur,ti} 
\end{cases}
\]

\( (45) \)

where \( ti \) represents the given time step and \( dur \) represents the duration of rainfall. To define flash flood events, consecutive hours with FFO equal to 1 were considered as a single flash flood event. A minimum inter-arrival time of 12 hours was also imposed for flash flood events. If a period between two FFO events was less than this minimum inter-arrival time, the second event was considered a continuation of the first event.
The results presented consider flash flood occurrence thus defined using a 3-hour rainfall duration. It is expected that within 3 hours the majority of the small basins identified in this study have responded to precipitation input with a flow increase at the basin outlet.

5.6.2 Historical flash flood occurrence frequency

The historical flash flood occurrence frequency is computed as a simple average of the number of flash flood events per year. Figure 5.16 presents the spatial distribution of historical flash flood occurrence frequency (N/year) based on the FFSM model using the GRIDOBS precipitation model forcing. The period of analysis consists of wet seasons only from October 1949 to April 2005. It is noted that the small watersheds without shading had a computed flash flood frequency of 0.02 or less (≤ 1 event over period of record). The model shows highest frequency in the eastern and central parts of the Transverse Range and slightly elevated frequency in the Peninsular Range. This follows the spatial pattern observed in the average runoff (i.e., Figure 5.13) and higher quantile soil saturation results. The range in occurrence frequency is 0.036 to 4.8 events per wet season.

The reciprocal of the frequency is indicative of the average return period, or number of years between events. Using the indicated range in frequency yields return periods of flash flooding in the range of 0.2 to 28 years. The very long return periods are defined by the period of record, but are unexpectedly long. Literature on the return period for bankfull flows, up on which the SRI is defined, generally cites return periods of 1.5 - 3 years (e.g., Leopold, 1994; Castro and Jackson, 2001), although cases of long
recurrence intervals have been noted (e.g., Williams, 1978). The analysis of the frequency of exceeding estimated bankfull discharge based on observed streamflow for select streams in southern California presented in Chapter 3 found the frequency to be in the range 0.15 to 3.3 years. The observational analysis suggested a higher frequency (or lower return interval) for streams along the Transverse Range. This is consistent with the spatial distribution shown in Figure 5.16.

The multi-model ensemble approach provides an indication of the uncertainty in estimating the flash flood occurrence frequency. As an example, Figure 5.16a presents the occurrence frequency for the ensemble member from the SIMOROP-FFSM model. There is a notable difference in the spatial extent of watersheds with estimated frequency greater than the 0.02 threshold: the SIMOROP-FFSM model produces a reduced extent in the number of watersheds with estimated frequency reaching this threshold. The reduction in extent is consistent with and resulting from the difference in the simulation of precipitation between the two forcing models, (e.g., Figures 5.11 and 5.12; Chapter 4). The GRIDOBS-FFSM model produces a relatively smooth transition in frequency values for closely located basins. The SIMOROP-FFSM based frequencies show an increase in values on a more localized basis, with the same general locations of higher frequency of occurrence as noted previously. Comparison with the frequencies estimated using the SAC model also indicates a reduction in the spatial extent of watershed meeting the minimum frequency threshold relative to the FFSM model. (This is likely resulting from the increase in total soil moisture storage capacity parameterized for the SAC model). With a lesser extent and magnitude, the watersheds with higher frequency of occurrence are again noted for the same general regions.
Figure 5.17 presents a comparison of estimated frequency from the four models for individual watersheds. The x-axis is the frequency as computed by the SIMOROP-SAC model ensemble for the watersheds with frequency greater than 0.02 events per year (1 event in 50 years). The y-axis presents the corresponding watershed estimated frequency (events per year) for the other ensemble members. Since the y-axis is scaled differently, the black line represents equal frequency (45° line). The figure is intended to convey the uncertainty in estimating the magnitude of flash flood occurrence frequency under this modeling approach with the selected models and parameterizations. Clearly model selection affects the magnitude of the estimated frequency, and for this study case, with great variation for low frequency events.

Improvement through calibration (if calibration were possible for all small watersheds) or in the regional adjustments may result in a change in the frequency magnitude, or through comparison with frequencies computed from observations at limited locations (e.g., as computed in Chapter 3 for bankfull flow), and this modeling approach could provide additional insight in the spatial variability at small scales.

Section 5.4 discussed the spatial variation of soil saturation and postulated the role of soil saturation on conditions with high likelihood for flash flood occurrence. Figure 5.18 quantifies the level of soil saturation during simulated flash flood occurrence. The figure compiles the average soil saturation during flash flood events (as defined above) for each of the small watershed and presents the cumulated distribution of soil saturation in both the upper soil layer and lower soil layer (the FFSM model is shown for this example). All models show that the upper soil layer is near saturation, with a saturation fraction > 0.95 for all events and for all watersheds. The
lower layer soil saturation has a greater range. The GRIDOBS-FFSM shows the highest frequency of lower soil moisture; however, only about 10% of cases have lower soil saturation less than about 0.7. The SAC model produces higher lower layer soil saturation (10% of the cases have lower layer soil saturation less than nearly 0.8). As indicated in Figure 5.18b, the SIMOROP forcing produces the highest levels of soil saturation.

5.6.3 Potential changes in flash flood occurrence frequency

The developed modeling approach facilitates expanding the study of flash flood occurrence changes to include potential land-use and climate changes. In this section, initial results of applying the modeling approach developed to the climate change problem mentioned are presented, using CCSM climate scenarios (Collins et al., 2006). The CCSM model is selected as a state-of-the-art climate model that affords high temporal resolution (6-hourly) and 3-dimensional detail in atmospheric variables necessary to develop a comparable simulation of flash flood occurrence as just described for the historical analysis. In addition, a recent report (HRC-GWRI, 2010) has shown that the model reproduces well the historical climate in terms of 500hPa geopotential height and surface precipitation in the region of northern California as compared to the NCEP-NCAR reanalysis. CCSM climate model output including wind speed and direction, temperature and humidity were used to force the orographic precipitation model to generate high resolution precipitation estimates for each of two thirty year periods. The CCSM-forced orographic precipitation was then used to drive the hydrology and flash flood occurrence simulation models as described in this chapter. The first period is based on the CCSM control climate period of 1970-1999.
The second period is based on the CCSM A1B climate change scenario and focused at the end of the 21st century for the years 2070-2099. The flash flood frequency was computed for each period and the end of 21st century frequency was compared to the frequency obtained by the control scenario.

The spatial distribution of flash flood occurrence frequency from the CCSM-driven runs show similar patterns to that presented for the historical runs in Fig. 16. As seen before, there is a limited extent of watersheds with flash flood occurrence frequency > 0.03 (1 event in 30 years) driven by the extent of orographically-force precipitation in the model. The control run period (1970-1999) shows a range in occurrence frequency from 0.06 to 6.7 events per year for the FFSM model, which is similar to that computed for the historical analysis (Figure 5.16b). Control period simulation for the SAC model reaches a peak of 2.3 events per years, and as noted in Section 5.6.2, and there are fewer watersheds with flash flood occurrence frequency exceeding the given minimum threshold.

Figure 5.19 illustrates the precipitation forcing differences between the control and future periods as simulated with the SIMOROP model. This is a compilation of all hours with precipitation greater than 0.025mm/hour and for watersheds which had flash flood occurrence frequency greater than 0.03 (1 in 30 years) during both simulation periods. The Figure shows a general shift to higher precipitation rates over the watersheds over a range of the frequency distribution. It also indicates the total number of hours of precipitation included in the distribution plot. Overall, there is a 3% reduction in the total number of hours with simulated precipitation for all watersheds during the 21st Century.
Finally, Figure 5.20 compares the 1970-1999 control run flash flood occurrence frequencies with the 2070-2099 A1B climate change scenario frequencies for the watersheds meeting the minimum frequency threshold. Each point represents a single watershed. Both models indicate an increase in the flash flood occurrence frequency at the end of the 21st century relative to the 20th century control run. The slope of the linear regression line is shown for the two models to indicate the average increase for the future scenario. This increase is 49% under the SAC model, while for the FFSM model the slope is lower with an increase of ~30%.

This increase in the occurrence frequency is interesting given the relatively small changes in overall precipitation distribution shown in Figure 5.19. Further investigation into the changes of precipitation and hydrologic conditions (e.g., soil moisture) is reserved for future work, although Figure 5.19 is suggestive that, at least in terms of forcing, changes may be found in the intensity of precipitation rather than the number of periods with precipitation.

These results suggest a potential increase in flash flood occurrence in southern California at the end of the 21st century under the A1B future climate scenario as inferred from a state-of-the-art climate model. Various caveats and potential modifications to these specific results exist, including possible incorporation of land use, evapotranspiration, or geomorphologic changes to the model for the 21st century. However, this initial application does demonstrate the capability of the modeling approach for application to the study of such scientific issues.
5.7 Conclusions

This chapter represents the culmination in a multi-disciplinary modeling effort aimed at examining historic flash flood occurrence potential over a large region with high spatial resolution. The effort synthesizes elements of geomorphology, orographic precipitation modeling, and hydrologic modeling issues including parameterization and spatial scaling, to estimate flash flood occurrence frequency at scales of ~30km$^2$ over the southern California mountainous region. The flash flood occurrence potential is defined through accounting of soil moisture condition throughout the region and a comparison of precipitation estimates with the amount required do produce flooding (bankfull flow) at the outlet of small streams over the wet seasons for the historic time period from 10/1948 – 4/2005. This chapter has focused on the hydrologic modeling and flash flood occurrence simulation through a multi-model ensemble approach utilizing two models for precipitation forcing over small watersheds and two hydrologic models for soil moisture accounting and runoff generation.

The main results of this chapter show:

- The two hydrologic model reasonably reproduced observed streamflow at both hourly and daily timescales utilizing a priori parameters based on soil property information and with regional adjustments for select watersheds ranging in drainage area from ~40 to 400 km$^2$.

- The two hydrologic models, although both representing the soil column with two conceptual layers but with varying complexity in terms of parameterizations, produce variation in hydrologic response across the region on the small watershed scale of 30km$^2$;
- Flash flood occurrence potential defined and occurrence frequency estimated in terms of average number of (potential) flash flood events occurring per year for over 950 small subbasins within the region and for four multi-model ensemble members. Variability in frequency estimates for a given basin is relatively small for low frequency magnitudes ( < 0.5 events per year), and is large as the occurrence frequency increases;

- Flash flood occurrence found only when models show high degree of soil saturation in both upper and lower soil layers;

- Spatial pattern and relative magnitude in occurrence frequency is consistent among models, indicating spatial patterns may hold or can be interpreted if different estimates of occurrence frequency magnitude are developed.

- Application of the modeling approach under climatic change scenario utilizing both hydrologic models demonstrates utility of methodology, and indicated increase in flash flood occurrence potential at the end of the 21st century under the A1B climate change scenario.

Development and demonstration of this interdisciplinary modeling approach has been the first step toward increased understanding of flash flood occurrence frequency at small spatial scale in the region and opens opportunities for further detailed research. Recommended next research steps include:

- examination of historical periods with wide-spread flash flood occurrence and identify characteristics including large scale atmospheric forcing that may differ for regional versus sub-regional flash flood occurrence;
- examine similarity and differences in hydrologic states and atmospheric forcing under periods with high flash flood occurrence potential versus high precipitation occurrence (e.g., Chapter 2);
- explore role of uncertainty in model parameterizations (hydrologic, precipitation, geomorphologic) on flash flood occurrence frequency estimation (as in Carpenter and Georgakakos, 2004b, 2006).
- consider other high spatial and temporal resolution precipitation databases, such as radar rainfall or commensurate resolution mesoscale numerical prediction models (e.g., WRF) over shorter time periods to further explore uncertainty in precipitation forcing impacts on hydrologic response and flash flood occurrence frequency estimation.

5.8 References


pp. + appendices. Current dataset available via:


Table 5.1. SAC-SMA model states and parameters.

<table>
<thead>
<tr>
<th>Model States</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>X₁</td>
<td>upper zone tension water content (mm)</td>
</tr>
<tr>
<td>X₂</td>
<td>upper zone free water content (mm)</td>
</tr>
<tr>
<td>X₃</td>
<td>lower zone tension water content (mm)</td>
</tr>
<tr>
<td>X₄</td>
<td>lower zone free primary water content (mm)</td>
</tr>
<tr>
<td>X₅</td>
<td>lower zone free supplemental water content (mm)</td>
</tr>
<tr>
<td>X₆</td>
<td>additional impervious water content (mm)</td>
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<table>
<thead>
<tr>
<th>Model Parameters</th>
<th>Description</th>
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<tr>
<td>X₁₀</td>
<td>upper zone tension water capacity (mm)</td>
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<tr>
<td>X₂₀</td>
<td>upper zone free water capacity (mm)</td>
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<td>X₃₀</td>
<td>lower zone tension water capacity (mm)</td>
</tr>
<tr>
<td>X₄₀</td>
<td>lower zone free primary water capacity (mm)</td>
</tr>
<tr>
<td>X₅₀</td>
<td>lower zone free supplemental water capacity (mm)</td>
</tr>
<tr>
<td>dᵤ</td>
<td>upper zone withdrawal rate (1/Δt)</td>
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<td>dₜₚ</td>
<td>lower zone primary withdrawal rate (1/Δt)</td>
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<td>ε</td>
<td>percolation function constant</td>
</tr>
<tr>
<td>θ</td>
<td>percolation function exponent</td>
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<tr>
<td>pₚ</td>
<td>fraction of percolation which goes to lower zone free storage</td>
</tr>
<tr>
<td>μ</td>
<td>deep recharge parameter</td>
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<tr>
<td>β₁</td>
<td>fraction of area which becomes impervious as tension storage filled</td>
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<tr>
<td>β₂</td>
<td>fraction of permanently impervious area</td>
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Table 5.2. Physical soil properties for various soil texture classifications.

<table>
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<tr>
<th>Texture*</th>
<th>$\theta_s$ (m$^3$/m$^3$)</th>
<th>$\theta_f$ (m$^3$/m$^3$)</th>
<th>$\theta_w$ (m$^3$/m$^3$)</th>
<th>$K_s$ (mm/h)</th>
<th>c1</th>
<th>C</th>
<th>mp</th>
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<tbody>
<tr>
<td>S</td>
<td>0.37</td>
<td>0.15</td>
<td>0.04</td>
<td>634.6</td>
<td>3.4</td>
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<td>LS</td>
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<td>0.19</td>
<td>0.05</td>
<td>562.6</td>
<td>3.5</td>
<td>12.0</td>
<td>1.5</td>
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<tr>
<td>SL</td>
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<td>0.27</td>
<td>0.09</td>
<td>124.8</td>
<td>3.6</td>
<td>19.0</td>
<td>1.6</td>
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<tr>
<td>SIL</td>
<td>0.47</td>
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<td>25.0</td>
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<tr>
<td>SI</td>
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<td>0.11</td>
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<td>35.0</td>
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<td>L</td>
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<tr>
<td>SC</td>
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<td>6.6</td>
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<tr>
<td>SIC</td>
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<td>4.6</td>
<td>7.5</td>
<td>205.0</td>
<td>3.5</td>
</tr>
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</table>

* Soil texture classifications are: sand (S), loamy-sand (LS), sandy-loam (SL), silty-loam (SIL), silt (SI), loam (L), sandy-clay-loam (SCL), silty-clay-loam (SICL), clay-loam (CL), sandy-clay (SC), silty-clay (SIC), and clay (C).

1 Values for $\theta_s$, $\theta_f$, $\theta_w$, and $K_s$ are adopted from Anderson et al (2006).

2 Values of $c1$ are adopted from Bras, (1990, Table 8.1) and interpolated from 4 texture classes for sand, sandy-loam, silty-loam, and clay.

3 Values of percolation parameters adopted from Anderson (2002) and interpolated from 3 texture classes for sand, silt, and clay.
Table 5.3. Validation statistics for a priori hourly simulations.

<table>
<thead>
<tr>
<th>Station</th>
<th>11128250 A. Pintado</th>
<th>11111500 Sespe</th>
<th>11063510 Cajon</th>
<th>11047300 San Juan</th>
<th>11046530 A. Trabuco</th>
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</thead>
<tbody>
<tr>
<td>County</td>
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<td>Ventura</td>
<td>San Bernardino</td>
<td>Orange</td>
<td>Orange</td>
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<tr>
<td>Area (km²)</td>
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**Sacramento Model – A priori**

<table>
<thead>
<tr>
<th>Bias</th>
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<tbody>
<tr>
<td>Corr.</td>
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<td>0.81</td>
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**FFSM Model – A priori**

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Table 5.4. Validation statistics for a priori daily simulations.

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<th>11111500 Sespe</th>
<th>11063510 Cajon</th>
<th>11047300 San Juan</th>
<th>11046530 A. Trabuco</th>
</tr>
</thead>
</table>

**Sacramento Model – A priori**

<table>
<thead>
<tr>
<th>Bias</th>
<th>9.32</th>
<th>2.40</th>
<th>1.77</th>
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<tbody>
<tr>
<td>Corr.</td>
<td>0.63</td>
<td>0.81</td>
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**FFSM Model – A priori**

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<tbody>
<tr>
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<td>0.63</td>
<td>0.80</td>
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</table>

<table>
<thead>
<tr>
<th>Station</th>
<th>11124500 Santa Cruz</th>
<th>11098000 A/ Seco</th>
<th>11042400 Temecula</th>
<th>11015000 Sweetwater</th>
<th>11012500 Campo</th>
</tr>
</thead>
<tbody>
<tr>
<td>County</td>
<td>Santa Barbara</td>
<td>Los Angeles</td>
<td>Riverside</td>
<td>San Diego</td>
<td>San Diego</td>
</tr>
<tr>
<td>Area (km²)</td>
<td>191.6</td>
<td>41.4</td>
<td>339.1</td>
<td>117.5</td>
<td>220.0</td>
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**Sacramento Model – A priori**

<table>
<thead>
<tr>
<th>Bias</th>
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<th>0.77</th>
<th>2.54</th>
<th>2.07</th>
<th>2.03</th>
</tr>
</thead>
<tbody>
<tr>
<td>Corr.</td>
<td>0.68</td>
<td>0.70</td>
<td>0.76</td>
<td>0.60</td>
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**FFSM Model – A priori**

<table>
<thead>
<tr>
<th>Bias</th>
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<th>2.20</th>
<th>2.70</th>
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<tbody>
<tr>
<td>Corr.</td>
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<td>0.62</td>
<td>0.87</td>
<td>0.60</td>
<td>0.43</td>
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</table>
Table 5.5. Validation statistics for adjusted parameters model runs for all focus watersheds and for both hourly and daily simulations.

<table>
<thead>
<tr>
<th>Station</th>
<th>11128250 A. Pintado</th>
<th>11111500 Sespe</th>
<th>11063510 Cajon</th>
<th>11047300 San Juan</th>
<th>11046530 A. Trabuco</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>HOURLY RUN: SAC-SMA Model – Adjusted</strong></td>
<td></td>
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<td></td>
</tr>
<tr>
<td>Bias</td>
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<td>0.64</td>
<td>0.86</td>
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<tr>
<td><strong>HOURLY RUN: FFSM Model – Adjusted</strong></td>
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</tr>
<tr>
<td>Bias</td>
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<td>0.31</td>
<td>0.05</td>
<td>-0.01</td>
<td>-0.01</td>
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<tr>
<td>Correlation</td>
<td>0.45</td>
<td>0.86</td>
<td>0.58</td>
<td>0.68</td>
<td>0.81</td>
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<table>
<thead>
<tr>
<th>Station</th>
<th>11124500 Santa Cruz</th>
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<th>11042400 Temecula</th>
<th>11015000 Sweetwater</th>
<th>11012500 Campo</th>
</tr>
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<tbody>
<tr>
<td><strong>DAILY RUN: SAC-SMA Model – Adjusted</strong></td>
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<td>0.88</td>
<td>0.73</td>
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<td>0.84</td>
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<td><strong>DAILY RUN: FFSM Model – Adjusted</strong></td>
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</tr>
<tr>
<td>Bias</td>
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<tr>
<td>Correlation</td>
<td>0.75</td>
<td>0.87</td>
<td>0.70</td>
<td>0.79</td>
<td>0.81</td>
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</table>

<table>
<thead>
<tr>
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<td>0.72</td>
<td>0.71</td>
<td>0.58</td>
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</table>
Figure 5.1. Southern California study region. Major cities and county boundaries indicated.
Figure 5.2. Characteristics of small watersheds defined region: (a) location, (b) channel slope, (c) soil surface texture, and (d) average depth to bedrock. Latter two derived from STATSGO soils database.
Figure 5.3. Geologic (a) and land cover (b) characteristics of southern California.
Figure 5.4. Depiction of Sacramento Soil Moisture Accounting model (SAC-SMA).
Figure 5.5. Depiction of Flash Flood-Soil Moisture model (FFSM).
Figure 5.6. Location of CIMIS stations and CNRFC watershed outlets with reference evapotranspiration data (a), along with example seasonal cycles for selected stations in (b) Santa Barbara County and (c) San Diego County.
Figure 5.7. Variation of CIMIS station reference ET: (a) normalized seasonal cycle and (b) variation with peak month (July) average precipitation with station location.
Figure 5.8. Estimated peak month (July) average ET Demand for small watersheds.
Figure 5.9. Location of selected focus watersheds for hydrologic model application. Darker shaded basins have hourly discharge records; lighter basins have daily records.
Figure 5.10. Example cumulative distributions of Box-Cox transformed flows for the Alamo Pintado Creek and Sweetwater River focus watersheds.
Figure 5.11. Hydrologic model output for small watersheds based on SAC-SMA model with GRIDOBS model forcing. Monthly average values for month of February for (a) MAP, (b) runoff, (c) average soil saturation fraction in upper layer, (d) standard deviation of upper layer soil saturation, (e) average soil saturation fraction in lower layer, (d) standard deviation of lower layer soil saturation.
Figure 5.12. As in Figure 5.11 but for the SIMOROP-SAC model
Figure 5.13. As in Figure 5.11, but for the GRIDOBS-FFSM model.
Figure 5.14. As in Figure 5.11 for the SIMOROP-FFSM model.
Figure 5.15. Illustration of the relationship between flash flood guidance (FFG), surface response index (SRI), and soil moisture deficit.
Figure 5.16. Historical flash flood threat occurrence using the GRIDOBS-FFSM model.
Figure 5.17. Variation in multi-model ensemble member estimates of historic flash flood occurrence frequency for selected watersheds.
Figure 5.18. Distribution of average soil saturation (upper and lower soil layers) during flash flood occurrence for four multi-model ensemble members.
Figure 5.19. Cumulative distribution of precipitation (MAP > 0.025 mm/hr) for select watersheds under historical control and future climate simulation periods.
Figure 5.20. Comparison of flash flood occurrence frequencies for select watersheds between end of 20th and end of 21st centuries under A1B climate change scenario.