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Relationships between Runoff, Land Cover and Climate in the Semi-Arid Intermountain Region of the Western U.S.A.

Ibrahim Nourein Mohammed
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RELATIONSHIPS BETWEEN RUNOFF, LAND COVER AND CLIMATE IN THE SEMI-ARID
INTERMOUNTAIN REGION OF THE WESTERN U.S.A

by

Ibrahim Nourein Mohammed

A dissertation submitted in partial fulfillment
of the requirements for the degree

of

DOCTOR OF PHILOSOPHY

in

Civil and Environmental Engineering

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2012
ABSTRACT

Relationships between Runoff, Land Cover and Climate in the Semi-Arid Intermountain Region of the Western U.S.A.

by

Ibrahim Nourein Mohammed, Doctor of Philosophy
Utah State University, 2012

Major Professor: Dr. David G. Tarboton
Department: Civil and Environmental Engineering

Land cover and climate change with their associated impacts on runoff are among the pressing areas of research within the western United States. In the first paper of this dissertation, we identified a total of 39 watersheds draining to U.S. Geological Survey (USGS) streamflow gauges, chosen either from the USGS Hydroclimatic Climatic Data Network of gauges that are minimally impacted by anthropogenic alterations, or because they have long, relatively continuous streamflow records and are representative of large areas within the study region in Utah. In each of these watersheds we examined trends in precipitation, temperature, snow, streamflow and runoff ratio as well as land use and land cover information. In addition, we developed a water balance model to quantify the sensitivity of runoff to changes in vegetation based on differences in evapotranspiration from different land cover types.

The second paper addressed runoff sensitivity to land cover changes in a spatially explicit way by performing detailed simulations using a Regional Hydro Ecological Simulation
System (RHESSys) model applied to the Weber River near Oakley watershed (USGS gauge # 10128500). Our runoff sensitivity results suggest that during winter reduced Leaf Area Index (LAI) decreases canopy interception, which tends to increase snow accumulations, and hence snow available for runoff during the early spring melt season. Increased LAI during spring melt season tends to delay the snow melting process due to reduced radiation beneath high LAI surfaces relative to low LAI surfaces.

The last paper examined the sensitivity of the Great Salt Lake level to changes in streamflow input or changes in climate that manifest as changes in air temperatures over the lake. We quantified this sensitivity by examining an elasticity measure defined as the ratio of the variability of streamflow, precipitation, evaporation, area and salinity to the variability in historic volume changes. We also developed a mass balance model to simulate lake level and volume driven by stochastic precipitation, streamflow and climate inputs. We showed that fluctuation in streamflow is the dominant factor in lake level fluctuations, but that fluctuations in lake area, which modulates evaporation and precipitation directly on the lake, are also important.
PUBLIC ABSTRACT

Relationships between Runoff, Land Cover and Climate in the Semi-Arid Intermountain Region of the Western U.S.A.

by

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Hydrologic science is broadly defined as the geoscience that describes and predicts the occurrence, circulation and distribution of the water of the earth and its atmosphere. The need to predict the effects of land cover and climate changes with their associated impacts on hydrology are among the pressing areas of research, especially within the western United States. The main objective of this dissertation is to provide a better knowledge of land use or land cover along with climate change impacts on streamflow generated from watersheds in the semi-arid intermountain region in and around Utah. This objective was addressed using analyses of historical observations, as well as statistical and physically based computer models. The results show trends in streamflow and other associated climate quantities in the Utah watersheds studied. The dissertation also presents a quantification of the sensitivity of streamflow to changes in land cover related to watershed management. The Great Salt Lake, a closed basin lake that receives inputs from its surrounding watershed, has a level that fluctuates in response to streamflow and other
inputs. A model for fluctuations in Great Salt Lake volume and level was used to quantify the sensitivity of these fluctuations to changes in streamflow inputs, and climate conditions over the lake that drive evaporation from the lake.
To my mother Bothyna, my father Nourein, my wife Khadiga, my daughter Qinwan, my brother Mohammed and my sisters Nisreen, Nora, Nuha and Najwan. This dissertation is dedicated to all of them.

Verily! In the creation of the heavens and the earth and in the alternation of night and day and the ships which sail through the sea with that which is of use to mankind and the water (rain) which Allah (SWT) sends down from the sky and makes the earth alive therewith after its death and the moving (living) creatures of all kinds that He has scattered therein and in the veering of winds and clouds which are held between the sky and the earth, are indeed Ayât (proofs, evidences, signs, etc.) for people of understanding. Quran 2:164
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With pleasure and deep sense of gratitude, I express my sense of indebtedness to my major advisor, Dr. David G. Tarboton. I have benefitted from his invaluable guidance, collaborations, suggestions, thoughts, critiques and scientific opinions. His care that spanned through my graduate study in Utah State has shaped my life. Every endeavor has its infrastructure and research is no exception. Also, it is a great pleasure to thank my committee members for their discussions and contributions which have truly improved and enhanced the quality of this dissertation. Dave, my committee members, as well as other people that I collaborated with during my study in U.S., have made this research possible.

My sincere gratitude goes to my wife, Khadiga, whom I truly thank for her strong stand beside me during my study. Her devotion, love and care have made things true. Without her, I can say I could not have accomplished what I have done.

My gratitude is to my father – Nourein; my mother – Bothyna; my sisters – Nisreen, Nora, Noha, Najwan; and my brother – Mohammed whom I love for their strong stand and encouragement during my study. This success is accredited to their wishes and prayers.

I would like also to voice my gratitude to all my relatives and especially my maternal uncle Talaat for his sincere care. I do appreciate his encouragement and best wishes that helped me to reach this point.

Life has another dimension through knowing friends. I would like to thank all my friends who have been ever supportive. I take this opportunity to thank my colleagues in the hydrology lab and back home in Sudan who stood behind me. They shared with me my
feelings, discussed with me this work, listened to my presentations and talks and prayed for me to accomplish this achievement.

Financial support for this work was from the State of Utah Governor's Public Lands Policy Coordination Office and the Utah Water Research Laboratory. I am grateful for this support that helped me accomplish this work.

Above all, I would like to thank Allah (SWT) for his greatest bounties and blessings that He bestows upon me and my family.

Ibrahim Nourein Mohammed
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CHAPTER 1

INTRODUCTION

Land cover and climate changes with their associated impacts on the availability of fresh water are among the pressing areas of research within western United States. One of the challenges stated by the Consortium of Universities for the Advancement of Hydrologic Science (CUAHSI) science plan is predicting the effects of climate change and human population on water resources [CUAHSI, 2002]. In order to address that challenge the science plan proposed more comprehensive, systematic understanding of continental water dynamics. In the light of that, understanding the basic hydrologic processes and water balance components and how they affect each other is vital. The water cycle study group at the US Global Change Research Program (USGCRP) cited that the most promising scientific approach for water management is predictive modeling [USGCRP, 2001]. Using models to understand the physical mechanisms that control the variability of water cycle components is a goal for much current research.

Changes in land use, land cover or water management systems result in complex interactions of various processes which in turn affect the runoff response. Climate change has a pronounced effect on the amount, timing and form of precipitation which in turn alters the runoff response. The central theme of this work is examining how watershed management and land use or land cover along with climate change impact water production, through analysis of historical observations and statistical and physical based computer models.

The overarching goal of this work is to be able to quantify the sensitivity of hydrology to land cover, land use and climate change. Specifically, we are interested in estimating the
runoff from a watershed as driven by precipitation and snowmelt and how this runoff might change as land use, land cover or climate changes. Since runoff is a major water balance input for the Great Salt Lake (GSL), we are also interested in predicting the level of the GSL when runoff from the surrounding GSL watersheds is impacted by climate and land use changes. The specific questions this work addressed were:

A. How watershed management and land use impacts water production from watersheds in Utah?

B. How does vegetation change in a mountain environment impact runoff?

C. When considering the effects associated with management decisions in watersheds contributing to the Great Salt Lake for 30 years into future, how would changes in streamflow input to the Great Salt Lake or climate conditions over the lake impact the Great Salt Lake level?

This dissertation is made up of five chapters including this introduction and a summary chapter (Chapter 5). The middle three chapters forming the core of this dissertation were written in the format of papers intended for publication as separate journal articles. Each is outlined in the following paragraphs. It worth noting that, chapter 2 has been submitted to the Utah Governor Public Land Office as a part of a study requested by the Six County Association of Governments (SCAOG) to Utah State University to provide support for the SCAOG’s response to the socioeconomic impact analyses in the Richfield Bureau of Land Management (BLM) Resources Management plan.

Chapter 2 focuses on the general hydrology of state of Utah and address the first part of our objectives. In Chapter 2, we examined factors that impact the production of runoff from selected Utah watersheds, focusing on factors related to land and watershed
management. Within each of the study watersheds, we examined trends in precipitation, temperature, snow, streamflow, runoff ratio as well as land use and land cover information. To model the water balance sensitivity to landscape vegetation changes, we developed a water balance model to quantify the sensitivity of runoff production to changes in vegetation based on differences in evapotranspiration from different land cover types. The water balance model focused around water balance principles where precipitation water inputs are ultimately partitioned between outflow, either surface or subsurface and evapotranspiration. The model approach was drawn upon the climatological ideas of Budyko [1974] who found an empirical function that at a broad scale approximates water balance partitioning based on an index of dryness.

Since the empirical watershed runoff modeling approach introduced in Chapter 2 has a number of specific deficiencies which are: i) the model results depend directly on assumed land cover coefficients, ii) the model depends upon the assumption that evapotranspiration is driving differences in water yield (in semi-arid regions differences in water yield may be more due to retention relative to snowmelt and rainfall intensity) and iii) the simplification of considering land cover change without including the topographic setting. In order to address these deficiencies and to evaluate runoff sensitivity to land cover changes in a spatially explicit way, we performed detailed simulations using the Regional Hydro Ecological Simulation System (RHESSys) model [Band et al., 1993; 1996; Tague and Band, 2001; 2004] in Chapter 3. The RHESSys simulations are intended to address the dependence between the distribution of land cover types and topography missing in the empirical watershed runoff model and calculate evapotranspiration and water balance directly rather than relying on empirical coefficients. RHESSys also simulates evapotranspiration and retention, so can be
used to evaluate the extent to which changes in potential evapotranspiration drive differences in water yield.

Chapter 4 focuses on the Great Salt Lake volume change sensitivities to variables and factors that are related to the system dynamics. Analysis of the GSL volume change sensitivity to streamflow fluctuations draining to the lake, precipitation changes on the lake, lake area changes, evaporation from the lake and lake salinity conditions in terms of the ratio of variabilities was presented. The sensitivities presented here offer a way to model the lake volume changes and help in understanding the lake dynamics. In addition, Chapter 4 presented a mass balance model developed to generate representative realizations of future lake level from the climate and streamflow input records simulated using the $k$-nearest neighbor method [Lall and Sharma, 1996]. The model incorporated lake salinity impacts on the GSL evaporation [Mohammed, 2006]. The forecast times presented in the model were monthly up to 30 years ahead initialized on 2010. Uncertainty analysis for the GSL level forecasts based upon different historic input data was provided. The overall understanding gained from the GSL level forecast model results would facilitate better management of the lake resources and better assessment of hazards due to lake level fluctuations.

This dissertation provides a knowledge base for i) integrative quantitative procedure for understanding relations among watershed management practices and water balance quantities, ii) Hydrological processes sensitivities to climate and land cover changes and iii) Physical and chemical changes effects on closed basin hydrology, i.e. the Great Salt Lake.

References


Executive Summary

The amount of water produced from a watershed depends on the climate, soils, geology, land cover and land use. Precipitation water inputs in the form of rain or snow are partitioned by the watershed into evapotranspiration, runoff and groundwater recharge. This study has examined factors that may impact the production of runoff from Utah watersheds, focusing on factors related to land and watershed management. Specifically we are interested in how land use changes, such as afforestation, deforestation, agricultural, urban, industrial and mining development, impact runoff. The scale of interest is regional subbasins at the USGS cataloging unit 8-digit Hydrologic Unit Code (HUC) scale (http://water.usgs.gov/GIS/huc.html). Twelve 8-digit HUCs in Utah, with an average area of 4500 km\(^2\) were selected for this study. Within these subbasins we identified a total of 39 watersheds draining to USGS streamflow gages, chosen either from the USGS Hydroclimatic Climatic Data Network of gages that are minimally impacted by anthropogenic alterations, or to be representative of large areas within the chosen HUCs with long relatively continuous streamflow records. In each of these watersheds we examined trends in precipitation, temperature, snow, streamflow and runoff ratio. Runoff ratio is the fraction of precipitation that becomes streamflow. We also examined land use and land cover information for these

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watersheds from the national land cover dataset, southwest regional GAP analyses and the Utah division of water resources water related land use inventory.

The most consistent trend noted was in temperature which is increasing. We did not note any significant trends in precipitation. Fourteen of the 39 watersheds examined had significant decreasing trends in streamflow and runoff ratio. We were unable to find definitive causes for these streamflow and runoff ratio trends, though we do have indications that some of them are associated with human development, storage in reservoirs and land cover and land use changes.

In analysis of the land cover data we found that unequivocal interpretation of land cover changes was confounded by differences in methodology and technology used to determine land cover over time. We were consequently unable to derive relationships from the data as to how land cover and land use affect water production.

So as to provide some information helpful for land management policy making and economic analyses we developed a water balance approach that quantifies sensitivity of runoff production to changes in land cover based on differences in evapotranspiration from different land cover types. The coefficients that quantify the potential evapotranspiration from each land cover type in this analysis are based on our judgment and information from the literature. In coming up with these coefficients we also endeavored to reconcile them with precipitation, streamflow and runoff ratio data for the Utah study watersheds. This water balance approach provides predictions of how water production from these Utah watersheds may change with land cover changes. By considering a range of water balance model parameters we provide water balance derived bounds on how streamflow could change given land cover changes. However, we caution that in the use of these results, the
sensitivities depend directly upon the coefficients that quantify the potential evapotranspiration from each land cover type. This represents a fairly gross simplification. In semi-arid settings the vegetation water use is often limited by water availability, rather than potential evapotranspiration, and differences in water yield may relate more to factors such as the timing and rate of water inputs (precipitation intensity and snowmelt). Vegetation also depends strongly on topographic setting, due to factors such as elevation, aspect and solar radiation exposure. Changes in the proportioning of land cover in a watershed therefore should consider the control that topographic setting has on land cover.

2.1. Introduction

Watershed development and management require an understanding of basic hydrologic processes, i.e. water balance components, and how they affect each other. Current concerns that are motivating the study of water production in arid regions include climate change, impacts of land management and management of water supplies. Changes in land use or management systems result in complex interactions of various processes which in turn affect runoff. The objective of this study was to address the broad question as to how watershed management and land use impacts water production from watersheds in Utah. To address this general question, the following specific questions were considered:

1. What is the water input to Utah watersheds?
2. What is the natural runoff from Utah watersheds?
3. How does land use and management impact runoff?
4. What is the economic value or impact of changes in water availability resulting from land use and management of watersheds?
2.2. Literature Review

Watershed management, land cover, land use and climate change may have both immediate and long-lasting impacts on terrestrial hydrology, altering the balance between rainfall and evapotranspiration and the resultant runoff. The impact of vegetation cover change on the hydrological cycle in the past has been studied through paired catchments. Paired catchments have helped determine the magnitude of water yield changes in response to changes in vegetation. The main categories of paired catchment studies are afforestation experiments, deforestation experiments, re-growth experiments and forest conversion experiments. These field experiments quantify the consequences of land use changes on annual runoff, flood and low flow response and water quality. Hibbert [1967] reviewed 39 studies of the effect of altering forest cover on water yield and concluded that the reduction of forest cover increases water yield and in contrast the establishment of forest cover on sparsely vegetated land decreases water yield. He indicated a practical upper limit of yield increase of 4.5 mm (0.18 in) per year for each percentage reduction in forest cover, although the increases were considerably less in the western U.S. (Colorado) data he reviewed. Bosch and Hewlett [1982] compiled data from 94 catchment experiments (watersheds from the U.S., New Zealand, Japan and Australia, including one, Beaver Creek, in Utah) that were essentially consistent with Hibbert’s findings. Bosch and Hewlett [1982] concluded that “Coniferous forest, Deciduous hardwood, Bush and Grass cover have (in that order) a decreasing influence on water yield from the source areas in which these covers are manipulated.” They noted that, on average, there is approximately 40 mm (1.6 in) increase in annual water yield for 10% reduction in Coniferous forest cover. For Deciduous hardwood forest they found on average 25 mm (1 in) increase in annual water yield for 10% reduction
in forest cover, while 10% reduction of bush and grassland generates on average a 10 mm (0.4 in) water yield increase.

Stednick [1996] assessed water yield changes after vegetation removal by analyzing 95 catchments studies in U.S. with one catchment study in Utah [Chicken Creek, UT, Johnston, 1984]. Stednick [1996] noted that in the Rocky Mountain/Inland Intermountain region studies the annual water yield increase, when 50% of the catchment was harvested, ranged from 25 mm (1 in) to 250 mm (10 in). For complete harvesting (100%), the annual water yield increase ranged from zero to over 350 mm (14 in). The regression that Stednick [1996] fit to data from the Rocky Mountain/Inland intermountain region had a slope indicating 9.4 mm (0.4 in) yield increase per 10% of area harvested. Stednick [1996] indicated that streamflow variation in response to vegetation conversion depends both on the region's annual precipitation and on the precipitation for the year under treatment. Johnston [1984] reported results from a paired catchment study in Chicken Creek watershed, in the head waters of Farmington Canyon about 14 miles northeast of Salt Lake City. Johnston reported that removing aspen from 13% of the watershed had no significant effect on streamflow yield. However, Stednick [1996] included Johnston's results in his study and reported a yield rate of 24.5 mm (1 in) per 10% of area harvested. Troendle et al. [2001] demonstrated that water yield augmentation technology, developed from research on small experimental watersheds, would work well at an operational scale. After removal of forest from 23.7% of Coon Creek watershed, a 1673 ha catchment on the Upper East Fork of the Encampment River, Wyoming, seasonal streamflow (April–October) increased on an average 76 mm (3 in) for the first five years after harvest.
Another approach for examining the effect of land use changes on a watershed’s hydrological response is to use physically based and spatially distributed ecosystems, land surface and hydrological models [Abbott et al., 1986; Refsgaard, 1987; Bathurst and O’Connell, 1992; VanShaar et al., 2002; Calder et al., 2003; Bathurst et al., 2004]. VanShaar et al. [2002] selected four catchments within the USA portion of the Columbia River Basin (ranging from 27 to 1033 km$^2$) to simulate the hydrological effects of changes in land cover using the DHSVM model [Wigmosta et al., 1994; Wigmosta and Lettenmaier, 1999]. VanShaar et al. [2002] indicated that lower leaf area, i.e. decreased vegetation extent, has led to increased snow accumulation, increased streamflow and reduced evapotranspiration. They also noted that streamflow changes are greatest during spring snowmelt runoff and evaporation changes are greatest when soils are moister (i.e., spring and early summer).

Calder et al. [2003] examined different types of vegetation and their possible impacts on water resources due to a proposed doubling of woodland area within United Kingdom by the year 2045. Observations in grass, heath, oak and pine were used with the water use model HYLUC, [Calder, 2003], to derive predictions of the impacts of different vegetation types on recharge at Clipstone Forest, Nottinghamshire, United Kingdom. The results from this study, which was conducted in a relatively dry region of Britain, demonstrated the extreme sensitivity of recharge plus runoff to vegetation covers. Calder et al. [2003] found that oak woodland is predicted to have a significant impact through its reduction of recharge plus runoff by almost one half when compared to grassland.

Land cover changes often impact evapotranspiration which in turn affects runoff. Extensive field work has been done in United Kingdom to observe the effects of land use changes on runoff through evapotranspiration changes by Calder [1986; 1993; 1998; 2003]
and Calder et al. [2003]. These studies have developed many models to assess the effect of afforestation and deforestation on water production. The approach taken to develop annual evaporation models appropriate to assess the effects of land cover and land use changes on runoff involved partitioning evaporation into two components, transpiration and interception. Interception was estimated by relation to the amount and duration of precipitation, whilst transpiration was determined in relation to a reference evaporation estimate [Calder, 1990]. Estimating the transpiration fraction, $\beta$, which is the ratio between the actual annual evaporation and annual reference potential transpiration estimate for different types of vegetation has been well tested in many experiments within U. K. [Calder, 1990].

Changes in water yield due to land cover changes can be addressed by considering variability in climate and water balance components. The water balance components of a watershed are: 1) Precipitation water input which is comprised of snow and rain detonated as, $P$; 2) the streamflow that leaves the watershed, $Q$; 3) evapotranspiration that leaves the watershed, $E$; 4) change in storage water within the watershed. Milly [1994] hypothesized that the long-term water balance is determined only by the local interaction of fluctuating water supply and demand mediated by the water storage in the soil. This hypothesis uses the concept of water holding capacity to summarize the role played by the land-water environment in hydrologic response, while ignoring many of the details of soil water flow, thereby providing a practical way to model the system when information on detailed variability of hydrological processes is limited. Milly [1994] suggested that partitioning of precipitation into runoff and evapotranspiration is determined by seven dimensionless numbers. These numbers are the ratio of annual potential evapotranspiration to annual
precipitation (index of dryness), the ratio of water holding capacity to annual mean precipitation, the mean number of precipitation events per year, the ratio of seasonal fluctuations to annual means of precipitation, storm arrival rate, potential evapotranspiration and the spatial variability of water holding capacity.

*Budyko* [1974] presented a physical, practical and meaningful explanation for climate variability through the climate index ratio which is the ratio between mean annual evaporation and mean annual precipitation (E/P). A low climate index ratio means a wet climate while a high climate index ratio means a dry climate. The *Budyko* [1974] curve is suggested to describe the geographical variation of E/P as a function of the ratio of the mean annual potential or reference evaporation (surrogate for the net radiant energy) and annual precipitation, R/P and serves as a practical tool available to explain some of the variability seen in hydrological processes. Spatial variability of the relationship between annual runoff and annual precipitation is credited to *L’vovich* [1979], who explained the geographical variations of the relationship between annual runoff and annual precipitation by presenting different climates, soils and vegetation including the way that vegetation adapts to water stress (leaf shedding & deep rooting) and how they would affect the spatial variability of the relationship between runoff and precipitation. *Sivapalan* [2005] surmised that runoff variability predictors include climate, catchment area and shape, river network, soil properties, geology, topography and vegetation.

A considerable body of work has examined trends and changes in hydrological variables in the Western United States, where streamflow is snowmelt driven. *Cayan et al.* [2001] documented the early onset of spring in the western United States by examining changes in the blooming of plants (lilac and honeysuckle bushes) and the timing of spring
snowmelt pulses. *McCabe and Wolock* [2002] observed a step increase in streamflow in the contiguous United States over the period 1941-99, with pronounced increases in the eastern United States after 1970. *Aguado et al.* [1992] and *Dettinger and Cayan* [1995] reported that increasing winter temperature, as observed in several parts of the western United States, reduces the amount of snow in a basin (e.g., more precipitation falling as rain than snow). *Mote* [2003] studied trends of Snow Water Equivalent (SWE) in the Pacific Northwest and observed strong declines in 1 April SWE, in spite of increases in precipitation, which is consistent with an increase in spring temperature. *Regonda et al.* [2005] analyzed streamflow, snowpack, temperature and precipitation in snowmelt-dominated river basins in the western United States. They found that significant declines in monthly SWE and increases in winter precipitation are evident for many stations in the western United States. The largest declines are occurring in the Pacific Northwest region, the northern parts of Idaho, Utah, Wyoming and the Sierra Nevada region. In addition, they found an indication of an advance in the timing of peak spring season flows over the past 50 years. They argued that the trends in SWE can be influenced by both temperature and precipitation. They also noted that during recent decades more precipitation is coming as rain rather than snow. *Mote et al.* [2005] extended the *Mote* [2003] study by incorporating the entire Western US from the Continental Divide to the pacific and from central British Columbia, Canada south to southern Arizona and New Mexico. In addition, they augmented the long-term monthly manual observations of snow with a more recent dataset of daily-telemetered snow observations. Moreover, they corroborate the analysis of snow data using a hydrological model [the Variable Infiltration Capacity model, VIC, *Liang et al.*, 1994] with observed daily temperature and precipitation data. Their findings are generally consistent with the earlier
work reviewed above. Overall, this body of work shows that widespread declines in spring time SWE have occurred in much of the North American West over the period 1925-2000, especially since mid-century [Aguado et al., 1992; Cayan et al., 2001; Mote, 2003; Regonda et al., 2005].

Summarizing the discussion above, the study of land cover changes impact on streamflow has been addressed by analyzing observations in paired catchments and using physically based and conceptual models. Runoff variability predictors include climate, catchment area and shape, river network, soil properties, geology, topography and vegetation. Streamflow increase after vegetation removal has been addressed in many studies in the Rocky mountains region. Notable among these studies are Wagon Wheel Gap, Fool Creek, Deadhorse Creek and Fraser Experimental Forest (FEF) in central Colorado [Bates and Henry, 1928; Troendle and King, 1985; 1987; Van Haveren, 1988; Troendle and Olsen, 1994; Troendle and Reuss, 1997]. Reduction of forest cover decreases evapotranspiration which increases water yield, while in contrast the establishment of forest cover on sparsely vegetated land decreases water yield. However, there are studies that indicate increased snow accumulation in areas with lower vegetation density, which may counter this effect.

With respect to water availability in the western United States, significant declines in monthly snow water equivalent (SWE) and increases in winter rain, rather than snow are evident for many watersheds. The largest declines in SWE are in the Pacific Northwest region, the northern parts of Idaho, Utah, Wyoming and the Sierra Nevada region. In addition, there is indication of an advance in the timing of peak spring season flows over the past 50 years.
2.3. Data

2.3.1. Streamflow

39 watersheds were selected across Utah to study the trends in and sensitivity of streamflow, Q, to land cover changes. The hydrological team at the State Engineer office provided a list of USGS cataloging unit subbasins (HUC 8) which are of interest to the state. Within these subbasins we identified a total of 39 watersheds draining to USGS streamflow gages, chosen either from the USGS Hydroclimatic Climatic Data Network (HCDN) (http://pubs.usgs.gov/wri/wri934076/1st_page.html) of gages that are minimally impacted by anthropogenic alterations, or to be representative of large areas within the chosen HUCs with long relatively continuous streamflow records. The delineated study watersheds are mapped in Figure 2.1 and listed in Table 2.1 which gives the United State Geological Survey (USGS) streamflow station at the outlet as well as drainage area for each watershed. The 4-digit Watershed ID in the first column of this table is used to identify watersheds in this study. HCDN stream gages [1993] are stream gages deemed to be relatively free of controls, diversion, or human impacts and are therefore suitable for the study of surface water conditions and climate studies. Five HCDN stations were used in this study. The streamflow dataset for the remaining gages was retrieved from USGS surface water data for Utah website (http://nwis.waterdata.usgs.gov/ut/nwis/sw).

2.3.2. Precipitation and temperature

Long term precipitation (P) and temperature (T) data were obtained from the Surface Water Modeling group at the University of Washington (http://www.hydro.washington.edu/Lettenmaier/Data/gridded/index_hamlet.html). The development of this gridded dataset is described by Hamlet and Lettenmaier [2005]. This
dataset includes daily 1/8–degree resolution gridded meteorological data for 1 Jan 1915–31 Dec 2003, grouped into the Northwest and Columbia, California, Great Basin and Colorado River regions. We extracted the data for our study watersheds from the data sets for the Great Basin and Colorado River regions.

The Parameter-elevation Regressions on Independent Slopes Model (PRISM) is an analytical method that uses point data, a digital elevation model and other spatial data sets to generate gridded estimates of monthly, yearly and event-based climatic parameters, such as precipitation, temperature and dew point (http://www.prism.oregonstate.edu/). The PRISM Group was established at Oregon State University (OSU) to provide spatial climate research datasets. Figure 2.2 gives the 30-year (1971–2000) average annual precipitation over the state of Utah retrieved from the PRISM Group website. This gives a general sense of the variability of mean annual precipitation across the study watersheds. When these values are aggregated over the study watersheds, annual precipitation averages range about 700 mm (28 in) in the highest elevation watersheds to about 180 mm (7 in) in the drier watersheds.

2.3.3. Snow

Snow water equivalent (SWE) datasets were obtained from the Natural Resources Conservation Service (NRCS) automated SNOTEL system (http://www.wcc.nrcs.usda.gov/snotel/Utah/utah.html). All SNOTEL sites in the 39 study watersheds were used in this study. An average time series of maximum and April 1st SWE values in each watershed was calculated by averaging the individual SNOTEL station maximum and April 1st values. These averages were adjusted to account for bias due to
different lengths of record at sites that have differing average SWE. This adjustment procedure from Mohammed [2006] is given in the Appendix.

Snow covered area data was retrieved from the National Operational Hydrologic Remote Sensing Center (NOHRSC) (http://www.nohrsc.noaa.gov/). The number of snow covered days within each year when the percentage of the snow covered area is more than 50% of the watershed was used to examine snow covered area trends.

2.3.4. Land Cover and Land Use

We investigated land cover and land use changes using multiple data sources: the Southwest Regional Gap Analysis Project (SWReGAP) (http://earth.gis.usu.edu/landcover.html), the GAP Analysis Program datasets (http://gapanalysis.nbii.gov/portal/server.pt), the National Land Cover Dataset (NLCD) (http://gisdata.usgs.net/website/MRLC/viewer.php) and water related land use files from the Utah Division of water resources. The SWReGAP is a multi–institutional cooperative effort coordinated by the U.S. Geological Survey Gap Analysis Program. The primary objective of the SWReGAP is to use a coordinated mapping approach to create detailed, seamless GIS maps of land cover, all native terrestrial vertebrate species, land stewardship and management status for the five-state region encompassing Arizona, Colorado, Nevada, New Mexico and Utah [Lowry et al., 2005]. The SWReGAP product gives the land cover and land use characteristics in 2004 while the GAP product is for 1995.

2.4. Trend and Data Analysis

Streamflow and precipitation trends were analyzed for all the study watersheds. These are presented in a series of figures (Figure 2.3 through Figure 2.41). Data was
aggregated for all the water years (October to September) of record for each watershed. In each of these figures, the first panel plots per unit area streamflow (Q/A) versus precipitation. Q/A is also referred to as runoff. The units used for both precipitation and runoff are meters as these are volume per unit area or depth quantities. Different symbols are used to partition the data into three time periods: 1916–1979, 1980–1995 and 1995–2003 so as to see whether the runoff production function may have changed over time. The second panel gives the time series of per unit area streamflow (Q/A). The third panel gives the time series of annual runoff ratio (Q/AP). This is the ratio of annual per unit area streamflow to precipitation and quantifies the fraction of the input precipitation that leaves the watershed as streamflow. To the extent that streamflow represents usable water, runoff ratio quantifies the water that is "produced" in the watershed as a fraction of precipitation. The fourth panel gives the time series of precipitation. Trends in these figures are visualized using LOWESS evaluated using the default parameters in the R software package [Cleveland, 1981; R Development Core Team, 2008].

Mann Kendall trend analysis [Helsel and Hirsch, 2002, chapter 8 & 12] was used to examine whether any trends were statistically significant. No trends in precipitation were found to be statistically significant (results not shown). Table 2.2 gives trend analysis results for the runoff ratio (Q/AP). This table includes the mean annual runoff ratio, μ, the annual standard deviation, σ, the coefficient of variation, CV, the lag 1 correlation, ρ1, the Kendall’s tau correlation coefficient, τ, as well as the p-value associated with the Mann Kendall test. Table 2.2 shows that there are 14 stations with significant (p < 0.05) decreasing trends in runoff ratio. Five of these are highly significant (p < 0.001), namely Weber River near Plain City, Virgin River at Virgin, Rock Creek near Mountain Home, Duchesne River near Tabiona
and Sevier River at Hatch. There is one station, the Jordan River and surplus canal at Salt Lake City that shows a highly significant increasing trend. This station is highly impacted by managed releases from Utah Lake.

SNOTEL Snow Water Equivalent was averaged across each watershed using the bias correction procedure described in the appendix. Figure 2.42 and Figure 2.43 give maps of the 2006 maximum and April 1\textsuperscript{st} snow water equivalent as an illustration of the spatial pattern of snow across the study watersheds. Snow trend analyses are presented in Figure 2.44 through Figure 2.50, for the watersheds with significant snow. Panels in these figures give the maximum, April 1\textsuperscript{st} and the mean snow water equivalent (calculated over 12 months) time series as well as the number of days with snow covered area more than 50% from the NOHRSC data. These figures show a declining trend of SWE represented in maximum, April 1\textsuperscript{st} and mean across most of the watersheds studied. This is consistent with the findings given in the literature review of widespread declines in SWE across much of the North American West over the period 1925–2000, especially since mid-century. However we should caution that the interpretation of trends from SNOTEL data suffers from the fact that records are short and some SNOTEL stations are impacted by other external influences [Julander and Bricco, 2006].

Air temperature trend analyses are presented in Figure 2.51 through Figure 2.89. These use the University of Washington gridded air temperature data averaged for each watershed and then averaged for water year (October to September) in panel 1 and for three "seasons" in panels 2 to 4, (Winter: November, December, January and February; Spring: March, April, May and June; and Summer: July, August, September and October). These
graphs show that there is an increasing air temperature trend in most of the watersheds studied.

Land cover information from the National Land Cover Dataset (NLCD) for 1992 and 2001 was summarized for each study watershed. However when we started looking at changes we found that different land cover classifications had been used in each dataset and upon deeper investigation noted that the NLCD 1992 and NLCD 2001 are not designed for direct comparison \cite{Homer et al., 2007, and http://www.epa.gov/mrlc/change.html}. This difference in methods used to produce nationally available land cover datasets makes them unsuitable for detecting the change in land cover and land use. Work is reportedly underway in the federal agencies involved with land cover monitoring (USGS, EPA, NOAA) to resolve these differences and produce National data appropriate for change monitoring \cite{http://www.epa.gov/mrlc/nlcd-2006.html}. Such data was however not available at the time of this study.

In order to detect the change in land cover and land use between 1995 and 2004 we modified the 1995 GAP classification to be consistent with the 2004 SWReGAP classification using a cross bridge classification analysis provided by the Utah State University RS/GIS laboratory (Ramsey, personal communication, 2008). This modification was necessary because the classification schemes in the two products are not the same (SWReGAP 2004 classes are different from 1995 GAP classes). Table 2. 3 groups land cover into five broad categories used in this study and presents the area percentage of each for the study watersheds for 1995 & 2004 from the modified 1995 GAP and 2004 SWReGAP studies. This serves as a preliminary analysis of land cover and land use change in the study watersheds within the State of Utah. Despite using the modified 1995 GAP, we still have some concerns
as to how many of the changes indicated in this table are real, versus methodological differences. Nevertheless we attempted a broad classification of the changes and noted that there appear to be four predominant change classes:

1. Increasing Barren area – 7 watersheds (1401, 1501, 1800, 2100, 2200, 2201, 2202).
2. Increasing Deciduous Forest, mostly together with a reduction in Range/Shrub/Other – 13 watersheds (1201, 1202, 1203, 1204, 1301, 1302, 1400, 1403, 1700, 1803, 1804, 2000, 2202).
3. Decreasing Deciduous Forest – 4 watersheds (1102, 1900, 1901, 1902).
4. Decreasing Coniferous Forest, mostly to Range/Shrub/Other and some to Deciduous Forest – 8 watersheds (1100, 1500, 1600, 2101, 2102, 2103, 2104, 2105, 2202).

Note that watershed 2202 appears in multiple categories.

Comparing these change classes to the runoff ratio trends identified in Table 2.2, we note that four of the seven watersheds with increasing barren area (1800, 2200, 2201, 2202) have decreasing runoff trends. Four of the 13 watersheds with increasing deciduous have decreasing runoff trends (1201, 1301, 1803, 2202) and four of the nine watersheds with decreasing coniferous forest have decreasing runoff trends (2101, 2104, 2105, 2202). There are also four watersheds with runoff trends that are not identified as having land use changes (1200, 1402, 1802, 2001). These patterns are sometimes counter to hydrologic understanding (decreasing runoff with decreasing coniferous forest) and there are a comparable number of watersheds in each land cover change class that do not have significant runoff ratio trends, compared to those that do. These patterns therefore do not appear to have a consistency that could allow them to be used for prediction.
Table 2.4 gives the water related land use from the Utah Division of water resources for the study watersheds. The table shows irrigated agricultural lands, non irrigated agricultural lands, residential urban areas, riparian lands, other urban lands (not residential which includes commercial, industrial, etc) and areas of open water (reservoirs). The columns $A_{86}$, $A_{96}$ etc. give the area associated with the indicated land use in km$^2$ in the year corresponding to the subscript. These do not add up to the total area of the watershed, because water related land use only covers a portion (sometimes a small portion) of each watershed. Urban growth can be seen from the data in Table 2.4 in watersheds where the areas designated as RES (residential) or URB (other urban) have increased. This is most notable in the Weber River watershed where urban area has doubled, but this table confirms that urban growth is occurring across the state. In Table 2.4, increased areas designated as open water occur in a majority of the study watersheds. This we take to be indicative of water development such as diversions and reservoirs that are likely to influence streamflow and may be responsible for some of the streamflow trends observed.

2.5. Water Balance Sensitivity Model

A water balance sensitivity model was developed to quantify the sensitivity of runoff production to changes in land cover based on differences in evapotranspiration from different land cover types. The approach assumes that potential or reference evapotranspiration is a function of land cover type and that relative differences can be quantified by a set of land cover coefficients for reference evapotranspiration from each land cover type. The set of coefficients used in this study were based upon our judgment but with reference to the literature. The average water balance partitioning function introduced by Budyko [1974] was used to estimate the basin average reference evapotranspiration within a
watershed. Then when land cover is changed this reference evapotranspiration is adjusted based on the land cover reference evapotranspiration coefficients. This is fed back in to the Budyko curve to estimate actual basin average evapotranspiration and streamflow for the changed conditions. This procedure was used to estimate the sensitivity of streamflow from each watershed to change in each land cover type.

The water balance of a watershed may be stated as:

\[ P = q + E + \Delta S \]  \hspace{1cm} (1)

where \( P \) is the precipitation input, \( q \) is the runoff that leaves the watershed, \( E \) is the evaporation and transpiration that leaves the watershed and \( \Delta S \) is the change in storage of water within the watershed. In this equation we have used the lowercase notation \( q \) to represent runoff, i.e. streamflow on a per unit area basis, to keep this distinct from \( Q \) used earlier as a volume. These are related through, \( q = Q/A \). All quantities in equation (1) are expressed in depth units. Equation (1) quantifies the proportioning of precipitation into runoff, evaporation and storage. Over long time scales the change in storage may often be neglected, such as if the time scale is several years and there is no net increase or decrease in subsurface or reservoir storage during this period or if the equation is interpreted as quantifying the ultimate disposition of water input. Then this equation can be written as:

\[ q = P - E \]  \hspace{1cm} (2)

This expresses the fact that runoff is the difference between \( P \) and \( E \) and that both variability in \( P \) and \( E \) impact runoff. Land use and watershed management changes have some direct impacts on runoff, \( q \), but the most significant impacts of land use and watershed management are often on evaporation and transpiration, namely \( E \). For example reduction of forest cover is generally presumed to reduce \( E \), while increases in forest cover increase \( E \).
Implicit here is the assumption that the land cover that replaces forests demands less water.

Budyko [1974] presented a semiempirical expression for average water balance partitioning as a function of the relative magnitudes of water and energy supply rates. He stated the relation as:

\[
\frac{\bar{E}}{\bar{P}} = \varphi \left( \frac{\bar{R}}{\bar{P}} \right)
\]

(3)

where \( \bar{E} \) is the average annual evaporation, \( \bar{P} \) is the average annual precipitation, \( \bar{R} \) is the mean annual potential or reference evapotranspiration (surrogate for the net radiant energy) and \( \varphi \) is a general partitioning function. Budyko [1974] suggested the following partitioning function, \( \varphi \), based upon fitting to data:

\[
\varphi(x) = \varphi_0(x) \equiv [x \, \tanh(x^{-1}) \left(1 - \cosh(x) + \sinh(x)\right)]^{1/2}
\]

(4)

Based upon the identification of net radiation as a control on evaporation under condition of available water supply many relationships have been proposed similar to equation (4) [Brutsaert, 1982; Choudhury, 1999]. Choudhury [1999] presented a family of functions that can be used to represent \( \varphi(x) \) as:

\[
\varphi(x) = \varphi_1(x) \equiv [1 + x^{-\nu}]^{-\nu/2}
\]

(5)

where \( \nu \) is a curve parameter. Milly and Dunne [2002] refer to equation (5) as the generalized Turc-Pike relation [Turc, 1954; Pike, 1964]. Moreover, Milly and Dunne [2002] indicate that the generalized Turc-Pike relation closely approximates the Budyko relation, equation (4), when \( \nu = 2 \).

The approach presented in this study has the following assumptions:
1. The potential evapotranspiration from each specific land cover type, $E_{plc}$, is a function of the reference evapotranspiration and can be written as:

$$E_{plc} = R^* \times r_{lc}$$  \hfill (6)

where $R^*$ is the regional reference evapotranspiration based on energy available and $r_{lc}$ is the land cover relative potential evapotranspiration coefficient.

2. The watershed average land cover adjusted reference evapotranspiration, $R$, is calculated as:

$$R = \sum E_{plc} \times P_{lc} = R^* \sum P_{lc} \times r_{lc}$$  \hfill (7)

where $P_{lc}$ is the proportion of land cover area for each specific land cover type.

3. The generalized Budyko function is taken to be applicable at watershed scale resulting in $E/P = \varphi(R/P)$. Thus, changes in land cover result in changes in $R$ which through this equation results in changes in $E$ and runoff, $q = P-E$.

Table 2.5 shows the values of the relative potential evapotranspiration coefficients, $r_{lc}$, for land cover types used in this study. These coefficients represent our judgment based upon reading the literature [e.g. Calder, 1993; Federer et al., 1996; Dingman, 2002] as to the differences in potential evapotranspiration for different land covers. In arriving at the coefficients used in Table 2.5 we considered reported values for leaf conductance, leaf area index, albedo, vegetation height and vegetation density. These coefficients are the principle determinants of the differences in runoff that we calculate.
The approach used to quantify the sensitivity of runoff to land cover changes within a watershed for the runoff production model can be expressed through the following steps:

1. From annual average estimates of precipitation, $P$ and runoff, $q$ we estimate the annual actual evapotranspiration through equation (2).

2. Using the actual evapotranspiration ratio, i.e. $E/P$, we find $R$, the watershed average land cover adjusted reference evapotranspiration, by solving equation (5).

3. The regional reference evapotranspiration, $R^*$, is then estimated from

$$R^* = R/\sum P_i \times r_i$$

(8)

4. For new land cover proportions, $P'_{lc}$, the new watershed average land cover adjusted reference evapotranspiration is then calculated as:

$$R = R' \sum P'_{lc} \times r_i = R \frac{\sum P'_{lc} \times r_i}{\sum P_{lc} \times r_i}$$

(9)

This assumes that $R^*$ remains the same and that $R$ is adjusted based on the coefficients $r_i$ and changing $P_{lc}$. The second expression of (9) above bypasses equation (8).

5. The new average annual actual evapotranspiration, $E'$, is then found by solving equation (5) with $R'$. In other words, $E'/P = \phi(R'/P)$.

6. A new estimate of streamflow, $q'$, is found through solving the mass balance equation (equation 2).
2.6. Sensitivity Results

Table 2.6 gives water balance estimates for streamflow, precipitation and evapotranspiration for the study watersheds. $Q/A$ and $P$ are based upon data described above and minimum, mean, median and maximum across the years of record are reported (in units of meters). The actual evapotranspiration column is calculated from mass balance using the mean $P$ and $Q/A$. The potential evapotranspiration is the watershed average land cover adjusted reference evapotranspiration inferred from the Budyko [1974] relation with $\nu = 2$ using mean $P$ and $Q/A$.

We evaluated the sensitivity of streamflow to land cover change for each watershed for some potential land cover changes of interest. We also used a range of Budyko curve parameters ($\nu = 1.5, 2, 10$) to explore the sensitivity of the findings to this parameter. We evaluated the sensitivity to changes in a specific land cover type by increasing it and reducing some other land cover types in proportion to their land cover fractions while holding remaining land cover proportions constant.

We found that the changes in streamflow were in general close to being linearly proportional to the changes in land cover, so have expressed sensitivity in terms of a derivative that was evaluated numerically using a 10% land cover fraction increase. For example for a watershed with 40% of a particular land cover type, the sensitivity to change of that specific land cover type would be calculated as:

$$S_{i,j}^{k \rightarrow i} = \frac{\text{Runoff}_{50\%} - \text{Runoff}_{40\%}}{0.5 - 0.4}$$

A decrease in streamflow is reflected in a negative sensitivity coefficient. Streamflow in units of acre-ft/mi$^2$/yr was used, so the sensitivity values reported are acre-ft/mi$^2$/yr per fraction change in land cover proportion.
In Table 2.7 we report the sensitivity to changes in the proportion of Coniferous land cover holding the proportion of Agricultural land cover constant and allowing other land cover proportions to change. This was done twice, first for the relative potential evapotranspiration coefficients in Table 2.5 and then for switched Coniferous and Deciduous land covers relative potential evapotranspiration coefficients. The relative potential evapotranspiration coefficients used are given in the header to the corresponding columns in Table 2.7. Switching the Coniferous and Deciduous coefficients was done, because although in Table 2.5 we have indicated greater water use from Coniferous forests there have been some studies [LaMalfa, 2007] that suggest higher water use from Aspen. We wanted to evaluate the sensitivity of water production to this question.

In Table 2.8 we report the sensitivity to changes in the proportion of Range/Shrub/Other land cover. This was done first allowing all other land cover proportions to adjust and for both the Coniferous and Deciduous coefficients from Table 2.5 and switched. Then we considered Range/Shrub/Other land cover being replaced by Forest (Deciduous and Coniferous land cover types), holding the proportion of Agricultural and Barren land covers constant, again also presenting results for coefficients from Table 2.5 and switched Coniferous and Deciduous land covers relative potential evapotranspiration coefficients.

Examination of Table 2.7 and Table 2.8 indicates that increasing Coniferous forests decreases streamflow, while increasing Range/Shrub/Other increases streamflow. The reduction in streamflow with increasing Coniferous forest, regardless of whether the relative potential evapotranspiration coefficient for Deciduous or Coniferous is greater is due to the generally large area of Range/Shrub/Other that is displaced and has smaller water use in this
sensitivity model. Streamflow reductions are calculated to be larger for the case when the relative potential evapotranspiration coefficient for Coniferous is greater. The increase in streamflow with increasing Range/Shrub/Other is similarly due to the generally large area of Forest (either Deciduous or Coniferous) that is displaced and has greater water use in this sensitivity model. Using the coefficients calculated in Table 2.7 for the set of relative potential evapotranspiration coefficients with Coniferous greater than Deciduous (Table 2.5) and \( \nu = 2 \), we found that reducing 50% of the Coniferous area present in the study watersheds, resulted in streamflow increases that ranged from 1 to 80 ac–ft/mi\(^2\)/year. Using the coefficients in Table 2.8 for the set of relative potential evapotranspiration coefficients with Coniferous greater than Deciduous (Table 2.5) and \( \nu = 2 \), we found that 50% reduction of the area present that is Range/Shrub/Others with this area transitioning to forest results in streamflow decreases that ranged from 8 to 78 ac–ft/mi\(^2\)/year.

Five watersheds were selected to show the streamflow sensitivity to changes in Coniferous and Range/Shrub/Other land cover types. From Southern Utah we selected the Virgin River near Virgin watershed (watershed_ID = 1800); from Central Utah, we selected the Sevier River at Hatch (watershed_ID = 2201); from the Uintah Basin, we selected the Duchesne River near Tabiona (watershed_ID = 2104); from the Wasatch front, we selected the Red Butte Creek at Fort Douglas near Salt Lake City (watershed_ID = 2000); and from Northern Utah, we selected the Blacksmith Fork near Hyrum watershed (watershed_ID = 1900). Figure 2.90 to Figure 2.119 give streamflow sensitivity in the above watersheds for changes in Coniferous and Range/Shrub/Other land covers. Figure 2.90 to Figure 2.94 give the transition of Coniferous land cover to Barren, Range/Shrub/Other and Deciduous with the fraction that is Agriculture held fixed and relative evapotranspiration coefficient for
Coniferous greater than Deciduous (Table 2.5). Figure 2.95 to Figure 2.99 give the transition of Coniferous land cover to Barren, Range/Shrub/Other and Deciduous with Agriculture fixed and relative evapotranspiration coefficient for Deciduous greater than Coniferous (i.e. switched from Table 2.5). Figure 2.100 to Figure 2.104 give the transition of Range/Shrub/Other land cover to Barren, Coniferous, Deciduous and Agriculture with relative evapotranspiration coefficient for Coniferous greater than Deciduous. Figure 2.105 to Figure 2.109 give the transition of Range/Shrub/Other land cover to Barren, Coniferous, Deciduous and Agriculture with relative evapotranspiration coefficient for Deciduous greater than Coniferous. Figure 2.110 to Figure 2.114 give the transition of Range/Shrub/Other land cover to Coniferous, Deciduous only, i.e. Agriculture and Barren are fixed with relative evapotranspiration coefficient for Coniferous greater than Deciduous. Figure 2.115 to Figure 2.119 give the transition of Range/Shrub/Other land cover to Coniferous, Deciduous only, i.e. Agriculture and Barren are fixed with relative evapotranspiration coefficient for Deciduous greater than Coniferous. The hatched area shown in the sensitivity of streamflow figures, expressed in streamflow per unit area, \( q \) (ac-ft/mi²/yr), represents the family of solutions to the Choudhury [1999] partitioning function, i.e. from \( \nu = 1.5 \) to \( \nu = 10 \) where \( \nu \) is the curve parameter. The red line, i.e. \( \nu = 2 \), represents the Budyko semi empirical expression for the average water balance partitioning function. The slopes shown in Figure 2.90 to Figure 2.119 give the sensitivity of runoff to change in percentage land cover (equation 10) for \( \nu = 2 \) at the existing land cover percentages, with \( \pm \) range indicated corresponding to \( \nu \) in the range from 1.5 to 10. These figures show how with this model increasing Coniferous areas leads to reducing streamflow while increasing Range/Shrub/Other leads to increasing streamflow. Switching the Coniferous and Deciduous
relative potential evapotranspiration coefficients generally reduces the streamflow sensitivity when the change is from Coniferous to other land covers, but because of the generally large area that is Range/Shrub/Other and Barren lands compared to Deciduous, the direction of the sensitivity is not changed.

2.7. Discussion and Conclusions

Mann Kendall runoff ratio trend analysis results revealed that there is a significant decreasing trend for 14 of the study watersheds, with this trend being highly significant (statistically) in Weber River near Plain City, Virgin River at Virgin, Rock Creek near Mountain Home, Duchesne River near Tabiona and Sevier River at Hatch watersheds. Analysis of the annual as well as seasonal temperature records revealed that there are increasing temperature trends for most of the watersheds studied (Figure 2. 51 through Figure 2. 89). No significant trends in precipitation were seen in any study watersheds. A decreasing trend in runoff ratio ($Q/AP$) means that less of the precipitation leaves the watershed in the form of streamflow. In the watersheds where there are significant decreases in $Q/AP$, these decreases may be directly due to diversions, storage and water use or due to increases in evapotranspiration due to land use changes or temperature changes. Five of the 39 watersheds examined were HCDN watersheds deemed to be relatively free from direct effects of diversions and use. One of these watersheds, the Sevier River at Hatch (#2201) had a decreasing streamflow and runoff ratio trend (Figure 2. 40). The cause for this is not known, although one reviewer noted that the Sevier River at Hatch has at least four major diversions in it, calling in to question its inclusion in the USGS HCDN network of relatively unimpacted streams.
We looked for patterns relating land cover change to trends in runoff ratio. Watersheds with decreasing runoff ratio trends occurred in three of the four predominant land cover change classes that we identified, as well as in watersheds where land cover was not changing. However there are a comparable number of watersheds in each predominant land cover change class that do not have significant runoff ratio trends as do have significant trends. We are consequently unable to find consistent relationships between land cover change and runoff ratio trends.

The specific causes for the decreasing trends in runoff ratio for 14 of the study watersheds is consequently not known. All potentially include areas with significant diversions and water use that has not been quantified in this study. The reductions are likely due to a combination of these diversions as well as increases in evapotranspiration related to temperature and land cover changes.

In the analysis of land cover data we found that unequivocal interpretation of land cover changes was confounded by differences in methodology and technology used to determine land cover over time. The NLCD 1992 and NLCD 2001 are not designed for direct comparison [Homer et al., 2007]. This is clearly seen in the classification scheme and the way each map was recoded (for example forest and range types). Detecting the change in land cover and land use through the state remains a challenge because of this inconsistency in methods used to produce nationally available land cover datasets, although we have been told that the USGS is testing a new tool that provides the connection between NLCD 1992, NLCD 2001 and the new product of NLCD 2006 that may resolve some of these difficulties.

Given that we were unable to derive relationships from the data as to how land cover and land use affect water production and so as to provide some information helpful for
land management policy making and economic analyses we developed the water balance approach that quantifies sensitivity of runoff production to changes in land cover based on differences in evapotranspiration from different land cover types. The coefficients that quantify the potential evapotranspiration from each land cover type in this analysis are based on our judgment and information from the literature. In coming up with these coefficients we also endeavored to reconcile them with precipitation, streamflow and runoff ratio data for the Utah study watersheds. The water balance approach was used to analyze the sensitivity of water production to land cover changes for five land cover types for the Utah study watersheds.

Physical understanding of the interactions between hydrology, climate and land cover changes is important for understanding and predicting the potential hydrological consequences of existing land use practices. The results of this work developed an integrative quantitative procedure for understanding relations among watershed management practices and water balance quantities. The theoretical approach taken in this study is simple and general and could be applied to a wide range of watersheds throughout the State. However it depends directly upon the relative evapotranspiration coefficients. It is therefore important to consider future work to better quantify the relative impact of land cover on evapotranspiration and streamflow production. Computer modeling and observations may both be required to advance knowledge in this area.

It is important to note that using the Budyko relation as a single valued function ignores the scatter that typically occur around a Budyko curve when water balance partition observations are plotted. This scatter has many causes such as seasonality and variability in hydrologic processes not captured by this aggregate model. By using a range of values for
the parameter $\nu$, some sensitivity to this uncertainty was assessed. Also, the approach presented above uses the Budyko relation in a relative sense so some of these uncertainty effects balance off against each other. This was outlined through steps 2 and 5 in which we estimated in step 2 the watershed average land cover adjusted reference evapotranspiration, $R$, while in step 5 we solve the Budyko relation to find the updated evapotranspiration given the new watershed average land cover adjusted reference evapotranspiration, $R'$. 

We caution that in the use of these sensitivity results, the sensitivities depend directly upon the coefficients that quantify the potential evapotranspiration from each land cover type. This represents a fairly gross simplification. In semi-arid settings the vegetation water use is often limited by water availability, rather than potential evapotranspiration, and differences in water yield may relate more to factors such as the timing and rate of water inputs (precipitation intensity and snowmelt). Vegetation also depends strongly on topographic setting, due to factors such as elevation, aspect and solar radiation exposure. Changes in the proportioning of land cover in a watershed therefore should consider the control that topographic setting has on land cover.

References


Table 2.1 USGS gauging stations at the outlet of each study watershed.

<table>
<thead>
<tr>
<th>Watershed ID</th>
<th>USGS #</th>
<th>Station Name</th>
<th>Drainage Area (km²)</th>
<th>Drainage Area (mile²)</th>
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<td>1100</td>
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<td>728.60</td>
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<td>262.54</td>
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<td>215.64</td>
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<td>2202</td>
<td>10173450</td>
<td>*Mammoth Creek ABV West Hatch Ditch, near Hatch</td>
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*HCDN station
## Table 2.2. Runoff ratio (Q/AP) Mann Kendall trend analysis.

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<th>Station Name</th>
<th>µ</th>
<th>σ</th>
<th>CV</th>
<th>ρ1</th>
<th>τ (tau)</th>
<th>p-value</th>
<th>Trend</th>
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<td>0.065</td>
<td>0.171</td>
<td>0.264</td>
<td>0.0765</td>
<td>0.38705</td>
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<td>0.049</td>
<td>0.575</td>
<td>0.520</td>
<td>-0.0124</td>
<td>0.90801</td>
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<td>0.575</td>
<td>0.634</td>
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<td>SOUTH FORK OGDEN RIVER NEAR HUNTSVILLE</td>
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<td>0.359</td>
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* HCDN station
Table 2.2. Continued
Key variables explanations

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<th>µ: arithmetic mean</th>
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<td>σ: unbiased standard deviation</td>
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<tr>
<td>CV: coefficient of variation (σ/µ)</td>
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<tr>
<td>(\rho_1): Lag 1 autocorrelation</td>
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<tr>
<td>(\tau) (tau): Kendall's tau correlation coefficient</td>
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<tr>
<td>p-value: 2 sided-test</td>
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<tr>
<td>Trend: (p \leq 0.05) Significant; (p \leq 0.001) Highly Significant</td>
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Table 2. Watershed land cover classification using the GAP 1995 & the SWReGAP 2004 datasets.

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<th>Alpine and Barren</th>
<th>Conifer Forest</th>
<th>Deciduous Forest</th>
<th>Range, Shrub, Others</th>
<th>Year</th>
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<td>1400</td>
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<td>35.57</td>
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* HCDN station
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*HCDN station

**Note:** The data represents water-related land use from Utah Division of Water Resources. The columns denote different types of land use: A_{total} (total area), A_{ir} (irrigated area), A_{ni} (non-irrigated area), A_{res} (residential area), A_{ip} (industrial area), A_{urb} (urban area), and A_{water} (water area). The values are in hectares (ha).
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* HCDN station
### Table 2.4. Continued

**Key variables explanations.**

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Table 2.5. Land covers relative potential evapotranspiration coefficients.

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* HCDN station.
**Table 2. 7. Runoff sensitivity to changes in Coniferous land cover percentage (acre-ft/mi²/yr)**

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*HCDN station; **v=2 i.e. Budyko relation; *C=Confident, D=decisive, R=Range/Shrub/Others, B=Barren, and A=Agriculture.
Figure 2.1. Study watersheds.
Figure 2.2. Utah 30 year (1971-2000) average annual precipitation from PRISM.
Figure 2.3. Precipitation and streamflow trend analysis for the Bear River near Utah-Wyoming state line watershed (watershed_ID = 1100).
Figure 2. 4. Precipitation and streamflow trend analysis for the Bear River near Randolph watershed (watershed_ID = 1101).
Figure 2.5. Precipitation and streamflow trend analysis for the Big Creek near Randolph watershed (watershed_ID = 1102).
Figure 2.6. Precipitation and streamflow trend analysis for the Weber River at Gateway watershed (watershed_ID = 1200).
Figure 2.7. Precipitation and streamflow trend analysis for the Weber River near Plain city watershed (watershed_ID = 1201).
Figure 2.8. Precipitation and streamflow trend analysis for the South Fork Ogden River near Huntsville watershed (watershed_ID = 1202).
**Watershed ID = 1203**

Figure 2.9. Precipitation and streamflow trend analysis for the Centerville Creek near Centerville watershed (watershed ID = 1203).
Figure 2. 10. Precipitation and streamflow trend analysis for the Ogden River below Pineview near Huntsville watershed (watershed_ID = 1204).
Figure 2. 11. Precipitation and streamflow trend analysis for the Salt Creek at Nephi watershed (watershed ID = 1300).
Figure 2.12. Precipitation and streamflow trend analysis for the Currant Creek near Mona watershed (watershed_ID = 1301).
Figure 2.13. Precipitation and streamflow trend analysis for the West Canyon Creek near Cedar Fort watershed (watershed_ID = 1302).
Figure 2.14. Precipitation and streamflow trend analysis for the Fish Creek near Scofield watershed (watershed_ID = 1400).
Figure 2. 15. Precipitation and streamflow trend analysis for the Price River at Woodside watershed (watershed_ID = 1401).
Figure 2. 16. Precipitation and streamflow trend analysis for the White River below Tabbyune Creek near Soldier Summit watershed (watershed_ID = 1402).
Figure 2.17. Precipitation and streamflow trend analysis for the Mud Creek below Winter Quarters Canyon at Scofield watershed (watershed_ID = 1403).
Figure 2. 18. Precipitation and streamflow trend analysis for the Fremont River near Bicknell watershed (watershed_ID = 1500).
Figure 2. 19. Precipitation and streamflow trend analysis for the Fremont River near Caineville watershed (watershed_ID = 1501).
Figure 2. 20. Precipitation and streamflow trend analysis for the East Fork Sevier River near Kingston watershed (watershed_ID = 1600).
Figure 2.21. Precipitation and streamflow trend analysis for the Coal Creek near Cedar City watershed (watershed_ID = 1700).
Figure 2.22. Precipitation and streamflow trend analysis for the Virgin River at Virgin watershed (watershed_ID = 1800).
Figure 2.23. Precipitation and streamflow trend analysis for the Santa Clara River at Gunlock watershed (watershed_ID = 1801).
Figure 2.24. Precipitation and streamflow trend analysis for the East Fork Virgin River near Glendale watershed (watershed_ID = 1802).
Figure 2.25. Precipitation and streamflow trend analysis for the North Fork Virgin River near Springdale watershed (watershed_ID = 1803).
Figure 2.26. Precipitation and streamflow trend analysis for the Santa Clara River above Baker near Central watershed (watershed_ID = 1804).
Figure 2.27. Precipitation and streamflow trend analysis for the Blacksmith Fork near Hyrum watershed (watershed_ID = 1900).
Figure 2. 28. Precipitation and streamflow trend analysis for the Logan River near Logan watershed (watershed_ID = 1901).
Figure 2. 29. Precipitation and streamflow trend analysis for the Little Bear River at Paradise watershed (watershed_ID = 1902).
Figure 2. 30. Precipitation and streamflow trend analysis for the Red Butte Creek at Fort Douglas near Salt Lake City watershed (watershed_ID = 2000).
Figure 2. 31. Precipitation and streamflow trend analysis for the Little Cottonwood creek at Jordan River near Salt Lake City watershed (watershed_ID = 2001).
Figure 2.32. Precipitation and streamflow trend analysis for the Jordan River & Surplus Canal at Salt Lake City watershed (watershed_ID = 2002).
Figure 2. 33. Precipitation and streamflow trend analysis for the Whiterocks River near Whiterocks watershed (watershed_ID = 2100).
Figure 2. 34. Precipitation and streamflow trend analysis for the Rock Creek near Mountain Home watershed (watershed_ID = 2101).
Figure 2. Precipitation and streamflow trend analysis for the Yellowstone River near Altonah watershed (watershed_ID = 2102).
Figure 2. 36. Precipitation and streamflow trend analysis for the Lake Fork River below Moon Lake near Mountain Home watershed (watershed_ID = 2103).
Figure 2. Precipitation and streamflow trend analysis for the Duchesne River near Tabiona watershed (watershed_ID = 2104).
Figure 2. 38. Precipitation and streamflow trend analysis for the Duchesne River above Knight Diversion near Duchesne watershed (watershed_ID = 2105).
Figure 2. 39. Precipitation and streamflow trend analysis for the Sevier River near Kingston watershed (watershed_ID = 2200).
Figure 2. 40. Precipitation and streamflow trend analysis for the Sevier River at Hatch watershed (watershed_ID = 2201).
Figure 2. 41. Precipitation and streamflow trend analysis for the Mammoth Creek above west Hatch ditch near Hatch watershed (watershed_ID = 2202).
Figure 2. 2006 Maximum Snow Water Equivalent for study watersheds.
April Snow Water Equivalent 2006 (mm)

- Yellow: 23 - 150
- Light Green: 151 - 300
- Green: 301 - 450
- Light Blue: 451 - 600
- Medium Blue: 601 - 750
- Dark Blue: 751 - 900
- Medium Green: 901 - 1050
- Dark Green: 1051 - 1200
- Dark Blue: 1201 - 1350
- Medium Green: 1351 - 1500

Figure 2. 43. 2006 April 1st Snow Water Equivalent for study watersheds.
Figure 2. Snowfall analyses for the Weber River near Plain City watershed (watershed_ID = 1201). The analyses shown above based upon water year give the maximum, the April 1st, and the mean snow water equivalent. The time span for snow water equivalent products is 1979-2006. The lower right panel gives the number of days which have greater than 50% snow covered area per year.
Figure 2. 45. Snowfall analyses for the White River below Tabbyune Creek near Soldier Summit watershed (watershed_ID = 1402). The analyses shown above based upon water year give the maximum, the April 1st and the mean snow water equivalent. The time span for snow water equivalent products is 1979-2006. The lower right panel gives the number of days which have greater than 50% snow covered area per year. Note: the snow covered area data is taken from Price River – Scofield Res. – near Scofield.
Figure 2. Snowfall analyses for the Virgin River at Virgin watershed (watershed_ID = 1800). The analyses shown above based upon water year give the maximum, the April 1st and the mean snow water equivalent. The time span for snow water equivalent products is 1979-2006. The lower right panel gives the number of days which have greater than 50% snow covered area per year.
Figure 2. 47. Snowfall analyses for the Little Cottonwood creek at Jordan River near Salt Lake City watershed (watershed_ID = 2001). The analyses shown above based upon water year give the maximum, the April 1\textsuperscript{st} and the mean snow water equivalent. The time span for snow water equivalent products is 1979-2006. The lower right panel gives the number of days which have greater than 50% snow covered area per year.
Figure 2. Snowfall analyses for the Rock Creek near Mountain Home watershed (watershed ID = 2101). The analyses shown above based upon water year give the maximum, the April 1\textsuperscript{st} and the mean snow water equivalent. The time span for snow water equivalent products is 1979-2006. The lower right panel gives the number of days which have greater than 50% snow covered area per year.
Figure 2. 49. Snowfall analyses for the Duchesne River near Tabiona watershed (watershed_ID = 2104). The analyses shown above based upon water year give the maximum, the April 1st and the mean snow water equivalent. The time span for snow water equivalent products is 1979-2006. The lower right panel gives the number of days which have greater than 50% snow covered area per year.
Figure 2. 50. Snowfall analyses for the Sevier River at Hatch watershed (watershed_ID = 2201). The analyses shown above based upon water year give the maximum, the April 1st and the mean snow water equivalent. The time span for snow water equivalent products is 1979-2006. The lower right panel gives the number of days which have greater than 50% snow covered area per year.
Figure 2. Temperature analyses for the Bear River near Utah-Wyoming state line watershed (watershed_ID = 1100). Panel (i) gives the average water year temperature time series over the watershed. Panel (ii) gives the annual average temperature in winter season (November, December, January and February) time series. Panel (iii) gives the annual average temperature in spring season (March, April, May and June) time series. Panel (iv) gives the annual average temperature in summer season (July, August, September and October) time series.
Figure 2. 52. Temperature analyses for the Bear River near Randolph watershed (watershed_ID = 1101). Panel (i) gives the average water year temperature time series over the watershed. Panel (ii) gives the annual average temperature in winter season (November, December, January and February) time series. Panel (iii) gives the annual average temperature in spring season (March, April, May and June) time series. Panel (iv) gives the annual average temperature in summer season (July, August, September and October) time series.
Figure 2. 53. Temperature analyses for the Big Creek near Randolph watershed (watershed_ID = 1102). Panel (i) gives the average water year temperature time series over the watershed. Panel (ii) gives the annual average temperature in winter season (November, December, January and February) time series. Panel (iii) gives the annual average temperature in spring season (March, April, May and June) time series. Panel (iv) gives the annual average temperature in summer season (July, August, September and October) time series.
Figure 2. Temperature analyses for the Weber River at Gateway watershed (watershed_ID = 1200). Panel (i) gives the average water year temperature time series over the watershed. Panel (ii) gives the annual average temperature in winter season (November, December, January and February) time series. Panel (iii) gives the annual average temperature in spring season (March, April, May and June) time series. Panel (iv) gives the annual average temperature in summer season (July, August, September and October) time series.
Figure 2. 55. Temperature analyses for the Weber River near Plain City watershed (watershed_ID = 1201). Panel (i) gives the average water year temperature time series over the watershed. Panel (ii) gives the annual average temperature in winter season (November, December, January and February) time series. Panel (iii) gives the annual average temperature in spring season (March, April, May and June) time series. Panel (iv) gives the annual average temperature in summer season (July, August, September and October) time series.
Figure 2. Temperature analyses for the South Fork Ogden River near Huntsville watershed (watershed_ID = 1202). Panel (i) gives the average water year temperature time series over the watershed. Panel (ii) gives the annual average temperature in winter season (November, December, January and February) time series. Panel (iii) gives the annual average temperature in spring season (March, April, May and June) time series. Panel (iv) gives the annual average temperature in summer season (July, August, September and October) time series.
Figure 2. 57. Temperature analyses for the Centerville Creek near Centerville watershed (watershed_ID = 1203). Panel (i) gives the average water year temperature time series over the watershed. Panel (ii) gives the annual average temperature in winter season (November, December, January and February) time series. Panel (iii) gives the annual average temperature in spring season (March, April, May and June) time series. Panel (iv) gives the annual average temperature in summer season (July, August, September and October) time series.
Figure 2. 58. Temperature analyses for the Ogden River below Pineview near Huntsville watershed (watershed_ID = 1204). Panel (i) gives the average water year temperature time series over the watershed. Panel (ii) gives the annual average temperature in winter season (November, December, January and February) time series. Panel (iii) gives the annual average temperature in spring season (March, April, May and June) time series. Panel (iv) gives the annual average temperature in summer season (July, August, September and October) time series.
Figure 2. 59. Temperature analyses for the Salt Creek at Nephi watershed (watershed_ID = 1300). Panel (i) gives the average water year temperature time series over the watershed. Panel (ii) gives the annual average temperature in winter season (November, December, January and February) time series. Panel (iii) gives the annual average temperature in spring season (March, April, May and June) time series. Panel (iv) gives the annual average temperature in summer season (July, August, September and October) time series.
Figure 2. 60. Temperature analyses for the Currant Creek near Mona watershed (watershed_ID = 1301). Panel (i) gives the average water year temperature time series over the watershed. Panel (ii) gives the annual average temperature in winter season (November, December, January and February) time series. Panel (iii) gives the annual average temperature in spring season (March, April, May and June) time series. Panel (iv) gives the annual average temperature in summer season (July, August, September and October) time series.
Figure 2. 61. Temperature analyses for the West Canyon Creek near Cedar Fort watershed (watershed_ID = 1302). Panel (i) gives the average water year temperature time series over the watershed. Panel (ii) gives the annual average temperature in winter season (November, December, January, and February) time series. Panel (iii) gives the annual average temperature in spring season (March, April, May, and June) time series. Panel (iv) gives the annual average temperature in summer season (July, August, September, and October) time series.
Figure 2.62. Temperature analyses for the Fish Creek near Scofield watershed (watershed_ID = 1400). Panel (i) gives the average water year temperature time series over the watershed. Panel (ii) gives the annual average temperature in winter season (November, December, January and February) time series. Panel (iii) gives the annual average temperature in spring season (March, April, May and June) time series. Panel (iv) gives the annual average temperature in summer season (July, August, September and October) time series.
Figure 2.63. Temperature analyses for the Price River at Woodside watershed (watershed_ID = 1401). Panel (i) gives the average water year temperature time series over the watershed. Panel (ii) gives the annual average temperature in winter season (November, December, January and February) time series. Panel (iii) gives the annual average temperature in spring season (March, April, May and June) time series. Panel (iv) gives the annual average temperature in summer season (July, August, September and October) time series.
Figure 2. Temperature analyses for the White River below Tabbyune Creek near Soldier Summit watershed (watershed_ID = 1402). Panel (i) gives the average water year temperature time series over the watershed. Panel (ii) gives the annual average temperature in winter season (November, December, January and February) time series. Panel (iii) gives the annual average temperature in spring season (March, April, May and June) time series. Panel (iv) gives the annual average temperature in summer season (July, August, September and October) time series.
Figure 2. Temperature analyses for the Mud Creek below Winter Quarters Canyon at Scofield watershed (watershed ID = 1403). Panel (i) gives the average water year temperature time series over the watershed. Panel (ii) gives the annual average temperature in winter season (November, December, January and February) time series. Panel (iii) gives the annual average temperature in spring season (March, April, May and June) time series. Panel (iv) gives the annual average temperature in summer season (July, August, September and October) time series.
Figure 2.66. Temperature analyses for the Fremont River near Bicknell watershed (watershed_ID = 1500). Panel (i) gives the average water year temperature time series over the watershed. Panel (ii) gives the annual average temperature in winter season (November, December, January and February) time series. Panel (iii) gives the annual average temperature in spring season (March, April, May and June) time series. Panel (iv) gives the annual average temperature in summer season (July, August, September and October) time series.
Figure 2. 67. Temperature analyses for the Fremont River near Caineville watershed (watershed_ID = 1501). Panel (i) gives the average water year temperature time series over the watershed. Panel (ii) gives the annual average temperature in winter season (November, December, January and February) time series. Panel (iii) gives the annual average temperature in spring season (March, April, May and June) time series. Panel (iv) gives the annual average temperature in summer season (July, August, September and October) time series.
Figure 2. 68. Temperature analyses for the East Fork Sevier River near Kingston watershed (watershed_ID = 1600). Panel (i) gives the average water year temperature time series over the watershed. Panel (ii) gives the annual average temperature in winter season (November, December, January and February) time series. Panel (iii) gives the annual average temperature in spring season (March, April, May and June) time series. Panel (iv) gives the annual average temperature in summer season (July, August, September and October) time series.
Figure 2. Temperature analyses for the Coal Creek near Cedar City watershed (watershed ID = 1700). Panel (i) gives the average water year temperature time series over the watershed. Panel (ii) gives the annual average temperature in winter season (November, December, January and February) time series. Panel (iii) gives the annual average temperature in spring season (March, April, May and June) time series. Panel (iv) gives the annual average temperature in summer season (July, August, September and October) time series.
Figure 2. 70. Temperature analyses for the Virgin River at Virgin watershed (watershed_ID = 1800). Panel (i) gives the average water year temperature time series over the watershed. Panel (ii) gives the annual average temperature in winter season (November, December, January and February) time series. Panel (iii) gives the annual average temperature in spring season (March, April, May and June) time series. Panel (iv) gives the annual average temperature in summer season (July, August, September and October) time series.
Figure 2. 71. Temperature analyses for the Santa Clara River at Gunlock watershed (watershed_ID = 1801). Panel (i) gives the average water year temperature time series over the watershed. Panel (ii) gives the annual average temperature in winter season (November, December, January and February) time series. Panel (iii) gives the annual average temperature in spring season (March, April, May and June) time series. Panel (iv) gives the annual average temperature in summer season (July, August, September and October) time series.
Figure 2. 72. Temperature analyses for the East Fork Virgin River near Glendale watershed (watershed_ID = 1802). Panel (i) gives the average water year temperature time series over the watershed. Panel (ii) gives the annual average temperature in winter season (November, December, January and February) time series. Panel (iii) gives the annual average temperature in spring season (March, April, May and June) time series. Panel (iv) gives the annual average temperature in summer season (July, August, September and October) time series.
Figure 2. 73. Temperature analyses for the North Fork Virgin River near Springdale watershed (watershed_ID = 1803). Panel (i) gives the average water year temperature time series over the watershed. Panel (ii) gives the annual average temperature in winter season (November, December, January and February) time series. Panel (iii) gives the annual average temperature in spring season (March, April, May and June) time series. Panel (iv) gives the annual average temperature in summer season (July, August, September and October) time series.
Figure 2. Temperature analyses for the Santa Clara River above Baker near Central watershed (watershed_ID = 1804). Panel (i) gives the average water year temperature time series over the watershed. Panel (ii) gives the annual average temperature in winter season (November, December, January and February) time series. Panel (iii) gives the annual average temperature in spring season (March, April, May and June) time series. Panel (iv) gives the annual average temperature in summer season (July, August, September and October) time series.
Figure 2. 75. Temperature analyses for the Blacksmith Fork near Hyrum watershed (watershed_ID = 1900). Panel (i) gives the average water year temperature time series over the watershed. Panel (ii) gives the annual average temperature in winter season (November, December, January, and February) time series. Panel (iii) gives the annual average temperature in spring season (March, April, May, and June) time series. Panel (iv) gives the annual average temperature in summer season (July, August, September, and October) time series.
Figure 2. Temperature analyses for the Logan River near Logan watershed (watershed_ID = 1901). Panel (i) gives the average water year temperature time series over the watershed. Panel (ii) gives the annual average temperature in winter season (November, December, January and February) time series. Panel (iii) gives the annual average temperature in spring season (March, April, May and June) time series. Panel (iv) gives the annual average temperature in summer season (July, August, September and October) time series.
Figure 2. 77. Temperature analyses for the Little Bear River at Paradise watershed (watershed_ID = 1902). Panel (i) gives the average water year temperature time series over the watershed. Panel (ii) gives the annual average temperature in winter season (November, December, January and February) time series. Panel (iii) gives the annual average temperature in spring season (March, April, May and June) time series. Panel (iv) gives the annual average temperature in summer season (July, August, September and October) time series.
Figure 2. Temperature analyses for the Red Butte Creek at Fort Douglas near Salt Lake City watershed (watershed_ID = 2000). Panel (i) gives the average water year temperature time series over the watershed. Panel (ii) gives the annual average temperature in winter season (November, December, January and February) time series. Panel (iii) gives the annual average temperature in spring season (March, April, May and June) time series. Panel (iv) gives the annual average temperature in summer season (July, August, September and October) time series.
Figure 2. 79. Temperature analyses for the Little Cottonwood creek at Jordan River near Salt Lake City watershed (watershed_ID = 2001). Panel (i) gives the average water year temperature time series over the watershed. Panel (ii) gives the annual average temperature in winter season (November, December, January and February) time series. Panel (iii) gives the annual average temperature in spring season (March, April, May and June) time series. Panel (iv) gives the annual average temperature in summer season (July, August, September and October) time series.
Figure 2. 80. Temperature analyses for the Jordan River & Surplus Canal at Salt Lake City watershed (watershed_ID = 2002). Panel (i) gives the average water year temperature time series over the watershed. Panel (ii) gives the annual average temperature in winter season (November, December, January and February) time series. Panel (iii) gives the annual average temperature in spring season (March, April, May and June) time series. Panel (iv) gives the annual average temperature in summer season (July, August, September and October) time series.
Figure 2. 81. Temperature analyses for the Whiterocks River near Whiterocks watershed (watershed_ID = 2100). Panel (i) gives the average water year temperature time series over the watershed. Panel (ii) gives the annual average temperature in winter season (November, December, January and February) time series. Panel (iii) gives the annual average temperature in spring season (March, April, May and June) time series. Panel (iv) gives the annual average temperature in summer season (July, August, September and October) time series.
Figure 2. 82. Temperature analyses for the Rock Creek near Mountain Home watershed (watershed_ID = 2101). Panel (i) gives the average water year temperature time series over the watershed. Panel (ii) gives the annual average temperature in winter season (November, December, January and February) time series. Panel (iii) gives the annual average temperature in spring season (March, April, May and June) time series. Panel (iv) gives the annual average temperature in summer season (July, August, September and October) time series.
Figure 2. 83. Temperature analyses for the Yellowstone River near Altonah watershed (watershed_ID = 2102). Panel (i) gives the average water year temperature time series over the watershed. Panel (ii) gives the annual average temperature in winter season (November, December, January and February) time series. Panel (iii) gives the annual average temperature in spring season (March, April, May and June) time series. Panel (iv) gives the annual average temperature in summer season (July, August, September and October) time series.
Watershed_ID = 2103

(i) Year

(ii) Winter

(iii) Spring

(iv) Summer

Figure 2. 84. Temperature analyses for the Lake Fork River below Moon Lake near Mountain Home watershed (watershed_ID = 2103). Panel (i) gives the average water year temperature time series over the watershed. Panel (ii) gives the annual average temperature in winter season (November, December, January and February) time series. Panel (iii) gives the annual average temperature in spring season (March, April, May and June) time series. Panel (iv) gives the annual average temperature in summer season (July, August, September and October) time series.
Figure 2.85. Temperature analyses for the Duchesne River near Tabiona watershed (watershed ID = 2104). Panel (i) gives the average water year temperature time series over the watershed. Panel (ii) gives the annual average temperature in winter season (November, December, January and February) time series. Panel (iii) gives the annual average temperature in spring season (March, April, May and June) time series. Panel (iv) gives the annual average temperature in summer season (July, August, September and October) time series.
Figure 2. 86. Temperature analyses for the Duchesne River above Knight Diversion near Duchesne watershed (watershed_ID = 2105). Panel (i) gives the average water year temperature time series over the watershed. Panel (ii) gives the annual average temperature in winter season (November, December, January and February) time series. Panel (iii) gives the annual average temperature in spring season (March, April, May and June) time series. Panel (iv) gives the annual average temperature in summer season (July, August, September and October) time series.
Figure 2. 87. Temperature analyses for the Sevier River Kingston watershed (watershed_ID = 2200). Panel (i) gives the average water year temperature time series over the watershed. Panel (ii) gives the annual average temperature in winter season (November, December, January and February) time series. Panel (iii) gives the annual average temperature in spring season (March, April, May and June) time series. Panel (iv) gives the annual average temperature in summer season (July, August, September and October) time series.
Figure 2. 88. Temperature analyses for the Sevier River at Hatch watershed (watershed_ID = 2201). Panel (i) gives the average water year temperature time series over the watershed. Panel (ii) gives the annual average temperature in winter season (November, December, January and February) time series. Panel (iii) gives the annual average temperature in spring season (March, April, May and June) time series. Panel (iv) gives the annual average temperature in summer season (July, August, September and October) time series.
Figure 2. Temperature analyses for the Mammoth Creek above west Hatch ditch near Hatch watershed (watershed_ID = 2202). Panel (i) gives the average water year temperature time series over the watershed. Panel (ii) gives the annual average temperature in winter season (November, December, January and February) time series. Panel (iii) gives the annual average temperature in spring season (March, April, May and June) time series. Panel (iv) gives the annual average temperature in summer season (July, August, September and October) time series.
Figure 2. 90. Runoff sensitivity to change in Coniferous land cover percentage in the Duchense River near Tabiona watershed. Percentage of Agriculture held fixed while percentages of other land covers adjust proportionately. The slope gives the runoff sensitivity at $\nu = 2$ with ± bounds from $\nu = 1.5$ to $\nu = 10$. The relative potential evapotranspiration coefficients are: Coniferous 0.9, Deciduous 0.8, Range/Shrub/Other 0.6, Barren 0.5 and Agriculture 1.0 (fixed). The watershed area is 356.37 $\text{mi}^2$. 
Figure 2. Runoff sensitivity to change in Coniferous land cover percentage in the Virgin River near Virgin watershed. Percentage of Agriculture held fixed while percentages of other land covers adjust proportionately. The slope gives the runoff sensitivity at $\nu = 2$ with ± bounds from $\nu = 1.5$ to $\nu = 10$. The relative potential evapotranspiration coefficients are: Coniferous 0.9, Deciduous 0.8, Range/Shrub/Other 0.6, Barren 0.5 and Agriculture 1.0 (fixed). The watershed area is 948.32 mi$^2$. 

slope = -74.1 + (-83.0, +13.8)
Figure 2. Runoff sensitivity to change in Coniferous land cover percentage in the Blacksmith Fork near Hyrum watershed. Percentage of Agriculture held fixed while percentages of other land covers adjust proportionately. The slope gives the runoff sensitivity at $\nu = 2$ with ± bounds from $\nu = 1.5$ to $\nu = 10$. The relative potential evapotranspiration coefficients are: Coniferous 0.9, Deciduous 0.8, Range/Shrub/Other 0.6, Barren 0.5 and Agriculture 1.0 (fixed). The watershed area is 262.54 mi$^2$. 

```
existing land percentages

slope = -133.7 + (-172.8, +25.7)

acre - ft/ft$^3$/yr
fraction change in area proportion

Coniferous %
Watershed_ID=1900
```
Figure 2. Runoff sensitivity to change in Coniferous land cover percentage in the Red Butte creek at Fort Douglas near SLC watershed. Percentage of Agriculture held fixed while percentages of other land covers adjust proportionately. The slope gives the runoff sensitivity at $\nu = 2$ with $\pm$ bounds from $\nu = 1.5$ to $\nu = 10$. The relative potential evapotranspiration coefficients are: Coniferous 0.9, Deciduous 0.8, Range/Shrub/Other 0.6, Barren 0.5 and Agriculture 1.0 (fixed). The watershed area is 7.24 mi$^2$. 
Figure 2. 94. Runoff sensitivity to change in Coniferous land cover percentage in the Sevier River near Hatch watershed. Percentage of Agriculture held fixed while percentages of other land covers adjust proportionately. The slope gives the runoff sensitivity at $\nu = 2$ with ± bounds from $\nu = 1.5$ to $\nu = 10$. The relative potential evapotranspiration coefficients are: Coniferous 0.9, Deciduous 0.8, Range/Shrub/Other 0.6, Barren 0.5 and Agriculture 1.0 (fixed). The watershed area is 335.21 mi$^2$. 
Figure 2.95. Runoff sensitivity to change in Coniferous land cover percentage in the Duchense River near Tabiona watershed. Percentage of Agriculture held fixed while percentages of other land covers adjust proportionately. The slope gives the runoff sensitivity at $\nu = 2$ with ± bounds from $\nu = 1.5$ to $\nu = 10$. The relative potential evapotranspiration coefficients are: Coniferous 0.8, Deciduous 0.9, Range/Shrub/Other 0.6, Barren 0.5 and Agriculture 1.0 (fixed). The watershed area is 356.37 mi$^2$. 

\begin{align*}
\text{Coniferous} & \quad |0.8| \\
\text{Deciduous} & \quad |0.9| \\
\text{Range/Shrub/Other} & \quad |0.6| \\
\text{Barren} & \quad |0.5| \\
\text{Agriculture} & \quad |1.0| \\
\end{align*}

$\nu = 2$

existing land percentages

slope = $-81.8 + (\pm 88.4, +15.6)$

acre-ft/m$^2$/yr

fraction change in area proportion

Watershed ID = 2104

The table below shows the runoff (acre-ft/m$^2$/yr) for different percentages of land cover:

<table>
<thead>
<tr>
<th>Coniferous %</th>
<th>Runoff (acre-ft/m$^2$/yr)</th>
</tr>
</thead>
<tbody>
<tr>
<td>9.83</td>
<td>25.97</td>
</tr>
<tr>
<td>7.86</td>
<td>20.77</td>
</tr>
<tr>
<td>5.9</td>
<td>15.58</td>
</tr>
<tr>
<td>3.93</td>
<td>10.39</td>
</tr>
<tr>
<td>1.97</td>
<td>5.19</td>
</tr>
<tr>
<td>0</td>
<td>0</td>
</tr>
</tbody>
</table>

Table entries for other land cover percentages are similarly provided in the text.
Figure 2. Runoff sensitivity to change in Coniferous land cover percentage in the Virgin River near Virgin watershed. Percentage of Agriculture held fixed while percentages of other land covers adjust proportionately. The slope gives the runoff sensitivity at $\nu = 2$ with $\pm$ bounds from $\nu = 1.5$ to $\nu = 10$. The relative potential evapotranspiration coefficients are: Coniferous 0.8, Deciduous 0.9, Range/Shrub/Other 0.6, Barren 0.5 and Agriculture 1.0 (fixed). The watershed area is 948.32 mi$^2$. 
Figure 2. 97. Runoff sensitivity to change in Coniferous land cover percentage in the Blacksmith Fork near Hyrum watershed. Percentage of Agriculture held fixed while percentages of other land covers adjust proportionately. The slope gives the runoff sensitivity at $\nu = 2$ with $\pm$ bounds from $\nu = 1.5$ to $\nu = 10$. The relative potential evapotranspiration coefficients are: Coniferous 0.8, Deciduous 0.9, Range/Shrub/Other 0.6, Barren 0.5 and Agriculture 1.0 (fixed). The watershed area is 262.54 mi$^2$. 

Watershed_ID=1900
Figure 2. Runoff sensitivity to change in Coniferous land cover percentage in the Red Butte creek at Fort Douglas near SLC watershed. Percentage of Agriculture held fixed while percentages of other land covers adjust proportionately. The slope gives the runoff sensitivity at $\nu = 2$ with $\pm$ bounds from $\nu = 1.5$ to $\nu = 10$. The relative potential evapotranspiration coefficients are: Coniferous 0.8, Deciduous 0.9, Range/Shrub/Other 0.6, Barren 0.5 and Agriculture 1.0 (fixed). The watershed area is 7.24 mi$^2$. 
Figure 2. 99. Runoff sensitivity to change in Coniferous land cover percentage in the Sevier River near Hatch watershed. Percentage of Agriculture held fixed while percentages of other land covers adjust proportionately. The slope gives the runoff sensitivity at ν = 2 with ± bounds from ν = 1.5 to ν = 10. The relative potential evapotranspiration coefficients are: Coniferous 0.8, Deciduous 0.9, Range/Shrub/Other 0.6, Barren 0.5 and Agriculture 1.0 (fixed). The watershed area is 335.21 mi².
Figure 2. 100. Runoff sensitivity to changes in Range/Shrub/Other land cover percentage in the Duchense River near Tabiona watershed. Other land cover percentages adjust proportionately. The slope gives the runoff sensitivity at ν = 2 with ± bounds from ν = 1.5 to ν = 10. The relative potential evapotranspiration coefficients are: Coniferous 0.9, Deciduous 0.8, Range/Shrub/Other 0.6, Barren 0.5 and Agriculture 1.0. The watershed area is 356.37 mi².
Figure 2. Runoff sensitivity to changes in Range/Shrub/Other land cover percentage in the Virgin River near Virgin watershed. Other land cover percentages adjust proportionately. The slope gives the runoff sensitivity at $\nu = 2$ with $\pm$ bounds from $\nu = 1.5$ to $\nu = 10$. The relative potential evapotranspiration coefficients are: Coniferous 0.9, Deciduous 0.8, Range/Shrub/Other 0.6, Barren 0.5 and Agriculture 1.0. The watershed area is 948.32 mi$^2$. 

Watershed_ID=1800
Figure 2. Runoff sensitivity to changes in Range/Shrub/Other land cover percentage in the Blacksmith Fork near Hyrum watershed. Other land cover percentages adjust proportionately. The slope gives the runoff sensitivity at $\nu = 2$ with $\pm$ bounds from $\nu = 1.5$ to $\nu = 10$. The relative potential evapotranspiration coefficients are: Coniferous 0.9, Deciduous 0.8, Range/Shrub/Other 0.6, Barren 0.5 and Agriculture 1.0. The watershed area is 262.54 mi$^2$. 

+ Watershed_ID=1900

$\text{slope} = 143.4 + (-28.1, +173.0)$

$\frac{\text{acre-ft/mi}^2/\text{yr}}{\text{fraction change in area proportion}}$
Figure 2. Runoff sensitivity to changes in Range/Shrub/Other land cover percentage in the Red Butte creek at Fort Douglas near SLC watershed. Other land cover percentages adjust proportionately. The slope gives the runoff sensitivity at \( \nu = 2 \) with \( \pm \) bounds from \( \nu = 1.5 \) to \( \nu = 10 \). The relative potential evapotranspiration coefficients are: Coniferous 0.9, Deciduous 0.8, Range/Shrub/Other 0.6, Barren 0.5 and Agriculture 1.0. The watershed area is 7.24 mi\(^2\).
Figure 2. 104. Runoff sensitivity to changes in Range/Shrub/Other land cover percentage in the Sevier River near Hatch watershed. Other land cover percentages adjust proportionately. The slope gives the runoff sensitivity at \( \nu = 2 \) with ± bounds from \( \nu = 1.5 \) to \( \nu = 10 \). The relative potential evapotranspiration coefficients are: Coniferous 0.9, Deciduous 0.8, Range/Shrub/Other 0.6, Barren 0.5 and Agriculture 1.0. The watershed area is 335.21 mi\(^2\).
Figure 2.105. Runoff sensitivity to changes in Range/Shrub/Other land cover percentage in the Duchense River near Tabiona watershed. Other land cover percentages adjust proportionately. The slope gives the runoff sensitivity at $\nu = 2$ with $\pm$ bounds from $\nu = 1.5$ to $\nu = 10$. The relative potential evapotranspiration coefficients are: Coniferous 0.8, Deciduous 0.9, Range/Shrub/Other 0.6, Barren 0.5 and Agriculture 1.0. The watershed area is 356.37 mi$^2$. 

slope = 136.7 + (-26.3, +138.0) 

acre-ft/m$^2$/yr 

fraction change in area proportion

Watershed_ID=2104
Figure 2. Runoff sensitivity to changes in Range/Shrub/Other land cover percentage in the Virgin River near Virgin watershed. Other land cover percentages adjust proportionately. The slope gives the runoff sensitivity at $\nu = 2$ with $\pm$ bounds from $\nu = 1.5$ to $\nu = 10$. The relative potential evapotranspiration coefficients are: Coniferous 0.8, Deciduous 0.9, Range/Shrub/Other 0.6, Barren 0.5 and Agriculture 1.0. The watershed area is 948.32 mi$^2$. 

Existing land percentages: 

- Barren: 9.65, 7.72, 5.79, 3.86, 1.93, 0
- Agriculture: 23.88, 19.1, 14.33, 9.55, 4.78, 0
- Coniferous: 28.31, 22.65, 16.98, 11.32, 5.66, 0
- Deciduous: 38.16, 30.53, 22.9, 15.27, 7.63, 0
- Range/Shrub/Other: 38.16, 30.53, 22.9, 15.27, 7.63, 0

Runoff (acre-ft/mi$^2$/yr) 

- Range and Shrub %: 58, 88, 118, 148, 177
- Watershed_ID=1800

Runoff (acre-ft/yr) 

- fraction change in area proportion
- slope = 41.9 + (-8.7, +74.4)

Watershed_ID=1800 

Figure 2. 106. Runoff sensitivity to changes in Range/Shrub/Other land cover percentage in the Virgin River near Virgin watershed. Other land cover percentages adjust proportionately. The slope gives the runoff sensitivity at $\nu = 2$ with $\pm$ bounds from $\nu = 1.5$ to $\nu = 10$. The relative potential evapotranspiration coefficients are: Coniferous 0.8, Deciduous 0.9, Range/Shrub/Other 0.6, Barren 0.5 and Agriculture 1.0. The watershed area is 948.32 mi$^2$. 
Figure 2. 107. Runoff sensitivity to changes in Range/Shrub/Other land cover percentage in the Blacksmith Fork near Hyrum watershed. Other land cover percentages adjust proportionately. The slope gives the runoff sensitivity at $\nu = 2$ with $\pm$ bounds from $\nu = 1.5$ to $\nu = 10$. The relative potential evapotranspiration coefficients are: Coniferous 0.8, Deciduous 0.9, Range/Shrub/Other 0.6, Barren 0.5 and Agriculture 1.0. The watershed area is 262.54 mi$^2$. 
Figure 2. 108. Runoff sensitivity to changes in Range/Shrub/Other land cover percentage in the Red Butte creek at Fort Douglas near SLC watershed. Other land cover percentages adjust proportionately. The slope gives the runoff sensitivity at $\nu = 2$ with $\pm$ bounds from $\nu = 1.5$ to $\nu = 10$. The relative potential evapotranspiration coefficients are: Coniferous 0.8, Deciduous 0.9, Range/Shrub/Other 0.6, Barren 0.5 and Agriculture 1.0. The watershed area is 7.24 mi$^2$. 
Figure 2. 109. Runoff sensitivity to changes in Range/Shrub/Other land cover percentage in the Sevier River near Hatch watershed. Other land cover percentages adjust proportionately. The slope gives the runoff sensitivity at $\nu = 2$ with ± bounds from $\nu = 1.5$ to $\nu = 10$. The relative potential evapotranspiration coefficients are: Coniferous 0.8, Deciduous 0.9, Range/Shrub/Other 0.6, Barren 0.5 and Agriculture 1.0. The watershed area is 335.21 mi$^2$. 

\[ \text{slope} = 88.8 + (-18.9, +135.6) \]
Figure 2. 110. Runoff sensitivity to change in Range/Shrub/Other percentage in the Duchense River near Tabiona watershed. The percentages of Agriculture and Barren are held fixed while percentages of Coniferous and Deciduous adjust proportionately. The slope gives the runoff sensitivity at ν = 2 with ± bounds from ν = 1.5 to ν = 10. The relative potential evapotranspiration coefficients are: Coniferous 0.9, Deciduous 0.8, Range/Shrub/Other 0.6, Barren 0.5 (fixed) and Agriculture 1.0 (fixed). The watershed area is 356.37 mi².
Figure 2. Runoff sensitivity to change in Range/Shrub/Other percentage in the Virgin River near Virgin watershed. The percentages of Agriculture and Barren are held fixed while percentages of Coniferous and Deciduous adjust proportionately. The slope gives the runoff sensitivity at \( \nu = 2 \) with ± bounds from \( \nu = 1.5 \) to \( \nu = 10 \). The relative potential evapotranspiration coefficients are: Coniferous 0.9, Deciduous 0.8, Range/Shrub/Other 0.6, Barren 0.5 (fixed) and Agriculture 1.0 (fixed). The watershed area is 948.32 mi\(^2\).
Figure 2. Runoff sensitivity to change in Range/Shrub/Other percentage in the Blacksmith Fork near Hyrum watershed. The percentages of Agriculture and Barren are held fixed while percentages of Coniferous and Deciduous adjust proportionately. The slope gives the runoff sensitivity at $\nu = 2$ with $\pm$ bounds from $\nu = 1.5$ to $\nu = 10$. The relative potential evapotranspiration coefficients are: Coniferous 0.9, Deciduous 0.8, Range/Shrub/Other 0.6, Barren 0.5 (fixed) and Agriculture 1.0 (fixed). The watershed area is 262.54 mi$^2$. 
Figure 2.113. Runoff sensitivity to change in Range/Shrub/Other percentage in the Red Butte creek at Fort Douglas near SLC watershed. The percentages of Agriculture and Barren are held fixed while percentages of Coniferous and Deciduous adjust proportionately. The slope gives the runoff sensitivity at $\nu = 2$ with ± bounds from $\nu = 1.5$ to $\nu = 10$. The relative potential evapotranspiration coefficients are: Coniferous 0.9, Deciduous 0.8, Range/Shrub/Other 0.6, Barren 0.5 (fixed) and Agriculture 1.0 (fixed). The watershed area is 7.24 mi$^2$. 
Figure 2. Runoff sensitivity to change in Range/Shrub/Other percentage in the Sevier River near Hatch watershed. The percentages of Agriculture and Barren are held fixed while percentages of Coniferous and Deciduous adjust proportionately. The slope gives the runoff sensitivity at $\nu = 2$ with ± bounds from $\nu = 1.5$ to $\nu = 10$. The relative potential evapotranspiration coefficients are: Coniferous 0.9, Deciduous 0.8, Range/Shrub/Other 0.6, Barren 0.5 (fixed) and Agriculture 1.0 (fixed). The watershed area is 335.21 mi$^2$. 
Figure 2. Runoff sensitivity to change in Range/Shrub/Other percentage in the Duchense River near Tabiona watershed. The percentages of Agriculture and Barren are held fixed while percentages of Coniferous and Deciduous adjust proportionately. The slope gives the runoff sensitivity at $\nu = 2$ with ± bounds from $\nu = 1.5$ to $\nu = 10$. The relative potential evapotranspiration coefficients are: Coniferous 0.8, Deciduous 0.9, Range/Shrub/Other 0.6, Barren 0.5 (fixed) and Agriculture 1.0 (fixed). The watershed area is 356.37 m$^2$. 

\[
\text{Range and Shrub %} \quad \nu = 2 \quad \text{existing land percentages}
\]

\[
\text{slope} = 158.6 + (-30.4, +166.1)
\]

\[
\text{acre} - \text{ft}^2 / \text{m}^2 / \text{yr}
\]

\[
\frac{\text{fraction change in area proportion}}{\text{acre} - \text{ft}^2 / \text{yr}}
\]

\[
\text{Runoff (acre - ft}^2 / \text{yr)}
\]

\[
\text{Runoff (acre - ft}^2 / \text{yr)}
\]

\[
\text{Watershed ID}=2104
\]
Figure 2. Runoff sensitivity to change in Range/Shrub/Other percentage in the Virgin River near Virgin watershed. The percentages of Agriculture and Barren are held fixed while percentages of Coniferous and Deciduous adjust proportionately. The slope gives the runoff sensitivity at $v = 2$ with bounds from $v = 1.5$ to $v = 10$. The relative potential evapotranspiration coefficients are: Coniferous 0.8, Deciduous 0.9, Range/Shrub/Other 0.6, Barren 0.5 (fixed) and Agriculture 1.0 (fixed). The watershed area is 948.32 mi$^2$. 

Watershed ID=1800
Figure 2. Runoff sensitivity to change in Range/Shrub/Other percentage in the Blacksmith Fork near Hyrum watershed. The percentages of Agriculture and Barren are held fixed while percentages of Coniferous and Deciduous adjust proportionately. The slope gives the runoff sensitivity at ν = 2 with ± bounds from ν = 1.5 to ν = 10. The relative potential evapotranspiration coefficients are: Coniferous 0.8, Deciduous 0.9, Range/Shrub/Other 0.6, Barren 0.5 (fixed) and Agriculture 1.0 (fixed). The watershed area is 262.54 mi$^2$. 
Figure 2. Runoff sensitivity to change in Range/Shrub/Other percentage in the Red Butte creek at Fort Douglas near SLC watershed. The percentages of Agriculture and Barren are held fixed while percentages of Coniferous and Deciduous adjust proportionately. The slope gives the runoff sensitivity at $\nu = 2$ with ± bounds from $\nu = 1.5$ to $\nu = 10$. The relative potential evapotranspiration coefficients are: Coniferous 0.8, Deciduous 0.9, Range/Shrub/Other 0.6, Barren 0.5 (fixed) and Agriculture 1.0 (fixed). The watershed area is 7.24 mi².
Figure 2. Runoff sensitivity to change in Range/Shrub/Other percentage in the Sevier River near Hatch watershed. The percentages of Agriculture and Barren are held fixed while percentages of Coniferous and Deciduous adjust proportionately. The slope gives the runoff sensitivity at $\nu = 2$ with ± bounds from $\nu = 1.5$ to $\nu = 10$. The relative potential evapotranspiration coefficients are: Coniferous 0.8, Deciduous 0.9, Range/Shrub/Other 0.6, Barren 0.5 (fixed) and Agriculture 1.0 (fixed). The watershed area is 335.21 mi$^2$. 

Watershed_ID=2201
CHAPTER 3

SIMULATED WATERSHED RESPONSES TO LAND COVER CHANGES USING THE REGIONAL HYDRO-ECOLOGICAL SIMULATION SYSTEM (RHESSYS)

Abstract

In this work, we used the Regional Hydro-Ecological Simulation System (RHESSys) model to examine runoff sensitivity to land cover changes in a mountain environment. Two independent experiments were evaluated where we conducted simulations with multiple vegetation cover changes that include conversion to grass, no vegetation cover and deciduous/coniferous cover scenarios. The model experiments were performed at two hillslopes within the Weber River near Oakley, Utah watershed (USGS gauge # 10128500). Daily precipitation, air temperature and wind speed data as well as spatial data that include a digital elevation model with 30 meter grid resolution, soil texture map and vegetation and land use maps were processed to drive RHESSys simulations. Observed runoff data at the watershed outlet were used for calibration and verification. Our runoff sensitivity results suggest that during winter reduced Leaf Area Index decreases canopy interception resulting in increased snow accumulations, and hence snow available for runoff during the early spring melt season. Increased Leaf Area Index during the spring melt season tends to delay the snow melting process due to reduced radiation beneath high LAI surfaces relative to low LAI surfaces. The model results suggest that annual runoff yield after removing deciduous vegetation is on average about 7% higher than with deciduous vegetation cover, while annual runoff yield after removing coniferous vegetation is on average as about 2% higher than that produced with coniferous vegetation cover. These simulations thus help quantify the sensitivity of water yield to vegetation change.
3.1. Introduction

The hydrology of the western United States mountains is mainly dominated by snowmelt. Precipitation in these snowmelt dominated watersheds is stored in the form of snow, lost due to evaporation, sublimation and transpiration, or released as snowmelt driven runoff or infiltrated to groundwater that sustains baseflow.

A National Research Council [2008] report presented the current understanding of forest hydrology, connections between forest management and attendant hydrologic effects and suggested directions for future research to sustainably manage water resources from forested landscapes. General principles deduced from paired watershed studies since the 1940s are summarized in this report as: 1) partial or complete removal of the forest canopy decreases interception and increases net precipitation arriving at the soil surface, 2) partial or complete removal of the forest canopy reduces transpiration, 3) reductions in interception and transpiration increase soil moisture and water yield, 4) increased soil moisture and loss of root strength reduces slope stability, 5) increases in water yield after forest harvesting are transitory and decrease over time as forests regrow and 6) when young forests with higher annual transpiration losses replace older forests with lower transpiration losses, this change results in reduced water yield as the new forests grow to maturity, [Table S-1, National Research Council, 2008]. The report called for future research in forest hydrology to move from principles to prediction. This call to move from principles to prediction is because the science community needs to understand the indirect and interacting hydrologic responses to changes in forested landscapes associated with climate change, forest disturbances, forest species composition and structure and land development and ownership, and how these changes will affect water quantity and quality downstream and over long time scales.
Notable among paired watershed studies in the intermountain Rocky Mountain region are Wagon Wheel Gap, Fool Creek, Deadhorse Creek and Fraser Experimental Forest (FEF) in central Colorado [Bates and Henry, 1928; Troendle and King, 1985; 1987; Van Haveren, 1988; Troendle and Olsen, 1994; Troendle and Reuss, 1997]. The main categories of paired studies are afforestation, deforestation, re-growth and forest conversion experiments. These field experiments quantify the consequences of land use changes on runoff, flood and low flow response and water quality.

Meeting water supply needs is becoming more difficult because elevated water demand is occurring simultaneously with changes in climate, human population growth and development and land use. Therefore, understanding the hydrologic effects of land cover, climate as well as land use changes is an urgent challenge for hydrologic science.

In this work, we used the Regional Hydro-Ecological Simulation System (RHESSys) model to examine how vegetation change in a mountain environment impacts runoff. RHESSys model simulations were intended to address the dependence of the distribution of land cover types on topography and climate and to examine differences in runoff generated from different vegetation types. The model was also intended to assess interactions between runoff and climate represented by precipitation, air temperature and plant water use in different vegetation types to assess how these interactions may vary at seasonal time steps. The watershed selected for this paper is located at the headwater of the Weber River, Utah (USA). The Weber River is an essential water resource to the state of Utah since its water is used for municipal, irrigation, industrial, power generation and wildlife purposes.

The RHESSys framework [Band et al., 1993; 1996; Tague and Band, 2001; 2004] has been used in multiple studies to assess the impact of climate and land use change on
hydrology. RHESSys provides a detailed representation of snow, runoff, soil and vegetation processes important for addressing our questions. The RHESSys model uses a hierarchical spatial framework that allows different processes to be modeled at their most representative scale. RHESSys uses the MTN-CLIM model [Running et al., 1987] to drive spatially variable climate inputs in mountainous regions. This is important in mountain study areas. RHESSys’s flexibility in its modeling element size (tessellation) and representation of vegetation processes are also attractive. RHESSys routes water using an explicit routing model adapted from the DHSVM model [Wigmosta et al., 1994; Wigmosta and Lettenmaier, 1999] and includes an evapotranspiration calculation procedure that has a higher sensitivity to leaf area index.

The RHESSys model was first calibrated to daily streamflows at the watershed outlet during the 1994 water year. Monte Carlo simulation was used to generate 5000 sample parameter sets from independent uniform distributions over the feasible parameter ranges determined from the literature. Each was used as input to the model and a group of behavioral parameter sets was selected in terms of the Nash-Sutcliffe metric on daily flows, Nash-Sutcliffe metric on log daily flows and total annual flow error. Variability over the range of calibrated parameter values from this behavioral group was used to quantify sensitivity of runoff to parameter uncertainty.

RHESSys was then used to quantify runoff sensitivity to land cover change in two independent numerical experiments. The first experiment examined the runoff sensitivity at a hillslope with coniferous cover to vegetation changes. These vegetation change scenarios include change to grass, no vegetation and deciduous cover. The second experiment was similar to the first experiment but differed in that the hillslope vegetation cover was
deciduous not coniferous. In this second experiment we changed deciduous vegetation cover to grass, no vegetation and coniferous cover. Each experiment was done separately while holding everything else constant.

Our runoff sensitivity results suggest that reducing leaf area index which tends to decrease transpiration rates, increases wintertime snow accumulation and hence increases runoff during the spring melt season. The model results suggest that annual runoff yield after removing deciduous vegetation is on average about 7% higher than with deciduous vegetation cover, while annual runoff yield after removing coniferous vegetation is on average about 2% higher than that produced with coniferous vegetation cover. These simulations thus help quantify the sensitivity of water yield to vegetation change.

We also found that coniferous and deciduous vegetation at our study watershed (Upper Weber basin, HUC 16020101 in Utah, USA) behave similarly in terms of having evapotranspiration rates limited by available energy only with no limitation due to water availability. This is unusual in semiarid Utah but it is due to the elevation and precipitation. We think that the results presented in this work should be interpreted as best estimates that serve as hypotheses for how actual ecosystems will respond based on the knowledge and understanding that is embodied in the model. Given that a model is an idealized approximation of reality, there is a need for monitoring programs to verify model predictions. In what follows, we first review literature on runoff sensitivity to vegetation and land use changes as well as vegetation responses to climate in mountain environments. We then describe our study site and input data. We then proceed with model description, analysis and calibration, followed by results and discussion.
3.2. Literature Review

The effect of land use changes on a watershed’s hydrological response has been examined by applying physically based and spatially distributed ecosystem, land surface and hydrological models [Abbott et al., 1986; Refsgaard, 1987; Bathurst and O’Connell, 1992; Matheussen et al., 2000; VanShaar et al., 2002; Calder et al., 2003; Bathurst et al., 2004; Tague et al., 2004; Hamlet et al., 2007; Christensen et al., 2008]. It is worth noting here that a physically based model always implies a numerical discretization in one or more space coordinates. Therefore, by physically based model we mean a description which is based on a scientific physical understanding of the processes involved at a scale consistent with the adopted level of numerical discretization [Jensen and Mantoglou, 1992]. In this section different studies that examined runoff sensitivities to land cover and climate changes are discussed.

Christensen et al. [2008] used RHESSys to assess the sensitivity of transpiration rates to elevation across the Upper Merced River watershed, Yosemite Valley, California, USA. Their model results suggest that elevational differences in vegetation water use and sensitivity to climate were significant. Those elevational transpiration sensitivities to climate were noted as: 1) low elevations (1200 - 1800 m) showed little interannual variation in transpiration due to topographically controlled high soil moisture, 2) both middle and high elevations (1800 - 2600 m) showed high correlation between precipitation and transpiration. Christensen et al. sensitivity results follow the snowpack orographic effect that influences the start of vegetation growing season, largely through its effect on soil temperature.

Tague et al. [2004] examined the sensitivity of streamflow and soil moisture to land use change in a Mediterranean climate (cool wet winters and long warm dry summers). They
compared monthly and annually streamflow predictions from RHESSys and MIKE-SHE [Refsgaard and Storm, 1995] models to stream gauge data for a 34 km² watershed near Santa Barbara, California, USA. Their results show that both models were able to capture the significant temporal seasonal variability in streamflow adequately but differ significantly in terms of estimates of soil moisture patterns and sensitivity of those patterns to the scale of landscape tessellation used to derive spatially distributed elements. Tague et al. [2004] argued that these differing soil moisture patterns produced by the two models suggest that a better process based understanding of the relationship between topography and soil moisture distribution is needed and that clear distinction between patterns predicted by the two models (and different landscape tessellations) could be used to design an efficient field measurement campaign that would provide critical information for constraining model space.

VanShaar et al. [2002] selected four catchments within the USA portion of the Columbia River Basin (ranging from 27 to 1033 km²) to simulate the hydrological effects of changes in land cover using the Distributed Hydrology Soil Vegetation Model (DHSVM) [Wigmosta et al., 1994; Wigmosta and Lettenmaier, 1999] and the Variable Infiltration Capacity model (VIC) [Liang et al., 1994]. VanShaar et al. [2002] showed that lower leaf area, i.e. decreased vegetation extent, has led to increased snow accumulation, increased streamflow and reduced evapotranspiration. They also mentioned that streamflow changes are greatest during spring snowmelt runoff and evaporation changes are greatest when soils are moister (i.e., spring and early summer). VanShaar et al.'s [2002] comparison of results between the topographically explicit DHSVM and the macroscale VIC models revealed that the trend in snow water equivalent, stream flow and evapotranspiration changes is similar
for both models. They discussed how DHSVM is more sensitive than VIC to predict runoff when land-cover changes. They attributed that difference of runoff prediction between the two models to longer periods of soil moisture stress in VIC than in DHSVM and to differences in the parameters used in the evapotranspiration formulations. They further suggested that more explicit representation of saturation excess in DHSVM, differences in the calculation of net radiation and VIC’s use of architectural resistance, i.e., the aerodynamic resistance between the leaves and the canopy top, used to account for an imperfectly ventilated canopy in the evapotranspiration calculations [Ducoudré et al., 1993], have led to the higher DHSVM sensitivity of runoff to leaf area index (LAI) changes.

Matheussen et al. [2000] evaluated the magnitude and extent of long-term changes in streamflow attributable to long term vegetation change in the Columbia River Basin. They presented streamflow sensitivity analysis results in which climate effects have been held constant so as to focus on the effect of vegetation change. They used the Variable Infiltration Capacity (VIC) model of Liang et al. [1994; 1996] and Lohmann et al. [1998a; b] to simulate streamflow scenarios that would have occurred under two vegetation cover conditions. The effect of vegetation cover on snow accumulation and melt is represented internally within the VIC model via a coupled snow interception and two-layer energy balance approach. The VIC snow model calculates energy exchange at the snow-air interface. Also the VIC snow model allows for snow to be intercepted by the canopy, to fall through it, or to completely cover low-lying vegetation and for bare areas to fall to the ground surface. Matheussen et al. [2000] noted that the Columbia River Basin forest cover change has a general tendency toward decreased vegetation maturity and/or species conversion with the effect of generally reducing leaf area indices which is similar to VanShaar
et al. [2002] findings. Matheussen et al.’s findings were from simulations performed for a 10-year period (1979 to 1989) using both the 1990 and the historic 1900 vegetation parameters. Regarding the hydrological effects, Matheussen et al. [2000] indicated that changes in forest characteristics have affected the basin water balance by: i) reduced leaf area index tends to increase maximum wintertime snow accumulations and hence snow is available for runoff during the spring melt season. ii) reduced leaf area indices result in lower annual evapotranspiration.

Hamlet et al. [2007] evaluated long-term trends of evapotranspiration, runoff and soil moisture over the western United States for the period 1916–2003 using VIC [Liang et al., 1994]. Hamlet et al.’s [2007] results show that trends in evapotranspiration in spring and summer are determined primarily by trends in precipitation and snowmelt that determine water availability. They further added that trends in the seasonal timing of evapotranspiration are modest, but during the period 1947–2003 when temperature trends are large, they reflect a shift of evapotranspiration from midsummer to early summer and late spring. Regarding trends in the annual runoff ratio, the authors mentioned that the trends are determined primarily by trends in cool season precipitation, rather than changes in the timing of runoff or evapotranspiration. Hamlet et al. [2007] found that the signature of temperature-related trends in runoff and soil moisture is strongly keyed to mean midwinter [December–February (DJF)] temperatures and that areas with warmer winter temperatures show increasing trends in the runoff fraction as early as February and colder areas as late as June. Hamlet et al. [2007] further added that increasing trends in soil moisture on 1 April are evident over much of the western U.S.
Different studies assessed climate change impacts on vegetation response within mountain environments [Ryan, 1991; Law et al., 2000; Royce and Barbour, 2001; Boisvenue and Running, 2006; Soulé and Knapp, 2006; Atkin et al., 2008]. Model simulations were performed and observations collected in these studies with the intent to a better understanding of the coupling between vegetation and hydrology.

Ryan [1991] argued that most changes postulated for global warming (increased temperature, increased CO$_2$, altered precipitation, increased pollutants) would cause direct or indirect changes in the functional components of respiration. He also added that increasing temperature would likely increase respiration, but the magnitude of the increase cannot be predicted. Ryan [1991] concluded that, with increased temperature and CO$_2$, forest ecosystems may grow faster, mature earlier and die younger. He further added that forests, particularly coniferous with their low species diversity and simple stand structure, may have poor ability to acclimate to temperature changes.

Boisvenue and Running [2006] presented a review of the impacts of climate change trends on forest productivity since the middle of the 20$^{th}$ century. Boisvenue and Running suggested that the climatic changes in the last 55 years seem to have a generally positive impact on forest productivity on sites where water is not strongly limiting. They argued that many interacting factors preclude the identification of one factor causing these changes as each site has specific and possibly unique combinations of factors, that correspond to reported changes in temperature, precipitation and radiation. They suggested that the potential limits to vegetation net primary production in the western United States are mainly water and temperature. Boisvenue and Running reported the major contributors to their inability to build reliable evidence or to agree on the evidence they had on the impacts of
climate change on forests were due to: 1) a lack of reliable data for below ground net primary production (NPP), 2) an incomplete understanding of mechanistic processes in forests and between forest and the atmosphere.

Law et al. [2000] conducted an ecosystem carbon and water vapor exchange study in an old growth ponderosa pine forest in the Pacific Northwest region of the United States (study site was located on the east side of the Cascade Mountains near Sisters, Oregon). Law’s objective was to generate an understanding of the controls on ecosystem processes across seasonal and annual cycles from a combination of fine-scale process modeling ecophysiological measurements and carbon and water vapor fluxes measured by the eddy covariance method. Law et al. [2000] found that soil water deficits and high atmospheric demand for water appeared to have a large impact on canopy as an increased frequency of stomatal closure.

Soulé and Knapp [2006] explored whether gradually increasing levels of atmospheric CO$_2$, as opposed to step increases, have a positive impact on radial growth rates of ponderosa pine in natural environments. They were also interested in determining the spatial extent and variability of the pine growth enhancement associated with higher atmospheric CO$_2$. Soulé and Knapp [2006] suggested that ponderosa pine would respond to gradual increases in atmospheric CO$_2$ over the post 50 years and that response would be most apparent during drought stress and on environmentally harsh sites. They also found that site stress (harshness) is closely and positively related to enhanced radial growth rates of pine. They concluded that within the interior Pacific Northwest there is likely a water-use efficiency enhancement in ponderosa pine associated with higher atmospheric CO$_2$. 
In summary, examining the effect of land use changes on a watershed’s hydrological response has been studied by applying multiple physically based and spatially distributed ecosystem models. Findings from climate change impacts studies on vegetation response within mountain environments could be summarized as: There is a balance between (1) interacting physiological responses that include increased respiration rates with high temperatures [Ryan, 1991; Atkin et al., 2008], (2) potentially higher rates of gross photosynthesis in temperature limited areas [Boisvenue and Running, 2006], (3) increased frequency of stomatal closure under increased soil-water deficits and high vapor pressure deficits [Law et al., 2000; Royce and Barbour, 2001] and (4) increased water-use efficiency associated with higher atmospheric CO₂ [Soulé and Knapp, 2006].

3.3. Methods

3.3.1. Study Site

The Weber River near Oakley watershed upstream of USGS gauge # 10128500 has been selected to be the study watershed for this paper. The location of this streamflow gauge within Summit County, UT is 40° 44’ 14” North and 111° 14’ 50” West referenced from the North American datum of 1927 within the Uinta mountains. The watershed drainage area at this gauge is about 422 km² (Figure 3.1). The Weber River near Oakley watershed, located within the Upper Weber basin (HUC 16020101), is important to the state of Utah since it is the headwater of the Weber River. The Weber River is an important water resource to the Wasatch front metropolitan area. Vegetation within the study watershed is primarily coniferous forest. The streamflow record at this gauge is available since October 1904. Streamflow data was retrieved from the USGS portal.
Mean annual runoff, $Q$, measured at the outlet of the study watershed is about 443 mm.

3.3.2. Spatial Data

A digital elevation model (DEM) with 30 meters grid resolution for the study watershed was obtained from the National Elevation Dataset, NED, (http://seamless.usgs.gov/website/seamless/viewer.htm) and was used to derive the slope and aspect grids for the model inputs. The elevation of the study watershed ranges from 2000 to 3640 meters (Figure 3.2) with mean elevation of 2758 meters.

The watershed soil's texture map was obtained from the Soil Survey Geographic (SSURGO) dataset through the Geospatial Data Gateway (http://datagateway.nrcs.usda.gov/). The watershed soil texture is mainly loam with small areas of clay loam. The mean soil depth is about 2.8 meters.

Vegetation and land use information was obtained from the National Land Cover Dataset (NLCD) (http://gisdata.usgs.net/website/MRLC/viewer.php). We grouped vegetation and land use into eight categories: coniferous forest, deciduous forest, shrub, mixed forest, grass, no vegetation, agriculture and urban. The dominant types of vegetation within this watershed are coniferous forests (~49%), shrub (~21%) and deciduous forests (~17%) (Figure 3.2). The watershed is mostly undeveloped lands, i.e. forest, with few agricultural lands that are close to small urban areas within valleys close to the watershed outlet.

3.3.3. Climate Data

Long term climate data in the form of daily precipitation ($P$), minimum and maximum air temperature ($T$) and wind speed ($w$) were obtained from the Surface Water
Modeling group at the University of Washington (http://www.hydro.washington.edu/Lettenmaier/Data/gridded/index_hamlet.html). The development of this gridded dataset was described by Hamlet and Lettenmaier [2005]. This dataset includes daily $1/8^{th}$ degree resolution gridded meteorological data for 1 Jan 1915–31 Dec 2003. We extracted the data for our study watershed from the Great Basin region group in this dataset (Figure 3.3). There is no wind speed data prior to 1949 so long term averages are used. July is the hottest month during the year at this watershed with maximum daily air temperature that varies between 17 and 30 °C.

Mean annual precipitation on the study watershed is about 830 mm. We summarize the precipitation, runoff and evapotranspiration ($\bar{E}$) annual information for the study watershed during 1921-2003 water years in Table 3.1. The annual average actual evapotranspiration was calculated using mass balance ($\bar{E} = \bar{P} - \bar{Q}$), while potential evapotranspiration ($\bar{E}_p$) was obtained from Vörösmarty et al. [1998].

Budyko [1974] uses ($\bar{E}_p / \bar{P}$) as a dryness index to classify the hydro-climate. He suggests that when $\bar{P}$ is large relative to $\bar{E}_p$ for a watershed then water is in abundant supply and evapotranspiration from this watershed is limited only by energy. While, when precipitation is short relative to $\bar{E}_p$ ($\bar{E}_p / \bar{P}$ is large) the evapotranspiration from this watershed is limited only by water availability. Budyko [1974] developed an empirical function $\bar{E}/\bar{P} = \phi(\bar{E}_p / \bar{P})$ that partitions $\bar{P}$ into $\bar{E}$ and $\bar{Q}$. For our study watershed $\bar{E}_p / \bar{P} = 0.7$ and $\bar{Q}/\bar{P} = 0.53$. These indicate that this is an energy limited watershed, something unusual for Utah that is perceived to be generally semi-arid. This is due to the
high elevation of this watershed that receives considerable snow and where $\bar{E}_\rho$ is somewhat reduced due to reduced air temperature with elevation.

3.3.4. Model Description

The Regional Hydro-Ecologic Simulation System (RHESSys) model described by Band et al. [1993; 1996] and Tague and Band [2001; 2004] is a GIS-based, hydro-ecological modeling framework designed to simulate carbon, water and nutrient fluxes. As a hydrologic model, RHESSys is intermediate in terms of its complexity as compared to more complex process-based hydrologic models such as MIKE-SHE [Refsgaard and Storm, 1995]. RHESSys combines both a set of physically based process models and a methodology for partitioning and parameterizing the landscape over spatially variable terrain ranging from ten meters to hundreds of kilometers. The version of RHESSys used for this work (5.14.4) includes both surface and subsurface storage routing and a deep groundwater store [Tague et al., 2008]. The RHESSys model is able to simulate interactions between carbon, water and nutrient fluxes and climate patterns within a mountainous environment. Water is explicitly routed between spatial patches, representing spatial heterogeneity in soil moisture and lateral water flux to the stream. The RHESSys hydrologic process models have been adapted from several pre-existing models and they include: snow accumulation and melt, interception, infiltration, transpiration, soil and litter interception, evaporation and shallow and deep groundwater subsurface lateral flow. RHESSys uses a hierarchical spatial framework that allows different processes to be modeled at their most representative scale. Specific algorithms within these original models have been modified to reflect various developments in the associated literature or to fit within the RHESSys modeling framework. Most processes
are run at a daily time step. RHESSys uses the Penman Monteith [Monteith, 1965] method for evaporation and sublimation of intercepted water, transpiration and soil and litter evaporation processes. RHESSys uses the Jarvis model for stomatal conductance calculations based on air temperature, vapor pressure deficit, wind speed and other environmental factors (such as light and CO₂) [Jarvis, 1976]. The full details of all process modules in RHESSys are documented by Tague and Band [2004]. The original process models in RHESSys include the following:

a) The MTN-Clim model [Running et al., 1987] which uses topography and user supplied base station information to extrapolate input climate variables such as radiation over topographically varying terrain.

b) An ecophysiological model adapted from BIOME-BGC [Running and Coughlan, 1988; Running and Hunt, 1993] that estimates carbon, water and potentially nitrogen fluxes from different canopy cover types.

c) Distributed hydrologic models – The original RHESSys utilized a single approach, TOPMODEL, to model soil moisture redistribution and runoff production. At present, RHESSys includes two approaches:

i. TOPMODEL [Beven and Kirkby, 1979] a quasi-distributed model that distributes hillslope soil moisture based on a topographically defined wetness index.

ii. An explicit routing model adapted from DHSVM [Wigmosta et al., 1994] which models saturated subsurface interflow and overland flow via explicit connectivity. An important modification from the grid-based routing in DHSVM is the ability to route water between arbitrarily shaped surface
elements. This allows greater flexibility in defining surface patches and varying shape and density of surface tessellation.

For this work, we used the explicit routing approach to route the water horizontally. RHESSys partitions the landscape into distributed elements hierarchically organized into basin (watershed), zone, hillslope, patch and stratum. In this work, zones representing climate information have been partitioned following the 1/8th degree climate grid from Hamlet and Lettenmaier [2005]. There are eight different zones spanning our study watershed (Figure 3.4). Hillslopes were generated using the watershed analysis routine (r.watershed) in GRASS [GRASS Development Core Team, 2010] with contributing area threshold of 0.16 km² resulting in 2,318 hillslopes (Figure 3.4). We obtained the stream network contributing area threshold objectively from a stream drop test following theory described in Tarboton et al. [1991; 1992]. Each hillslope was treated as a single model element (i.e., patch). Stratum is used for canopy information and inherits the patch spatial setting (i.e., hillslope in this work).

3.3.5. Model Parameters and Calibration

It is well known that properties of natural earth materials are highly variable in space. This brings to our attention that one of the problems in distributed hydrological models is that they attempt to provide a deterministic description of flow processes. Theoretically, these problems, i.e. heterogeneity problems, have been argued that they can be addressed by allowing parameter values to accommodate the physical characteristics of the flow processes when different parameters vary from grid element to grid element according to measurements [Abbott et al., 1986; Bathurst, 1986]. Moreover, meteorological variables tend to have a large temporal and spatial variation. These parameter requirements
imply in practice impossible field measurements that are required to fulfill all grid elements in a deterministic model application over a certain scale. Hence, considerable limitations to distributed hydrological models arise when we interpret their results. It is important to draw the reader attention to Beven [2001] discussion on rainfall runoff models parameter estimation and predictive uncertainty. Beven summarizes the key points on rainfall runoff models parameter estimation and predictive uncertainty as: 1) it is most unlikely that there will be one right answer, 2) calibrated parameter values may only be valid inside the particular model structure used, 3) the model results will be much more sensitive to changes in the values of some parameters than to changes in others and 4) different performance measures will usually give different results in terms of both the “optimum” values of parameters and the relative sensitivity of different parameters. Beven [2001] has also summarized different methods of model calibration that are available. He classified these model calibration methods into three classes. The first model calibration methods class assumes an optimum parameter set and ignores the estimation of predictive uncertainty around that optimum parameter set [Sorooshian and Gupta, 1995]. The second model calibration methods class assumes an optimum parameter set and estimate predictive uncertainty around that optimum parameter set [Melching, 1995], while the third model calibration methods class rejects the idea that there is an optimum parameter set in favor of the idea of equifinality of models [Gupta et al., 1998; Yapo et al., 1998]. Equifinality is a concept that there may be many models of a catchment that are acceptably consistent with the observation available.

Mathematical representations of the key controls on ecosystem processes are embedded in RHESSys in the form of models. RHESSys uses many parameters to describe
typical soil, vegetation and land use characteristics. Literature based estimates have been used to compile parameters for common vegetation and soil types. A substantial effort has been made to reduce the number of calibrated parameters within RHESSys to four hydrologic parameters which are: 1) the decay of hydraulic conductivity with depth ($m$), 2) saturated soil hydraulic conductivity at the surface ($k$), 3) the fraction of recharge that bypasses the shallow subsurface flow system to deeper groundwater storage ($gw1$) and 4) the drainage rate of deeper groundwater store ($gw2$). In this work, the model was calibrated to daily streamflows at the watershed outlet during the 1994 water year. The 1994 water year is relatively a dry year with annual precipitation of 638 mm. Monte Carlo simulation was used to generate 5000 sample parameter sets from independent uniform distributions over the feasible parameter ranges determined from the literature ranges for $m$, $K$, $gw1$ and $gw2$ parameters. Each was used as input to the model and model performance was assessed using the Nash-Sutcliffe metric on daily flows, Nash-Sutcliffe metric on log daily flows and total annual flow error (Figure 3.5). From these simulations we selected a group of behavioral parameter sets. We note that model results are more sensitive to changes in values of $gw1$ and $gw2$ parameters than to changes in the other parameters (Figure 3.5).

We used 1921-2003 water years’ runoff record for model verification.

3.4. Experiment

The experiment we conducted for this paper comprised two independent parts with the goal of examining the sensitivity of runoff to land cover changes. The first part of the experiment was selection of a hillslope within our study watersheds that is dominated with coniferous cover (area 0.222 km$^2$, mean elevation 2621 m and mean slope 28°) and conducted multiple simulations with different vegetation covers (see hillslope labeled (1) in
Multiple vegetation covers that include conversion to grass, deciduous cover, as well as no vegetation cover scenarios were evaluated. Comparison between runoff generated under each vegetation cover was examined. Analysis of relative evapotranspiration coefficients, runoff ratio $Q/P$ sensitivity to vegetation change and runoff prediction models has been evaluated. By relative evapotranspiration coefficient we mean the ratio of actual evapotranspiration to potential evapotranspiration. Actual evapotranspiration is the rate of evapotranspiration from a surface or vegetation canopy to the atmosphere under the prevailing meteorological conditions and water availability. Potential evapotranspiration is the rate of evapotranspiration from a surface or vegetation canopy with no limitation due to water availability.

The second part of this work experiment was similar to the first part but differs in the selection of a hillslope. The hillslope selected in the second part is dominated with deciduous vegetation cover (area 0.314 km$^2$, mean elevation 2600 m and mean slope 25°) (see hillslope labeled (2) in Figure 3. 6). This means that the last vegetation change scenario examined for runoff sensitivity to vegetation change implies the change from deciduous to coniferous cover. This two-part experiment should be viewed separately with no intent of characterizing coniferous and deciduous trees differences since the hillslopes chosen appear to have different shape and physical characteristics.

3.5. Results

We generated results from each parameter set in the behavioral group and found that patterns and trends of runoff obtained from the different sets were all essentially the same. Hence, in what follows we have chosen one parameter set to present the results and illustrate the sensitivities that we are interested in. This parameter set is given Table 3. 2 and
results with this parameter set are able to capture about 82% of the variability seen in observed daily runoff during the calibration year. Annual simulated runoff had a 5.97% error when compared with annual runoff observed. Figure 3.7 gives daily simulated versus observed runoff (mm) for the study watershed during calibration year. In Figure 3.7, we also give cumulative simulated runoff, evapotranspiration and storage (offset to be 0 at the start) as well as observed precipitation and runoff during the calibrating year. Nash-Sutcliffe efficiency for daily and log transformed daily runoff were 0.82 and 0.73, respectively. The model did a good job in capturing the variability seen in daily runoff during spring, but with less degree during summer time. This is shown with the Nash-Sutcliffe efficiency of log transformed daily runoff value of 0.73. We feel that this level of accuracy is acceptable to pursue this modeling exercise given the objectives we had of examining the sensitivity of runoff to land cover change.

Figure 3.8 gives daily simulated runoff for the study watershed in verification of RHESSys model during 1921-2003 years. In general, the model did well at capturing timing of onset and end of seasonal runoff, but was slightly off in some estimates of peak flows. The model was able to capture on average about 70% of the variability seen in daily runoff, 75% of the variability in daily log transformed flows and had about 2.83% error in estimating total annual flows during 83 years. These simulation efficiencies have to be considered in examining the results of this work. The uncertainty and limitations seen are due to the nature of modeling that could be related to error in inputs, parameters and process representation. Calibration parameters with performance metric for both calibration and verification periods were summarized in Table 3.2.
Since we were interested in examining the sensitivity of runoff to land cover changes, examination of water use from different vegetation covers is an important component. We analyzed 83 years of RHESSys simulations from the two selected hillslopes that had different vegetation covers that include runoff, storages and evapotranspiration estimates at each hillslope. We found that the relative evapotranspiration coefficient, $\bar{r}_{c}$; $\bar{r}_{c} = \overline{E/\text{PET}}$, in deciduous and coniferous trees at this watershed is quite similar (Table 3.3). Our results suggest that the mean annual relative potential evapotranspiration coefficient for deciduous trees is about 0.922 and for coniferous trees is about 0.964. This suggests that coniferous and deciduous vegetation at our study watershed behaves similarly in terms of having evapotranspiration rates limited to available energy only with no limitation due to water availability. This also suggests that the model predictions align with our mass balance observations stated earlier (i.e., energy limited watershed). The RHESSys model predicts that mean annual actual evapotranspiration from coniferous trees is about 351 mm, while the mean annual actual evapotranspiration from deciduous trees is about 354 mm. Precipitation on both selected hillslopes is the same. The RHESSys calculations of PET in Table 3.3 are less than PET from Vörösmarty et al. [1998]. These differences may be due to differences in leaf conductance and canopy resistance, or climate data resolution and do not affect our findings because it is the relative differences between vegetation that we are interested in more than absolute values. In Table 3.3, we give a summary of model evapotranspiration prediction information from the two study hillslopes with deciduous and coniferous vegetation covers as well as estimates of the relative evapotranspiration coefficients. Annual minimum and maximum, mean and standard deviation information for deciduous and coniferous vegetation covers is presented (Table 3.3).
The specific question of this paper is how vegetation changes in a mountain environment impact runoff. Figure 3.9 and Figure 3.10 give sensitivity of runoff to vegetation results from the two part experiment explained earlier. Mean monthly averages for 83 years were evaluated to produce both figures. We give seasonal average leaf area index (panel a), seasonal average snow water equivalent (panel b) and seasonal average runoff (panel c) sensitivities to vegetation change on the coniferous hillslope in Figure 3.9 (experiment part 1). This is repeated for the deciduous hillslope in Figure 3.10 (experiment part 2).

RHESSys modeled leaf area index as expected in that no vegetation cover had zero leaf area index while deciduous cover had higher leaf area index during the late spring-summer seasons. RHESSys predicts increased snow water equivalent in areas of decreased leaf area index as a result of no canopy and litter interception for non-vegetated areas. Panel b which gives snow water equivalent predictions indicates higher snow amounts in non-vegetated and grass covers and lower snow amounts for coniferous and deciduous covers. As a result of these differences in snow accumulated during winter and evapotranspiration during late spring and summer seasons, RHESSys predicts noticeable changes in runoff for the different vegetation changes scenarios (Figure 3.9). Let’s look at the case where we converted coniferous cover to no vegetation cover (green line in Figure 3.9). We realize that this vegetation cover change scenario has indicated a higher runoff with respect to other vegetation cover change scenarios tested. Our interpretation is that changes in vegetation cover have affected the water balance in two ways. First, reduced leaf area index results in lower annual evapotranspiration. This tends to increase runoff relative to evapotranspiration. Second, during winter reduced LAI decreases canopy interception...
that leads to increased wintertime snow accumulations, and hence snow available for runoff during the early spring melt season. Increased LAI during spring melt season tends to delay the snow melting process due to reduced radiation under high LAI surfaces relative to low LAI surfaces.

In Figure 3.10, we see that higher runoff amounts were obtained when we changed the deciduous cover to anything else since deciduous cover had the highest leaf area index than other land cover change scenario tested (going from high LAI to low LAI). Figure 3.10 or experiment part 2 helped us to reach the runoff sensitivities to vegetation changes at deciduous hillslope covers. The information obtained from both Figure 3.9 and Figure 3.10 was aggregated to annual average which is shown in Figure 3.11.

In Figure 3.11, we give runoff information from the two selected hillslopes that had deciduous and coniferous vegetation covers with existing and removing vegetation conditions. The x-axis gives the existing vegetation yield ($Q$), while y-axis gives yield after removing vegetation cover ($Q'$). Model simulations suggest that annual runoff after removing vegetation can be approximated as:

$$Q'_{\text{deciduous}} = 1.067Q_{\text{deciduous}}$$
$$Q'_{\text{coniferous}} = 1.021Q_{\text{coniferous}}$$

Both these relations have $R^2$ value of about 0.98 (Figure 3.11). We feel that these runoff prediction models are useful in examining changes in runoff generated from different vegetation types. Table 3.4 gives summary of runoff ratio information produced at the two selected hillslopes under the three conditions examined (existing vegetation cover, conversion to grass cover and conversion to no vegetation cover). The runoff increases in both deciduous and coniferous covers (6.7% in deciduous and 2.1% in coniferous) associated
with vegetation removal can be seen in runoff ratio changes (Table 3.4). Annual minimum, mean, median, maximum, standard deviation and coefficient of variation runoff ratio information is also presented (Table 3.4).

3.6. Conclusions and Discussion

This paper examined the sensitivity of runoff to land cover changes. RHESSys simulations from two hillslopes with different vegetation covers were evaluated. Our results suggest that during winter reduced LAI decreases canopy interception that results in increased wintertime snow accumulations, and hence snow available for runoff during the early spring melt season. Increased LAI during spring melt season tends to delay the snow melting process due to reduced radiation under high LAI surfaces relative to low LAI surfaces. The model results suggest that annual runoff yield after removing deciduous vegetation is on average about 7% higher than with deciduous vegetation cover, while annual runoff yield after removing coniferous vegetation is on average as about 2% higher than that produced with coniferous vegetation cover. The contribution of this work primarily lies on the examination of water use sensitivity to plants functions in a mountain environment using numerical simulations. These simulations thus help quantify the sensitivity of water yield to vegetation change.

In this work, our goal was to answer the specific question of how does vegetation change in mountain environment impact runoff. The approach was to conduct a numerical modeling experiment using the RHESSys model to examine the sensitivity of runoff to vegetation change. Our use of two adjacent hillslopes was not intended to characterize coniferous and deciduous trees differences since the hillslopes chosen appear to have different shape and physical characteristics. Rather, we used these two hillslopes as if we are
doing two experiments, experiment number 1 which was done for coniferous cover hillslope to examine the runoff sensitivity to changes in vegetation covers holding everything else constant while experiment number 2 which was done for deciduous cover hillslope to examine the runoff sensitivity to changes in vegetation covers and again holding everything else constant. We are aware that there is a concern regarding our approach of using the model that we have calibrated at the watershed scale to examine runoff sensitivity at a hillslope scale, but the lack of observed data at hillslope level in our study watershed was the only reason for not calibrating our model at hillslope scale. We caution the reader that the results presented in this work should be interpreted as best estimates based on the knowledge incorporated in RHESSys.

Results suggest that the mean annual relative potential evapotranspiration coefficient for deciduous cover is about 0.922 and for coniferous cover is about 0.964. This suggests that coniferous and deciduous vegetation at our study watershed behaves similarly in terms of having evapotranspiration rates limited to available energy only with no limitation due to water availability. This watershed behavior is also supported by the fact that $\frac{\bar{E}_p}{\bar{P}} < 1$ and $\frac{\bar{Q}}{\bar{P}}$ is so large, that suggests this is an energy limited watershed from concepts discussed by Budyko [1974]. The RHESSys model uses the standard Penman-Monteith [Monteith, 1965] methods to estimate the potential evapotranspiration rate, we looked at other methods and sources to find potential evapotranspiration rate estimates [Deichmann and Eklundh, 1991; Allen et al., 1998; Dingman, 2002]. We found that generally RHESSys estimates of potential evapotranspiration rates were aligned with other work that uses the Penman-Monteith method. Table 3.3 summarizes evapotranspiration information
from the two study hillslopes with deciduous and coniferous vegetation covers as well as estimates of relative evapotranspiration coefficients.

Our results match observations that suggest in regions where watersheds are dominated by snowmelt, peak flows increase as a result of increased snow accumulations in clearings, as compared with forested areas and more rapid snowmelt owing to enhanced turbulent energy transfer in harvested areas [Bosch and Hewlett, 1982; Stednick, 1996]. Our results need to be interpreted with the physical characteristics associated with our particular hillslopes examined (area in the range of 0.3 km$^2$, mean annual precipitation about 830 mm, elevation about 2600 meters and 40.73° in latitude) and with the approximation we have since we are using a model that has not been calibrated at a hillslope level.

It is important to mention that we have examined the sensitivity of our runoff results discussed earlier to uncertainty in calibrated parameters. A group of behavioral parameter sets was selected in terms of the Nash-Sutcliffe metric on daily observed flows, Nash-Sutcliffe metric on log daily observed flows and total annual flow error. Variability over the range of calibrated parameter values from this behavioral group was used to quantify sensitivity of runoff to parameter uncertainty. The different model solutions examined gave slightly different values of runoff estimates with the different vegetation types examined but the patterns and trends suggested by these different model solutions were essentially the same.

The results of this work cannot substitute for direct field measurements, since models are often uncertain. We agree with Christensen et al. [2008] in that model results such as the ones presented in this work should be thought as tools used to efficiently guide field measurements. These models should be considered as best estimate of reality given the knowledge we have about a rich area of research in hydrology, which is hydrological
processes. We feel that understanding changes in water yield requires a better knowledge of
plant area index, canopy conductance, interception and evapotranspiration. In order to
obtain that knowledge in these hydrological processes, extensive field programs intended to
examine their sensitivities to climate and land cover changes is needed.

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Table 3.1. Weber River near Oakley, UT (USGS # 10128500) annual water balance estimates (1921-2003). Runoff (Q) from the USGS national water information system (http://waterdata.usgs.gov/nwis), precipitation (P) from Hamlet and Lettenmaier [2005], mean annual evapotranspiration (E) estimated from both mass balance and potential evapotranspiration obtained from Vörösmarty et al. [1998].

<table>
<thead>
<tr>
<th>P [mm]</th>
<th>Q [mm]</th>
<th>E [mm]</th>
</tr>
</thead>
<tbody>
<tr>
<td>Min</td>
<td>Mean</td>
<td>Median</td>
</tr>
<tr>
<td>438</td>
<td>829</td>
<td>821</td>
</tr>
<tr>
<td>Min</td>
<td>Mean</td>
<td>Median</td>
</tr>
<tr>
<td>164</td>
<td>443</td>
<td>447</td>
</tr>
</tbody>
</table>
Table 3.2. Selected RHESSys calibration parameters used to drive model simulations. Nash-Sutcliffe (NS) performance metric for daily simulated and observed runoff (also used with log flows, NSlog) and percent error (Qerr) between daily simulated and observed runoff for calibration (1993-1994) and verification (1921-2003) time periods were used to select the model solution.

<table>
<thead>
<tr>
<th>Period</th>
<th>m (meter)</th>
<th>k (m/day)</th>
<th>gw1 (%)</th>
<th>gw2 (%)</th>
<th>NS</th>
<th>NSlog</th>
<th>Qerr (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Calibration</td>
<td>0.82</td>
<td>0.73</td>
<td>5.97</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Verification</td>
<td>0.70</td>
<td>0.75</td>
<td>2.83</td>
<td>9.75</td>
<td>0.40</td>
<td>23.10</td>
<td>30.40</td>
</tr>
</tbody>
</table>

Table 3. Annual evapotranspiration information for study hillslopes that have deciduous and coniferous vegetation covers. Evapotranspiration and potential evapotranspiration results from RHESSys model simulations.

<table>
<thead>
<tr>
<th>Vegetation</th>
<th>$\mu_E$</th>
<th>$\overline{PET}$</th>
<th>$\min (r_{lc})$</th>
<th>$\bar{r}_{lc}$</th>
<th>$\max (r_{lc})$</th>
<th>$\sigma(r_{lc})$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Deciduous</td>
<td>354</td>
<td>383</td>
<td>0.844</td>
<td>0.922</td>
<td>0.962</td>
<td>0.030</td>
</tr>
<tr>
<td>Coniferous</td>
<td>351</td>
<td>364</td>
<td>0.881</td>
<td>0.964</td>
<td>0.997</td>
<td>0.028</td>
</tr>
</tbody>
</table>

Key variables explanations:
- $\mu_E$: evapotranspiration arithmetic mean in millimeters.
- $\overline{PET}$: potential evapotranspiration arithmetic mean in millimeters.
- $\bar{r}_{lc}$: mean relative potential evapotranspiration coefficient, where $r_{lc} = E/PET$.
- $\min (r_{lc})$: minimum relative potential evapotranspiration coefficient.
- $\max (r_{lc})$: maximum relative potential evapotranspiration coefficient.
- $\sigma(r_{lc})$: relative potential evapotranspiration coefficient unbiased standard deviation.
Table 3.4. Runoff ratio (Q/P) annual sensitivity analysis for study hillslopes to vegetation change. Runoff results from RHESSys simulations. 3 vegetation change conditions that include existing vegetation, conversion to grass and conversion to no vegetation are shown. sd is standard deviation and CV is coefficient of variation (CV = sd/mean). Mean annual precipitation on the study hillslopes is 732 mm.

<table>
<thead>
<tr>
<th></th>
<th>Existing Condition</th>
<th>Conversion to grass</th>
<th>Conversion to no vegetation</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>min</td>
<td>mean</td>
<td>median</td>
</tr>
<tr>
<td>Deciduous</td>
<td>0.189</td>
<td>0.243</td>
<td>0.240</td>
</tr>
<tr>
<td>Coniferous</td>
<td>0.197</td>
<td>0.254</td>
<td>0.253</td>
</tr>
</tbody>
</table>
Figure 3.1. Weber River near Oakley watershed (USGS # 10128500) is located in north eastern Utah within the Uinta Mountains (left lower). The watershed drainage area is about 422 km$^2$. 
Figure 3.2. Spatial input data. a) Digital Elevation Model (DEM) (30 meter grid size), b) Land Cover. Land Cover from the NLCD 1992 dataset.
Figure 3.3. Monthly time series input data over the study watershed and measured at the watershed outlet during 1915-2003. Mean monthly values of maximum, minimum air temperature and wind speed. Precipitation and runoff amounts were summed monthly. Climate data from Hamlet and Lettenmaier [2005] and runoff measured at USGS gauge # 10128500.
Figure 3.4. Hillslopes for the Weber River near Oakley watershed used as modeling units (2,318 patches) in RHESSys framework. 8 zones, displayed as rectangles with dotted lines, were used for climate processing from 8 grids [Hamlet and Lettenmaier, 2005].
Figure 3.5. 5000 Monte Carlo simulation sample parameter sets over the literature parameter ranges for RHESSys calibration parameters ($m$, $K$, $gw1$ and $gw2$). Nash-Sutcliffe efficiency metric on daily flows ($NSE$), Nash-Sutcliffe efficiency metric on log daily flows ($NSE$ log-transformed) and annual flow error ($Qerr$) metrics were used to pick a reasonable set of RHESSys calibration parameters. 1994 water year observed runoff record was used for calibration.
Figure 3.6. Two hillslopes, coniferous labeled with (1) and deciduous labeled with (2), used for examining the sensitivity of runoff to land cover change. The background image is ESRI base map with resolution of 15 cm. Dotted lines are elevation contours with 100 m intervals. The two hillslopes location within the study watershed (USGS#10128500) is shown in the lower left (insert).
Figure 3.7. Daily simulated versus observed runoff (mm) for the Weber River near Oakley watershed in calibration of RHESSys during October 1993 to September 1994. Modeled runoff had a 5.97% error and Nash-Sutcliffe efficiency for daily and log transformed daily runoff of 0.82 and 0.73 respectively. Right panel gives cumulative runoff (observed & simulated), evapotranspiration and storage simulated with observed precipitation during 1994 year.
Figure 3.8. Daily simulated runoff (mm) for the Weber River near Oakley watershed in verification of RHESSys during 1921-2003. Modeled runoff had a 2.83% error and Nash-Sutcliffe efficiency for daily and log transformed daily runoff of 0.70 and 0.75, respectively.
Figure 3.9. Leaf area index (LAI), snow water equivalent and runoff seasonal sensitivities to vegetation change at coniferous hillslope. a) monthly average leaf area index, b) monthly average snow water equivalent and c) monthly average runoff. 83 years simulation result. Vegetation change scenarios include: conversion from coniferous to grass (red), no vegetation (green) and deciduous (blue).
Figure 3. Leaf area index (LAI), snow water equivalent and runoff seasonal sensitivities to vegetation change at deciduous hillslope. 

(a) monthly average leaf area index, 
(b) monthly average snow water equivalent and 
(c) monthly average runoff. 83 years simulation result. 
Vegetation change scenarios include: conversion from deciduous to grass (red), no vegetation (green) and coniferous (black).
Figure 3. Annual runoffs with vegetation cover being converted from deciduous/coniferous cover to no vegetation cover.
CHAPTER 4
AN EXAMINATION OF THE SENSITIVITY OF THE GREAT SALT LAKE
TO CHANGES IN INPUTS

Abstract

The Great Salt Lake is a closed basin lake whose level and volume fluctuates due to differences between inflows and outflows. The only outflow is evaporation, which depends directly on lake area and salinity that both depend on lake volume. As a result, lake volume adjusts to, on average, balance precipitation and streamflow inflows by evaporation. In this paper, we examine the sensitivity of lake volume changes to precipitation, streamflow and evaporation and the interactions among these involving lake area, volume and salinity. A mass balance model is developed to generate representative realizations of future lake level from climate and streamflow inputs simulated using the $k$-nearest neighbor method. Climate and salinity are used to estimate evaporation from the lake using a Penman model adjusted for the salinity dependent saturation vapor pressure. Our results show that fluctuation in streamflow is the dominant factor in lake level fluctuations, but that fluctuations in lake area which modulates evaporation and precipitation directly on the lake are also important. The results quantify the sensitivity of lake level to changes in streamflow and air temperature inputs, predicting, for example, that a 25% decrease in streamflow would reduce lake level by about 66 cm (2.2 ft), while a $+4 \, ^\circ C$ air temperature increase would reduce lake level by about 34 cm (1.1 ft) on average. This sensitivity is important in evaluating the impacts of anthropogenically induced streamflow input changes, due, for example, to increased consumptive water use, on the level of the lake.
4.1. Introduction

The Great Salt Lake (GSL) is a remnant of freshwater Lake Bonneville, which existed from about 10 to 30 thousand years ago [Fenneman, 1931]. Management of the GSL system requires a comprehensive understanding of its behavior and the physical processes leading to its level changes. The underlying physical processes that collectively produce changes in the GSL lake level operate at different spatial and temporal scales. Moreover, the relations between these physical processes are often nonlinear. These complexities limit our ability to forecast the GSL level and make it important to understand the interactions between lake level, volume, area, evaporation and salinity in response to driving inputs of precipitation, streamflow and climate.

This paper examines the sensitivity of GSL level to changes in streamflow input or changes in climate that manifest as changes in air temperatures over the lake. We quantify this sensitivity by examining an elasticity measure defined as the ratio of the variability of streamflow, precipitation, evaporation, area and salinity to the variability in historic volume changes. We also developed a mass balance model to simulate lake level and volume driven by stochastic precipitation, streamflow and climate inputs. The model tracks lake volume, level, area and salinity and uses air temperature, wind speed, humidity and salinity to estimate evaporation from the lake using Penman's equation adjusted for the salinity dependent saturation vapor pressure. The model uses k-nearest neighbor resampling [Lall and Sharma, 1996] to reproduce the stochastic dependence of total annual inputs to the lake, drawing from input data compiled for 61 years (1950-2010).

This model was used to forecast the distribution of future lake levels over a time scale of 30 years, assuming that inputs remain statistically the same as the 1950-2010 period.
for which we have data. The model was then used to quantify the sensitivity of the
distribution of future lake levels to changes in streamflow or air temperature inputs.
Streamflow input to the lake may change due to water resources development and
consumptive use in the drainage basin, while air temperature may change due to changing
climate and this model helps quantify the impact of these changes on the lake. We found
that the time scale associated with the lake adjusting to input fluctuations is on the order of
5 years. Historic streamflow fluctuations at this 5-year time scale are about ±25% of the
annual streamflow input to the lake. The model predicts that for a 25% increase in
streamflow the GSL level would increase by about 55 cm (1.8 ft) on average, while for a 25%
decrease in streamflow input to the lake, level would reduce by about 66 cm (2.2 ft) on
average. The North American Regional Climate Change Assessment Program (NARCCAP)
indicates potential warming in the range of 3 °C to 4 °C for this region [Mearns et al., 2007;
2009]. We adjusted monthly minimum, maximum and dew point temperature by +4 °C in
model inputs to evaluate the sensitivity to, and potential impact of, this warming and found
that GSL level would decrease by about 34 cm (1.1 ft) on average. These sensitivities are
important in evaluating the impacts of water resources and climate induced streamflow and
temperature input changes on the GSL.

In this paper we first give background on the GSL and review literature on forecasting
the GSL volumes and levels introducing the concepts of elasticity and nearest neighbor
resampling that we use. We then describe how input data was assembled and analyzed. We
then describe the model we have developed, the GSL mass balance model (GSLMBM) and
present results from the analysis of historic data and model simulations.
4.2. Background

The GSL (latitude 40.7° to 41.7° N, longitude 111.9° to 113.1° W) is located in the northeast of the Great Basin and it is the fourth largest, perennial, closed basin lake in the world (Figure 4.1). The lake is shallow (average depth 4-6 m), with a large and variable surface area (3000 - 6000 km²) and its salinity ranges from 5% to 28%. Covering portions of northern Utah, southern Idaho and western Wyoming, the GSL basin has an area of about 55,000 km². The GSL level has fluctuated between the lowest recorded level of 1278.5 meters (4194.4 ft) in 1963 (15 October 1963) and highest level of 1283.7 (4211.6 ft) recorded in 1872, 1873 and in 1986 (3 June 1986) over the historic record available from 1847 (Figure 4.2). The annual 25th, 50th and 75th percentiles for the Great Salt Lake level during the 1847-2010 period are 1279.3 meters (4197.1 ft), 1280.4 meters (4200.9 ft) and 1281.2 meters (4203.4 ft) respectively. Figure 4.2 shows that GSL level fluctuations occur over time scales of 5 to 20 years and longer. Superimposed on these longer cycles is an annual scale with fluctuation of about 0.5 m (1 to 2 ft) rising during winter and spring runoff (November to June) then dropping during summer (July to October) when evaporation is high and inflows are low. The Great Salt Lake volume fluctuates with its level, resulting in concentration or dilution of the salt in the lake, affecting surface salinity. Water residence time in the lake is about 5 years as is the time scale implied by the historical range of volume changes (active volume) in comparison to mean inflows [Mohammed and Tarboton, 2011].

Fluctuations of the GSL’s level are of direct concern to industries and infrastructure along the shore, such as the Salt Lake City Airport, the Union Pacific Railroad, wastewater treatment plants and Interstate highway 80 [Lall et al., 1996]. They are also well correlated with regional water supply conditions. During 1983-1986 the Great Salt Lake rose rapidly to
its highest level in a hundred years and then declined quickly. A pumping project (the West Desert Project) that cost about $60 million was initiated due to that event and removed more than 3.08 km$^3$ (2.5 million acre-feet) of water and 695 million tons of salt from the lake from April 1987 to June 1989. From January 1990 to June 1992, 0.25 km$^3$ (200,000 acre-feet) of this pumped water and 94 million tons of salt was returned to the lake [Wold and Waddell, 1994; Loving et al., 2000]. The following two decades (1990-2010) have seen concern that the GSL might be drying up. A lower GSL level leads to higher salinity in the lake. Less water reduces shoreline perimeter and results in islands becoming connected to the mainland. Such outcomes have potentially serious ramifications for lake ecology and human health. Ecologically, a reduced shoreline means reduced bird habitat, increasing crowding and the risk of diseases. Human health effects include the contribution of exposed lake bed sediments to respirable dust concentrations. This issue is compounded by urban development creeping closer to lake shorelines [Bedford, 2009].

There has been a lot of research on forecasting GSL volumes and levels from a statistical and dynamical systems perspective [Lall and Mann, 1995; Mann et al., 1995; Abarbanel and Lall, 1996; Abarbanel et al., 1996; Lall et al., 1996; Sangoyomi et al., 1996; Asefa et al., 2005; Khalil et al., 2006; Lall et al., 2006; Moon et al., 2008; Wang et al., 2010]. In order to investigate the relationship between structured low-frequency climate variability, low-order dynamical behavior of the GSL and the enhanced long term predictability of the GSL volume Abarbanel et al. [1996] examined decadal and interannual signals in the GSL volume time series, western U.S. precipitation, northern hemisphere sea level pressure and air temperature using multivariate spectral analysis. They identified signals that represent 2 year, 3-5 year, 10-12 year and 15-20 year intermittent oscillations with slowly varying
amplitude and phase characteristics. They argued that the GSL volume responds with a small phase lag to regional precipitation and temperature anomalies, which are in turn forced by large-scale atmospheric circulation anomalies. Abarbanel et al. [1996] used the Global False Nearest Neighbors method to choose the embedding dimension $d$ appropriate for describing the GSL volume in multivariate state space. They found that an embedding dimension of 4 is sufficient to describe the GSL volume time series, suggesting that there are four degrees of freedom active in the Great Salt Lake volume record. They also suggested that physically based models of climate that are guided by low frequency spatial and temporal features observed in data and that reproduce the dynamical attributes of the corresponding time-series may be more useful for analysis of climate changes issues than high resolution models that lack such guidance. In this regard, long time-series such as that of the GSL volume that represents a spatially averaged hydro-climate may provide a useful baseline.

Moon [1995], Moon and Lall [1996] and Moon et al. [2008] studied the relationships between the time variability of the volume of GSL and selected atmospheric circulation indices. Moon [1995] developed and applied nonlinear measures of dependence between selected atmospheric circulation indices and the GSL volume at various lags (presuming that these indices are considered to lead the GSL volume). In addition, he forecast the volume of the GSL using selected atmospheric circulation indices. The indices considered in his study were the Southern Oscillation Index (SOI), The Pacific/North America (PNA) climatic pattern and the Central North Pacific (CNP) climatic pattern. Moon et al. [2008] applied a local weighted polynomials methodology discussed by Lall et al. [2006] to predict the volume of the GSL using both the above mentioned atmospheric indices and previous lake volumes. Moon et al. [2008] developed a nonlinear GSL forecasting model with locally weighted
polynomials and automatically and locally chosen parameters. The purpose of Moon et al. 's [2008] work was to examine improvements in GSL volume predictions by including information from atmospheric indices using the local polynomial regression approach of Lall et al. [2006]. Moon et al. [2008] suggested that predictions of GSL volumes using the local polynomial regression approach can lead to significant improvements in the predictability of lake volume.

Lall et al. [2006] presented an application of a multivariate, nonparametric regression approach to forecast the biweekly GSL volume time series (a short term forecast of 1 year). Lall et al. [2006] argued that their local polynomial regression scheme could be an alternative to multivariate adaptive regression splines (MARS) [Friedman, 1991]. Lall et al.'s [2006] approach that only considers a fraction of the full sample size (part of the time series) to estimate the parameters of the local regression model provides an improvement in algorithms that develop predictions from multivariate data structures.

Asefa et al. [2005] studied the GSL volume biweekly time series and applied the Support Vector Machines (SVM) method [Vapnik, 1995] to forecast future values. Asefa et al. [2005] forecast the GSL volume time series by splitting it into wet and dry periods. The selected dry period in their study was the five years 1924-1929, while the wet period was during the major rise of the GSL, 1983-1987. Asefa et al. [2005] presented SVM as an appropriate tool able to forecast the GSL dry period (4 month prediction) as well as the wet period (2 weeks prediction). Khalil et al. [2006] applied both the SVM and the Relevance Vector Machines (RVM) methods to forecast the GSL biweekly volume time series from 1848 to 2004. Khalil et al. [2006] argued that both methods were able to capture the variability of the GSL volume time series. Khalil et al. [2006] also assessed the uncertainty of the GSL
volume forecasts attained using the two methods due to model structure and model parameters. These studies [2005; 2006] focused primarily on the shorter time scales (weeks and months) and they do not capture the longer time scale variability that dominates the fluctuations in the level of the GSL.

Wang et al. [2010] presented a methodology to predict the precipitation variation in the Great Basin and the GSL level for subsequent years using the Pacific Quasi-Decadal Oscillation (QDO) index. They also discussed the physical link between the Pacific QDO index and the hydrological processes in the Great Basin region. Wang et al. [2010] found that the GSL level lags precipitation in the Great Basin by about 3 years and the GSL level takes an average of 6 years to respond to the Pacific QDO index. These associations with longer term patterns present an opportunity for longer term GSL level forecasts that would be important.

Schaake [1990] adapted the concept of elasticity used in economics to define the elasticity of runoff to change in precipitation:

$$\varepsilon_P(P, Q) = \frac{dQ/Q}{dP/P} = \frac{dQ}{dP} \frac{P}{Q}$$

(1)

where $P =$ Precipitation falling on a watershed and $Q =$ runoff leaving a watershed. This quantifies the sensitivity of runoff to changes in precipitation and is also referred to as a sensitivity factor by Dooge [1992] and Dooge et al. [1999] and magnification factor by Kuhnel et al. [1991]. A number of other studies have used and extended this concept [Yates and Strzepek, 1998; Sankarasubramanian et al., 2001; Arnell, 2002; Chiew, 2006; Fu et al., 2007b; 2007c]. Fu et al. [2007a] introduced a two parameter climate elasticity of streamflow index to assess climate change effects on annual streamflow. Fu et al.'s [2007a] elasticity index
quantifies the dependence of streamflow on precipitation also conditional on temperature.

The *Fu et al.* [2007a] two-parameter elasticity index is defined as:

\[
\varepsilon_{p,\delta T} = \left( \frac{Q_{p,\delta T} - \bar{Q}}{\bar{P} - \bar{P}} \right)
\]

where \(\delta T = (T - \bar{T})\) is the temperature departure. *Fu et al.* [2007a] applied the two parameter climate elasticity of streamflow to the Spokane River Basin (located in eastern Washington and northern Idaho, USA) where they found that the elasticity of streamflow index varies from 2.4 to 0.2, for a precipitation increase of 20%, as temperature varies from 1 °C lower to 1.8 °C higher than the long term annual mean. *Fu et al.*'s [2007a] results have the flexibility to be applied to other basins to inform planning of long term basin water management strategies taking into account global change scenarios.

The autocorrelation function (ACF) measures the amount of linear dependence between observations in a time series that are separated by lag \(k\) [Hipel et al., 1977]. The ACF helps in identifying the nature of short and long term persistence in time series analysis. The complexity of water resources management problems often requires stochastic models that can reproduce the statistical dependence such as quantified by the ACF of the historic data. One such model is the \(k\)-nearest neighbor (\(k\)-nn) model [Lall and Sharma, 1996]. \(K\)-nn is a nonparametric method that makes few assumptions about the underlying distribution of the data and is useful for the Monte Carlo simulation of hydrologic time series for water resources analysis, design and operation problems.

In summary, this prior research has examined several methods for forecasting GSL levels or volumes based in statistical or dynamical systems approaches, or measures of dependence between selected atmospheric circulation indices and the GSL volume. While
this dynamical systems approach has provided improved predictive capability, the empirical
nonlinear relationships that underlie the methods do not decompose the GSL system into its
component processes. Therefore, the approach is not intended or suited to evaluating the
sensitivity of how regional management practices (watershed changes such as consumptive
use, withdrawals, urbanization, deforestation, etc.) in the GSL basin would alter the lake level
nor is it able to address how lake level is impacted by climate change that leads to changes in
evaporation from the lake. The approach also does not provide capability to directly assess
impacts of lake management, such as mineral pond withdrawals and bathymetry alterations.

4.3. Analysis of Historic Data

The bathymetry of the GSL has been compiled by the USGS in tables that report the
volume and area of the lake for a range of levels. The GSL bathymetry data sources that we
are aware of are: 1) north and south arms: volume-area tables for levels 1271.3 to 1285 m
(4171.0 to 4216 ft) [Loving et al., 2000], 2) south arm: volume-area tables for levels 1270 to
1280 m (4167.0 to 4200 ft) but excluding Farmington and Bear River Bays [Baskin, 2005] and
3) north arm: volume-area tables for levels 1270 to 1280 m (4167.0 to 4200 ft) [Baskin,
2006]. In this work we used the Loving et al. [2000] bathymetry tables because they provide
estimates of volume and area for levels greater than 1280 m and cover the entire lake. Lake
levels together with the bathymetry were used to determine lake area and volume.

Three major rivers, the Bear, Weber and Jordan Rivers, flow into the GSL. The Bear
River has been gauged since 1902, the Weber River since 1907 and the Jordan since 1949.
The specific stations where streamflow data are available have changed over the years. A
detailed study of streamflow inputs to the GSL from the Bear River, the Weber River, the
Jordan River, as well as other minor streams was conducted by Loving et al. [2000], who
estimated the GSL streamflow inputs for 12 years from 1987-1998. Loving et al. [2000] present regression equations for estimating streamflow at locations where streamflow data is missing. We followed these regressions as closely as possible to compile the monthly streamflow inputs for the period 1 October 1949 to 30 September 2010 (Figure 4.3). However, we did need to extend these methods to be able to estimate streamflow into the GSL for the extended period. Where streamflow data was missing it was estimated using regression with a nearby station, preferably upstream in the same basin, or where correlation was best. Streamflow gaging stations used to estimate monthly surface water inflow to the Great Salt Lake are given in Table 4.1. The regression estimates of monthly surface inflows to the GSL (1950-2010) are presented in Table 4.2. Groundwater inflow to the GSL was estimated to be 0.093 km$^3$ (75,000 acre ft/yr) [Waddell and Barton, 1980; Loving et al., 2000]. The annual average streamflow input to the GSL is about 3 km$^3$ with about 58% of that coming from the Bear River (Table 4.3).

Direct precipitation on the lake and minimum, maximum and dew point temperatures over the lake were obtained from the Oregon State University, PRISM Climate Group (http://www.prism.oregonstate.edu/) [Daly et al., 2008] for the period October 1949 to September 2010. This data is reported monthly on a 2.5 arc min (~4 km) grid. Grid cells falling within the GSL were identified and data from these grid cells were averaged to produce time-series of monthly precipitation (Figure 4.4) and air temperature (Figure 4.5). Precipitation, streamflow, groundwater and west desert pumping withdrawals and return flow were summarized for the GSL system during the 1950-2010 time period (Figure 4.6).

Gridded meteorological data from the University of Washington for the Great Basin, as described by Hamlet and Lettenmaier [2005] was retrieved from
This is daily data at a 7.5 min grid (~15 km). Grid cells falling within each arm of the GSL were identified and data from these grid cells was averaged to produce time series of monthly wind speed for the period 1949-2010. Monthly wind speed for the period from January 2004 to September 2010 that is not available from the University of Washington was estimated using long term averages from the available years of University of Washington wind speed data.

Salinity observations made by the Utah Geological Survey (UGS) have included brine density, temperature, ion concentrations and total dissolved solids (TDS) recorded at multiple depths within both arms of the GSL. We obtained data for the period 1966-2007 from Erica Gaddis (unpublished data, 2011). These data were used to estimate the salinity in both arms as well as the GSL total salt load. The locations of the 17 salinity sites, where data were obtained, are shown in Figure 4.1. Brine density measurements were recorded at multiple depths within both arms of the lake so as to quantify the density and salinity stratification that occurs in the GSL [Loving et al., 2000]. The density/salinity measuring program has been such that data are collected sporadically at the sites indicated in Figure 4.1, seldom on the same day at multiple sites and ranging from 1 to 12 times per year. The GSL brine concentrations are related to densities of brine [2000] through:

$$\rho_{b20} = \rho_{w20} + 0.63C$$

where $$\rho_{w20}$$ is the density of fresh water [g/cm³] at 20 °C, $$\rho_{b20}$$ the density of brine [g/cm³] at 20 °C and C the brine concentration [g/cm³]. Waddell and Bolke [1973] relate the GSL brine density to fresh water density through temperature adjustment:
\[ \rho_b = \rho_{b20} \cdot \rho_w / \rho_{w20} \]  

where \( \rho_w = \left( 8T - T^2 + 132416 \right) / 132432 \) [g/cm\(^3\)] is the density of freshwater at temperature \( T \) (°C), \( \rho_b \equiv \) is the density of brine at any temperature.

On every day that density/salinity measurements were made at any site, we used that set of salinity records to estimate the load, \( L \), of the arm of the lake in which the site resides. To calculate salt load based on the range of density measurements over depth on any date, the lake was divided into layers (Figure 4.7) and each layer load was evaluated. Bathymetry and lake level were used to estimate the volume for each layer. Layer salt load estimate was obtained by the product of layer’s salinity and volume. Salt loads in each layer were summed to obtain load in the corresponding arm of the lake. Observed average concentration was then estimated using:

\[ C = \frac{L}{V} \]  

where \( L \) is the GSL salt load (in kg or tons) and \( V \) the volume of the arm of the lake.

To obtain a continuous salt load estimate in each arm of the lake, we smoothed the sporadic load estimates from all sites in that arm using loess [R Development Core Team, 2010], with span parameter 0.1. The smoothed salt loads from each arm were then summed to estimate total GSL salt load. Figure 4.8 shows that the total salt load estimated for each arm from observations based on brine density, bathymetry and level. The observations available for salinity from 17 sites scattered in both lake arms (see Figure 4.1) were used to calculate the salt load summing over layers. We only calculated load on days when density/salinity at 4 or more depths was available. Figure 4.8 also shows the beginning of month smoothed load in each arm and total load estimated by summing the smoothed load.
from each arm. Figure 4.8 also gives the dates of west desert pumping and west desert return flows. This figure shows that when the loads in each arm are added, the total salt load is close to constant but peaks at times when the lake is high and salinities well below saturation (1986). Following Loving et al. [2000], it can be assumed that at these high lake levels essentially all of the salt is dissolved. This suggests a total salt load in the GSL of 4.87 Billion US tons prior to the west desert pumping. Loving et al. estimated the total salt load to be 5.0 Billion US tons and Wold et al. [1997] estimated the total salt load to be 4.9 Billion US tons. These earlier studies used much of the same data as we have used so we assume the differences are due to slightly different methods (subdivision into layers) and are within the uncertainty of the data. When the total load is less than 4.87 Billion US tons the difference is presumed to be precipitated salt in the North arm where concentrations are frequently at the saturation level, which for the GSL is 355 g/L [Loving et al., 2000]. The salt load estimates in 1992 suggest a total load of 4.56 Billion US tons representing a net loss of about 0.31 Billion US tons due to the west desert pumping (Figure 4.8). This net loss is a bit less than the 0.6 or 0.5 Billion US tons loss to the west desert reported by Loving et al. [2000] but still within measurement uncertainty. In Figure 4.8 precipitated salt load was estimated as the difference between these high lake level totals (4.87 Billion US tons prior to 1987 and 4.56 Billion US tons post 1992) and the current total. There is a degree of approximation to this since we know that there has been mineral extraction that reduces total salt load in the lake, mostly post west desert pumping. Based on Figure 4.8 we take the average salt load in the south arms as 1.81e+12 kg (1.99 Billion US tons) and average north arm salt load as about 2.29e+12 kg or 2.52 Billion US tons. Figure 4.8 shows that overall a greater fraction of the load is held in the north arm and that this fraction increases when the lake is high.
In order to obtain continuous estimates of salinity beyond the period when salinity was measured for use in calculation of lake evaporation we calculated the salinity in each arm of the lake using equation (5), but bounded by the saturation level. This calculation assumed, to a first approximation, that the load in each arm was constant [2000]. Specifically we used:

\[ C = \min\left(\frac{(L_s + L_n)}{(V_s + V_n)}\right)^{355} \]  

prior to closure of the causeway. After the closure of the causeway we used:

\[ C_s = \min\left(\frac{L_s}{V_s}, 355\right); \quad C_n = \min\left(\frac{L_n}{V_n}, 355\right) \]  

where \( C \) refers to concentration, \( L \) salt load, \( V \) volume and subscripts \( N \) and \( S \) refer to the north arm and south arm of the lake. The causeway closure was taken to be on 1 January 1960. We treated the lake as a single water body prior to causeway closure. Figure 4.9 gives salinity estimates from these equations compared with observed concentrations. Note that in Figure 4.9 there are no salinity observations prior to and immediately after the causeway closure (observations started on 1966).

A water balance for the lake can be expressed as:

\[ \frac{\Delta V}{\Delta t} = P \cdot A + Q - E \cdot A - Pump + R \]  

where \( P \) = precipitation directly on the lake, \( Q \) = flow entering the lake, \( A \) = lake area, \( E \) = per unit area evaporation rate from the lake, \( Pump \) = pumping from the lake into the west desert and \( R \) = return water to the lake from west desert pumping. \( Q \) includes predominantly streamflow, but also the small component of groundwater. The variables \( Pump \) and \( R \) are 0 most of the time, but have been included here so as to be able to account for the
manipulations of the GSL level by the West Desert pumping project in 1987-1989 and return
flow to the GSL in 1990-1992. Recorded lake levels were used with bathymetry to evaluate
lake volume changes and area each month and given observations of $P$, $Q$, $Pump$ and $R$, to
infer historic evaporation volume, $E.A$ and depth, $E$, using a rearrangement of Equation [8].
Evaporation calculated this way is referred to as mass balance evaporation.

Evaporation from the GSL is sensitive to salinity which controls the saturation vapor
pressure above the lake's surface. Salinity decreases as volume increases and vice versa.
Stumm and Morgan [1981] defined the activity coefficient, $\beta$, of water with salinity, $C$, as the
ratio of vapor pressure over salt water to vapor pressure over fresh water at the same
temperature. This activity coefficient of water ($\beta$) in a solution of known chemical
composition can be calculated using a composite reduction factor obtained by summing the
weighted reduction in saturation vapor pressure due to each of the constituent salt ions.
Mohammed [2006] suggested that the evaporation from the GSL can be estimated from the
following modification to the Penman equation which adjusts for saturation vapor pressure
above a saline surface being less than saturation vapor pressure of fresh water at the same
temperature:

\[
E_{sal} = \frac{\Delta'}{\Delta' + \gamma} \cdot \frac{R_n}{\lambda_v \rho_v} + \frac{\gamma}{\Delta' + \gamma} \cdot \frac{\Lambda}{\rho} \cdot K \cdot V_s \cdot (e_s(T_a) - e_a) \cdot (T_a, C)
\]  

(9)

where

$E_{sal}$ = evaporation from a saline surface $[m \, day^{-1}]$,

$\Delta'$ = the gradient of the saturated vapor pressure for a saline surface $[kPa \, ^\circ C^{-1}]$, 

$\lambda_v$ = vapor pressure of water at $T_a$ $[kPa]$,

$\rho_v$ = density of water at $T_a$ $[kg m^{-3}]$,

$\rho$ = density of salt water at $T_a$ $[kg m^{-3}]$,

$K$ = composite reduction factor due to salinity,

$V_s$ = specific volume of saturated vapor $[m^3 kg^{-1}]$,

$e_s(T_a)$ = saturation vapor pressure at $T_a$ $[kPa]$,

$e_a$ = actual vapor pressure $[kPa]$.
\[ \gamma = \frac{K_u}{\lambda_v \rho_w K_E} = \text{psychometric constant [kPa °C}^{-1}]; \ K_u = \text{bulk sensible heat transfer coefficient [MJ m}^{-3} \cdot \text{C}^{-1}], \]

\[ \lambda_v = 2.5 \times 2.36 \times 10^{-3} T = \text{the latent heat of vaporization [MJ kg}^{-1}]; \ T \text{ in [°C]}, \]

\[ \rho_w = \text{density of water [kg m}^{-3}], \]

\[ K_E = \text{bulk latent heat transfer coefficient [kPa}^{-1}], \]

\[ R_n = \text{net energy available at the water surface [MJ m}^{-2} \cdot \text{day}^{-1}], \]

\[ \nu_a = \text{wind speed over the surface [m day}^{-1}], \]

\[ e_s(T_a) = \text{the saturation vapor pressure of a saline water surface at air temperature [kPa]}, \]

\[ e_a = \text{vapor pressure of the air [kPa]}, \]

\[ T_a = \text{air temperature [°C]}, \]

\[ \beta = \text{water activity coefficient}, \]

and \ C = \text{brine concentration [g L}^{-1}]. \text{ Equation [9] was used with monthly climate inputs to calculate a climate driven evaporation based on lake salinity, } C, \text{ from equations (6) and (7).} \]

\( e_a \) was determined from dew point temperature and \( T_a \) was taken as the average of monthly maximum and minimum temperatures. \text{ Daily wind speed data was averaged to months. Separate north and south arm climate-driven evaporation values were computed and evaporation from the lake as a whole was calculated by area weighted averaging of these values. We also calculated the climate driven evaporation for fresh water conditions setting } C = 0. \text{ Figure 4.10 compares these climate driven evaporation estimates with evaporation calculated from mass balance. We note that evaporation calculated from monthly climate inputs is generally higher by an average of about 0.1 m/year than the} \]
evaporation calculated using mass balance. The 0.1 m/year difference in evaporation corresponds to a volume of about 0.38 km$^3$/year that is about 9.45% of the total GSL inputs. This difference may reflect unquantified inflows, recalling that the mass balance evaporation is a closure quantity that absorbs errors and omissions. This difference may also reflect bias or uncertainty in the evaporation calculations noting that the Penman equation may not fully or properly account for apportioning of available energy into heat absorbed by the lake and the averaging of nonlinearities when applied at a monthly scale. Nonlinearity of the relationship between adjusted Penman calculated evaporation and salinity and the fact that we treated the lake as a single water body with average salinity, rather than separate arms with separate salinities may also contribute to some of this difference. This difference will carry through to the results that use evaporation calculated from climate inputs and result in lower predicted lake levels.

Equation [8] states that changes in precipitation on the lake, streamflow input to the lake and evaporation from the lake produce changes in the lake volume. In other words lake volume changes are sensitive to changes in precipitation, streamflow and evaporation. Evaporation and precipitation are also modulated by lake area that varies in response to the system dynamics involving the bathymetry relationships between volume, area and level [Mohammed and Tarboton, 2011]. The question that arises then: is lake volume change equally sensitive to precipitation, streamflow and evaporation or not? Which variable among precipitation, streamflow and evaporation could influence the lake volume the most? In response to these questions, we drew upon the concept of elasticity [Schaake, 1990; Sankarasubramanian et al., 2001; Fu et al., 2007a] to quantify sensitivity of changes in lake volume to input changes. The quantity we are interested in understanding is change in
volume, rather than volume per-se, and the change in volume may be 0, positive or negative. Therefore, rather than using quantities \( dV/V, dQ/Q, dP/P \) as in equations [1] and [2], we think that quantifying the sensitivity of lake volume change is best addressed by evaluating the ratio of standard deviation of these variables (precipitation, streamflow and evaporation) to the standard deviation of the lake volume change. This can be expressed as:

\[
\phi_p = \frac{\sigma_p}{\sigma_{\Delta V}}, \phi_Q = \frac{\sigma_Q}{\sigma_{\Delta V}}, \phi_E = \frac{\sigma_E}{\sigma_{\Delta V}}
\]

where \( \phi_p \) is precipitation sensitivity, \( \phi_Q \) is streamflow sensitivity and \( \phi_E \) is evaporation sensitivity. The units for precipitation, evaporation and streamflow are volume units and the time scale is annual. In Table 4.4, we present these sensitivity estimates based on 61 years of historic data for the GSL from 1950-2010. We have also included sensitivity measures for lake area \( A \) and have evaluated separately the sensitivity associated with precipitation and evaporation expressed as volume and depth quantities. In calculating these sensitivities, the volume units need to be balanced between numerator and dominator. The column \( (\phi \text{ formula}) \) in Table 4.4 gives the scaling we used to make these ratios dimensionless. We also report arithmetic mean \( (\mu) \), unbiased standard deviation \( (\sigma) \), coefficient of variation \( (CV = \sigma/\mu) \) and correlation \( (\rho) \) of each variable with change in lake volume. The statistics \( \mu, CV, \rho \) and \( \phi \) are not reported for \( \Delta V \) because the mean of \( \Delta V \) is in theory 0, so \( CV \) is undefined and \( \rho \) and \( \phi \) are being evaluated with respect to \( \Delta V \). The evaporation variables \( (E, E_v) \) were determined from climate and salinity, Equation [9], while \( Ef \) and \( Ef \) were determined from climate assuming no salinity \( (C=0) \).

Examining GSL volume change sensitivity values to variables studied in Table 4.4, we see that streamflow fluctuations have the highest sensitivity value of 0.83 with volume change. This is consistent with the high correlation (0.86) between streamflow and volume change.
change at an annual scale. In addition, streamflow input to the lake has the highest variability \((CV = 0.54)\). This indicates that streamflow fluctuations dominate the GSL volume change sensitivity. The variable with second highest sensitivity is evaporation volume with a value of 0.55 (or 0.51 when evaluated as fresh). However the correlation of evaporation volume \((E_v)\) with changes in lake volume is weak \((\rho = -0.07)\). Interestingly, the correlation of evaporation depth \((E)\) with lake volume changes is high in a negative sense \((\rho = -0.51)\). The obvious question is then why is lake volume changes highly correlated with evaporation depth \((E)\) while it is not correlated with evaporation volume \((E_v)\), but sensitivities to lake volume change are larger to evaporation volume than depth. Our interpretation is that evaporation volume is dominated by fluctuations in lake area which occur over scales of 5 to 20 years and even longer. This area effect on modulating evaporation is an important process in the overall sensitivity of lake volume change which is dominated by the longer time scales. On the other hand year to year fluctuations in climate driven evaporation correlate well with year to year lake volume change but play a lesser role in overall sensitivity. The variable with third highest sensitivity is precipitation (as depth) over the lake with a value of 0.30. Precipitation is also highly correlated with changes in lake volume changes \((\rho =0.73)\). We think that correlation overstates the sensitivity of direct precipitation on the lake because lake precipitation is highly correlated with precipitation in the watersheds that drives streamflow draining to the lake, whereas the sensitivity index correctly quantifies the relative importance of precipitation variability as a driver of lake volume changes. The variable with fourth highest sensitivity is GSL area with a sensitivity value of 0.23. GSL area through its association with evaporation volume acts as a stabilizing factor. The GSL volume change sensitivity to salinity correction factor is small with a value of
0.09. We can thus summarize the GSL volume change sensitivity analysis by saying that it is dominated by fluctuations in streamflow input to the lake and the stabilizing effect of lake area. Climate conditions over the lake play a lesser role, with precipitation having greater sensitivity than evaporation overall. These interpretations are based on the range of historic variability. While these are helpful in understanding the behavior of the lake and the relative dominance of the interacting processes, they do not directly quantify the sensitivity to shifts in the mean of input quantities. The next section uses a mass balance model to quantify sensitivity to shifts in input quantities.

4.4. Great Salt Lake Mass Balance Model (GSLMBM)

We have developed a model, the Great Salt Lake Mass Balance Model (GSLMBM) to track lake volume, level, area and salinity based on Equation [8]. The purpose of the model is to evaluate the sensitivity of GSL level in the future to changes in streamflow or climate inputs. The model is driven by inputs of precipitation on the lake, streamflow draining to the lake which includes groundwater and evaporation from the Lake’s surface based on climate and salinity conditions. The model outputs the change in lake volume. Monthly time steps are used.

The model was validated by comparing its output to historic lake levels values from October 1949 to September 2010 when driven by historic inputs. The lake level on 1 October 1949 of 1279.1 meters (4196.7 ft) was used to initialize the GSLMBM. Historical monthly inputs of precipitation, streamflow, wind speed and air temperature from 1949 to 2010 were used to drive the model. We used the north salt load of 2.29e+12 kg and south salt load of 1.81e+12 kg to calculate the lake salinity for each arm. The comparison of observed lake level (average of both arms weighted by area) and modeled lake level over this historic
period indicates the general ability of the model to track lake levels (Figure 4.11). The model explains about 81% of the variability seen in GSL level as quantified by the Nash-Sutcliffe efficiency. Note however that modeled lake levels are consistently lower than observed. This is due to the evaporation bias seen in Figure 4.10. Future predictions of GSL levels with this model need to be interpreted recognizing the presence of this bias.

The autocorrelation function (ACF) of total annual inputs (1950-2010) to the GSL indicates significant annual correlation with a decay in correlation to around zero after about 3-years (Figure 4.12). We used the $k$-nearest neighbor method [Lall and Sharma, 1996] to generate representative realizations of future climate and streamflow inputs to drive the model and reproduce this autocorrelation. We used block sampling at the annual scale based on total annual input to the lake. This involved for each simulation year identifying $k$ years from the historic record with total previous year annual input close to the annual input of the previous simulation year, then picking one of these at random using the $k$-nn kernel. All inputs (precipitation, streamflow, temperature for all months) from the historic year selected are taken as simulation inputs for the current year. This process is repeated for each simulation year. Figure 4.12 shows boxplots of ACF from 100 realizations of total annual inputs simulated this way, in comparison to the historic ACF. Note that the range of variability of the simulations encompasses the historic ACF well indicating that this approach satisfactorily reproduces the historic ACF. This block resampling of all inputs by year retains the statistical dependence among monthly inputs (except across years) and among different inputs (precipitation, temperature, streamflow) in a natural and simple way.

The average streamflow input to the GSL over the last 5 years (2005-2010) was about 25% less than the average streamflow input to the GSL over the full record (1950-2010). This
indicates that at a time scale comparable to the time scales of fluctuation of the GSL that variability in streamflow on the order of 25% is plausible. On the basis of this we evaluated the sensitivity of the GSL to changes in streamflow inputs by generating simulations where annual streamflow input to the lake was altered ±25%.

The North American Regional Climate Change Assessment Program (NARCCAP) future estimates for air temperature over the GSL area suggest that air temperature might increase in the ranges of 3 °C to 4 °C [Mearns et al., 2007; 2009]. On the basis of this we used a climate change scenario that comprised a 4.0 °C increase in air temperature and dew point to evaluate the sensitivity of the GSL to potential future warming.

An ensemble of 100 sequences, each consisting of 30 years drawn from the 61 years of historic streamflow, precipitation and evaporation was generated using the k-nearest neighbor resampling method and was used to drive the GSLMBM for the lake as it currently is. Predictions were initialized using lake levels recorded on 1 October 2010. In these simulations the climate driven evaporation calculated from temperature, wind and salinity was used. We examined the sensitivity of future predicted lake levels by assuming that either streamflow input to the GSL or air temperature over the lake (evaporation) would change. Figure 4. 13 gives the range of GSL level predictions over the 30 year and 100 realization simulations for streamflow drawn from the historic data and plus and minus 25%. The 3 lines give median lake level across the ensemble each scenario. We also depict ranges by showing bars that give the 25th and 75th percentiles for the ±25% streamflow scenarios. Our prediction results suggest that changes in streamflow input to the lake would manifest themselves significantly on lake level in 5 years and that lake level would stabilize in about 15 years around a new median. For example after 5 years, a 25% increase in streamflow input
to the GSL from contributing watersheds would increase the lake level by about 55 cm (1.8 ft) on average, while 25% decrease in streamflow input to the lake would reduce the lake level by about 66 cm (2.2 ft) on average.

The same ensemble of 100 sequences of 30 years each used for streamflow was used to examine the sensitivity to a potential 4.0 °C warming. The results indicate the degree to which lake level is lowered due to simulated increase in evaporation. After 5 years, 4.0 °C increase in air temperature would reduce the lake level by about 34 cm (1.1 ft) on average in comparisons with lake level conditions when we have no changes in air temperature (Figure 4.14). The 2 lines give lake level median predictions for no air temperature change and 4.0 °C increase. We also depict lake level ranges by showing bars that give the 25th and 75th percentiles. We also note here that lake level prediction under different air temperatures scenarios stabilize in about 15 years. This examination of the sensitivity of GSL levels to changes in air temperature helps understand how the lake would respond to potential future climate warming.

4.5. Discussion and Conclusions

This work has used analysis of historic data and modeling to identify the sensitivities of the Great Salt Lake volume to inputs and factors internal to the dynamics of the system. We found that changes in lake volume are most sensitive to fluctuations in streamflow input with sensitivity as quantified by the ratio of variabilities, $\phi_Q = \sigma_Q / \sigma_{AV}$, of 0.83. The index quantifying sensitivity to changes in evaporation from the lake, precipitation acting on the lake, fluctuations of lake area and lake salinity conditions has values of 0.55, 0.30, 0.23 and 0.09, respectively. These variables constitute the most influential system drivers and factors
with respect to GSL volume changes. Volume change sensitivities to precipitation and evaporation have been estimated using the average lake area. Quantification of the GSL volume change sensitivity to the various variables acting in the GSL system is important in support of robust decision making that addresses multiple GSL management issues.

This work has also developed a physically based mass balance model (GSLMBM) that forecasts the GSL levels for different input scenarios. The GSLMBM includes a physical evaporation component that accounts for salinity. We drove the GSLMBM with climate and streamflow inputs generated using the \( k \)-nearest neighbor method (\( k \)-nn) [Lall and Sharma, 1996]. This enables the evaluation of sensitivity within the context of the natural range of variability and dependence structure of the inputs. The results provide a distribution of possible lake levels for any time in the future. The differences between these distributions for different input scenarios quantify the sensitivity of lake level to streamflow and climate changes.

Consistent with the greater sensitivity to streamflow inferred from the elasticity like sensitivity index we find from the dynamic simulations a greater sensitivity to streamflow changes than the 4 °C warming change. The results also show that the range of variability remains as big as or bigger than the changes in median that are predicted and that the whole distribution shifts with the median.

Several factors interact and affect evaporation from the GSL. These include lake area, salinity and weather conditions. In evaluating lake sensitivity we used adjusted Penman evaporation calculations because they are physically based so are responsive to the climate (temperature) change inputs we wanted to evaluate. This is best for evaluating differences in a relative sense, but absolute values of predictions are subject to the
difference/bias we noted between mass balance and adjusted Penman evaporation calculations, both of which have uncertainty as discussed above. These uncertainties suggest that it would be valuable to measure evaporation from the lake directly (e.g. using Eddy Covariance) to bring better resolution to the difference noted.

In summary, this paper addresses the sensitivity of GSL volume changes to watershed management, climate and salinity. This quantification of the GSL volume change sensitivity to variables and factors related to the lake is part of our effort to understand the GSL dynamics and the different impacts associated with watershed changes along with anthropogenic use of lake resources (withdrawals for mineral extraction). Bear in mind that the prediction results presented in this paper reflects only single change impact (streamflow change to the lake/evaporation from the lake) and not a combined change impacts. GSL level predictions under combinations of changes in streamflow, air temperature and precipitation scenarios would be a real challenge in forecasting the GSL levels that needs to be pursued. The result of this work provides an improved understanding in the Great Salt Lake level prediction research and in general awareness of the Great Salt Lake system implications and risks associated with lake level changes.

References


Table 4.1. Streamflow gaging stations used to estimate monthly surface water inflow to the Great Salt Lake.

<table>
<thead>
<tr>
<th>USGS #</th>
<th>Name</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Period of Record</th>
<th>Basin</th>
<th>Notes</th>
</tr>
</thead>
<tbody>
<tr>
<td>1012710</td>
<td>Bear River outflow AC</td>
<td>41°32'44&quot;N</td>
<td>112°5'43&quot; W</td>
<td>Oct 1971—Sep 1986</td>
<td>Bear River</td>
<td>Primary GSL input</td>
</tr>
<tr>
<td>1012600</td>
<td>Bear River nr Corinne</td>
<td>41°34'35&quot; N</td>
<td>112°6'00&quot; W</td>
<td>Oct 1949—present, missing Oct 1957—Sep 1963</td>
<td>Bear River</td>
<td>Used to fill USGS 1012710</td>
</tr>
<tr>
<td>1011800</td>
<td>Bear River nr Collinston</td>
<td>41°50'3&quot; N</td>
<td>112°3'16&quot; W</td>
<td>Oct 1902—Sep 1996</td>
<td>Bear River</td>
<td>Used to fill USGS 1012600</td>
</tr>
<tr>
<td>1014100</td>
<td>Weber River nr Plain City</td>
<td>41°16'42&quot; N</td>
<td>112°5'28&quot; W</td>
<td>Oct 1907—present</td>
<td>Weber River</td>
<td>Primary GSL input</td>
</tr>
<tr>
<td>1017250</td>
<td>Jordan River @ Sth North @ Salt Lake City</td>
<td>40°46'49&quot; N</td>
<td>111°56'16&quot; W</td>
<td>Oct 1974—Sep 1989, missing Oct 1986—Jan 1989, supplemented from Salt Lake County Dec 1990—Sep 2006</td>
<td>Jordan River</td>
<td>Primary GSL input</td>
</tr>
<tr>
<td>1017250</td>
<td>City Creek nr Salt Lake City</td>
<td>40°47'33&quot; N</td>
<td>111°52'35&quot; W</td>
<td>Oct 1963—Sep 1980, supplemented from Salt Lake County to present</td>
<td>Jordan River</td>
<td>Used to fill USGS 1017250</td>
</tr>
<tr>
<td>1017050</td>
<td>Jordan River @ 1700 South @ Salt Lake City</td>
<td>40°44'1&quot; N</td>
<td>111°52'21&quot; W</td>
<td>Dec 1942—present</td>
<td>Jordan River</td>
<td>Used to fill USGS 1017250</td>
</tr>
<tr>
<td>10172630</td>
<td>Goggin drain nr Magna</td>
<td>40°49'0&quot; N</td>
<td>112°6'00&quot; W</td>
<td>Oct 1963—present, missing Oct 1984—May 2003</td>
<td>Jordan River</td>
<td>Used to construct USGS 10172630</td>
</tr>
<tr>
<td>10141040</td>
<td>Hooper Slough nr Hooper</td>
<td>41°11'26&quot; N</td>
<td>112°9'7&quot; W</td>
<td>Apr 1975—Sep 1984, missing Oct 1977—Aug 1978</td>
<td>Other surface inflow</td>
<td>Primary GSL input, missing part filled by USGS 10141000</td>
</tr>
<tr>
<td>10141400</td>
<td>Howard Slough at Hooper</td>
<td>41°8'25&quot; N</td>
<td>112°7'17&quot; W</td>
<td>1-Oct 1971—30-Sep 1984</td>
<td>Other surface inflow</td>
<td>Primary GSL input, missing part filled by USGS 10141000</td>
</tr>
<tr>
<td>10172640</td>
<td>Lee Creek nr Magna</td>
<td>40°46'50&quot; N</td>
<td>112°8'19&quot; W</td>
<td>Oct 1971—Apr 2008, missing Oct 1982—May 2006</td>
<td>Other surface inflow</td>
<td>Primary GSL input, missing part filled by mean monthly flow</td>
</tr>
<tr>
<td>10172650</td>
<td>Kennecott drain nr Magna</td>
<td>40°45'37&quot; N</td>
<td>112°10'14&quot; W</td>
<td>Oct 1963—Sep 1967, Oct 1971—Sep 1984, and Jul 2006—Apr 2008</td>
<td>Other surface inflow</td>
<td>Primary GSL input, missing part filled by mean monthly flow</td>
</tr>
<tr>
<td>10143500</td>
<td>Centerville Creek abv div nr Centerville</td>
<td>40°54'39&quot; N</td>
<td>111°57'44&quot; W</td>
<td>Oct 1949—present, missing Oct 1980—Apr 1999</td>
<td>Davis County</td>
<td>Primary GSL input</td>
</tr>
<tr>
<td>10172200</td>
<td>Red Butte Creek at Fort Douglas nr SLC</td>
<td>40°46'48&quot; N</td>
<td>111°48'19&quot; W</td>
<td>Oct 1963—present</td>
<td>Davis County</td>
<td>Used to fill USGS 10143500</td>
</tr>
<tr>
<td>10141500</td>
<td>Holmes Creek nr Kaysville</td>
<td>41°3'18&quot; N</td>
<td>111°53'40&quot; W</td>
<td>May 1950—Sep 1966</td>
<td>Davis County</td>
<td>Primary GSL input</td>
</tr>
<tr>
<td>10142000</td>
<td>Farmington Creek abv div nr Farmington</td>
<td>41°0'5&quot; N</td>
<td>111°52'21&quot; W</td>
<td>Oct 1949—present, missing Nov 1971—Sep 1975 and May 1985—Sep 2008</td>
<td>Davis County</td>
<td>Primary GSL input</td>
</tr>
<tr>
<td>10142500</td>
<td>Ricks Creek ab Diversions nr Centerville</td>
<td>40°56'25&quot; N</td>
<td>111°52'00&quot; W</td>
<td>May 1950—Sep 1966</td>
<td>Davis County</td>
<td>Primary GSL input</td>
</tr>
<tr>
<td>10143000</td>
<td>Parinn Creek ab Diversions nr Centerville</td>
<td>40°55'25&quot; N</td>
<td>111°51'50&quot; W</td>
<td>Oct 1949—Sep 1968</td>
<td>Davis County</td>
<td>Primary GSL input</td>
</tr>
<tr>
<td>10144000</td>
<td>Stone Creek ab Diversion nr Bountiful</td>
<td>40°53'40&quot; N</td>
<td>111°50'40&quot; W</td>
<td>May 1950—Sep 1966</td>
<td>Davis County</td>
<td>Primary GSL input</td>
</tr>
<tr>
<td>10145000</td>
<td>Mill Creek at Mueller Park nr Bountiful</td>
<td>40°51'50&quot; N</td>
<td>111°50'10&quot; W</td>
<td>May 1950—Sep 1968</td>
<td>Davis County</td>
<td>Primary GSL input</td>
</tr>
</tbody>
</table>

‡ USGS 10172630 has been discontinued and it was located close to USGS 10172500

† Used to fill USGS 10172500 and Davis County streams (USGS 10145000, 10145000, 10145000, 10145000, 10145000, 10145000)
Table 4.2. Regression estimates of monthly surface water inflow to the Great Salt Lake (1950-2010). Regression equation used in this table: Discharge at estimated site = A(Discharge at measured site) + B and R² is coefficient of determination.

<table>
<thead>
<tr>
<th>Basin</th>
<th>Site estimated (dependent variable)</th>
<th>Site measured (independent variable)</th>
<th>Period of regression</th>
<th>A (slope)</th>
<th>B (intercept)</th>
<th>R²</th>
<th>Months for which regression is valid</th>
<th>units</th>
</tr>
</thead>
<tbody>
<tr>
<td>Bear River</td>
<td>10126000</td>
<td>10118000</td>
<td>Oct 1957—Sep 1963</td>
<td>1.039</td>
<td>750</td>
<td>0.995</td>
<td>all month</td>
<td>ac-ft/month</td>
</tr>
<tr>
<td></td>
<td>10127100</td>
<td>10126000</td>
<td>972 — 1986</td>
<td>1.007</td>
<td>14264</td>
<td>0.999</td>
<td>all month</td>
<td>ac-ft/month</td>
</tr>
<tr>
<td>Jordan River</td>
<td>10172550</td>
<td>10170000+10172500</td>
<td>Oct 1974—Sep 1976</td>
<td>1.0897</td>
<td>4695</td>
<td>0.553</td>
<td>all month</td>
<td>ac-ft/month</td>
</tr>
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<td></td>
<td>10172500</td>
<td>10143500</td>
<td>Oct 1949—Sep 1960†</td>
<td>3.178</td>
<td>52.744</td>
<td>0.665</td>
<td>all month</td>
<td>ac-ft/month</td>
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<td></td>
<td>10170800</td>
<td>10170500</td>
<td>Oct 1963—Sep 1968</td>
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<td>-2657</td>
<td>0.913</td>
<td>Oct-Nov</td>
<td>ac-ft/month</td>
</tr>
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<td></td>
<td>10170800</td>
<td>10170500</td>
<td>Oct 1963—Sep 1968</td>
<td>0.858</td>
<td>706</td>
<td>0.976</td>
<td>Dec-Feb</td>
<td>ac-ft/month</td>
</tr>
<tr>
<td>Davis County</td>
<td>10143500</td>
<td>10172200</td>
<td>Oct 1985—Apr 1999</td>
<td>0.7808</td>
<td>17.273</td>
<td>0.82</td>
<td>all month</td>
<td>ac-ft/month</td>
</tr>
<tr>
<td></td>
<td>10141500</td>
<td>10143500</td>
<td>Oct 1949—Apr 1950, Oct 1966—Sep 2010</td>
<td>1.113</td>
<td>52.446</td>
<td>0.935</td>
<td>all month</td>
<td>ac-ft/month</td>
</tr>
<tr>
<td></td>
<td>10142500</td>
<td>10143500</td>
<td>Apr 1949—Apr 1950, Oct 1965—Sep 2010</td>
<td>0.9999</td>
<td>-17.956</td>
<td>0.946</td>
<td>all month</td>
<td>ac-ft/month</td>
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<tr>
<td></td>
<td>10143000</td>
<td>10143500</td>
<td>Oct 1966—Sep 2010</td>
<td>0.805</td>
<td>-30.838</td>
<td>0.967</td>
<td>all month</td>
<td>ac-ft/month</td>
</tr>
<tr>
<td></td>
<td>10144000</td>
<td>10143500</td>
<td>Oct 1949—Apr 1950, Oct 1966—Sep 2010</td>
<td>1.63</td>
<td>-55.762</td>
<td>0.967</td>
<td>all month</td>
<td>ac-ft/month</td>
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<tr>
<td></td>
<td>10145000</td>
<td>10143500</td>
<td>Oct 1949—Apr 1950, Oct 1966—Sep 2010</td>
<td>3.322</td>
<td>-120.674</td>
<td>0.964</td>
<td>all month</td>
<td>ac-ft/month</td>
</tr>
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<td>Other surface inflow</td>
<td>10141040</td>
<td>10141000</td>
<td>1975 — 1984</td>
<td>0.0062</td>
<td>8704</td>
<td>0.944</td>
<td>all month</td>
<td>ac-ft/year</td>
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<td>10141400</td>
<td>10141000</td>
<td>1972 — 1984</td>
<td>0.0171</td>
<td>12985</td>
<td>0.823</td>
<td>all month</td>
<td>ac-ft/year</td>
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</table>

Table 4.3. Annual average Great Salt Lake streamflow input (km$^3$).

<table>
<thead>
<tr>
<th></th>
<th>Bear</th>
<th>Weber</th>
<th>Jordan</th>
<th>Davis</th>
<th>Other</th>
<th>Total</th>
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</thead>
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<tr>
<td>Value</td>
<td>1.65</td>
<td>0.41</td>
<td>0.61</td>
<td>0.04</td>
<td>0.13</td>
<td>2.84</td>
</tr>
<tr>
<td>Percent</td>
<td>58%</td>
<td>15%</td>
<td>22%</td>
<td>1%</td>
<td>5%</td>
<td>100%</td>
</tr>
</tbody>
</table>
Table 4.4. Great Salt Lake sensitivity analysis based on annual data from 1950 to 2010. $\Delta V$ is change in lake volume in $\text{km}^3$, $Q$ is annual streamflow input to the lake, $P_v$ is volumetric precipitation on the lake, $P$ is depth precipitation on the lake, $E_v$ is saline volumetric evaporation from the lake, $Ef_v$ is fresh volumetric evaporation from the lake, $E$ is depth saline evaporation from the lake, $Ef$ is depth fresh evaporation from the lake, $A$ is lake area and $SCF$ is evaporation salinity correction factor calculated as $SCF = E/Ef$. $\mu$ is arithmetic mean, $\sigma$ is standard deviation, $CV$ is coefficient of variation of each quantity, $\rho$ is correlation with change in lake volume and $\phi$ is sensitivity index scaled as indicated in the formula column ($\phi$ formula).

<table>
<thead>
<tr>
<th>Variable</th>
<th>$\mu$</th>
<th>$\sigma$</th>
<th>$CV$</th>
<th>$\rho$</th>
<th>$\phi$</th>
<th>$\phi$ formula</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\Delta V$ ($\text{km}^3$)</td>
<td>-</td>
<td>1.86</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>$Q$ ($\text{km}^3$)</td>
<td>2.84</td>
<td>1.54</td>
<td>0.54</td>
<td>0.86</td>
<td>0.83</td>
<td>$\sigma(Q)/\sigma(\Delta V)$</td>
</tr>
<tr>
<td>$P_v$ ($\text{km}^3$)</td>
<td>1.07</td>
<td>0.41</td>
<td>0.38</td>
<td>0.61</td>
<td>0.22</td>
<td>$\sigma(P_v)/\sigma(\Delta V)$</td>
</tr>
<tr>
<td>$P$ (m)</td>
<td>0.53</td>
<td>0.15</td>
<td>0.29</td>
<td>0.73</td>
<td>0.30</td>
<td>$\frac{\sigma(P)}{\sigma(\Delta V)} \times \frac{\bar{A}}{A}$</td>
</tr>
<tr>
<td>$E_v$ ($\text{km}^3$)</td>
<td>4.40</td>
<td>1.03</td>
<td>0.23</td>
<td>-0.07</td>
<td>0.55</td>
<td>$\frac{\sigma(E_v)}{\sigma(\Delta V)}$</td>
</tr>
<tr>
<td>$Ef_v$ ($\text{km}^3$)</td>
<td>4.94</td>
<td>0.95</td>
<td>0.19</td>
<td>-0.06</td>
<td>0.51</td>
<td>$\frac{\sigma(Ef_v)}{\sigma(\Delta V)}$</td>
</tr>
<tr>
<td>$E$ (m)</td>
<td>1.14</td>
<td>0.05</td>
<td>0.04</td>
<td>-0.51</td>
<td>0.10</td>
<td>$\frac{\sigma(E)}{\sigma(\Delta V)} \times \frac{\bar{A}}{A}$</td>
</tr>
<tr>
<td>$Ef$ (m)</td>
<td>1.29</td>
<td>0.04</td>
<td>0.03</td>
<td>-0.56</td>
<td>0.07</td>
<td>$\frac{\sigma(Ef)}{\sigma(\Delta V)} \times \frac{\bar{A}}{A}$</td>
</tr>
<tr>
<td>$A$ ($\text{km}^2$)</td>
<td>3656.02</td>
<td>792.00</td>
<td>0.22</td>
<td>-0.15</td>
<td>0.23</td>
<td>$\frac{\sigma(A)}{\sigma(\Delta V)} \times \frac{\bar{A}}{A}$</td>
</tr>
<tr>
<td>$SCF$</td>
<td>0.88</td>
<td>0.04</td>
<td>0.04</td>
<td>-0.08</td>
<td>0.09</td>
<td>$\frac{\sigma(SCF)}{\sigma(\Delta V)} \times \bar{A} \times \bar{Ef}$</td>
</tr>
</tbody>
</table>
Figure 4. 1. Location of the Great Salt Lake, subbasins that drain to it and data collection sites used to estimate inflow, water surface altitude and salt load. The Great Salt Lake (latitude 40.7° N–41.7° N, longitude 111.9° W–113.1° W) is located in northeast Utah (insert). Green labeled surface water stations represent Davis County inflow to the Great Salt Lake, red labeled stations represent other surface inflow and black labeled stations represent all of the primary drainages (Bear, Weber and Jordan) as well as Great Salt Lake level stations.
Figure 4.2. Historic Great Salt Lake levels. The lake was divided into north and south arms by a railroad causeway in 1959. Lake level data was retrieved from the USGS [http://water.usgs.gov/data.html](http://water.usgs.gov/data.html) accessed on 13 June 2005 then updated from [http://waterdata.usgs.gov/nwis](http://waterdata.usgs.gov/nwis), accessed on 1 April 2011 for the south arm (USGS 10010000, Great Salt Lake at Saltair Boat Harbor) and the north arm (USGS 10010100, Great Salt Lake near Saline). These incorporate USGS benchmark corrections given at [http://ut.water.usgs.gov/greatsaltlake/elevations/gslcorrection.html](http://ut.water.usgs.gov/greatsaltlake/elevations/gslcorrection.html) [Loving et al., 2000]. The earliest level data available is from 18 October 1847. USGS lake level measurements were first made in 1875 with lake level values prior to this being estimates based on observer reports.
Figure 4.3. Monthly average streamflow input to the Great Salt Lake.
Figure 4. Annual total precipitation onto the Great Salt Lake from PRISM.
Figure 4.5. Annual averages of monthly minimum and maximum air temperature over the Great Salt Lake.
Figure 4.6. Great Salt Lake inputs summary.
Figure 4.7. Diagram of a hypothetical lake divided into 4 layers used to calculate lake salt load at a specific site. For the lake at level \( h \), the interfaces between layers are
\[ h - (z_1 + z_2)/2 \], \[ h - (z_2 + z_3)/2 \] and so on for remaining layers. The volume of layer (1) is then calculated from the bathymetry volume-level relationship as
\[ V(h) - V(h - (z_1 + z_2)/2) \]. The volume of layer (2) is calculated as
\[ V(h - (z_1 + z_2)/2) - V(h - (z_2 + z_3)/2) \] and so on. The total salt load is the sum of salt loads from each layer expressed by
\[ L = C_1V_1 + C_2V_2 + C_3V_3 + \cdots \].
Figure 4.8. GSL total salt load calculated from volume and depth averaged measurements. Data from Erica Gaddis (Utah Geological Survey, UGS). North arm sites are LVG4, NML, RD2, RT3 and ECN. South arm sites are FB2, AS2, RT2, RT4, NLN, AC3, SS, AC1, AC2, IS1, IS2, RT1. These ion concentration sites are shown in Figure 4.1. Loads here are reported in US or short tons. (1 US ton = 0.9072 metric tons).
Figure 4.9. Salinity computed from $C = \frac{L}{V}$ for each arm (lines) compared to observations at stations. North arm sites are LVG4, NML, RD2, RT3 and ECN while south arm sites are FB2, AS2, RT2, RT4, NLN, AC3, SS, AC1, AC2, IS1, IS2, RT1. North arm salt load used is $2.29 \times 10^{12}$ kg (2.52 Billion US tons) while south arm salt load used is $1.81 \times 10^{12}$ kg (1.99 Billion US tons).
Figure 4.10. The Great Salt Lake’s annual evaporation. Total lake evaporation calculated from mass balance, climate based on salinity and climate based on freshwater conditions.
Figure 4.11. Observed and modeled monthly Great Salt Lake level. Modeled monthly Great Salt Lake levels were initialized on 1 October 1949.
Figure 4. 12. Autocorrelation (ACF) boxplots of the total annual inputs to the Great Salt Lake using the $k$-nn method. The line gives the ACF of the historical total annual inputs to the GSL (1950-2010).
Figure 4.13. Great Salt Lake level predictions time series under different streamflow input change scenarios (25% decrease from annual streamflow input, no streamflow input changes and 25% increase from annual streamflow input). 100 Great Salt Lake level simulations were evaluated. Bars give the 25th and 75th percentiles for lake level predictions under streamflow changes. Lines give the median (50th percentile) lake level predictions. Predictions were initialized on 1 October 2010.
Figure 4. Great Salt Lake level predictions time series under air temperatures input change scenario of 4.0 °C increase from average monthly air temperatures. 100 Great Salt Lake level simulations were evaluated. Bars give the 25th and 75th percentiles for lake level predictions under air temperatures change. Lines give the median (50th percentile) lake level predictions. Predictions were initialized on 1 October 2010.
CHAPTER 5
SUMMARY, CONCLUSIONS AND RECOMMENDATIONS

This dissertation considered the relations between climate, land cover and runoff for the purpose of understanding the linkage between land cover change with emphasis on vegetation and runoff generation in watersheds under climate change conditions. This dissertation also examined terminal lake salinity impacts on level/volume dynamics. The results of this dissertation quantified the general sensitivity of water availability at the scale of regional subbasins to changes in land use, climate and watershed management. This dissertation addressed this problem using the following three approaches, each of which are presented in a chapter: 1) aggregate empirical model with water balance estimates, 2) detailed hydrological ecosystem model simulations and 3) sensitivity analysis and forecasts for the Great Salt Lake. Chapters 2 through Chapter 4 present the main results of this work. In this chapter we summarize important conclusions from these chapters, followed by recommendations for future research.

5.1. Summary and Conclusions

The overarching goal of this work was to be able to quantify the sensitivity of hydrology to land cover, land use and climate change. The specific questions this work addressed were:

A. How watershed management and land use impacts water production from watersheds in Utah?

B. How does vegetation change in a mountain environment impact runoff?
When considering the effects associated with management decisions in watersheds contributing to the Great Salt Lake for 30 years into future, how would changes in streamflow input to the Great Salt Lake or climate conditions over the lake impact the Great Salt Lake level?

The first paper (Chapter 2) is a report submitted to the Utah Governor Public Lands Office. In this report we identified a total of 39 watersheds draining to U.S. Geological Survey (USGS) streamflow gauges, chosen either from the USGS Hydroclimatic Climatic Data Network of gauges that are minimally impacted by anthropogenic alterations, or to be representative of large areas within the chosen HUCs with long relatively continuous streamflow records. In each of these watersheds we examined trends in precipitation, temperature, snow, streamflow and runoff ratio. Runoff ratio is the fraction of precipitation that becomes streamflow. We also examined land use and land cover information for these watersheds from the national land cover dataset, southwest regional GAP analyses and the Utah division of water resources water related land use inventory.

The most consistent trend noted was in temperature which is increasing. We did not note any significant trends in precipitation. Fourteen of the 39 watersheds examined had significant decreasing trends in streamflow and runoff ratio. We were unable to find definitive causes for these streamflow and runoff ratio trends, though we do have indications that some of them are associated with human development, storage in reservoirs and land cover and land use changes.

In addition, we developed a water balance approach that quantifies sensitivity of runoff production to changes in land cover based on differences in evapotranspiration from different land cover types. This water balance approach provides predictions of how water
production from these Utah watersheds may change with land cover changes. By considering a range of water balance model parameters we provided water balance derived bounds on how streamflow could change given land cover changes.

In the second paper (Chapter 3), we used the Regional Hydro-Ecological Simulation System (RHESSys) model to examine runoff sensitivity to land cover changes in a mountain environment. The purpose of these detailed simulations was to examine physically the empirical assumptions related to relative evapotranspiration coefficients stated in Chapter 2. Two numerical experiments were evaluated where we conducted simulations with multiple vegetation cover changes that include conversion to grass, no vegetation cover and deciduous/coniferous cover scenarios. The model experiments were performed at two hillslopes within the Weber River near Oakley, Utah watershed (USGS gauge # 10128500) that have coniferous and deciduous vegetation covers. Observed runoff data at the watershed outlet were used for calibration and verification. Our results suggest that during winter reduced LAI decreases canopy interception that leads to increased wintertime snow accumulations, and hence snow available for runoff during the early spring melt season. Increased LAI during spring melt season tends to delay the snow melting process due reduced radiation under high LAI surfaces relative to low LAI surfaces. These sensitivities have been evaluated as annual runoff ratio increase on average by 7% in case of deciduous cover being converted to no vegetation cover, and by 2% in case of coniferous cover being converted to no vegetation cover. Our results suggest that coniferous and deciduous vegetation at our study watershed behave similarly in terms of having evapotranspiration rates limited to available energy with no limitation due to water availability. This is unusual in semiarid Utah, USA but it is due to the elevation and precipitation. In order to examine
our runoff results to model solution used, we tried other model solutions using upper and lower bounds on RHESSys parameters \((m, k)\) from 5000 Monte Carlo simulation sample and used that as model solutions to evaluate the runoff sensitivities to vegetation changes. Similar results to what we have showed were found. The different model solutions examined gave slight different values of runoff estimates with the different vegetation types examined than what we presented in this work but the patterns and trends suggested by these different model solutions were all the same as what we have presented. The results presented in this paper should be interpreted as best estimates that serve as hypotheses for how actual ecosystems will respond based on the knowledge and understanding that is embodied in the model. Given that a model is an idealized approximation of reality, there is a need for monitoring programs to verify model predictions.

In the third paper (Chapter 4), we quantified the GSL volume change sensitivities to variables and factors that are related to the system dynamics. In addition, we developed a mass balance model (GSLMBM) to generate representative realizations of future Great Salt Lake (GSL) level from the climate and streamflow input realizations simulated using the \(k\)-nearest neighbor method. Analysis of the GSL volume change sensitivity to system drivers and factors quantified by the ratio of variabilities revealed that streamflow fluctuations, precipitation changes on the lake, lake area changes, evaporation from the lake and lake salinity conditions respectively are the influential system variables and factors. We showed that, if there is a streamflow change to the GSL, it will manifest significantly in 5 years and stabilize in 15 years. The utility of using a nonparametric method (\(k\) nearest neighbor) for generating realizations for future lake level provides answers to the call for addressing the sensitivity of GSL level to climate and watershed management practices. This paper
contributed to the efforts of understanding the sensitivity of the GSL level prediction under different regional management practices (streamflow input changes) with the inclusion of modeled evaporation that considers lake's climate and salinity variabilities. Management of the Great Salt Lake (GSL) system requires a comprehensive understanding of its behavior and the physical processes leading to its level changes. We examined the GSL salinity fluctuations in terms of total dissolved salt load. Lake volume fluctuates with lake level, resulting in concentration or dilution of the salt in the lake, affecting surface salinity. Thus, as level and volume increase, the salt concentration decreases. Increases in salinity reduce evaporation through the effect on saturation vapor pressure at the lake surface. We showed that the total salt load in the GSL is approximately constant.

The total output of this dissertation improved the capability for estimating the water availability in the intermountain region and assisted in managing the water resources given the land use, climate and watershed management changes that are occurring. This work will help to assist natural resources and land management planners and the hydrologic science community who have to anticipate: (a) how runoff will likely change in response to human induced activities and climate change, (b) how sensitive runoff is to changes in land cover and (c) how would changes in streamflow input to the GSL or climate conditions over the lake impact the GSL level. The interaction between multiple areas of research fields that involve hydrology, natural resources, climatology and statistics seen in this dissertation foster a collaborative effort to form research techniques suggested by recent literature to address scientific overarching questions in hydrology.
5.2. Recommendations

From a broader perspective this dissertation has attempted to develop an understanding of the sensitivity of runoff to land cover, land use and climate changes. It has also looked at how changes in variables and factors acting on the Great Salt Lake system affect the level of the Great Salt Lake. Given that, I see a number of avenues that can be foreseen for carrying ahead this research.

In Chapter 2, we developed a water balance approach that quantifies sensitivity of runoff production to changes in land cover based on differences in evapotranspiration from different land cover types. The coefficients that quantify the potential evapotranspiration from each land cover type in this analysis are based on our judgment and information from the literature. These coefficients are the key factors in producing runoff sensitivities, so conducting experiments with experimental watersheds that can approach that will be needed information for natural resources managers as well as the broader scientific community. I feel that future research on heterogeneity of hydrological processes and its implications on water availability in western United States would be more than needed. This call would also help in improving the knowledge of water use in semi arid vegetation setting. Within these lines, it is good to point out that the initiative within the hydrologic science community that call for the establishment of cooperative, large-scale hydrological observatories has argued that advancing the science of hydrology will require creation of new data networks and field experiments specifically designed to recognize the spatial and temporal heterogeneity of hydrologic processes [Hart and Martinez, 2006; Montgomery et al., 2007]. These hydrological observatories will be helpful frameworks that can utilized to explore a number of related new avenues for research in watershed science, including the
use of comparative analysis, classification, optimality principles and network theory, all with
the intent of defining, understanding and predicting watershed functions [Kirchner, 2006;
McDonnell et al., 2007]

In Chapter 3, we examined runoff sensitivity to land cover changes in a mountain
environment. In mountainous regions, quantifying spatial-temporal patterns of
evapotranspiration is key research challenge. This is because of the relative rarity of weather
stations, limited monitoring resources in remote areas and access difficulties [Diaz, 2005;
Van Lier et al., 2005]. Understanding changes in water yield requires a better knowledge of
plant area index, canopy conductance, interception and evapotranspiration. In order to
obtain that knowledge in these hydrological processes, I feel extending the modeling
exercise presented in Chapter 3 with field programs intended to examine the above
mentioned hydrological processes sensitivities to land cover changes is needed. Field
programs also should be used to address an important element missing in our physical
modeling exercise which is model calibration at hillslope outlet rather than watershed outlet
used. The information we get from examining the water balance components changes in
watershed hydrology from both modeling and experimenting is often vital.

Recent efforts have been made with in the Consortium of Universities for the
Advancement of Hydrologic Science (CUAHSI) (http://www.cuahsi.org) organization to
address improving hydrologic data access. One component of CUAHSI’s activity is a
Hydrologic Information System (HIS) project, which is developing infrastructure and services
to improve access to hydrologic data. We encourage efforts such as the Little Bear River
Experimental Watershed [Horsburgh et al., 2011] in to include data to support quantifying
plants water use. The knowledge of plants water use gained through the availability of such
data across a watershed in response to varying energy, water availability, atmospheric conditions, vegetation and topography is really helpful in identification of where ecosystem function within the watershed is most sensitive to land cover and climate changes.

In Chapter 4, we investigated the sensitivity of lake volume changes to precipitation, streamflow and evaporation. We also investigated the challenge of forecasting the Great Salt Lake (GSL) levels. The approach we undertook was estimating the ratio of variables to changes in lake volume variabilities. The sensitivity analysis approach we undertook considered the individual effect of variables on explaining the changes in lake volume and ignored the combined impact of variables on explaining the changes in lake volume. Looking at estimating the sensitivities of lake volume changes to combined variables acting on the system will be attractive. Also, the physically based modeling framework that we developed will be conducive to address practical questions related to the physical and management changes in the GSL system as well as future climate change scenarios. I feel physically based modeling approach that brings together considerations of watershed response, water and salt mass balance, bathymetry, salinity, atmospheric factors affecting evaporation, withdrawals (pumping) and other lake management actions, offers a promising path forward for further understanding of this system. To conclude, I do join the call for the GSL observatory because the Great Salt Lake Basin is a unique location ideally suited, both physically and in terms of infrastructure, to address fundamental hydrologic science questions in western mountain basin systems in an open community-driven hydrologic observatory [Johnson and Tarboton, 2004].
References


Johnson, W. P., and D. G. Tarboton (2004), Great Salt Lake basin hydrologic observatory prospectus, submitted to CUAHSI for consideration as a CUAHSI hydrologic observatory.


APPENDIX
This appendix describes the procedure used to adjust snow water equivalent averages to remove bias due to record length differences at individual sites. This procedure is from Mohammed (2006) and was developed to have an average over the full length of record that is comparable to the average from all sites. It is based on the idea that the average when a site does not have data (i.e. before it was established) should be adjusted by the ratio of the average of all sites to the average with that site left out, over the period where common data is available. The procedure is illustrated in Figure A-1 for four stations ranked by their period of record. Stations S1 and S2 have periods of record from year $n1=n2$ to present, p. Station S3 has period of record from $n3>n2$ to present, p and station S4 has period of record from $n4>n3$ to present, p. Stations S1 and S2 have full records, but S3 has a shorter record and S4 the shortest, in this illustrative example. For each year the unadjusted average is simply the mean across all stations with data. Thus the unadjusted average in year $i$ is represented by:

$$U(i) = \text{Ave}(S(1:4,i)) \text{ for } i \text{ ranging from } n4:p$$

$$U(i) = \text{Ave}(S(1:3,i)) \text{ for } i \text{ ranging from } n3:(n4-1)$$

$$U(i) = \text{Ave}(S(1:2,i)) \text{ for } i \text{ ranging from } n2:(n3-1) \text{ (recalling that } n1=n2)$$

Where Ave(.) denotes averaging and $S(s,y)$ denotes the specific end of month snow water equivalent values for a station (or range or stations), $s$, and year, $y$ and the unadjusted average in year $i$ is denoted by $U(i)$.

The adjusted average for the year $i$ will be denoted by $X(i)$. For the years $n4$ to $p$, no adjustments are needed so we have

$$X(i) = U(i) \text{  } i \text{ in } n4:p$$
For the years \( n_3 \) to \( n_4 - 1 \), i.e. the years when \( S_4 \) does not have a record the adjusted average is calculated as:

\[
X(i) = \text{Ave}(S(1:3,i)) \times \text{Ave}(S(1:4,n_4:p)) / \text{Ave}(S(1:3,n_4:p)) \\
= U(i) \times \text{Ave}(X(n_4:p)) / \text{Ave}(S(1:3,n_4:p)) \quad i \text{ in } n_3:(n_4-1)
\]

Similarly, the adjusted average for the year \( i \) in the range \( n_2 \) to \( n_3 - 1 \), i.e. the years when \( S_3 \) and \( S_4 \) do not have records, is calculated as:

\[
X(i) = U(i) \times \text{Ave}(X(n_3:p)) / \text{Ave}(S(1:2,n_3:p)) \quad i \text{ in } n_2:(n_3-1).
\]
Figure A-1. Illustration of the adjustment procedure used in calculating an average time series of maximum or April’s first snow water equivalent (SWE) values in each watershed. Four stations (S1, S2, S3, and S4) with varying length of record over the interval (1990-2000) are used for illustrative purpose.
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PROFILE  

EXPERIENCE  
Research Assistant, Utah Water Research Laboratory, Logan, Utah (2004-2011)  
Numerical modeling of Physical systems

Teaching Assistant, Civil & Environmental Eng., Utah State University, Logan, Utah (2004-2011)  
Taught Physical Hydrology, GIS in Water Resources

Reviewing structural designs, quantities calculation, and construction supervision

Executive Manager Assistant, MASA Group, Khartoum, Sudan (2001-2002)  
Administrative office work and served as Board of Directors Secretary

Teaching Assistant, Civil Eng. Department, University of Khartoum, Khartoum, Sudan (2000-2003)  

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WORKSHOP
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SKILLS
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- Digital Elevation Model Analysis
- Data Analysis Techniques (NetCDF)
- Statistical Modeling (Parametric and Nonparametric)
- Time Series Analysis
- Stochastic methods in Hydrology

COMPUTING
- Programming Languages (C & C++)
- Statistical and Mathematical Tools: R, MATLAB, Maple
- Spatial Data Analysis: ArcGIS (ArcMap, ArcCatalog, ArcToolbox, ArcGlobe) with Python scripting, GRASS
- Database Management: Access
- Remote Sensing Image Analysis: ERDAS IMAGINE
- CUAHSI tools: HydroSeek, HydroExcel, HydroGET, HydroDesktop
- Hydrologic Simulation Models (HEC series models)
- Geoscience Data Analysis: Integrated Date Viewer (IDV)
- Graphic 3D Modeling tools: Photoshop

AWARDS & HONORS
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Utah Water Research Laboratory (UWRL) Fellowship, Utah State University, Logan, UT 2004—2010

Ivanhoe Fellowship, The Ivanhoe Foundation, Ojai, CA—2004

Sir Edgar Horne prize for the best academic record, University of Khartoum, Khartoum, Sudan —2000

CONFERENCE PRESENTATIONS


Mohammed, I. N. and D. G. Tarboton, 2005, Modeling the dynamics of the Great Salt Lake as an integrator of regional hydrologic and climate processes?, EOS Trans. AGU, 86(52), Fall Meeting Supplemental, Abstract H41C-0422, 5th - 9th December, San Francisco, CA.


MEMBERSHIPS

National Society of Black Engineers, NSBE — 2007
Golden Key International Honor Society, GK — 2007
National Geographic Science, NG — 2006
American Geophysical Union, AGU — 2004
Geological Society of America, GSA — 2004
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