DISSERTATION

TRENDS IN ACCUMULATION AND MELT OF SEASONAL SNOW IN AND NEAR ROCKY MOUNTAIN NATIONAL PARK, COLORADO, USA, WITH EMPHASIS ON MONTHLY VARIATIONS

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ABSTRACT

TRENDS IN ACCUMULATION AND MELT OF SEASONAL SNOW IN AND NEAR ROCKY MOUNTAIN NATIONAL PARK, COLORADO, USA, WITH EMPHASIS ON MONTHLY VARIATIONS

The seasonal snowpack in and near Rocky Mountain National Park is undergoing changes that will pose challenges for water providers, natural resource managers, and winter recreation enthusiasts. Assessing long-term trends in measures of the seasonal snowpack, and in the climatic factors that influence its accumulation and melt, helps to characterize those challenges. In particular, evaluating the patterns of variation in those trends during the snow season provides new understanding as to the causes, specific ramifications, and likely future course of the trends. In addition, placing the current 35-year trends in the longer context of longer-term observational records, and paleoclimate tree-ring reconstructions provides useful comparisons of current and past trends. Finally, projections of future trends provided by linked climate and hydrologic models offer a sense of how these trends are likely to affect the snowpack of the future.

Some factors such as the high elevation of the study area help to preserve conditions favorable to development of the seasonal snowpack, and hence to limit trends toward greater warming-induced melt and less precipitation falling as snow. Nevertheless, traditional measures such as April 1 snow water equivalent (SWE) show consistent declining trends over the 35-year period of record for automated snow monitoring stations in the study area. The trends are not uniform throughout the snow season, but vary significantly by month. As a result, November

and March have warming and drying trends that retard the beginning of the winter snow season and reduce the traditional accumulation that formerly characterized the early spring. In contrast, the core winter months of December, January, and February have cooling and wetting trends that have been enhancing SWE during the heart of the winter. Mid-April to early May is another period during which cooling and wetting trends have been enhancing SWE, although these months also show more variability. This oscillating pattern helps to explain why there has not been a pervasive shift to earlier and lower annual peak SWE in the study area.

Tree-ring reconstructions of paleo-climate suggest that trends with similar rates of SWE loss have occurred in past centuries, but generally not of the same duration as the trends observed during 1981-2015. Linked climate and hydrologic models project that the observed trends are likely to continue, and that by 2050 measures such as April 1 SWE in the study area are likely to decrease by 25 percent.

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TABLE OF CONTENTS

LIST OF FIGURES

LIST OF TABLES

LIST OF ABBREVIATIONS

CMIP5	Coupled Model Inter-comparison Project, 5^{th} Phase
NRCS	Natural Resources Conservation Service
RMNP	Rocky Mountain National Park
SFE	Snowfall Equivalent
SNOTEL	Snowpack Telemetry
SWE	Snow Water Equivalent

CHAPTER 1. INTRODUCTION

1.1 The Seasonal Snowpack

The seasonal snowpack in mountainous regions is important to humans and to natural ecosystems for many reasons. Often known as "*nature's water tower*" (Viviroli *et al.*, 2007), the snowpack provides seasonal storage for most of the annual precipitation in the mountains, releasing the stored winter precipitation during the spring and early summer when water demand for reservoir storage and irrigation is higher than in the winter (Bales *et al.*, 2006). Snowmelt runoff also produces most of the groundwater recharge that takes place in the mountains and nearby plains (Barnett *et al.*, 2008). In most river basins of the western United States, water storage in the snowpack exceeds the storage capacity of man-made reservoirs (Mote *et al.*, 2005). In Colorado, more than 60 percent of annual precipitation falls as winter and spring snow (Serreze *et al.*, 1999). Spring snowmelt runoff is said to generate 80 percent of streamflow in Colorado (Colorado Climate Center, 2016).

As in most of the west, Colorado's economy is sensitive to changes in water availability, and hence to changes in the seasonal snowpack. One estimate of the marginal value of an acre-foot (1233 cubic meters) of water under conditions of changing streamflow is $34-46 in 1985 dollars (Harding and Payton, 1990). This equates to $75-102/acre-foot (6-8 cents/cubic meter) in 2016 dollars. Rocky Mountain National Park (RMNP) has an area of 265,761 acres (107,550 hectares), so a difference of 1 mm of snow water equivalent (SWE) over the whole park would equate to a marginal value for water of about $77,000 per year.

In addition to providing water supply, the seasonal snowpack represents an important recreational resource. During the decade 2005-2015, the economic impact of the downhill ski industry in Colorado grew from $2.5 billion to $4.8 billion (Denver Post, 2015). During the same period recreational visits to Rocky Mountain National Park during March more than doubled, from about 60,000 to over 130,000 (National Park Service, 2016a). This is the season for snow-related recreation in the park, including snowshoeing, cross-country skiing, sledding, and back-country skiing. In 2011, 63 percent of winter visitors to Rocky Mountain National Park participated in recreation involving snow (Papadogiannaki *et al.*, 2011).

Like the water-supply industry, the winter recreation industry is sensitive to changes in the seasonal snowpack. Potential impacts on the ski industry include not only loss of skiable snow (Gilaberte-Burdalo *et al.*, 2014), but also increased risk of wet avalanches (Lazar and Williams, 2008). Measures of the number of days per year with sufficient snow for enjoyable winter recreation, in terms of both observed trends and future projections, provide important information for recreation planners. The difference in the economic value of Colorado's ski industry in a low snow year compared with a high snow year has been estimated at $154 million (Burakowski and Magnusson, 2012). Assuming that the difference in April 1 SWE in the snow zone in a low snow year compared with a high snow year is typically about 500 mm, then a difference of 1 mm of April 1 SWE in Colorado's snow zone would equate to a marginal value, for recreation, of about $308,000 statewide.

In addition to its value for humans, the seasonal snowpack provides important benefits for ecosystems. It is the source of most of the water used by plants and animals in the snow zone and in riparian areas downstream. In addition to nourishing vegetation, it also enhances the ability of vegetation to resist wildfire (Westerling *et al.*, 2006). The snowpack also provides critical shelter and winter habitat for a variety of plants and animals (Jones *et al.*, 2001, chapters 5 and 6). Snow shelters tree branches from cold, dessicating winds that would otherwise kill the branch (Denver Botanical Garden, 2016). Similarly, snow shelters grass and forbs that provide food for elk, which are adapted to pawing through the snow to find sustenance. Animals such as ptarmigan, pikas, snowshoe hares, ermine, voles, shrews, and snow fleas are examples of animals that are adapted to seeking food and shelter in the snow (Jones *et al.*, 2001, chapter 5).

The seasonal snowpack also plays an important feedback role affecting winter climate. Snow has a much higher albedo than bare soil, rock, or vegetation. Once the seasonal snowpack forms, most of the incoming solar radiation is reflected back to the sky, leaving less to be absorbed by the land surface than when snow is not present. This helps to cool the terrain and the snowpack itself (Flanner *et al.*, 2011).

1.2 Trends in Snowpack Measures

In light of the importance of the seasonal snowpack and of changes in its patterns of accumulation and melt, it is important to examine long-term trends in measures of the snowpack. A number of studies have found indications that the seasonal snowpack in the western United States is changing over time (mostly diminishing), and that these changes reflect underlying trends in climate.

Accumulation and melt of snow are tightly linked with temperature and precipitation. Like most of the rest of the globe, Colorado has been experiencing a warming trend over the past century, and that warming trend has been accelerating since about 1970. The current rate of warming in Colorado, in terms of average annual temperature, is about 0.37˚C/decade statewide (Lukas *et al.*, 2014). The warming trends are not uniform in space and time. Warming has been found to be more rapid in the northern central mountains of Colorado, while a few stations in the southern San Juan Mountains have shown cooling trends (Ray *et al.*, 2008; Mote *et al.*, 2005).

The warming trends have also been shown to vary with elevation, with higher elevation sites showing more rapid warming trends (Diaz and Eischeid, 2007). Variation in time has also been noted for warming trends, with November and March, in particular, showing more rapid warming trends than other months (Rense, 2016; Ray *et al.*, 2008).

In contrast to temperature, researchers often report that there are no obvious trends in cold-season precipitation (Harpold *et al.*, 2012). As with temperature trends, however, precipitation trends are variable in space and time, reflecting the varying weather patterns that bring moisture to different parts of the state at different times (Ray *et al.*, 2008).

The most common measure of the seasonal snowpack is the snow water equivalent (SWE) on April 1. This measure has been declining in most of the western United States over the second half of the last century (Mote *et al.*, 2005). The period of snow cover was found to be declining in the highest elevations of the Colorado River Basin, with the caveat that significant spatial and temporal variability highlights the need for additional research into the mass and energy balance of the seasonal snowpack in the inter-mountain west (Harpold *et al.*, 2012). There is spatial and temporal variability of snowpack dynamics across the Southern Rocky Mountains in Colorado with different snowpack patterns in different parts of the state and in some locations changing with elevation (Fassnacht and Derry, 2010). Not only are accumulation patterns different, typically increasing with elevation (Fassnacht *et al.*, 2003), but melt rates are also highly variable (Fassnacht and Records, 2015). Trends in April 1 SWE in the area of this investigation for the period 1978-2007 were found to be level, as opposed to declining trends elsewhere in Colorado (Clow, 2010). Trends in SWE in the mountains of Colorado have also been found to vary by month (Rasmussen *et al.*, 2014).

Earlier snowmelt has been found to be associated with earlier spring runoff in streams in the western United States (Stewart *et al.*, 2004; Stewart *et al.*, 2005; Fritze *et al.*, 2011) and in Colorado (Clow, 2010). In the latter study, while statewide trends toward earlier snowmelt runoff averaged about two weeks earlier in 2007 compared with 1978, the trends toward earlier snowmelt runoff for the area of this investigation were among the weaker trends noted in the state (Clow, 2010). In a more recent study using slightly different methods, trends in snowmelt runoff timing in the Colorado Rockies during 1976-2015 were found to be mostly toward earlier runoff, but about one-quarter of the trends were toward later snowmelt runoff (Pfohl, 2016).

1.3 Problem Statement

The foregoing discussion suggests a need for additional research addressing trends in snowpack dynamics in specific areas using measures that take into account the variable nature of trends with time during the snow season. To characterize this intra-season variability in SWE trends, it is important to use measures with annual time series that are based on more than one measurement per year, such as only April 1st SWE. A more complete picture of these time-varying trends can emerge if the annual time series analyses are based on more frequent measurements, such as monthly during the winter (e.g., February 1st through May 1st) or from daily measurement that can yield sub-monthly changes in SWE or specific characteristics of the snow season, such as amount and timing of peak SWE, or length of season. The purpose of this investigation is to examine such intra-season variability in trends of snowpack dynamics for an area in and around Rocky Mountain National Park in the northern mountains of Colorado. Daily SWE measurements and interpolated daily SWE estimates will be used to evaluate patterns among snowpack trends. The trends will be examined in relation to corresponding trends in temperature, elevation of the average level of freezing temperatures, and precipitation, as well as terrain variables, such as elevation and location.

The time period of the trend analysis will encompass the 80-year period of snow-course measurements in the study area (1936-2015), with emphasis on the 35-year period of daily SNOTEL measurements (1981-2015). These trends will be examined in the context of the past, by using multi-century reconstructions of paleo-SWE to determine whether the observed trends are unique over a longer term. The trends will also be examined in the context of the future, by using projections of future SWE based on climate models, to anticipate future changes in snowpack dynamics. Finally, some ramifications of the results for managing park resources will be discussed.

1.4 Study Area

The area in and near Rocky Mountain National Park (Figure 1.1) is important in relation to snowpack dynamics and trends for several reasons. The park and surrounding national forest lands comprise one of the most significant areas of protected land in the Rocky Mountain region. A unique aspect of the park is its emphasis on high-elevation mountain terrain. One-third of the park (about 36,000 hectares) is within the alpine tundra biome, above timberline which is about 3,505 m above sea level (National Park Service, 2016b). Another major portion of the park is within the subalpine biome (2,895-3,505 m), typically the zone of maximum seasonal snow accumulation. The high elevation, memorable scenery, abundant recreation opportunities, accessibility, and proximity to the Denver metropolitan area make this a popular tourist destination, with over 3,000,000 visitors per year to the park (National Park Service, 2016c), and over 6,000,000 recreational visits per year to Arapahoe-Roosevelt National Forest (U.S. Forest Service, 2016).

Snowmelt runoff from the study area contributes significant portions of the flows of the Colorado, North Platte, Cache la Poudre, Big Thompson, and Little Thompson Rivers, and St. Vrain, Left Hand, and Boulder Creeks. These creeks and rivers provide water for irrigation, industry, municipal supply, recreation, and aquatic habitat for a large part of northern Colorado and adjacent states.

The high elevations in the study area keep temperatures cool. On the tundra, temperatures vary from lows of about -37°C in winter to highs of about 24°C in summer. At lower elevations in the park, temperatures are typically about 5°C warmer (National Park Service, 2016d).

Monthly variation in precipitation is relatively small, with lower monthly totals during the summer months of June through September, and highest monthly totals in April and May. Precipitation occurs mostly as snow, which can fall during any month of the year. Most of the snowfall occurs during October through May. The snowpack typically develops in October, and peaks during March, April, or May with SWE ranging from about 100 to about 1400 mm (Dressler *et al.*, 2006; Patterson and Fassnacht, 2014). While snowmelt may continue to produce runoff in decreasing amounts throughout the summer, the bulk of snowmelt runoff is complete by mid-June. Snow tends to collect in depressions in the alpine zone that are protected from sun and wind, where it may persist from one year to the next. Snow accumulation in these depressions is augmented by wind redistribution of snow that may fall on the windward side of the ridge. These firn fields of multi-year snow that remain at the end of the summer, often referred to as glaciers, are not considered part of the seasonal snowpack.

Figure 1.1. General location of Rocky Mountain National Park in Colorado, USA (from <*http://www.myrockymountainpark.com/places/*>).

CHAPTER 2. DATA SOURCES

Snowpack measurements have been made across the western United States for the past eight decades by the U.S. Natural Resources Conservation Service (NRCS), formerly known as the Soil Conservation Service (SCS). These data were collected to facilitate current-year forecasts of snowmelt runoff, underscoring the importance of year-to-year variations in snowpack dynamics as drivers of water-management decisions. The use of consistent locations and methods of data collection make the resulting multi-year datasets valuable for examining long-term trends in snowpack dynamics.

2.1 Snow Courses

Snow courses are designated locations, operated by the NRCS, where repeated monthly manual measurements of snow depth and SWE are made; snow density is calculated from SWE and depth. Most snow courses have 5 to 20 sampling points at regularly spaced intervals along a transect. The interval between sampling points is typically in the range of 20-100 ft (6.1-30.5 m). Samples are taken using a coring device (Federal Sampler) to determine depth, and are weighed to determine SWE (Natural Resources Conservation Service, 2016a). While there is some variation in sampling schedules, at most snow courses, measurements are made four times per year, on or within 5 days prior to the first day of February, March, April, and May. Snow courses were established in the study area starting in the late 1930s. For this investigation, 23 snow courses in and near Rocky Mountain National Park were selected for analysis based on proximity to the park and on length and completeness of record (Figure 2.1, Table 2.1). Of these 23, nine lie within the park.

2.2 Snowpack Telemetry Stations

Snowpack Telemetry (SNOTEL) stations are instrumented, automated sites, also operated by the NRCS, that measure SWE and precipitation. More recently they measure temperature and snow depth, as well as other variables (e.g., soil moisture and temperature) at some stations. The frequency of measurement is hourly year-round, and meteor-burst communications technology is used to transmit the data to receiving stations and the Internet in near-real time. SWE is measured using a snow pillow and pressure transducer, precipitation is measured using a weighing storage gauge, snow depth is measured using a sonic sensor, and temperature is measured using a shielded thermistor (NRCS, 2016b). The longest SNOTEL records are for SWE and precipitation (more than 35 years at some stations); sensors for temperature (last 25 years) and snow depth (last 10 years) were added later.

The hourly data are compiled into daily statistics. Both snow-course and SNOTEL data are available on the world-wide web at <*www.wcc.nrcs.usda.gov*> (NRCS, 2016c). SNOTEL stations were initially established in the study area in 1979, with various other longer term stations in operation by 1981. For this investigation, 13 SNOTEL stations were selected for analysis based on proximity to the park and on length and completeness of record (Figure 2.1, Table 2.1). Of these 13, five lie within the park. These 13 SNOTEL stations are the same stations used by Clow (2010) in the orange and purple clusters in his analysis of snowmelt and streamflow timing. In this investigation all 13 SNOTEL stations were used for snowpack analysis, and precipitation data were used from four stations (Phantom Valley, Joe Wright, Willow Park, and Copeland Lake).

2.3 Temperature Stations

Although the SNOTEL stations collect temperature data, changes in sensor technology have complicated the evaluation of long-term temperature trends at these stations (Oyler *et al.*, 2015). As well, temperature data have been collected for shorter time periods, with many of the thermistors being installed around 1990. Consequently, temperature records for this investigation were obtained from three non-SNOTEL data collection sites in and near Rocky Mountain National Park (Figure 2.1, Table 2.2). The temperature records for the period 1983-2015 from Loch Vale represent an update of the Loch Vale temperature data previously reported for the period 1983-2007 (Clow, 2010). Although there are issues with SNOTEL temperature data, temperature records from SNOTEL stations were used for some analyses in which the association with a particular SNOTEL station was important, and for which the absolute temperature values were less important.

2.4 Elevation of Freezing Temperature

Estimated monthly mean elevation of freezing temperatures (freezing level or 0-degree isotherm level) were obtained from the North American Freezing Level Tracker (Western Regional Climate Center, 2016). The definition of instantaneous freezing level for this application is:

"*The elevation above sea level in the free atmosphere at which a temperature of 0° C (or 32° F) is first encountered. The mean daily temperature profile used for this process is formed from the four six-hour averages available from Global Reanalysis*." (WRCC, 2016).

The spatial discretization of the freezing level tracker is 2.5 degrees of latitude and longitude, or a rectangle of about 320 x 288 km. This rectangle is centered at 40.35° north latitude, 105.72°

west longitude, just south of Trail Ridge Road in Rocky Mountain National Park. The temporal discretization is monthly. As the freezing level determination is simulated for the free atmosphere, terrain effects are ignored.

2.5 Paleoclimate SWE and Temperature Reconstructions

Estimates of multi-century patterns of SWE and temperature for locations near the study site were derived from tree-ring reconstructions in the general vicinity of the study area. These reconstructions generally are calculated based on relative ring-width indices (departures from normal). The normal is assumed to be without a long-term trend. Therefore, the reconstructions are not useful for determining the presence, direction, or magnitude of multi-century trends in SWE or temperature, if indeed there were any. However, the reconstructions can be useful in in examining trends over several decades. This investigation used four tree-ring reconstructions in for the evaluation of past trends (Table 2.3).

2.6 Projections of Future Climate and SWE

Climate scientists have used various models to project future trends in climatic and hydrologic variables that affect the seasonal snowpack. This investigation examines results from several of these strategies to compare trends from the past few decades with trends that are likely to occur over the rest of the 21st century according to the models. The primary group of model projections used are the multi-model ensemble projections made by the Coupled Model Inter-comparison Project-Phase 5 (CMIP5) (World Climate Research Programme, 2016). Regionally downscaled projections from these models offer estimates of future trends in temperature and precipitation on a monthly basis, for grid cells measuring 1/8 by 1/8 degree of latitude and longitude (Maurer *et al.*, 2007; <*http://gdo-dcp.ucllnl.org/downscaled_cmip_projections/*>)

The projections of future temperature and precipitation from the regionally downscaled models have been used to drive hydrologic models that provide projections of monthly values for hydrologic variables including SWE (U.S. Bureau of Reclamation, 2014). For this investigation both the ensemble mean of SWE projections was used, as well as selected individual model projections. The selection of individual models was based on how well a model projection of recent trend matched the observed trend. One model that matched the observed trend fairly well was the Institut Pierre Simon Laplace (IPSL) model from the IPSL Climate Modelling Centre (Dufresne *et al.*, 2013).

In addition to the regionally downscaled CMIP5 climate models and their coupled hydrologic models, this investigation also looked at projections of future snowfall and SWE in the Colorado River headwaters region from the Weather Research and Forecasting-Hydrology (WRF-Hydro) model, which was calibrated using observed SNOTEL data (Rasmussen *et al.*, 2011; Rasmussen *et al.*, 2014). Information about the climate model projections used in this investigation is summarized in Table 2.4.

Table 2.1. Snow course and SNOTEL stations included in the investigation. Data are available online at <http://www.wcc.nrcs.usda.gov/snow/>, accessed 7-9-16

Snow Courses	Elevation [m]	Station in/out of RMNP	SNOTEL stations	Elevation [m]	Cluster (Clow, 2010)
Lake Irene	3261	In	Lake Irene	3261	Orange
Longs Peak	3201	In	Willow Park	3261	Purple
University Camp	3140	Out	University Camp	3140	Purple
Cameron Pass	3135	Out	Deadman Hill	3116	Purple
Deadman Hill	3116	Out	Joe Wright	3085	Orange
Boulder Falls	3049	Out	Niwot	3021	Purple
Long Draw Reservoir	3042	In	Roach	2957	Orange
Milner Pass	2973	In	Lake Eldora	2957	Purple
Wild Basin	2915	In	Willow Creek Pass	2909	Orange
Willow Creek Pass	2909	Out	Bear Lake	2896	Purple
Ward	2896	Out	Phantom Valley	2752	Orange
Hidden Valley	2890	In	Stillwater Creek	2659	Orange
Hourglass Lake	2854	Out	Copeland Lake	2621	Purple
Bennett Creek	2804	Out			
Park View	2793	Out			
North Inlet Grand Lake	2744	In			
Red Feather	2744	Out			
McIntyre	2744	Out			
Deer Ridge	2743	In			
Chambers Lake	2743	Out			
Granby	2622	Out			
Copeland Lake	2621	In			
Big South	2621	Out			

Table 2.2. Temperature data collection stations used in this investigation. Data are from the National Oceanic and Atmospheric Administration (2016), William P. Rense (private monitoring), and U.S. Geological Survey (Colorado State University, 2016).

Station name (identification)	Data source	Elevation [m]	Latitude [N]	Longitude [W]	First year of record
Grand Lake 6SSW (USC00053500)	NWS (NOAA) Cooperative Weather Station	2526	40° 11’ 06”	105° 52’ 00”	1948
Allenspark	Private weather station operated by William P. Rense	2520	40° 11’ 17”	105° 30’ 05”	1960
Loch Vale Main and Remote Area Weather Stations	U.S. Geological Survey	3162	40° 17’ 17”	105° 39’ 46”	1983

Table 2.3. Tree-ring reconstructions used in this investigation

Name of reconstruction	Type of reconstruction	Dates of reconstruction	Reference
Gunnison, Colorado	SWE	1569-1999	Woodhouse, 2003
Upper Colorado River Basin (3 sites)	Hydrologic	1378-2000	Woodhouse and Lukas, 2006
Peak to Peak, Colorado (3 sites)	Hydrologic	1364-2000	Woodhouse and Lukas, 2006
Amalgre Mountain, Colorado Tree-ring and Isotopic	Temperature	1500-2000	Berkelhammer, 2010, LaMarche and Stockton, 1974

Table 2.4. Climate and hydrologic models from which projections were used in this investigation.

Name of model(s)	Description	Representative Concentration Pathway	Period of projection	Spatial resolution	Reference
CMIP5 Ensemble	Averaged results of 31 regionally downscaled global climate models, linked to hydrologic models	RCP 4.5	1951-2099	12 km	Maurer *et al.*, 2007
IPSL	Single model from the CMIP5 ensemble, linked to hydrologic model	RCP 4.5	1951-2099	12 km	Dufresne *et al.*, 2013
WRF-Hydro	High-resolution climate and hydrologic model	AR4 scenario A1B; equivalent to RCP 6.0	2006-2050	4 km	Rasmussen, 2011, Rasmussen 2014, Skamarock *et al.*, 2005

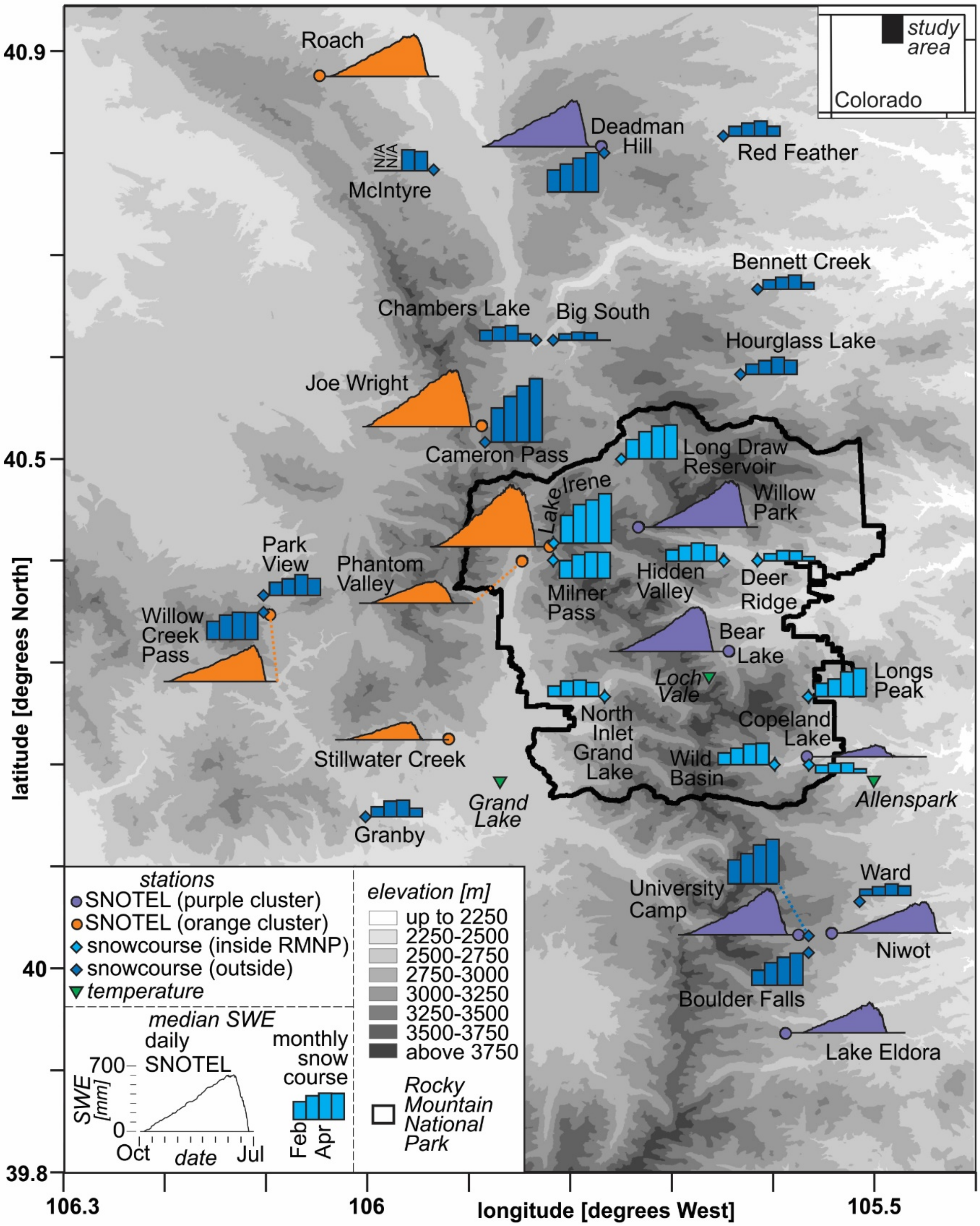

Figure 2.1. Locations of data collection sites used in this investigation. SNOTEL stations are designated by median niveographs scaled according to the example in the legend. Snow courses are designated by bar graphs showing median monthly SWE, on the same scale. Snow courses within Rocky Mountain National Park are shown in light blue; those outside the park are in dark blue. SNOTEL purple and orange clusters refer to those used by Clow (2010).

CHAPTER 3. METHODS

The data sources described in Chapter 2 provide a wealth of information that can be used to analyze trends of snowpack dynamics in the study area, and to relate the snowpack status and trends to associated characteristics such as elevation, location, temperature, precipitation, and freezing level. This chapter describes the approaches used to perform these analyses.

3.1 Snow Water Equivalent

Monthly records of SWE were tabulated for the 23 snow courses for their periods of record, covering up to eight decades (as early as 1936 through 2015). Daily records of SWE were tabulated for the 13 SNOTEL stations, covering the period 1981 to 2015. The 13 SNOTEL stations were considered as two groups based on proximity, according to the orange and purple groupings used by Clow (2010) (Figure 2.1). First of each month SWE (typically occurring within 3 days of the first of the month) for the snow season for snow courses (available over the long term from February 1 to May 1) and SNOTEL stations (October 1 to June 1) were compiled, and monthly change in SWE was computed.

Traditionally, the most commonly used measure of snowpack condition is April 1 SWE, based on the assumption that of the four monthly SWE measurements at most snow courses, April 1 SWE best approximates the annual peak SWE. In this study area, on average peak SWE occurs any time from mid-March through mid-May, but it can be earlier (1981, 2002, or 2012) or later (1995, 2011), depending on location, elevation, and the amount of winter precipitation in a given year. Using the assumption of peak SWE on April 1 can yield and under-estimation of peak SWE by as much as 12 percent (Bohr and Aguado, 2001). The daily SWE records provided by SNOTEL stations facilitate computation of additional measures to help characterize snowpack dynamics in greater detail. In this investigation these included measures such as annual peak SWE, difference between annual peak and first-of-month SWE values, date of peak SWE, number of days per year with accumulation greater than 5 mm and greater than 10 mm, and the total accumulation of SWE during a given period such as a month or a season (Figure 3.1). This last measure is roughly equivalent to "snowfall equivalent" as defined by Knowles *et al.* (2006): "*precipitation totals on days for which newly fallen snow was recorded*". Another measure of SWE that can be calculated using daily SNOTEL data is the change in SWE over a 15-day period. Evaluating the change in SWE over a specified period, such as a month or 15 days, and trends in the change in SWE, rather than just the amount of SWE present over the ground, can be an instructive way to emphasize the changes that are occurring to the pattern of snowpack dynamics. For example, if there is a trend toward more SWE during February, followed by a trend toward less SWE in March, then the negative change in SWE from February to March grows at a rate that may be greater than either of the monthly SWE trends. As stated in Chapter 1, winter recreation is increasingly popular in the study area. A reasonable index of sufficient snow depth for enjoyable winter recreation is a minimum of 100 mm of SWE. Therefore, the SWE measures analyzed also included number of days each year with SWE greater than 100 mm.

3.2 Trend Analysis

The primary objective of the investigation is to evaluate long-term trends in snowpack dynamics in the study area. Many of the SWE measures described above provide annual time series suitable for trend analysis. In this investigation trend analysis involves testing a time series of annual values to determine the likely presence of a long-term trend, and further estimation of the slope and statistical significance level of the trend, if present. Trend analysis of snow course SWE data covered two different periods. One was the entire period of record for the snow course through 2015, starting in the 1930s to 1950s. The second was the same period used for SNOTEL data, i.e., the 35-year period from 1981 to 2015. Trend analyses of SWE on early and late halves of the period of record for snow courses has shown varying patterns of sequential trends in northern Colorado (Fassnacht and Hultstrand, 2015). This variability is also evident in the snow course records used in this investigation, hence the emphasis on the recent 35-year period.

Some trend statistics are computed using one value per year, such as April 1 SWE or annual peak SWE. Since sub-seasonal variations in SWE and other variables are an important part of the pattern of snowpack dynamics, many of the trend statistics are computed for annual series of monthly values, such as SWE on the first of each month during the snow season. Some trend statistics are computed for annual series on a shorter, sub-monthly time step, such as 15-day change in SWE, or even shorter, such as daily for the snow season. The month-to-month trends exhibited irregular variability that could not be characterized as a seasonal cycle, hence, the Seasonal Kendall test (Helsel and Hirsch, 1995) was not used.

The SWE trends were further analyzed in relation to station characteristics such as location and elevation, and to other variables such as temperature, freezing level, and precipitation. For location, stations were grouped according to Clow's orange and purple clusters (Clow, 2010), representing locations on the western and eastern sides of the Front Range in north-central Colorado. For elevation, SWE trends were presented as a function of elevation. For temperature, freezing level, and precipitation, monthly SWE trends were compared with monthly trends in temperature and precipitation.

Given the large amount of natural inter-annual variability in many of the annual series of SWE data, occasional missing values, and the likelihood that the data are not normally distributed, a non-parametric approach was used for the trend analysis method. The method used here is the Mann-Kendall test for monotonic trend (Mann, 1945; Kendall, 1975; Gilbert, 1987), supplemented by the Theil-Sen estimate of slope of the linear trend (Theil, 1950; Sen, 1968; Gilbert, 1987; Helsel and Hirsch, 1995). Application of the combined method was made using the "MAKESENS" Excel macro (Salmi *et al.*, 2002).

3.3 Precipitation

Monthly precipitation values from the SNOTEL stations for the cold season (October-June) were used in two types of analysis. The first was a comparison of total monthly precipitation, total monthly SWE accumulation, and the fraction of total precipitation represented by the SWE accumulation. This comparison was made using precipitation records from four SNOTEL stations representing the range of station elevation, temperature, SWE, and precipitation totals for the study area. The stations were Joe Wright, Willow Park, Phantom Valley, and Copeland Lake (Figure 2.1). Joe Wright and Phantom Valley are in the orange cluster, and Willow Park and Copeland Lake are in the purple cluster. As mentioned above, the value for total SWE accumulation is roughly equivalent to "snowfall equivalent" (Knowles *et al.*, 2006). The second analysis, which used monthly precipitation and temperature data from all of the SNOTEL stations, was a three-way comparison of trends in total cold-season precipitation, average cold-season temperature, and April 1 SWE, to evaluate associations between SWE trends and precipitation or temperature trends.

3.4 Temperature

Although SNOTEL stations have temperature sensors, there are several reasons why the temperature records from SNOTEL stations are not the preferred records to use for studies of long-term temperature trends. The first reason is the shorter length of record; while the first SNOTEL stations were installed in 1979 across the study area, and temperature was not recorded until the late 1980s or thereafter. The second reason is inconsistency in the collection of temperature data at SNOTEL stations. During the early to mid-2000s the SNOTEL program switched to an extended range temperature sensor, installed a new radiation shield, instituted a new data collection protocol, and moved all temperature sensors to a new location so that each was above the snow pillow. These changes resulted in records with uncertainty about consistency over the long term (Oyler *et al.*, 2015). For this reason, most of the temperature data used in this investigation are from three stations (Allenspark, Grand Lake, and Loch Vale) that are not part of the SNOTEL network. Average monthly temperature data from the three weather stations were analyzed for magnitudes and trends for the period 1981-2015 (1983-2015 for Loch Vale).

This investigation did make use of temperature data from SNOTEL stations for analyses in which the association with particular SNOTEL stations was important. Two SNOTEL stations, the highest Lake Irene and the lowest Copeland Lake, were selected for an analysis of temperature records to characterize the amount of warmer-than-freezing weather at the station in a given month. Daily maximum temperature records were examined to identify days on which the maximum daily temperature was warmer than zero. The data homogenization method developed by Oyler *et al.* (2015) was applied, in an effort to improve data homogeneity. The quantity included in the comparison was the number of degrees Celsius, if any, warmer than zero, reported for each day, accumulated for each month. As a measure of total monthly duration of warmer-than-freezing weather that can help to melt snow, this quantity involves an assumption that the duration of warmer-than-freezing weather, on each day when it occurs, is similar from day to day, or that the variations over a month average out to the same durations from month to month. Without examining at the hourly records (some issues with the hourly data are summarized in Avanzi *et al.*, 2014), this assumption represents a reasonable approach to quantifying the amount of warmer-than-freezing weather in a month. Temperature data from SNOTEL stations were also used in an analysis of trends and variability in SWE and in October-June temperature and precipitation at SNOTEL stations, similar an analysis presented for the Pacific Northwest (Mote, 2003). For this analysis the patterns of trends in average October-June temperature and total October-June precipitation at each SNOTEL station were evaluated with respect to trends in SWE in order to assess influences of temperature and precipitation trends on SWE trends.

3.5 Freezing Level

The monthly freezing-level estimates from the North American Freezing Level Tracker <*www.wrcc.dri.edu/cwd/products/*> (Chapter 2) were compared with the elevation range of the SNOTEL stations at times when SWE decreases were occurring. Trends in monthly freezing level were also evaluated, to identify months with strong upward or downward trends over time. These trends were compared with those for temperature and SWE.

3.6 Estimation of Daily Niveographs from Snow Course Data

A niveograph is a plot of SWE versus time (Fassnacht and Patterson, 2013), usually for a full snow season. As mentioned above, the daily SWE records provided by SNOTEL stations provide significant added value for analysis of snowpack dynamics, compared with the four monthly SWE values provided by snow courses. However, snow courses have the advantage of a longer period of record compared with SNOTEL stations. Accordingly, it would be helpful to have a method for estimating daily niveographs based on monthly snow course data. Simply interpolating between the four monthly SWE values from the snow course often results in a poor estimate for the daily niveograph, as the annual peak frequently occurs after the final SWE measurement on May 1. Even for those cases when the annual peak SWE occurs prior to May 1, simple interpolation misses the peak (Bohr and Aguado, 2001).

Median values for daily SWE data from co-located or nearby SNOTEL stations can be used in estimating the daily niveograph at the snow course. In addition, the magnitude of the observed SWE values at the snow course can be used to characterize the snow year as low, moderate, or high in SWE. Further, such estimated daily niveographs can be used to estimate the magnitude and timing of the annual peak. This enhanced niveograph interpolation method was tested using data from two sites in Rocky Mountain National Park (Lake Irene and Willow Park SNOTEL stations used as snow courses), and found to produce improved accuracy in estimates of annual peak SWE compared with the assumption that the peak occurred on April 1 (Patterson and Fassnacht, 2014).

Four SNOTEL stations in the park were selected as index stations to provide daily niveographs. A fifth SNOTEL station, Willow Park, was treated as if it were a snow course, providing data only on the first days of February, March, April, and May. Its daily niveograph was used only as a check to test the assumption that daily niveographs from nearby index SNOTEL stations could be used to approximate daily niveographs for a snow course. For each day in each snow year median values of SWE from the four index stations were computed. Next, medians of these SWE values for all years of record were computed for each day in the snow year. The resulting combined median niveograph was then smoothed by computing 5-day running means for each daily value (Figure 3.2). Such a smoothed niveograph was developed for each of three groups of years representing high, medium, and low May 1 SWE values. The smoothed combined median niveographs served as the unadjusted models for the simulations. To adjust the smoothed combined median niveograph to approximate each annual niveograph at Willow Park, time-varying daily multipliers were applied to the unadjusted model. The daily multipliers were based on six calibration points coinciding with the first-of-month SWE values and the components of the niveograph (Figure 3.3). Multipliers at the calibration points were set according to the measured SWE on those dates. If SWE increased from April 1 to May 1 in a particular year, the interpolation routine was adjusted to extend the annual peak SWE into May. The resulting interpolated niveographs could then be tested against the actual daily niveographs from the Willow Park SNOTEL record to validate the technique.

3.7 Tree-ring Reconstructions of SWE and Temperature

A question that arises in a trend study is: Have there been other periods of similar duration during recent centuries, with indications of similar trends? This question can be addressed using tree-ring reconstructions of SWE and temperature. With such reconstructions it can be difficult or impossible to identify multi-century trends, because growth trends and autocorrelation have been removed from the record (Woodhouse and Lukas, 2006), and because the algorithms used to produce the reconstructions mute extreme values. As a result, "*reconstructions are usually a conservative estimate of past variability*" (Woodhouse, 2016). However, these reconstructions can be used to examine trends with durations of several decades. The reconstructions listed in Chapter 2 (Table 2.3) were examined visually to identify any multi-decade periods of similar duration to the 7-decade and 3.5-decade periods used in this investigation, during which trends in SWE and/or temperature were relatively consistent and monotonic, increasing or decreasing. The two hydrologic reconstructions were used because they have been found to be strongly correlated with April 1 SWE (Woodhouse and Lukas, 2006). To estimate the relation between tree-ring width index included in these hydrologic reconstructions, and SWE in the study area, a regression analysis was made of SWE at Wild Basin in Rocky Mountain National Park, on tree-ring width index for a recent period, 1986-2000, during which observed SWE records were available.

3.8 Projections of Future Climate and SWE

Another question that arises in trend studies is whether the identified trends are likely to continue in the future. One way to address this question is to examine future projections provided by climate models and linked hydrologic models that apply to the study area (Milly *et al.*, 2005; Stahl *et al.*, 2008). For this investigation, regionally downscaled projections of climate and SWE for the remainder of the 21st century, averaged from an ensemble of 31 CMIP5 climate models, were examined and compared with the observed trends (Maurer *et al.*, 2007; Brekke *et al.*, 2014). The selected representative concentration pathway was RCP4.5, representing an assumption for a low-to-moderate greenhouse-gas emissions scenario (Clarke *et al.*, 2007). The area selected for regionally downscaled results is a rectangle of 3/8 degree of latitude and 1/8 degree of longitude, roughly centered on Rocky Mountain National Park (Figure 3.4). In addition to examining the ensemble model projections, individual CMIP5 models were evaluated to select the one that, when linked to the Variable Infiltration Capacity (VIC) hydrologic model (Brekke *et al.*, 2014), most closely matched observed SWE data in the study area. Specifically this was the IPSL model mentioned in Chapter 2. Projections made using this model, linked to the VIC hydrologic model, were also examined and compared with observed trends in SWE.

The regionally downscaled CMIP5 climate models have a spatial resolution of 1/8 of latitude by 1/8 degree of longitude, or about 12 x 12 km (U.S. Bureau of Reclamation, 2014). Finer-resolution models have been shown to provide more accurate representation of complex topography in mountain areas, and hence of the complex weather patterns that influence the snowpack in these areas (Rasmussen *et al.*, 2011; Rasmussen *et al.*, 2014). The Weather Research and Forecasting-Hydrology (WRF-Hydro) model, which has a resolution of 4 x 4 km, was found to have close agreement with observed snowfall and SWE data observed at SNOTEL stations in the Colorado mountains (Rasmussen *et al.*, 2014). Projections of future snowfall and SWE for the headwaters region of the Colorado River Basin in Colorado, provided by the WRF-Hydro model, were also examined and compared with the observed temperature, precipitation, snowfall and SWE data.

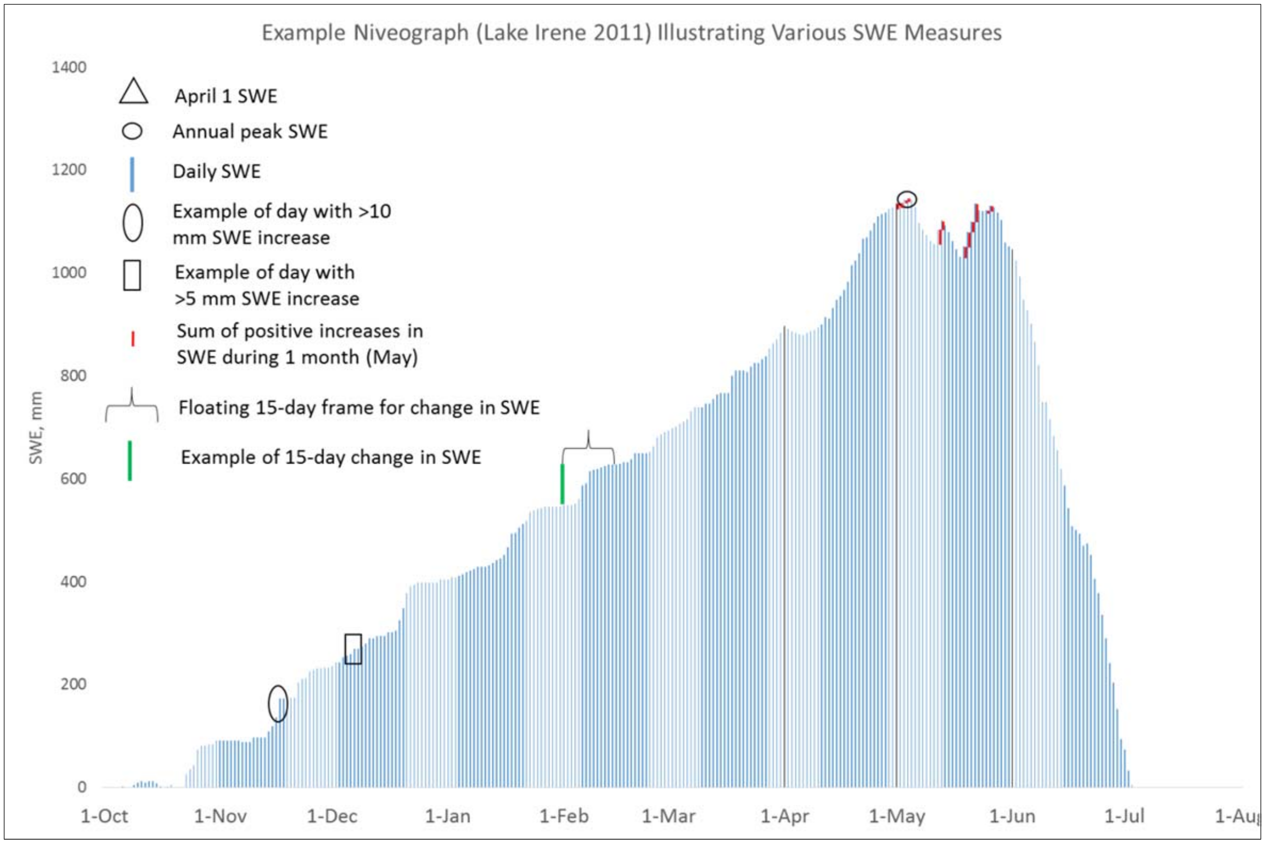

Figure 3.1 Example niveograph illustrating various SWE measures.

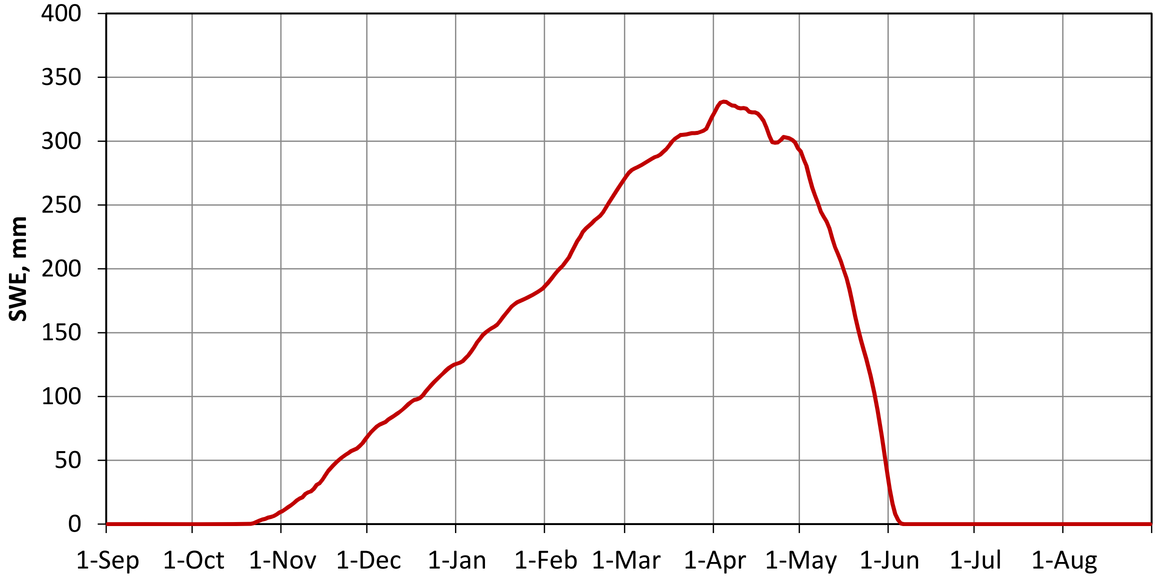

Figure 3.2 Smoothed median niveograph based on four index SNOTEL stations.

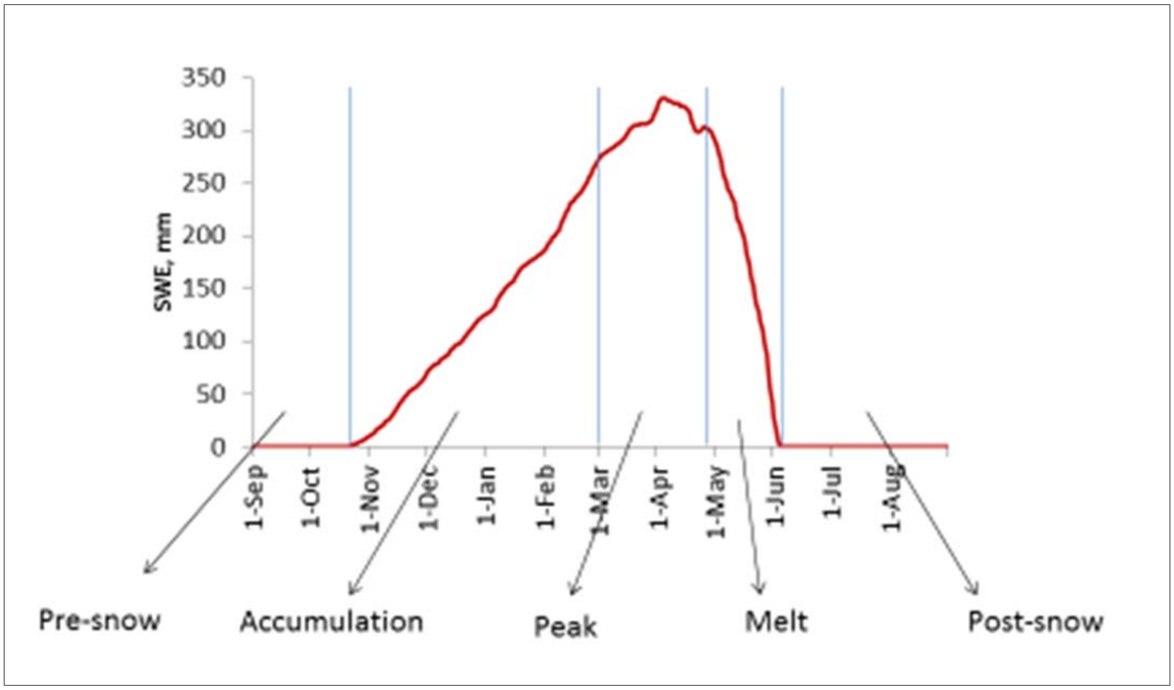

Figure 3.3. Separation of niveograph into components for estimation.

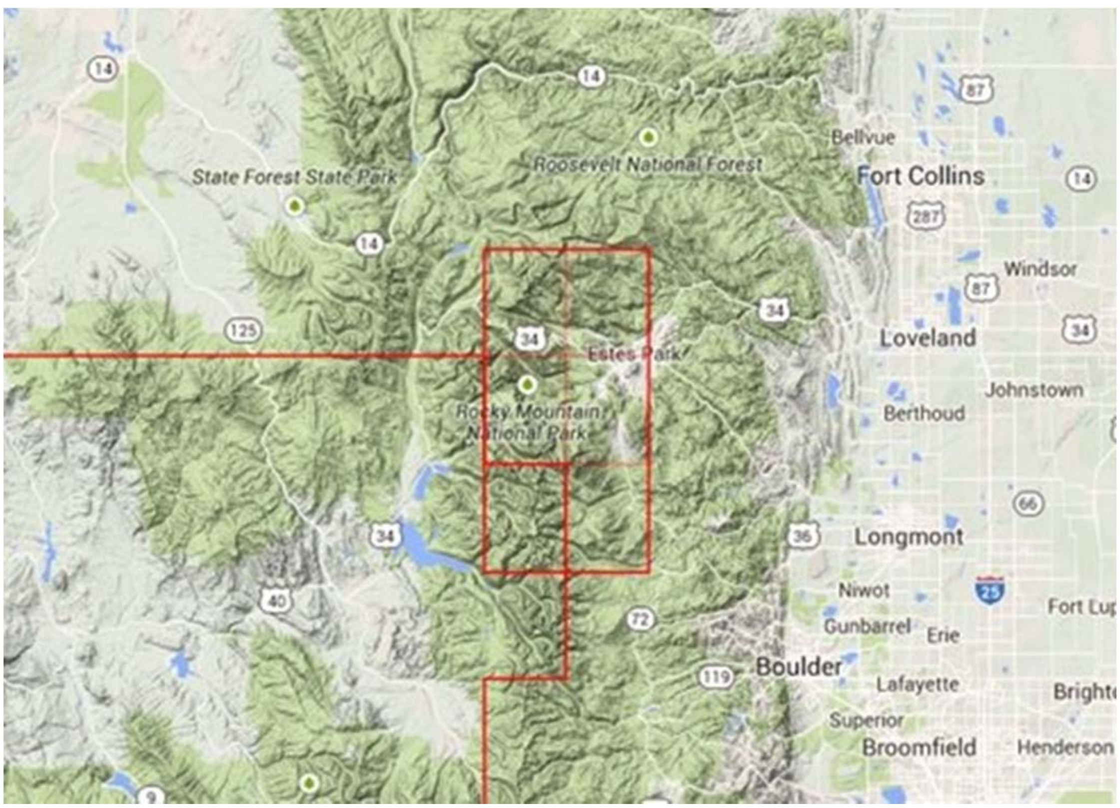

Figure 3.4. Area (red box) selected for regionally downscaled projections from the CMIP5 climate models.

CHAPTER 4. RESULTS

4.1 Trends in Snow Water Equivalent

4.1.1 Monthly SWE

SWE measures, such as first-of-month SWE (Figure 4.1) and peak annual SWE (Figure 4.2), exhibit strong inter-annual variability. This variability is related to variations in weather patterns from year to year. Hence, a very high year for April 1 SWE and peak SWE (2011) can be followed by a very low year (2012), even with decreasing trend in a 35-year period. Inter-annual variations in measures of SWE, such as April 1 SWE, are more closely related to variations in precipitation than to variations in temperature (Figure 4.3). Regression analyses of April 1 SWE at a representative SNOTEL station, Phantom Valley, on average October-June temperature as recorded at the SNOTEL station, and on total October-June precipitation, showed a much stronger correlation with precipitation (R^2 = 0.50, with a strong positive slope) than with temperature (R^2 = 0.04, with a slight negative slope).

Median SWE values computed over the 35-year analysis period illustrate the range of snowpack dynamics in the study area (Figure 2.1). Median April 1 and May 1 SWE values ranged from just over 100 mm at low-elevation snow courses on the eastern slope, such as Copeland Lake, Deer Ridge, Bennett Creek, Chambers Lake, Big South, and Red Feather, to 500 to 600 mm at higher elevation snow courses near the summit of the Front Range, such as Lake Irene and Cameron Pass. Several of the higher-elevation snow courses, such as Deadman Hill, Cameron Pass, Long Draw Reservoir, Lake Irene, Longs Peak, University Camp, and Boulder Falls, had median May 1 SWE values that exceeded those on April 1, indicating annual peaks that typically occurred after April 1. Median values for annual peak SWE at SNOTEL stations reflect similar variability, from 148 mm at Copeland Lake to 668 mm at Lake Irene. The median date of peak SWE at SNOTEL stations ranged from March 6 at Copeland Lake to May 6 at Joe Wright.

Across the study area, there is a preponderance of decreasing trends in SWE on the first of the month, meaning a trend toward less SWE with time (Figure 4.4). For example, the trend at the Lake Irene SNOTEL station for the period 1981-2015 is generally downward at a rate of about 31 mm/decade (Figure 4.1). These decreasing trends are not universal, but are pervasive. Magnitudes of the SWE trends varied from zero to gains of as much as 30 mm/decade during February, and losses of as much as 50 mm/decade during March (Figures 4.4 and 4.5). Of the 159 monthly trend values for the period 1981-2015, computed based on data from snow courses and SNOTEL stations, 94 (59 percent) were negative, while 65 (41 percent) were positive (Figures 4.4 and 4.5). Four of the increasing trends and none of the decreasing trends were statistically significant at the $p<0.05$ level; six of the increasing trends and four of the decreasing trends were statistically significant at the $p<0.10$ level. When sorted according to Clow's orange and purple clusters, the orange cluster (western side of the Front Range) had 80 percent decreasing trends, and the purple cluster (eastern side of the Front Range) had 52 percent decreasing trends.

When the SWE trends are sorted by month, the difference between upward trends in February and downward trends in March is again highlighted. The monthly SWE value with the most increasing trends (more SWE) was March 1, representing conditions during February (Figure 4.5a). The following month, April 1, representing conditions during March; as well as December, representing conditions during November, had the most decreasing trends. Comparing the monthly SWE values for the orange and purple SNOTEL clusters on the western and eastern sides, respectively, of the Front Range, shows that on March 1 and June 1, reflecting conditions during February and May, the purple cluster had more increasing trends than the orange cluster (Figure 4.5a).

Higher-elevation sites tended to have more decreasing trends in SWE, while lower-elevation sites tended to have more increasing trends in SWE (Figure 4.6). This pattern holds for all months, as indicated by the slopes of the colored best-fit lines in Figure 4.6. The positions of these best-fit lines indicates that the March 1 SWE trend, representing conditions during February, has the highest preponderance of increasing trends, while the April 1 SWE trend, representing conditions during March, has the highest preponderance of decreasing trends. This is a pattern that will recur throughout the results and discussion of this investigation. As indicated by the values for R^2 in Figure 4.6, the correlation between elevation and SWE trend tended to be stronger during November and December (on December 1 R^2=0.52 and on January 1 R^2= 0.66), moderate during January to March (on February 1 R^2=0.31, on March 1 R^2=0.27, and on April 1 R^2=0.32), and weaker during April (on May 1 R^2=0.16).

Nearly all of the trends in monthly change in SWE during December, January, and February were positive, meaning a trend toward greater gain in SWE during those months (Figure 4.5b). Nearly all of the trends during November and March were negative, meaning a trend toward less gain (or greater loss) in SWE during those months. April and May had mixed trends in monthly change in SWE.

While this investigation focuses on the 35-year period (1981-2015), some trends in SWE measures from snow courses were analyzed for longer periods. Most (66) of monthly SWE trends for the 23 snow courses over the entire period of record were decreasing, especially on April 1 and May 1, reflecting conditions during March and April (Figure 4.7). Twelve of the trends for February 1 and March 1st SWE outside the park (6 stations) were increasing, while only two stations inside the park had an increase on March 1st (Figure 4.7). In comparison to monthly SWE trends over the 35-year period 1981-2015, trends over the longer period tended to be more negative. Only two stations with decreasing trends over the short term had increasing trends over the long term (upper left quadrant in Figure 4.7). However, 25 stations with increasing trends over the short term had decreasing trends over the long term (lower right quadrant in Figure 4.7). This is especially true for SWE trends on February 1 and March 1. Trends in monthly change in SWE at snow courses were negative during March for both periods, with stronger negative values in the shorter period (figure 4.8). This reflects the largely decreasing trends for April 1 SWE in Figure 4.7. Trends in monthly change in SWE during April were mostly positive for both periods, especially the shorter period. This reflects the largely increasing trends for May 1 SWE in Figure 4.7. Trends during February were mixed and close to level for the longer period, and were mostly positive for the shorter period.

4.1.2 Trends Derived from Daily SWE

Trends in annual peak SWE at the 13 SNOTEL stations were mixed, with 7 declining trends, 5 increasing trends, and one station with no trend. When plotted in order of elevation, there was again a tendency for higher elevation stations to have trends that are more negative, and lower elevation stations to have trends that are more positive (figure 4.9). When grouped by cluster, the orange (west) cluster had more decreasing trends (4 down, 1 up, 1 zero), while the purple (east) cluster had slightly more increasing trends (3 down, 4 up).

For some stations, there was a large difference between the annual peak SWE estimated from the daily data and the monthly SWE. For the 13 SNOTEL stations in the study area, the average difference between annual peak SWE and April 1 SWE was 22 percent (Table 4.1). The average difference between annual peak SWE and the highest first-of-month SWE (often May 1) was 7 percent.

Trends in date of annual peak SWE at the 13 SNOTEL stations were also mixed, with 4 trends toward earlier date of peak, 8 trends toward later date of peak, and one station with no trend. There was a slight tendency for higher elevation stations to have trends toward later date of peak, and for lower elevation stations to have trends toward earlier date of peak (Figure 4.8.2). When grouped by cluster, the orange cluster had a relatively even mix, with 3 trends toward later peak, 2 trends toward earlier, and one zero trend. The purple cluster had more trends toward later peak, with 5 trends toward later peak, and 2 trends toward earlier peak.

For the Lake Irene SNOTEL station, all of the daily SWE trends were negative, varying from zero to -8 mm/year, while they were all positive for Copeland Lake (Figures 4.10a and b). The most pronounced decrease in daily SWE trends at Lake Irene occurred in mid to late May, but the trends during this period were not statistically significant. The most pronounced increase in daily SWE trends at Copeland Lake occurred during early March, with trends that were significant at the $p<0.05$ or $p<0.10$ level. The most significant trends occurred around the time of the median annual peak SWE. The variability in these daily trends often occurred on time scales of a few days or weeks, creating patterns of variability that could not be discerned using monthly SWE data. Significant trends occurred early in the season at Lake Irene (decreasing for 3 days in early January), as well as at Copeland Lake (increasing for 9 days in late December). Reversals in daily SWE trends took place over a few days at all four SNOTEL stations for which daily trends were computed (Figures 4.10 and 4.11). At Joe Wright, trends were nearly always negative, except for a few days in early May. Significant decreasing trends occurred at Joe Wright during much of November. At Bear Lake, the trends, which were not significantly significant, were negative in November and early December, positive from early January to late March, negative from late March to late April, then positive again for the rest of the snow season. At all four stations, whether the trends were negative or positive, they decreased from late February through early April, becoming less positive or more negative. At the three stations with niveographs extending later than early April, the trends increased again through early May, followed by a final decrease at the end of the snow season.

Accumulation, loss, and net change in SWE during a moving 15-day period were calculated on a daily basis for three representative SNOTEL stations with high, moderate, and low median niveographs: Willow Park, Phantom Valley, and Copeland Lake, respectively. Willow Park and Copeland Lake are in the purple cluster, and Phantom Valley is in the orange cluster (Clow, 2010). All three SNOTEL stations showed similar patterns of accumulation, loss, and net change, with progressively lower SWE values at the three stations (Figure 4.12). Fifteen-day accumulation was relatively consistent from October through May, with a slight increase in accumulation rate during late April to early May at Willow Park. SWE loss occurred in October and November when weather suitable for melting typically occurred. Then during December, January, February, and (at Willow Park and Phantom Valley) early March, 15-day SWE loss was virtually zero, as very little melt occurred during that period. From late March (early March at Copeland Lake) through the end of the melt season, 15-day SWE loss accelerated, and reached its greatest (negative) values. Combining accumulation and loss into a

15-day net change in SWE, a typical pattern emerges of positive 15-day net change in SWE during the accumulation season, and negative 15-day net change in SWE during the melt season.

Trends in these 15-day SWE changes for the period 1981-2015 were computed for every 20th day during the snow season. All three SNOTEL stations showed similar trends (Figure 4.13). During November the trend was toward greater net loss of SWE. November also had a trend toward less accumulation. From December through early March, the trend was toward greater net gain in SWE. This period also had a trend toward greater accumulation. During mid-March through early April, the trend shifted abruptly to greater net loss in SWE. This period also had a trend toward greater melt, and to some extent a trend toward less accumulation. During mid-April through mid to late May, the trend in net SWE change was again positive (or zero for Copeland Lake). This period also had a trend toward greater accumulation, and less melt. Finally, at Willow Park and Phantom Valley, where the snow season is sufficiently long, in late May through early June, the trend in net change in SWE shifted one more time to greater net loss. This period also had a trend toward greater loss.

Trends in number of days per year with over 100 mm of SWE at SNOTEL stations during the period 1981-2015 were mostly negative, with 9 decreasing trends, 3 increasing trends, and one zero trend (Figure 4.14). Higher elevation stations had uniformly decreasing trends, while lower elevation stations had mostly increasing trends. The orange and purple clusters had similar trends, mostly negative but one or two trends that were zero or positive.

4.1.3 Trends in Number of Days per Year of Accumulation and Melt

Trends in the number of days per year of SWE accumulation or melt in excess of 10, 5, and 0 mm, were also mixed (Table 4.2). The higher elevation stations tended to have trends toward fewer days with accumulation of over 10 mm, but more days with accumulation of lesser amounts. The lower elevation stations tended to have trends toward more days with accumulation exceeding all three amounts. Stations at all elevations tended to have trends toward more days with melt exceeding all three amounts. The trends in number of days with accumulation and melt in excess of the lowest amount (0 mm) were the most statistically significant. When grouped by cluster, the orange cluster had slightly more increasing trends in days with both accumulation and melt exceeding the thresholds. The purple cluster had a strong preponderance of increasing trends, especially for melt.

4.2 Trends in Precipitation

Precipitation in the study area is well distributed throughout the year, yet April, May, July, and August typically have higher monthly totals (Figure 4.15). There is a strong elevation dependency to the distribution of precipitation, with higher sites receiving more than twice as much precipitation as lower sites (Figure 4.16).

Monthly trends in precipitation at the four SNOTEL stations have the same general pattern (Figure 4.17). October tended to have weak trends for precipitation, snowfall equivalent (SFE), and fraction of precipitation represented by SFE. November had significant decreasing trends in precipitation and SFE, indicating trends toward warmer and drier conditions. December, January, and February had moderate to strong increasing trends in precipitation, SFE, and fraction of precipitation represented by SFE, indicating trends toward colder and wetter conditions. March had strong decreasing trends in precipitation, indicating a trend toward drier conditions, and no trend or decreasing trends in SFE. March had increasing trends in fraction of precipitation represented by SFE at three of the four stations, and a slight decreasing trend in this fraction at Willow Park. During April trends again became positive for precipitation (except at Phantom Valley), and for SFE, and were mixed for fraction, indicating a trend toward wetter conditions. During May trends were mixed with generally small trends.

4.3 Trends in Temperature

The three temperature stations exhibit a normal pattern of seasonal variation in average monthly temperature (Figure 4.18). They also exhibit a normal pattern of cooler temperatures at higher elevation, as seen at Loch Vale.

For the period 1981-2015 (1983-2015 at Loch Vale), the three temperature stations show warming trends ranging from 0 to 0.28°C/decade (Table 4.3, Figure 4.19). The annual time series show typical inter-annual variations, but the general warming trend appears to be relatively consistent throughout the 33-year period (Figure 4.19). The longer temperature record at Grand Lake enables us to evaluate the temperature trend at that station over a longer period from 1949 to 2015 (Figure 4.20). While it is possible to draw a Theil Sen's slope line, with an increasing trend, through these data points, the pattern of the data suggest a cooling trend during 1949 to 1973, and a warming trend from 1973 to 2015, as indicated by the linear best-fit lines.

Trends in monthly average temperature at these three stations over the period 1981-2015 show substantial variability among months (Figure 4.21). Consistent warming trends at all three stations were noted during eight months: September, October, November, January, March, June, July, and August. Rates were as high as 0.60°C/decade in November at Allenspark. Grand Lake had warming trends in every month, although those for December and February were the smallest, around 0.10°C/decade. Seven of the monthly warming trends were statistically significant. At Allenspark and Loch Vale, cooling trends were present during December, February, April, and May, with rates as high as -0.60°C/decade in May at Loch Vale. The decreasing trend in May at Loch Vale was statistically significant.

4.4 Trends in Freezing Level

Simulated free-atmosphere average monthly freezing-level elevations were retrieved from the North American Freezing Level Tracker for the seven coldest months of the year (November through May) for each year during the 35-year study period (1981-2015). The averages of these monthly simulated freezing levels show a typical seasonal trend of decreasing early in the winter then increasing freezing level later in the snow season (Figure 4.22). The average freezing levels for November and March both were about 500-600 m below the bottom of the elevation zone containing the SNOTEL stations. Freezing levels for December, January, and February were well below this level, and freezing levels for April and May were above this level. Trends in monthly freezing level, represented by the red line in Figure 4.22, were upward (warming) in all months except February, which had a slight downward (cooling) trend. The strongest upward trends, about 1600-1700 m/century, were in November and March.

Results of the analysis of warmer-then-freezing weather at Lake Irene and Copeland Lake showed consistent seasonal patterns for the two SNOTEL stations, with Copeland Lake having about 100-200 more maximum-daily degrees each month than Lake Irene (Figure 4.23). At both stations, November and March had similar values for this measurement. Analysis of trends in these monthly maximum-daily degree totals over their respective periods of record (1986-2015 for Lake Irene, 1989-2015 for Copeland Lake), showed that November had the strongest increasing trend at both stations (Figure 4.24). January, March, and July also had increasing trends, while October, February, and May had decreasing trends.

4.5 Estimation of Daily Niveographs from Snow Course Data

Testing of the niveograph estimation technique (Chapter 3, Section 3.6), resulted in interpolated niveographs for Willow Park for the period 1981-2012. The mean differences between modeled and observed peak SWE, date of peak SWE, and total SWE accumulation during the year small for the first two at 4.8 and 1.4% less, but greater for the latter at 22.0% less.

4.6 Tree-ring Reconstructions of SWE and Temperature

The four tree-ring reconstructions of SWE, hydrology, and temperature listed in Chapter 2 (Table 2.3) were examined for trends of similar magnitude and duration as those identified in this investigation. The longest period with a consistent trend in the Gunnison SWE reconstruction (Woodhouse, 2003) was a 13-year period from 1613-25, during which SWE reportedly decreased at a rate of about 115 mm/decade. This is a short duration of declining trend compared with the 80 years of decreasing SWE seen at the Wild Basin snow course (Figure 4.1). The slope of the decreasing trend in the Gunnison reconstruction is estimated to be steeper than the 19 mm/decade loss in April 1 SWE noted at the Wild Basin snow course, or the 31 mm/decade loss noted at the Lake Irene SNOTEL, for the period 1981-2015.

The longest period with a consistent trend in the combined Upper Colorado River Basin hydrologic reconstruction (Woodhouse and Lukas, 2006) was a 37-year period from 1557-93, during which the tree-ring width index declined from 1.3 to 0.6. According to the regression between tree-ring width index and April 1 SWE at Wild Basin, a decline in tree-ring width index of 0.1 is roughly equivalent to a decline in SWE of 10 mm. Therefore, the decline in tree-ring width index of 0.7 during 1557-93 would roughly correspond to a decline in SWE of about 70 mm, or about 19 mm/decade. This duration of a trend in declining SWE is still less than the duration seen in the current investigation, while the rate of SWE loss is estimated to be similar to that seen at the Wild Basin snow course.

The longest period with a consistent trend in the combined Peak to Peak hydrologic reconstruction (Woodhouse and Lukas, 2006) was a 35-year period from 1315-49, during which the tree-ring width index declined from 1.2 to -0.5. According to the Wild Basin SWE regression, this should roughly correspond with a decline in SWE of about 170 mm, or 49 mm/decade. This is a shorter duration, but similar rate of decline, to some of the faster rates of SWE decline seen in this investigation.

The temperature reconstruction from Amalgre Mountain, Colorado (LaMarche and Stockton, 1974; Berkelhammer, 2010) contains several multi-decade periods with consistent warming trends. The longest was a 50-year period from about 1840-90 when temperature anomalies were estimated to change from about -0.7 to +0.2°C, indicating a temperature increase of about 0.9°C, or 0.18°C/decade. The record also included a 31-year period from 1670-1700, with a temperature increase of about 0.4°C, or 0.13°C/decade; and a 26-year period from 1575-1600, with a temperature increase of about 0.3° C, or about 0.12°C/decade. These warming trends have similar durations to the 43-year warming trend noted in this investigation for Grand Lake, and similar rates to the warming trends noted at all three temperature stations.

4.7 Projections of Future Climate and SWE

Projections from climate models were examined to determine how closely they match observed trends in temperature, precipitation, and SWE during the study period, and to provide estimates of future conditions. The temperature projection of the ensemble average of CMIP5 climate models was made using projected values for monthly minimum temperature, as average temperatures were not available. The projected warming during the period 1983-2015 was about 1.2°C, or 0.34°C/decade. This is a faster rate of warming than the rates seen for average annual temperature at the stations used in the study, which ranged from 0.06 to 0.23°C/decade (Figure 4.19). The ensemble mean projection for future increase of the average of monthly minimum temperatures during 2015-2099 was about 2.0 degrees, or 0.24°C/decade. This projected warming rate is similar to observed rates at the lower elevation stations in the study area. In the model projections, all months had similar warming trends. In contrast, the observed temperature trends varied monthly (Figure 4.20).

The projection for total annual precipitation from the ensemble average of CMIP5 climate models for the period 1981-2015 was similar to the observed total annual precipitation at the centrally located Willow Park SNOTEL station, but the models did not display the inter-annual variability observed at the station (Figure 4.25). On a monthly basis, the patterns of simulated and observed average monthly precipitation values were similar (Figure 4.26). Trends in monthly precipitation also showed relatively agreement in terms of direction of change for most months, with the exceptions of November and March (Figure 4.27). In these two months the modeled trends projected more precipitation while the trends at the station were toward less precipitation.

Average annual precipitation trends projected into the future by the CMIP5 model ensemble for the period 2015-2099 showed slightly increasing trends, averaging about 0.60 mm/year. On a monthly basis, the projected monthly trends showed a slightly different pattern from the simulated 1981-2015 trends (Figure 4.28). These future trends emphasized increasing precipitation in the early part of the snow season through December, and much smaller increases, or decreases, in trend for the remainder of the snow season, through June.

Projections of SWE from the Variable Infiltration Capacity (VIC) hydrologic model linked to the CMIP5 climate models were generally similar to observed values. For example, the model ensemble projections for average monthly SWE for the period 1981-2015 showed a similar pattern to the observed average monthly SWE values at Willow Park. The difference was that the models simulated slower accumulation and a later peak SWE compared to observed SWE at Willow Park (Figure 4.29). Simulated April 1 SWE was much less than observed, by about 200 mm, with much less inter-annual variability (Figure 4.30). Monthly, there is little similarity between the simulated and observed trends in SWE as the observed trends were larger (Figure 4.31). October, November, and February were the only months were the trends were in the same direction, with the first two months decreasing and the last month increasing. Average annual SWE trends projected into the future by the CMIP5 model ensemble showed a continued level to slightly decreasing trend in April 1 SWE for the period 2015-2099, averaging only -2.0 mm/decade (Figure 4.32).

Examination of projections of individual CMIP5 models, rather than the ensemble mean of all of the models, showed that one model, IPSL model (see Chapter 2), computed April 1

SWE patterns that were similar to the declining trends noted at many of the snow courses and SNOTEL stations in this investigation, as illustrated over a 65 year period at the Longs Peak snow course (Figure 4.33). Since the observed trend in April 1 SWE at Longs Peak during 1951-2015 of -9.6 mm/decade is very similar to the average April 1st trend at all of the snow courses of -10.2 mm/decade, it was used to compare to IPSL results. Due to inter-annual difference, smoothing using a 5-year running mean, the observed April 1 SWE time series at Longs Peak was very similar to the IPSL model during the period from 1951 to 2015 (Figure 4.33). The trend simulated using the IPSL model has a slope of -7.2 mm/decade, according to the linear best-fit trend line. The IPSL model projection through 2099 shows a decrease in April 1st SWE starting around 2050 (Figure 4.33).

The monthly distribution of SWE trends for the period 1981-2015, simulated by the IPSL model, has a few important similarities with observed SWE trends from snow courses and SNOTEL stations (Figure 4.34). Both the simulated and observed monthly SWE trends show a general pattern of more SWE accumulation, or less SWE loss, during the early part of the snow season, and greater SWE loss during the latter part of the snow season, particularly March through May. IPSL model projections through 2099 suggest that trends toward SWE loss will intensify in all cold-season months and expand into the early part of the snow season (Figure 4.35).

Table 4.1. Percent difference between annual peak SWE and first-of-month SWE for SNOTEL stations in the study area, 1981-2015, expressed as percent of peak SWE

station	peak SWE vs. April 1st SWE				peak SWE vs. max 1st of the month SWE (Feb-May)				peak SWE vs. max 1st of the month SWE (Feb-June)			
	max	mean	std dev	min	max	mean	std dev	min	max	mean	std dev	min
Joe Wright	55.7	21.8	12.4	1.39	28.9	7.61	6.23	0	12.4	5.55	3.12	0
Lake Irene	36.6	12.8	8.76	0.59	14.6	5.70	3.89	0	11.2	5.20	3.56	0
Willow Park	58.7	20.8	13.2	1.23	27.9	7.63	6.03	0.39	19.1	6.62	4.84	0
Bear Lake	41.1	18.0	11.1	1.05	15.5	6.05	4.58	0	15.5	6.05	4.58	0
Copeland Lake	100.0	64.2	34.2	0	36.0	9.22	9.48	0	36.0	9.22	9.48	0
Phantom Valley	91.8	17.1	19.8	0	21.2	7.41	5.93	0	21.2	7.41	5.93	0
Stillwater Creek	85.5	20.3	23.3	0	23.3	7.66	5.68	0	23.3	7.66	5.68	0
Roach	44.1	14.8	10.6	1.04	28.2	6.91	5.50	0.56	14.0	6.11	3.91	0.56
Willow Creek Pass	52.8	21.4	10.6	2.38	17.6	6.06	4.32	0	14.9	5.70	3.84	0
Deadman Hill	59.7	22.8	11.3	4.62	37.5	6.89	6.56	0	11.5	5.58	3.38	0
University Camp	49.0	22.3	11.8	1.09	18.2	6.46	5.12	0.43	18.2	5.97	4.86	0.43
Niwot	51.4	16.9	14.6	0	22.6	7.10	5.63	0	22.6	6.66	5.52	0
Lake Eldora	60.0	14.9	14.6	0.57	30.6	8.28	7.97	0.56	30.6	8.28	7.97	0.56

Table 4.2. Trends in number of days with SWE accumulation and melt exceeding specified thresholds. Note: * is a significance level of p<0.05, and + is p<0.10

station	accumulation threshold [mm]			melt threshold [mm]		
	>10	>5	>0	<-10	<-5	<0
Joe Wright	-1.88	0	0.88*	-1.20	0	0.29*
Lake Irene	-2.50	-3.64	0.67*	0	0.14	0.63*
Willow Park	0	0.40	0.50+	0	0.33*	0.68*
Bear Lake	0.33	-1.43	0.32	0	0.21*	0.61*
Copeland Lake	1.11+	3.33	0.78*	2.58*	0.25+	0.40*
Phantom Valley	0	2.94	0.33+	0.71	0.20+	0.50*
Stillwater Creek	-0.95	0	0.50+	0	-0.13	-0.20
Roach	0	0	0.36+	-0.48	0.12	0.23*
Willow Creek Pass	0	0.83	0.62+	0.77	0.09	0.22+
Deadman Hill	2.73*	1.11	0.13	0	0.14	0.45*
University Camp	-1.43	0.48	0.91*	0	0.06	0.52*
Niwot	-1.43	0.77	0.73*	0	0.06	0.33*
Lake Eldora	0	1.94	0.82*	1.58*	0.32*	0.62*

Table 4.3. Trends in average annual temperature at three weather stations.

station	Trend in average annual temperature, degrees C/decade		Mann-Kendall statistical significance
	Linear	Theil-Sen	
Allenspark	0.17	0.20	none
Grand Lake	0.23	0.28	p<0.01
Loch Vale	0.06	0.00	none

Figures for Chapter 4

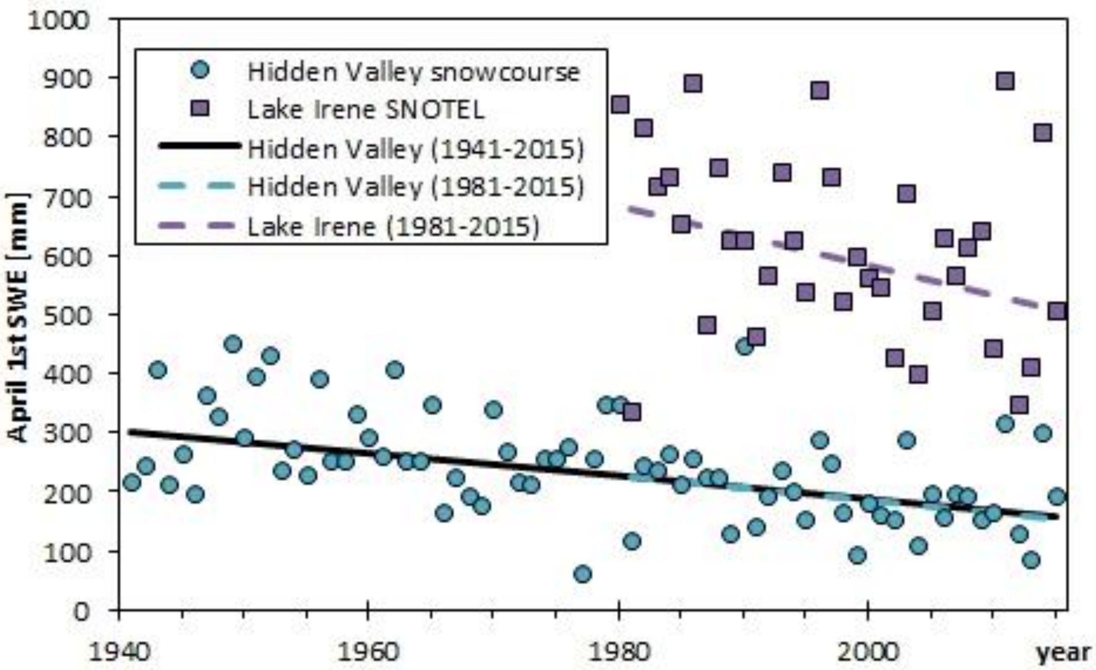

Figure 4.1. Example plot of annual series of April 1 SWE for a snow course and a SNOTEL station, showing strong inter-annual variability. Lines are linear best-fit trend lines.

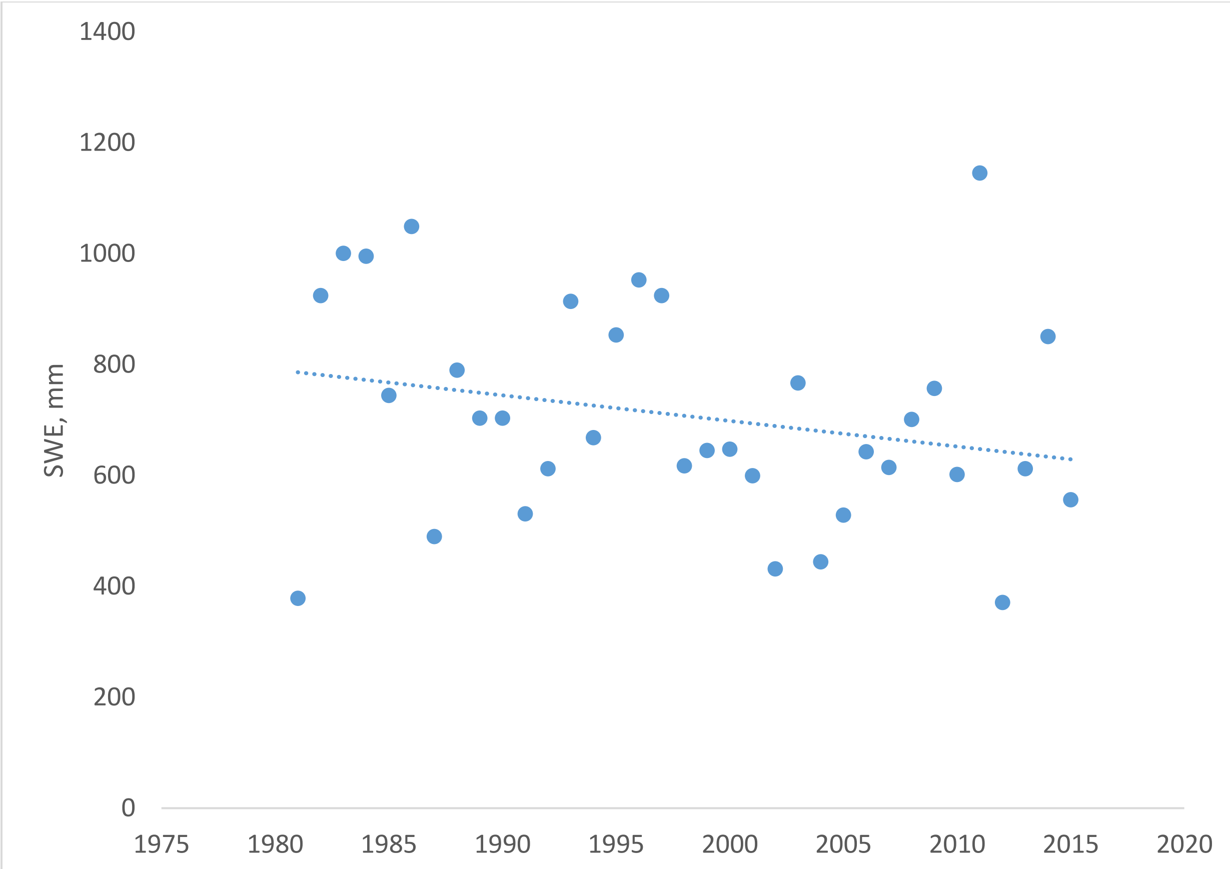

Figure 4.2. Example plot of annual peak SWE at Lake Irene SNOTEL station, 1981-2015, showing strong inter-annual variability. Dashed line is linear best-fit trend line.

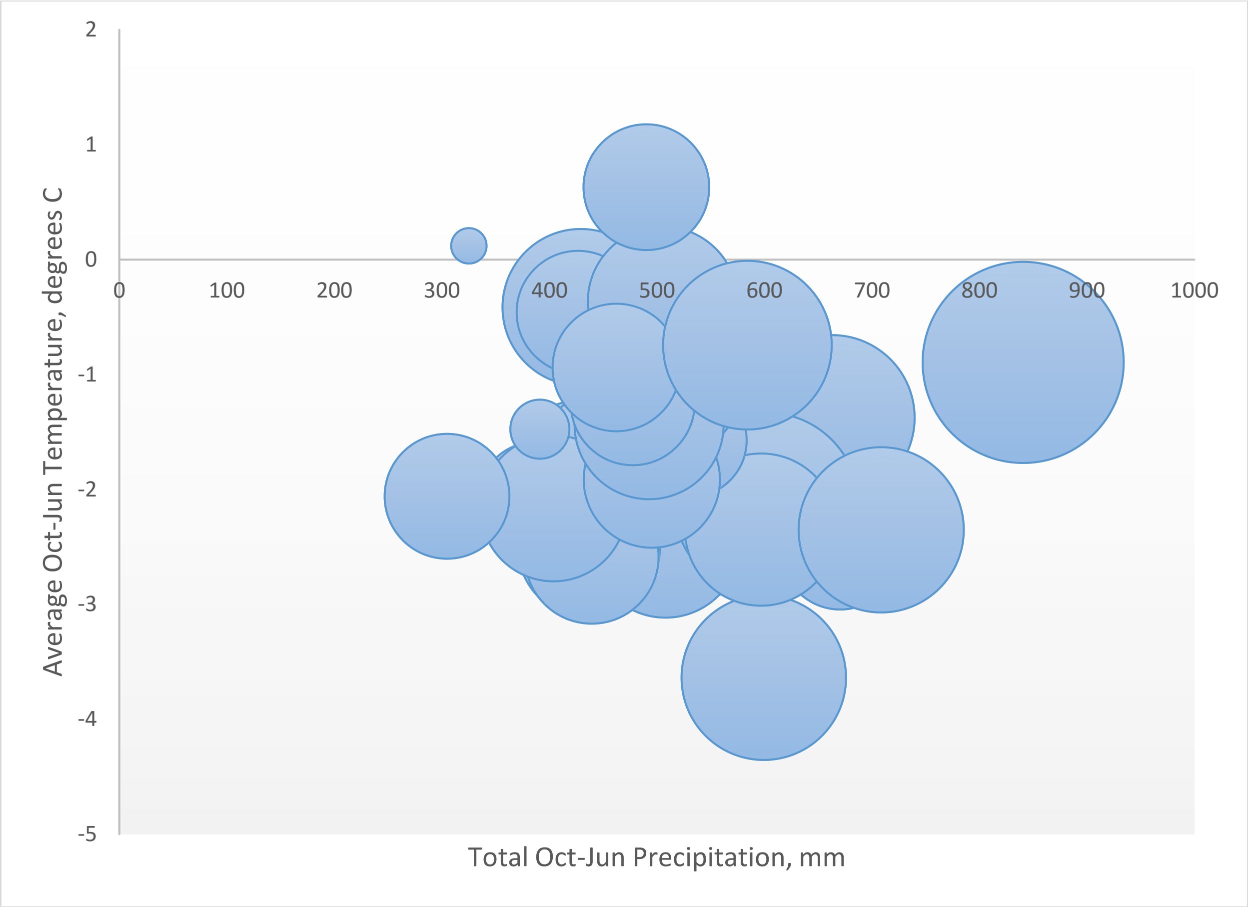

Figure 4.3. April 1 SWE for each snow year, 1986-2015, at Phantom Valley SNOTEL, in relation to annual variations in total October-June precipitation and average October-June temperature as recorded at the SNOTEL station.

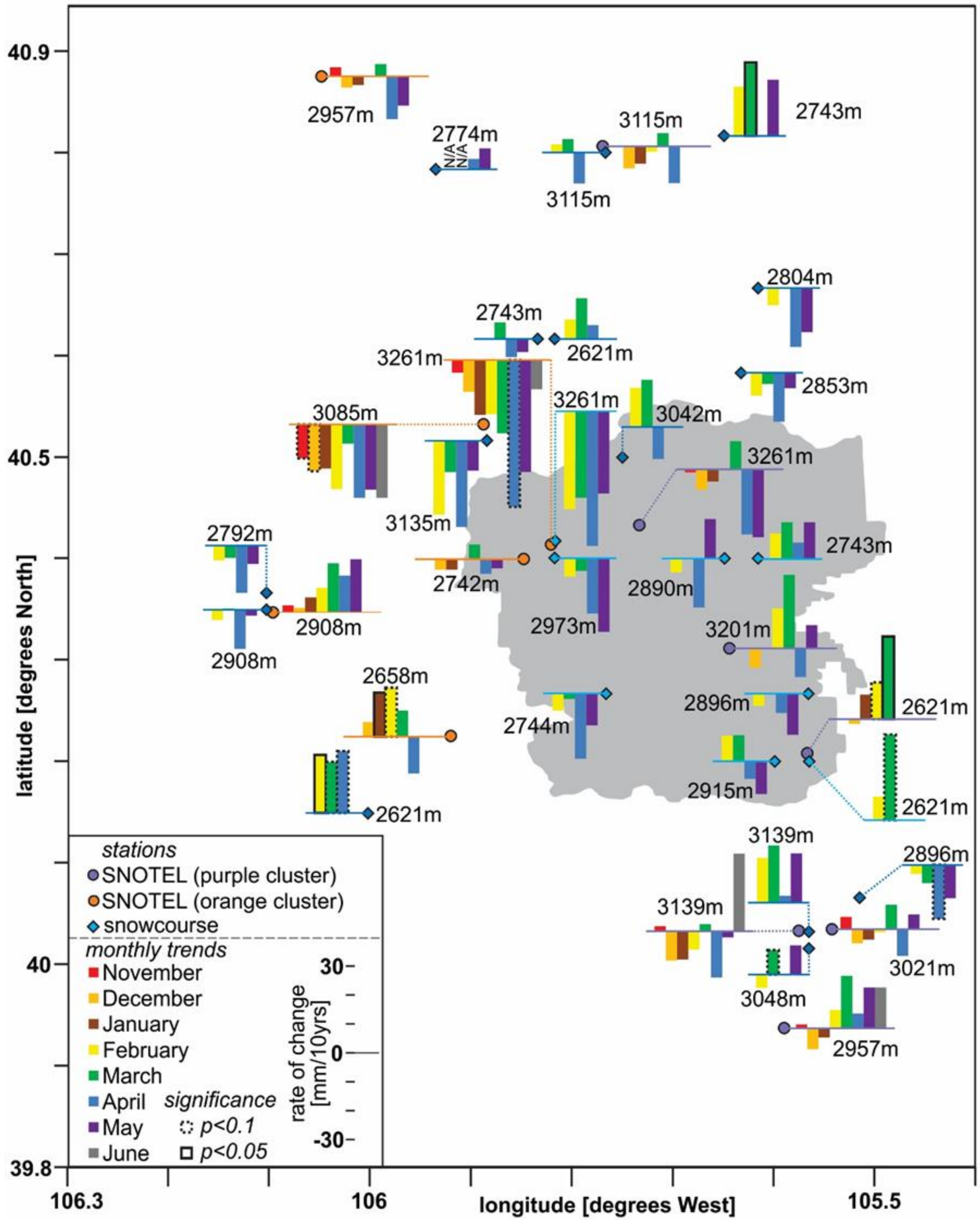

Figure 4.4. Trends in first-of-month SWE at snow courses and SNOTEL stations, 1981-2015. Orange and purple clusters refer to Clow's clusters (2010).

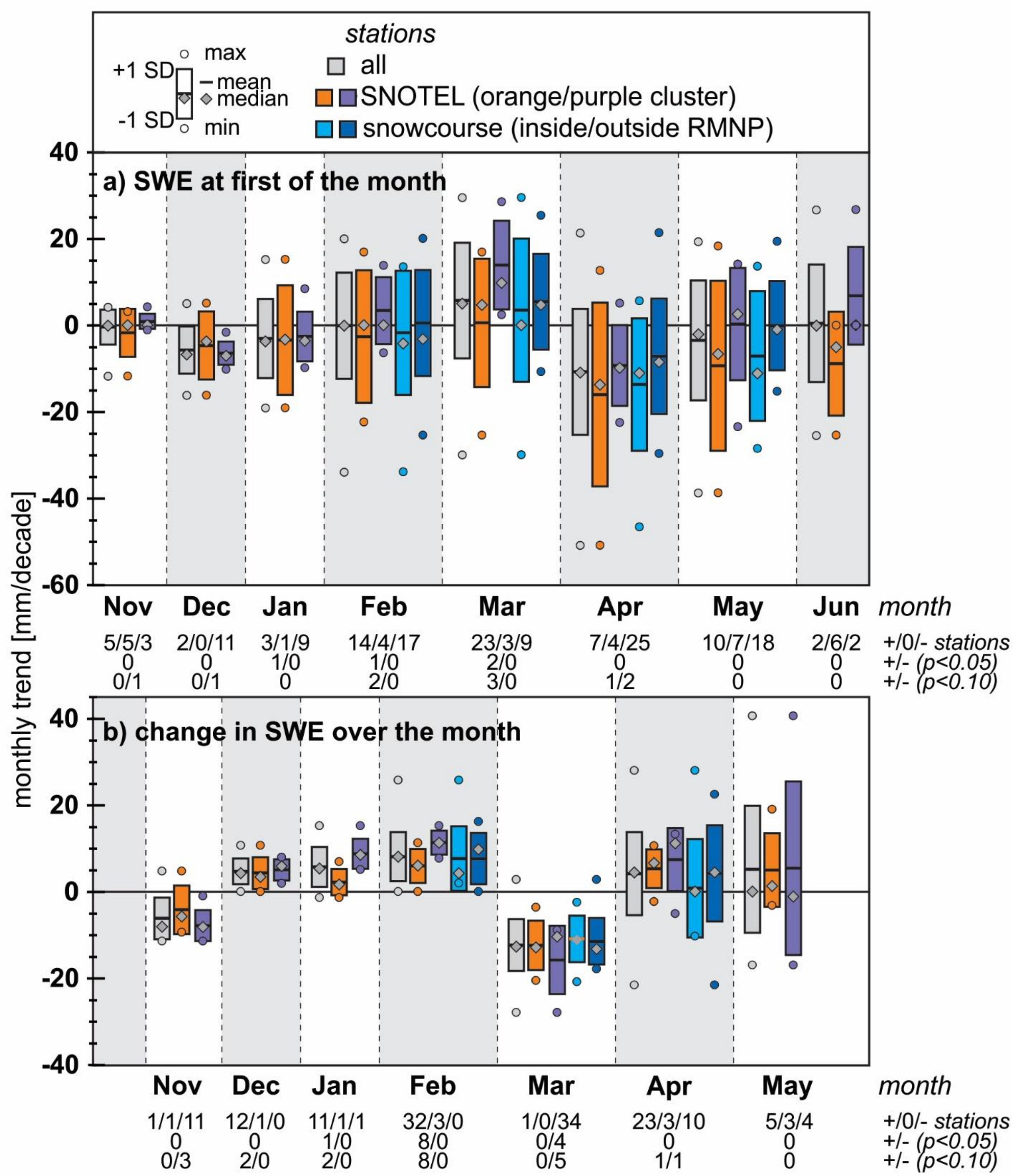

Figure 4.5.a) Distribution of monthly trends in SWE on the first of each month during the snow season for various groupings of snow courses and SNOTEL stations, 1981-2015, and b) distribution of trends in monthly change in SWE. Designation such as 23/3/9 means 23 trends were positive, 3 were level, and 9 were negative. Designation such as 2/0 means 2 positive trends and 0 negative trends were statistically significant.

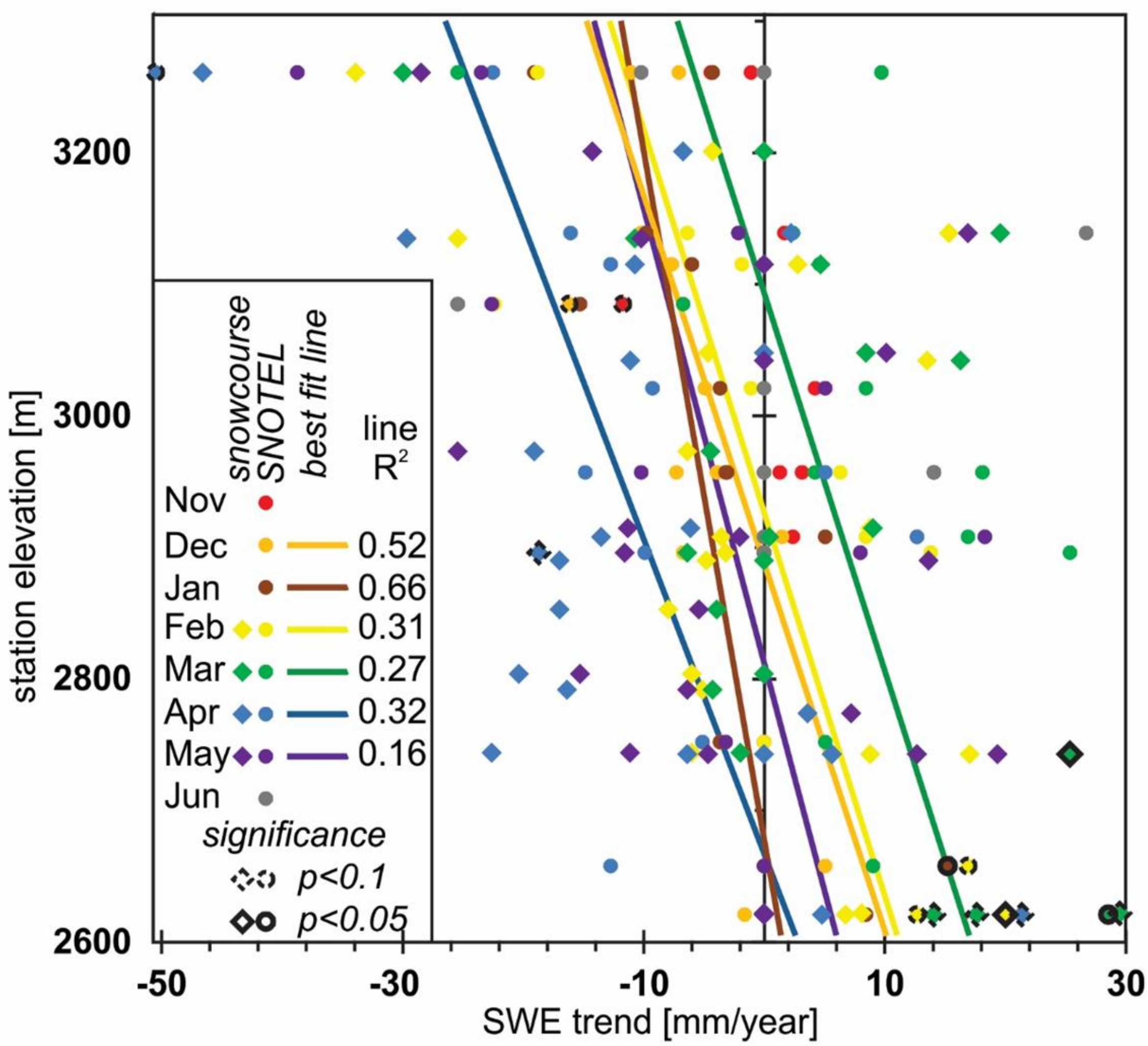

Figure 4.6. Trends in first-of-month SWE at snow courses and SNOTEL stations, 1981-2015, in relation to elevation.

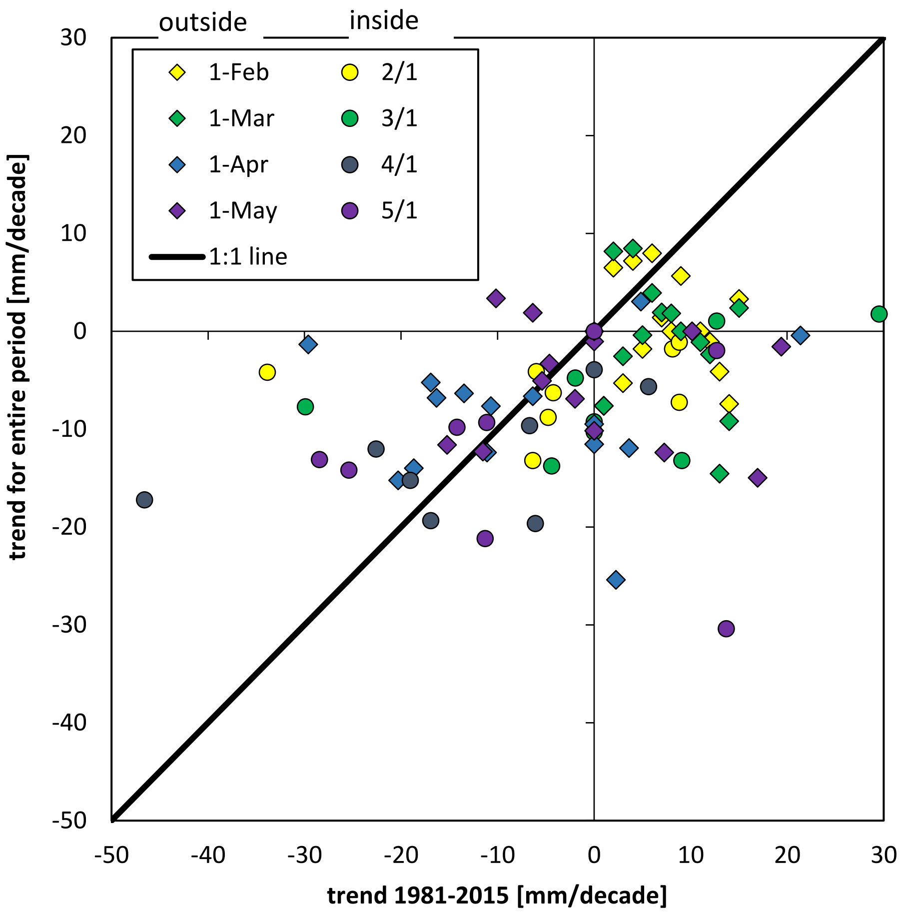

Figure 4.7. Comparison of trends in SWE on first of month at snow courses inside and outside Rocky Mountain National Park for two different periods: 1981-2015, and entire period of record.

Figure 4.8. Trends in monthly change in SWE at snow courses for two periods.

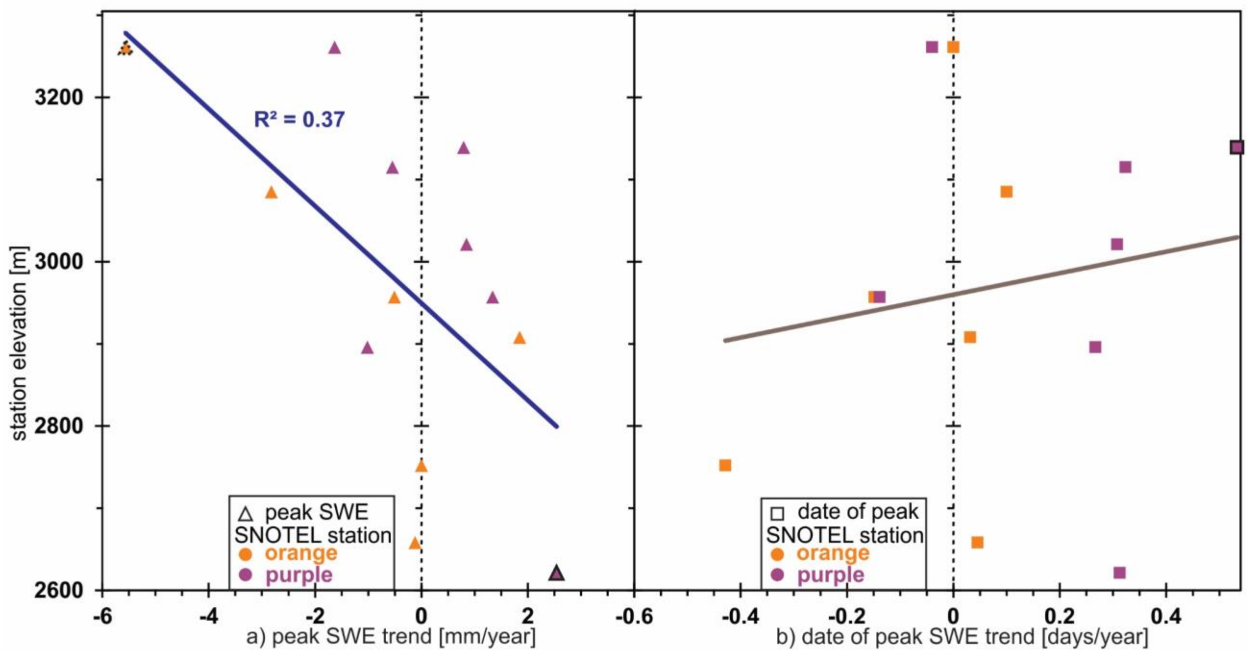

Figure 4.9. Trends in a) annual peak SWE and b) date of peak SWE at SNOTEL stations, 1981-2015, in relation to elevation.

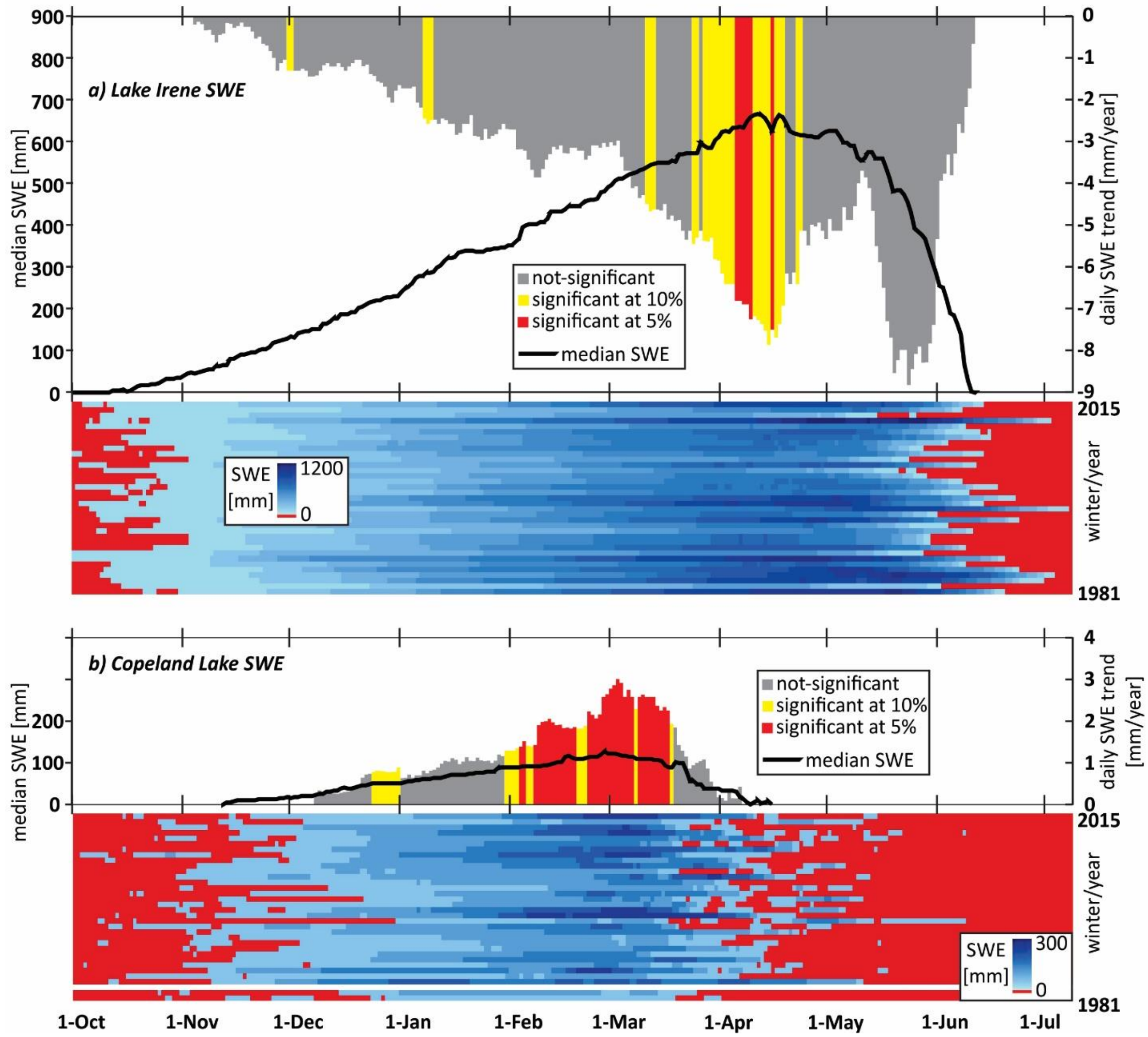

Figure 4.10. Examples of analyses of daily SWE trends for two SNOTEL stations, in relation to median daily SWE and to raster plots of daily SWE for each year, 1981-2015.

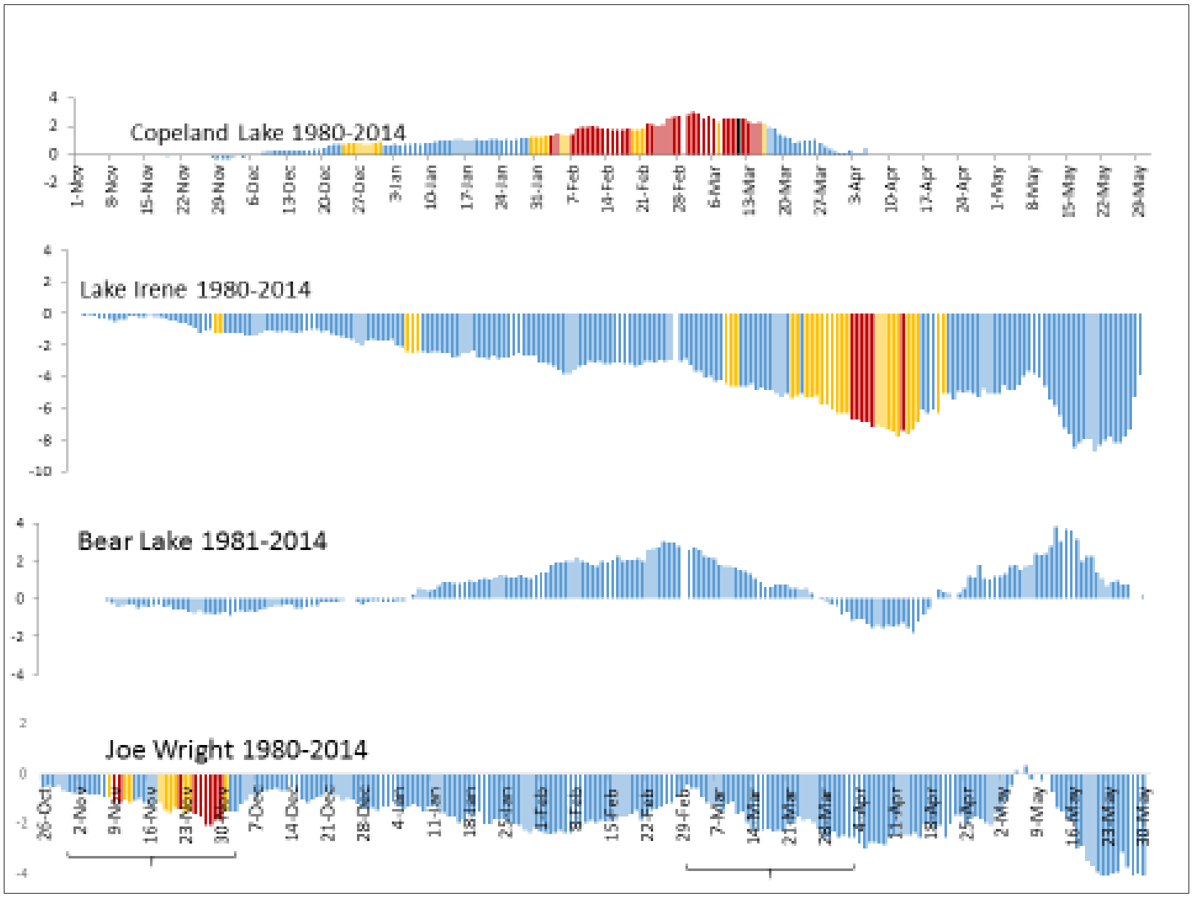

Figure 4.11. Daily SWE trend plots for additional SNOTEL stations.

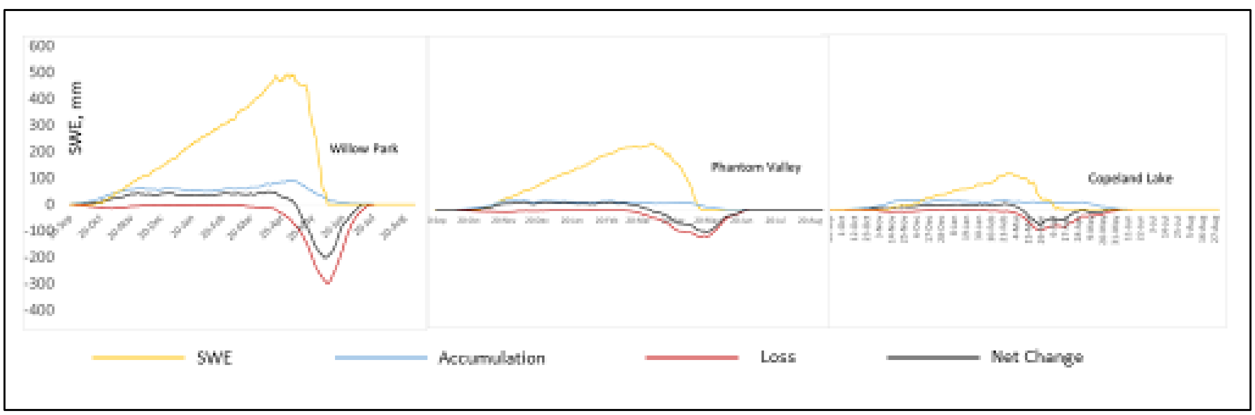

Figure 4.12. Fifteen-day changes in SWE, along with the average niveographs, for three SNOTEL stations.

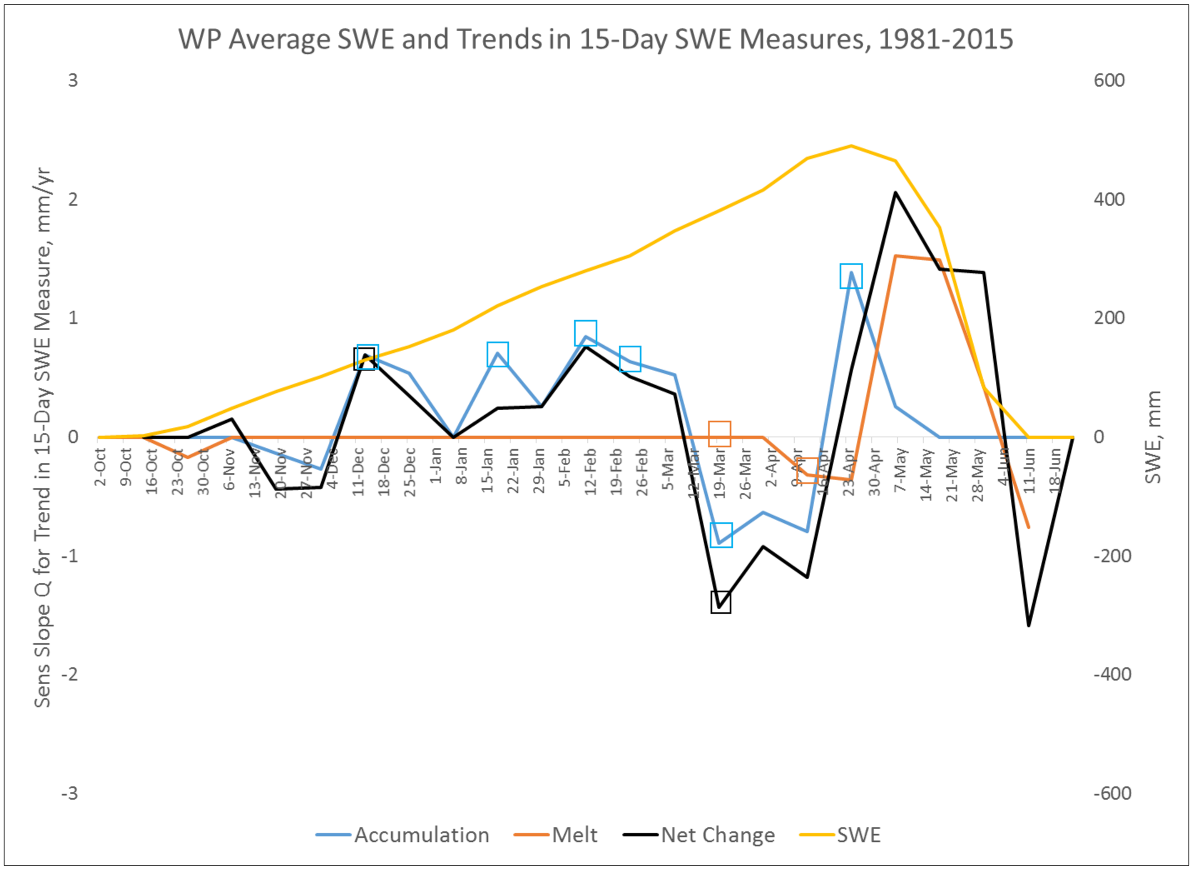

Figure 4.13a). Trends in 15-day changes in SWE measures at Willow Park SNOTEL, 1981-2015, computed at 20-day intervals.

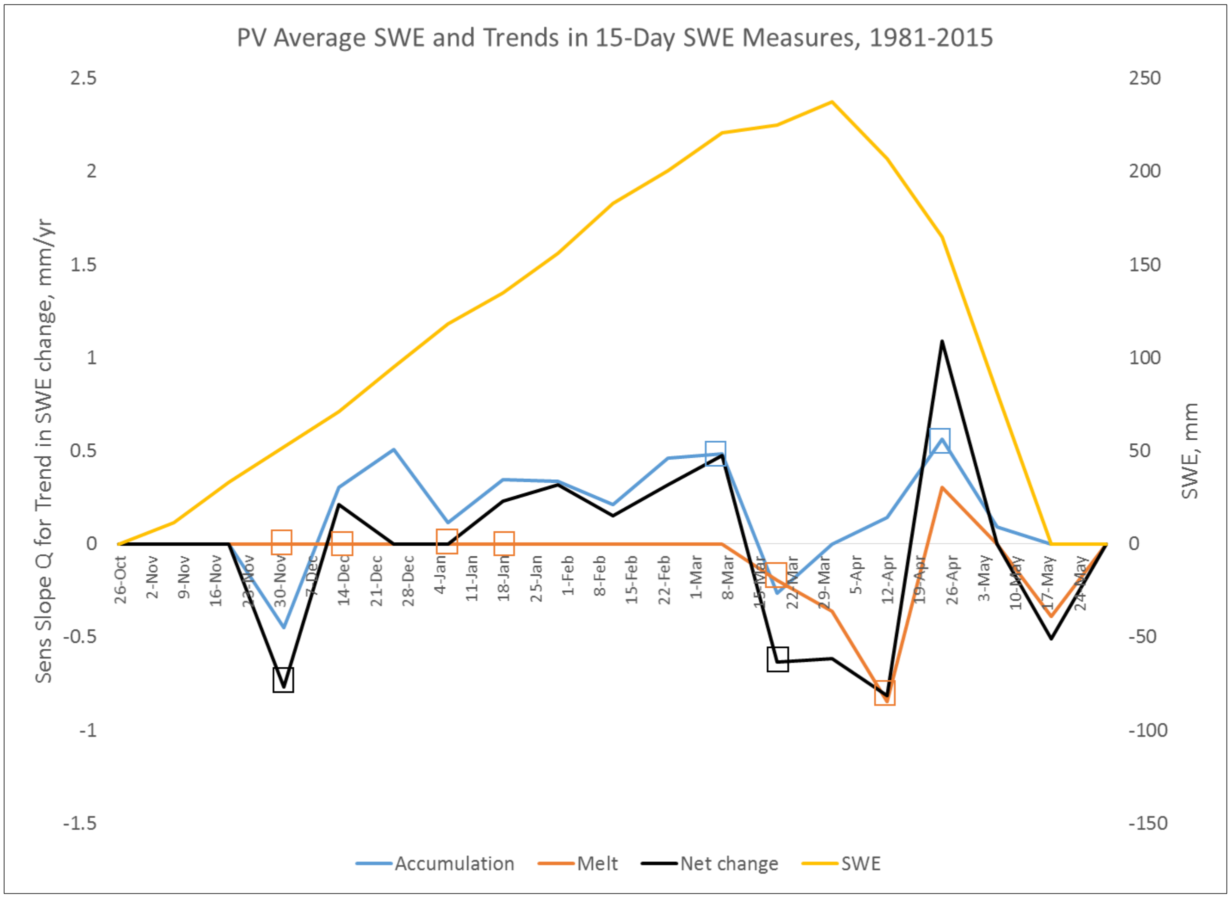

Figure 4.13b). Trends in 15-day changes in SWE measures at Phantom Valley SNOTEL, 1981-2015, computed at 20-day intervals.

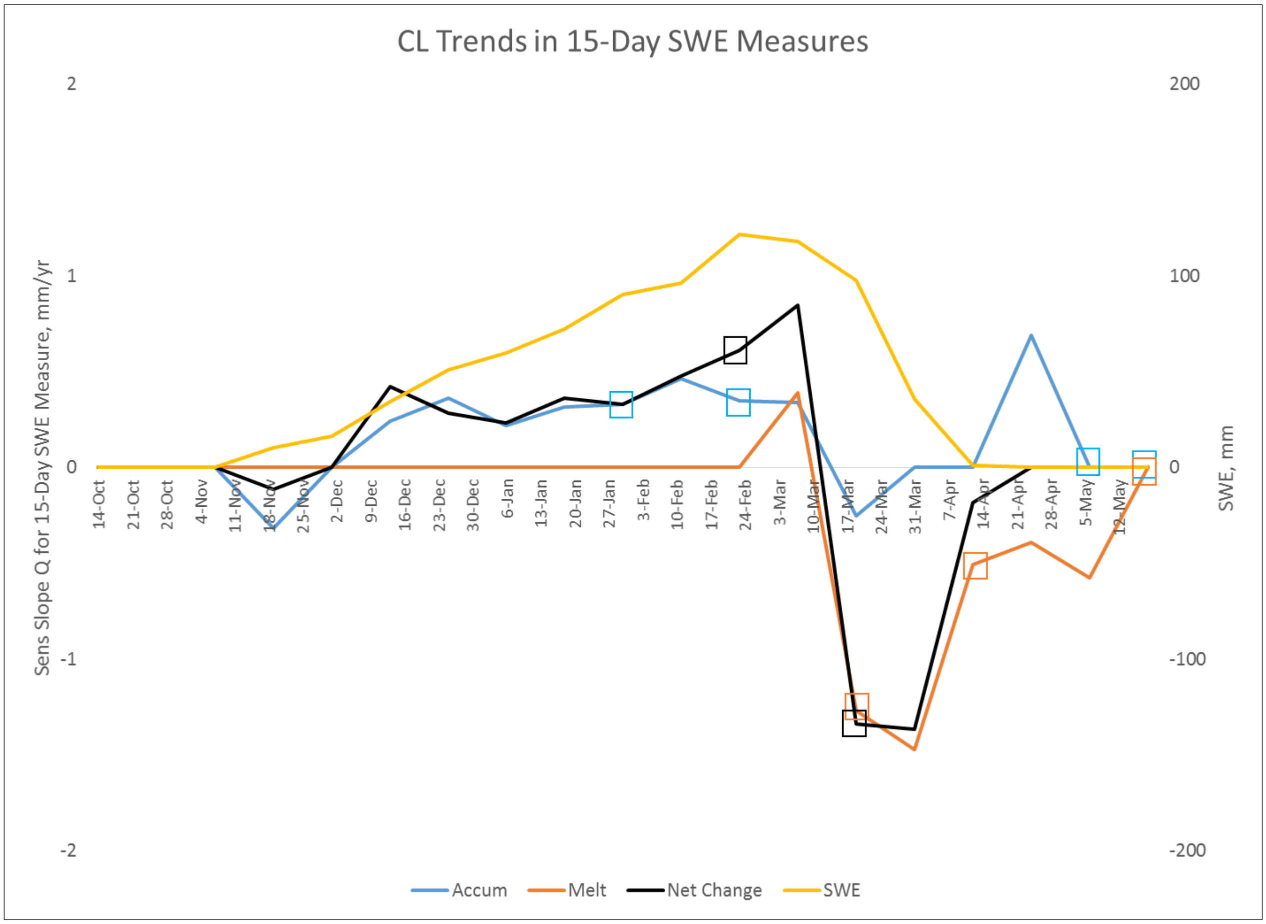

Figure 4.13c). Trends in 15-day changes in SWE measures at Copeland Lake SNOTEL, 1981-2015, computed at 20-day intervals.

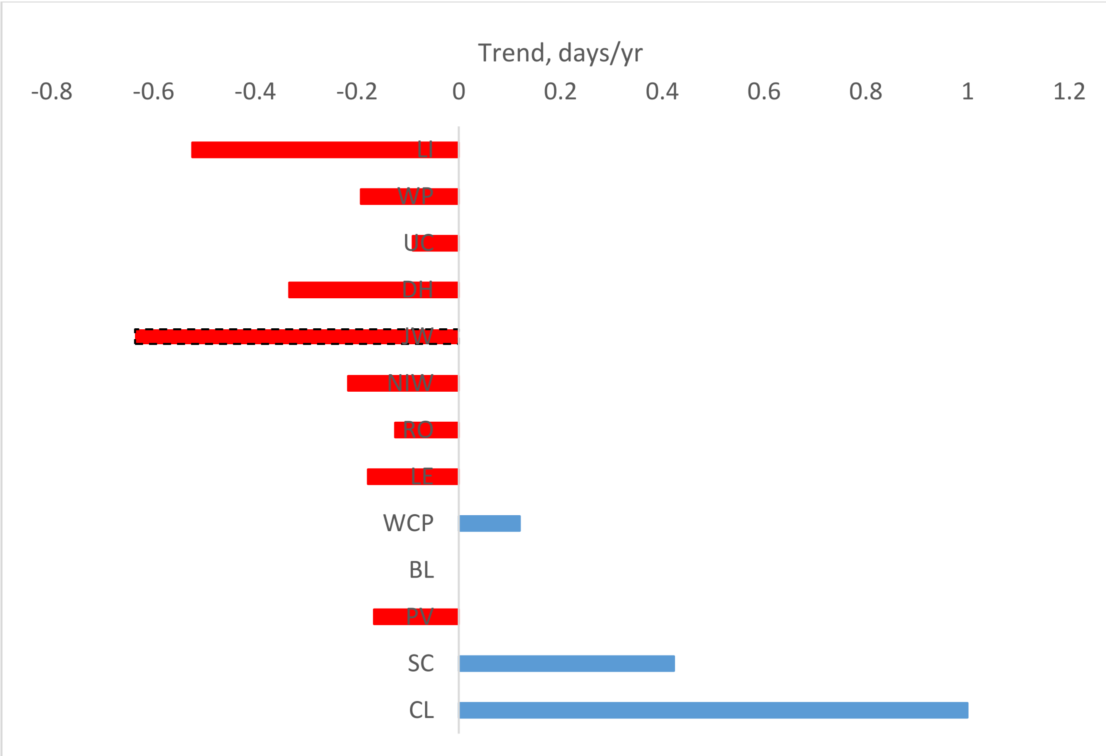

Figure 4.14 Trends in number of days per year with over 100 mm SWE at SNOTEL stations, 1981-2015.

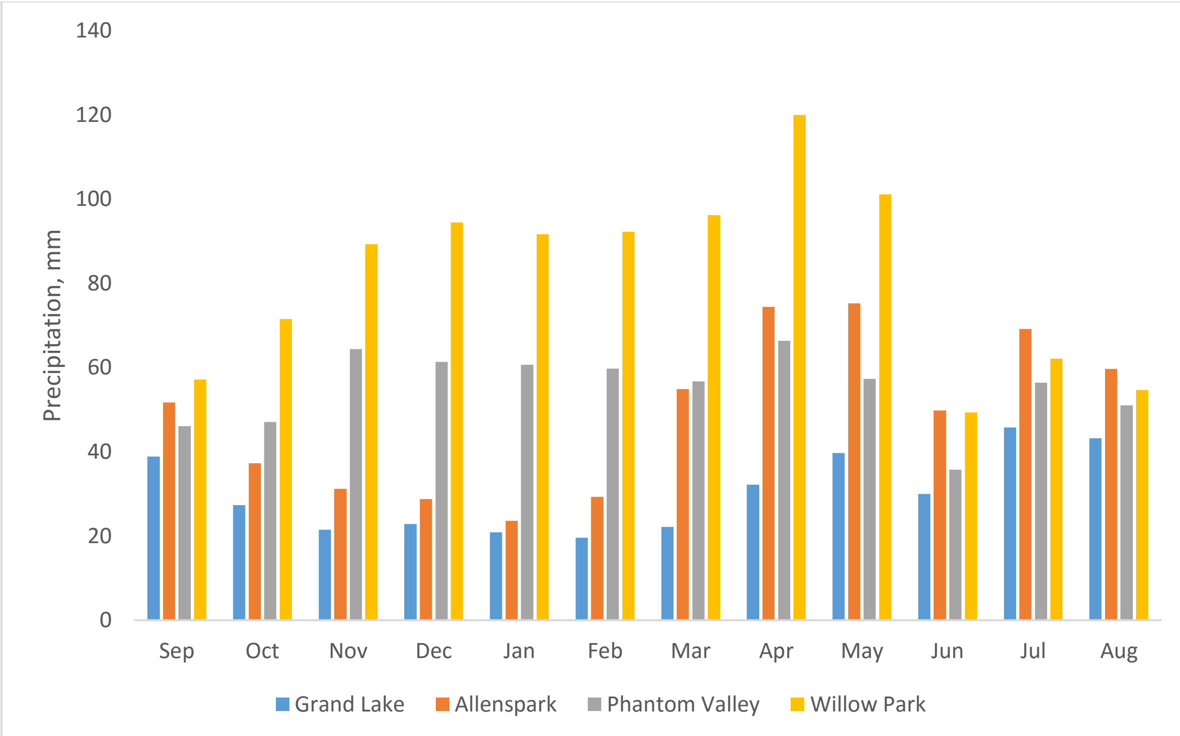

Figure 4.15. Average monthly precipitation at four sites, 1981-2015.

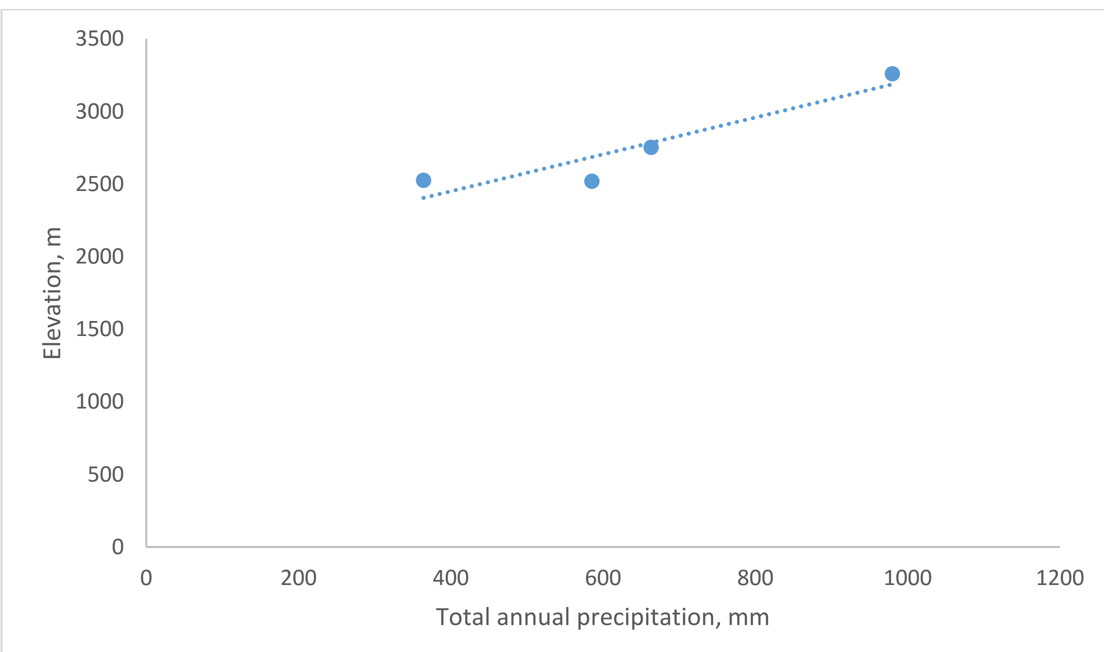

Figure 4.16. Average annual precipitation at four sites in relation to elevation.

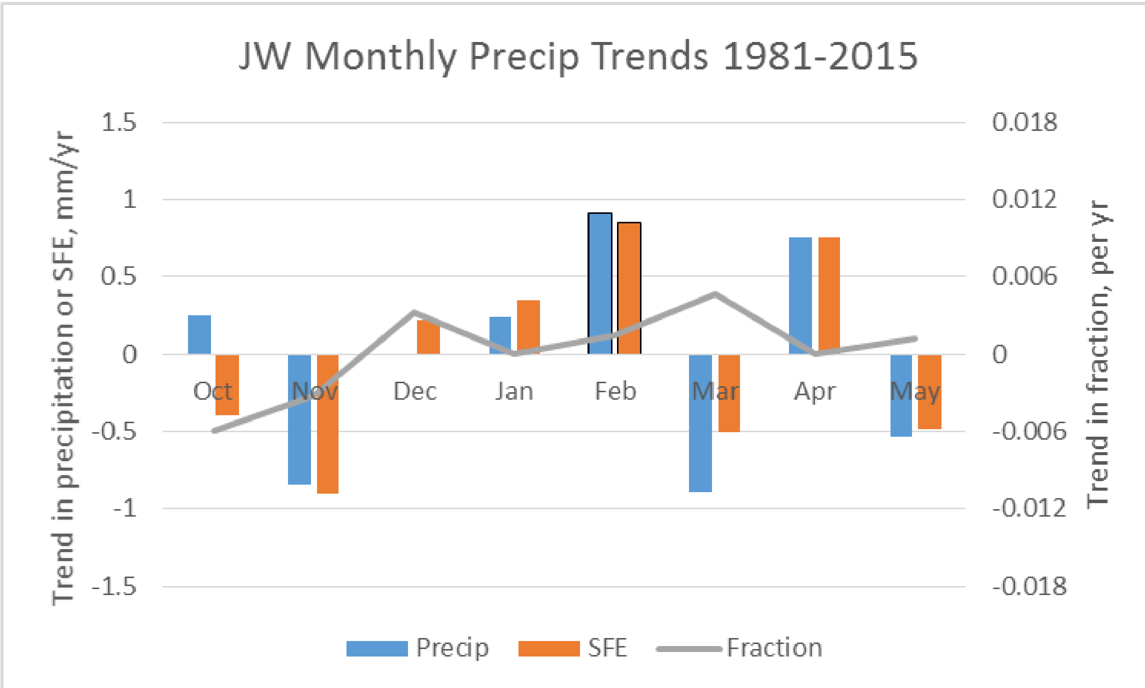

Figure 4.17(a). Monthly precipitation trends for Joe Wright SNOTEL, 1981-2015.

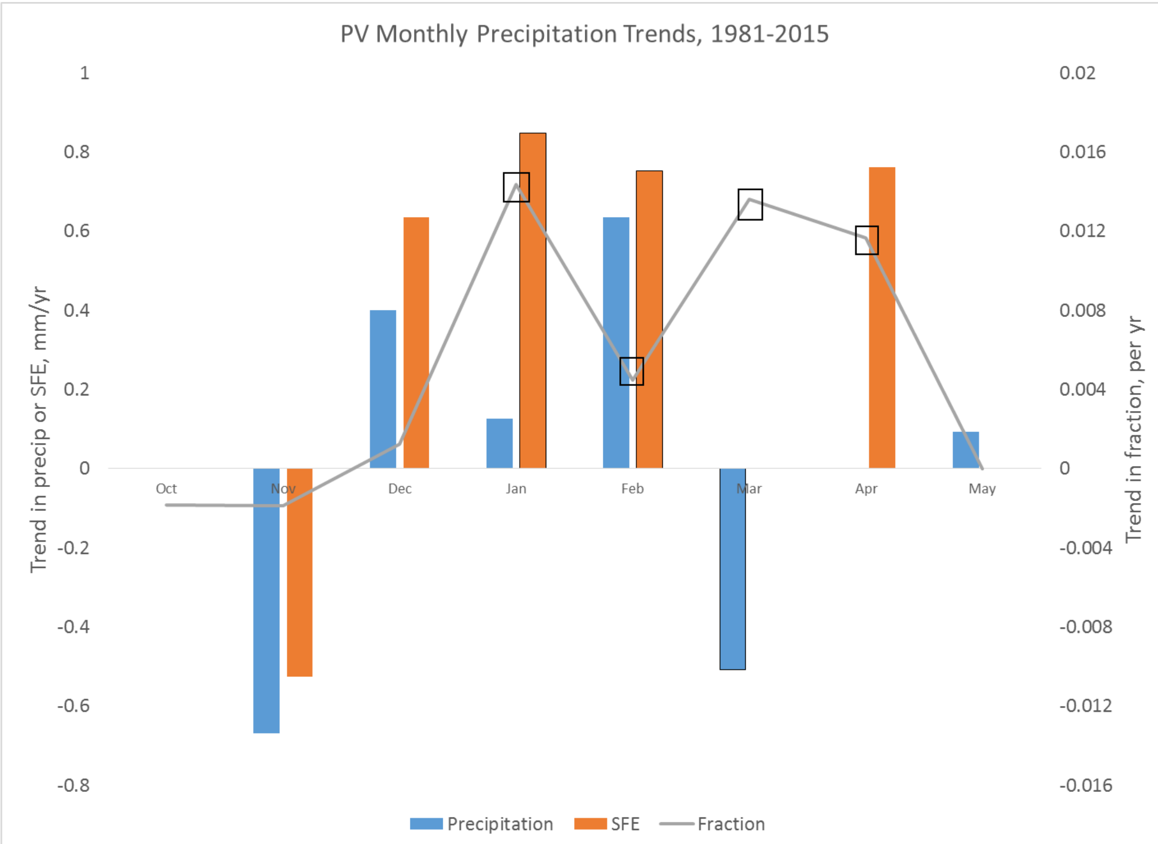

Figure 4.17(b). Monthly precipitation trends for Phantom Valley SNOTEL, 1981-2015.

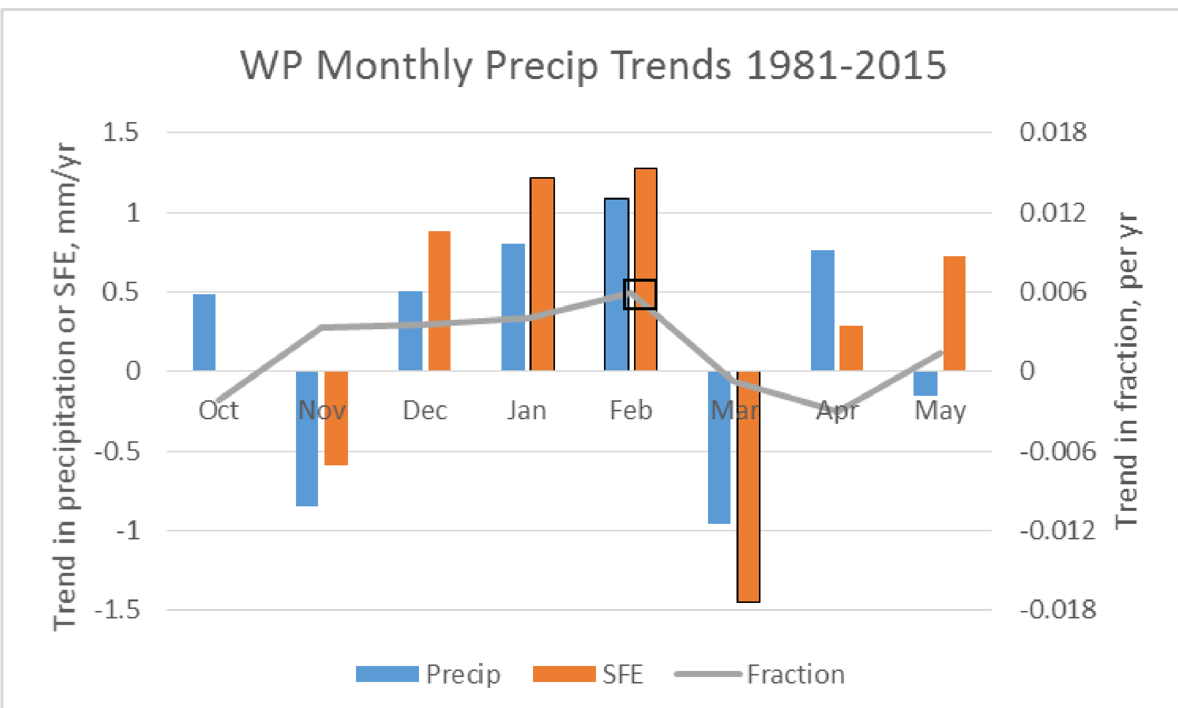

Figure 4.17(c). Monthly precipitation trends for Willow Park SNOTEL, 1981-2015.

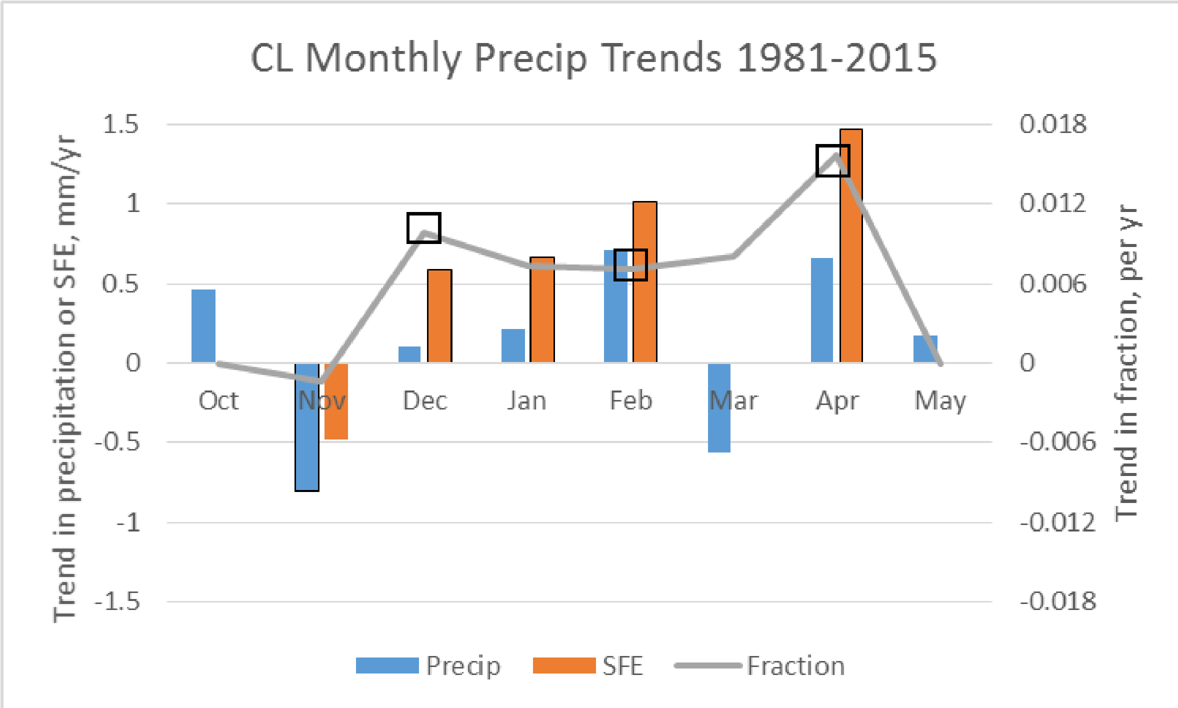

Figure 4.17(d). Monthly precipitation trends for Copeland Lake SNOTEL, 1981-2015.

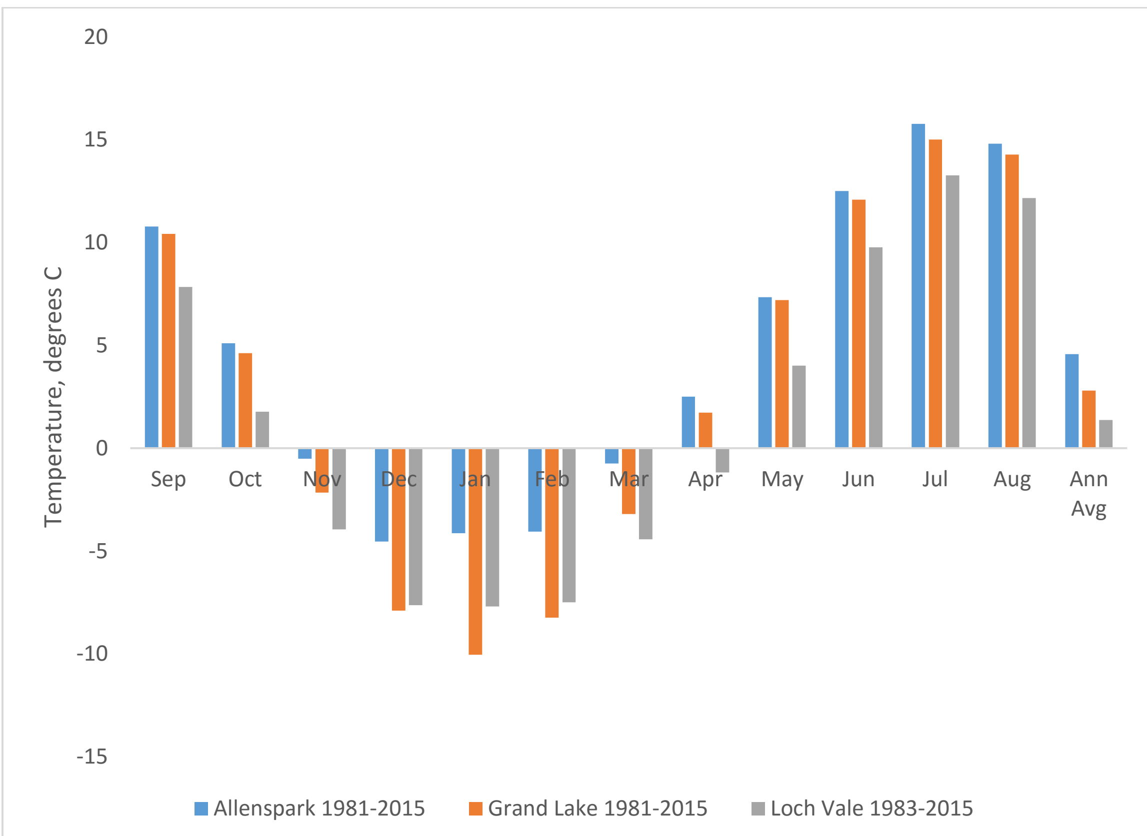

Figure 4.18. Average monthly temperatures at three sites, 1983-2015.

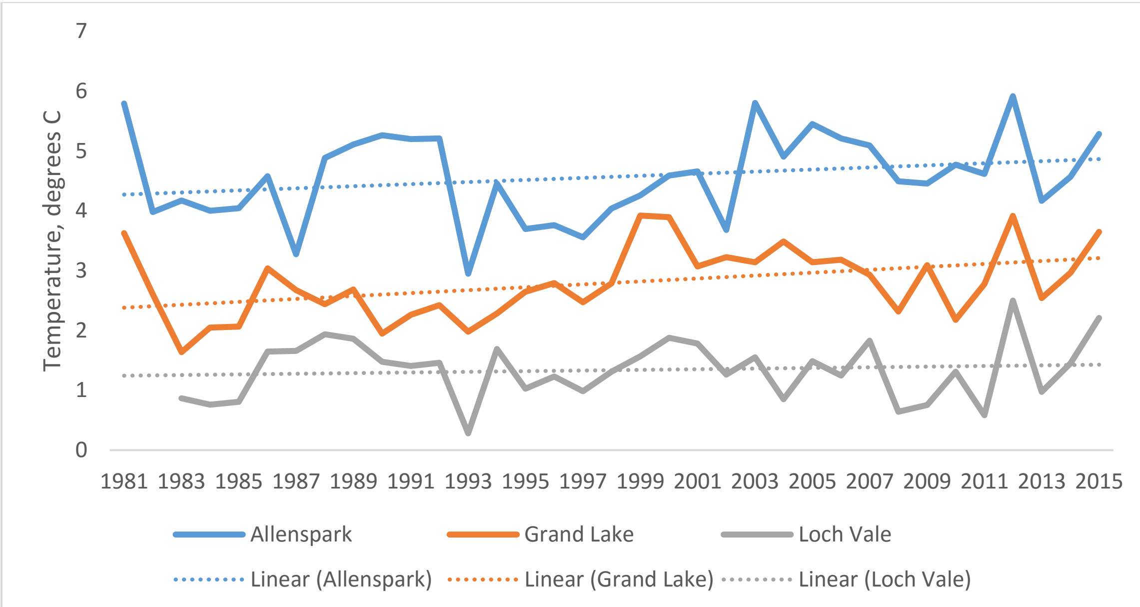

Figure 4.19. Average annual temperatures at three sites, 1981-2015.

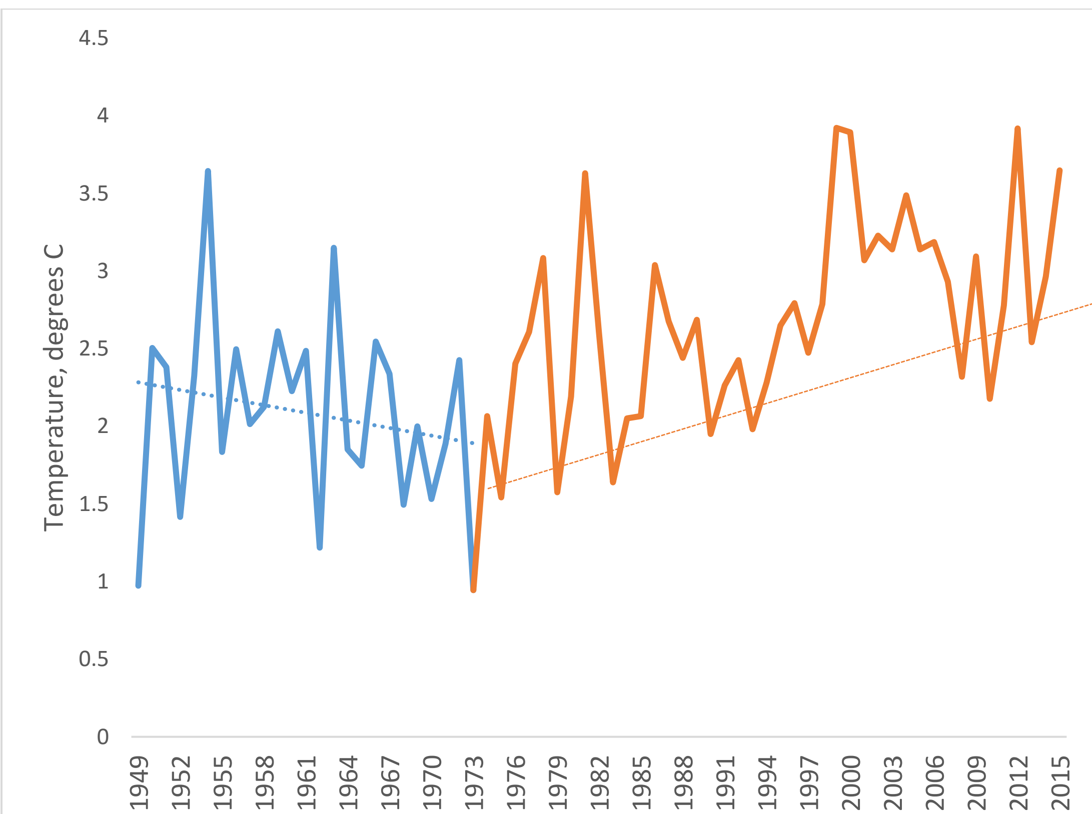

Figure 4.20. Average annual temperature at Grand Lake, 1949-2015.

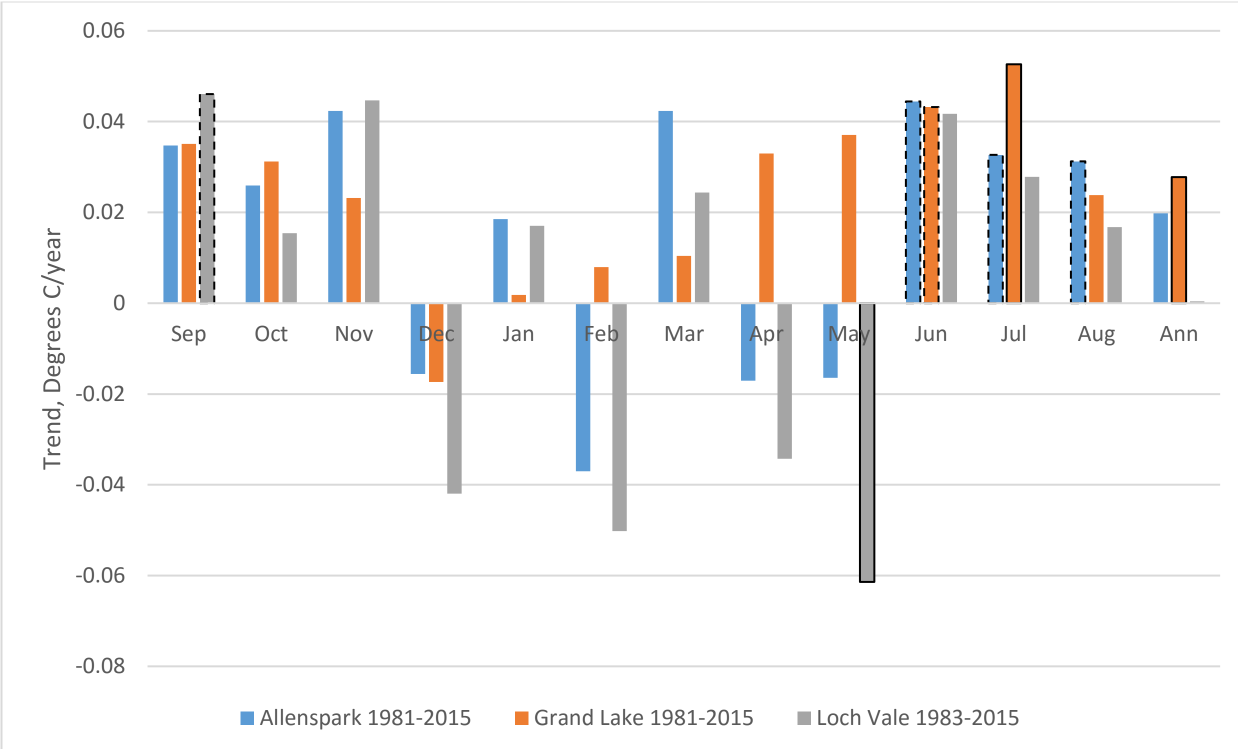

Figure 4.21. Trends in monthly average temperature at three sites, 1981-2015.

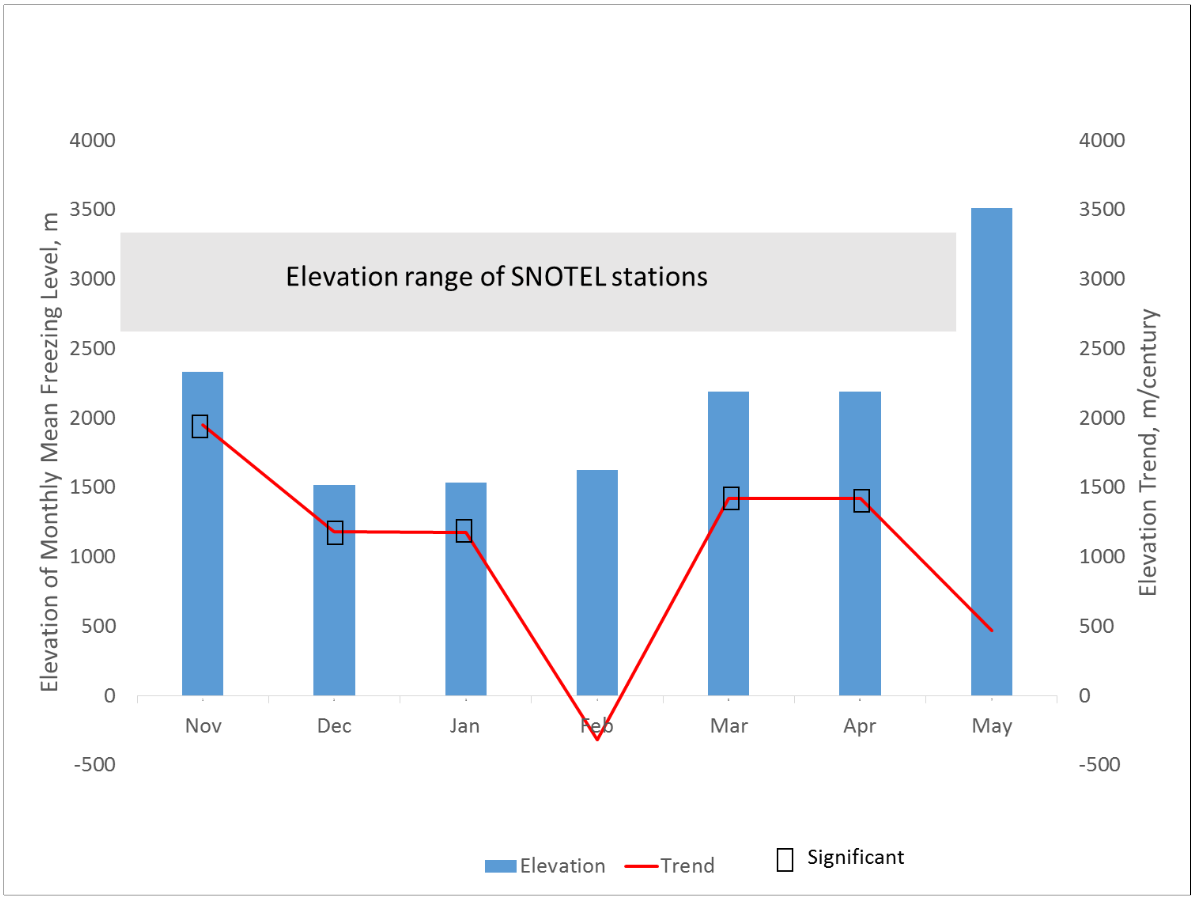

Figure 4.22. Monthly average elevation of freezing level, and monthly trend in freezing level elevation, 1981-2015.

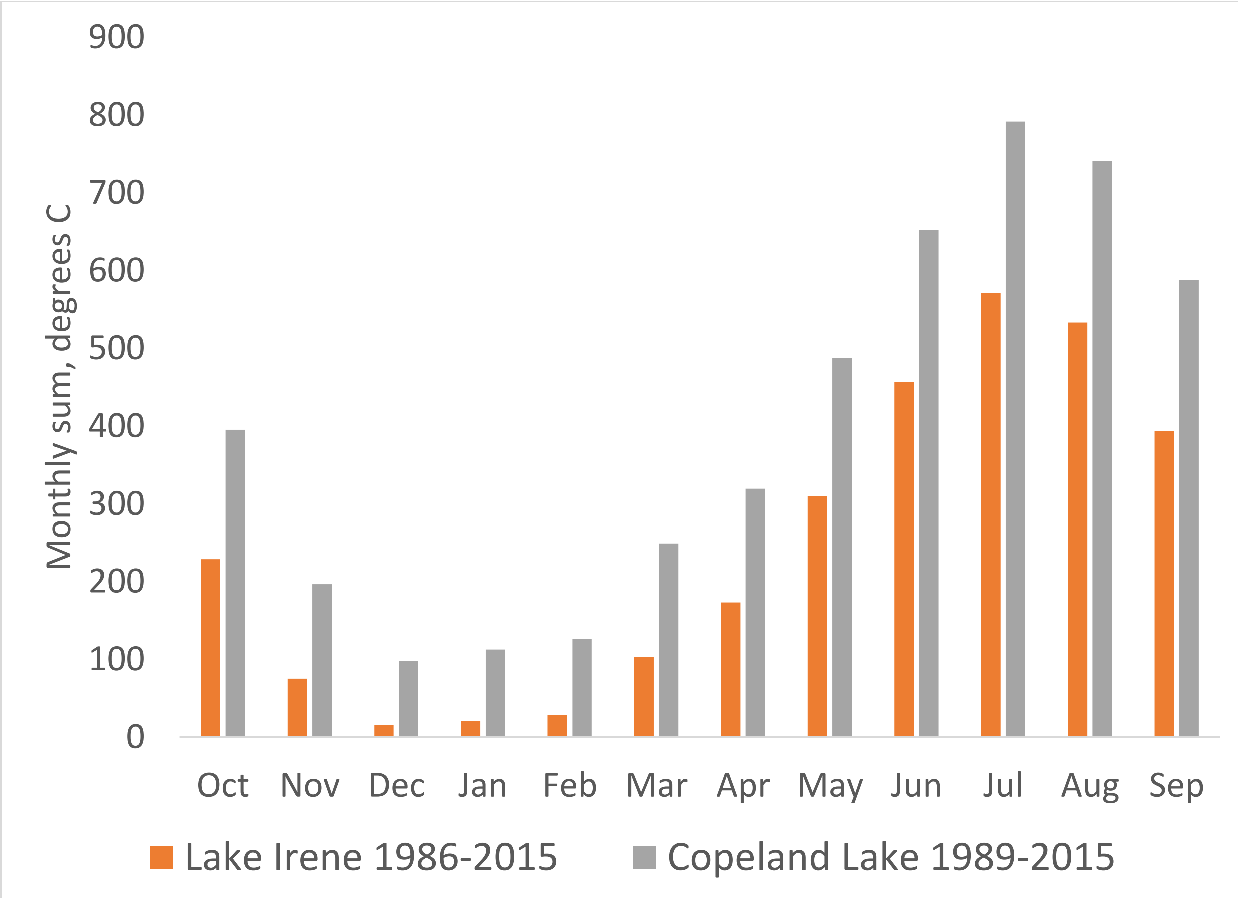

Figure 4.23. Monthly sum of maximum daily temperatures warmer than zero for Lake Irene and Copeland Lake.

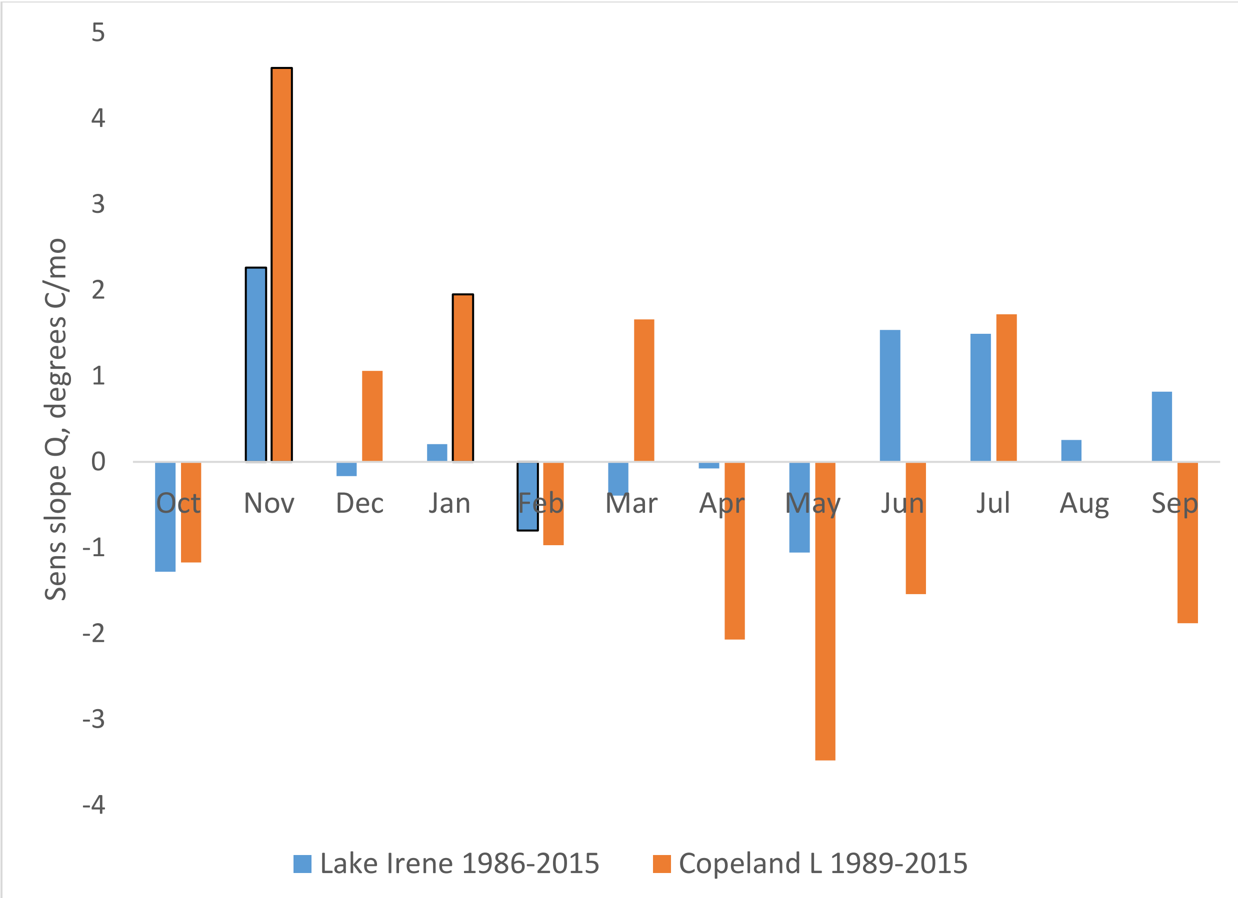

Figure 4.24. Trends in monthly sum of daily maximum temperatures warmer than zero for Lake Irene and Copeland Lake.

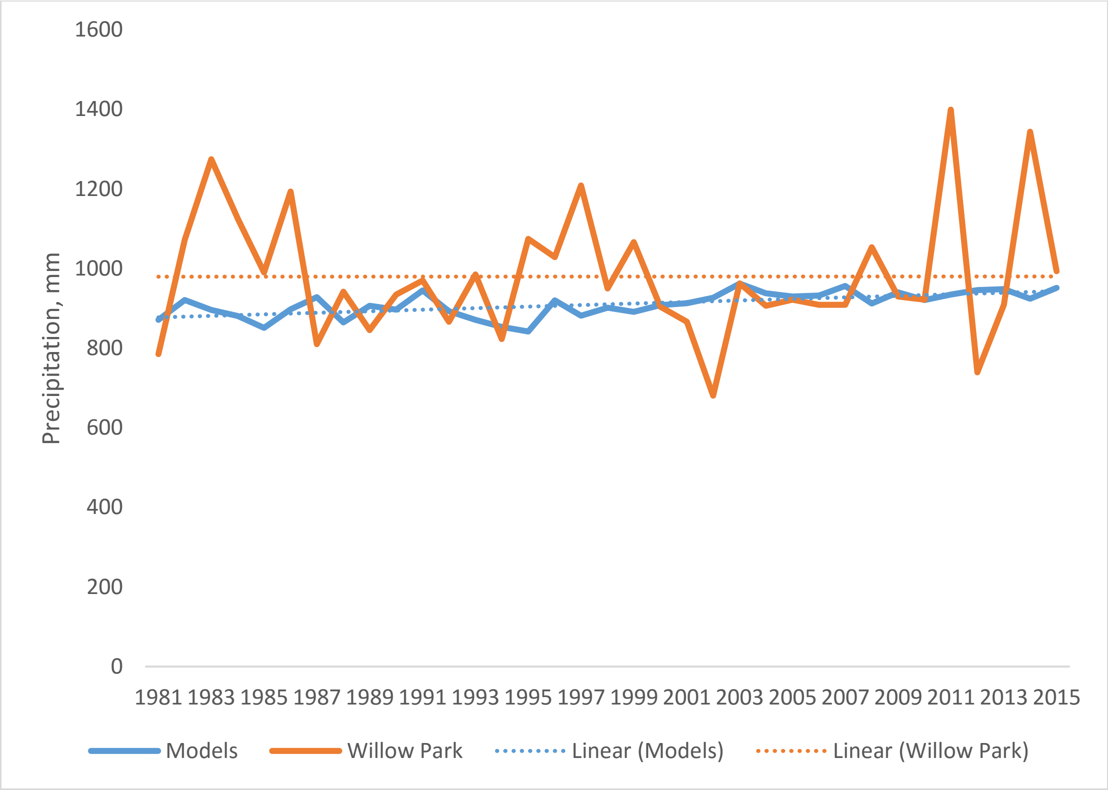

Figure 4.25. Total annual precipitation simulated by the CMIP5 model ensemble, and observed at Willow Park SNOTEL, 1981-2015.

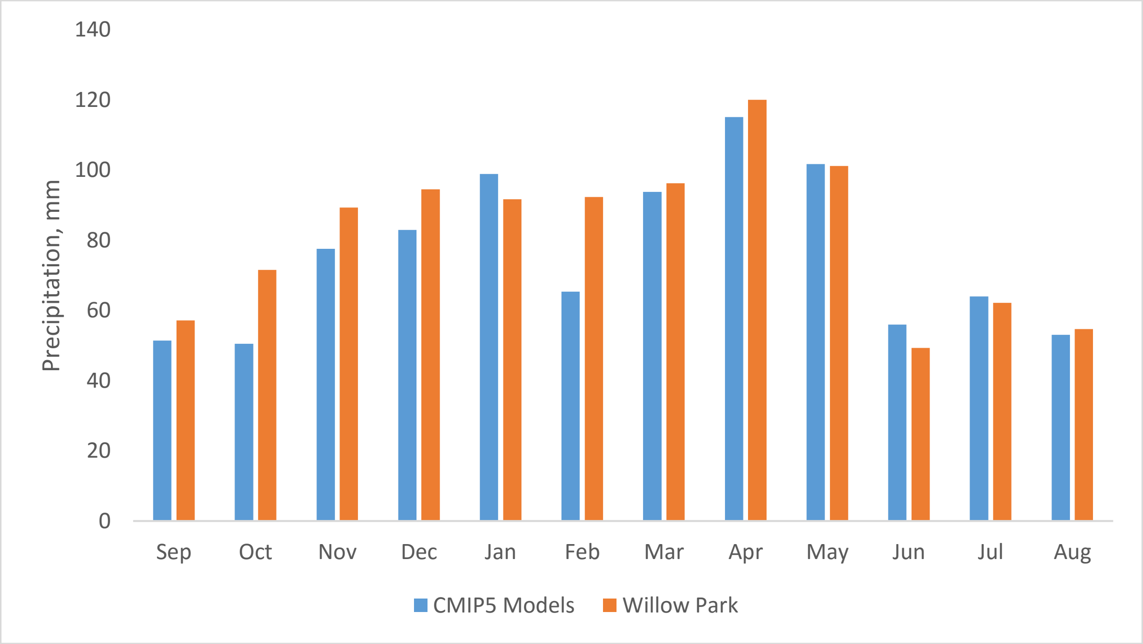

Figure 4.26. Average monthly precipitation simulated by the CMIP5 model ensemble, and observed at Willow Park SNOTEL, 1981-2015.

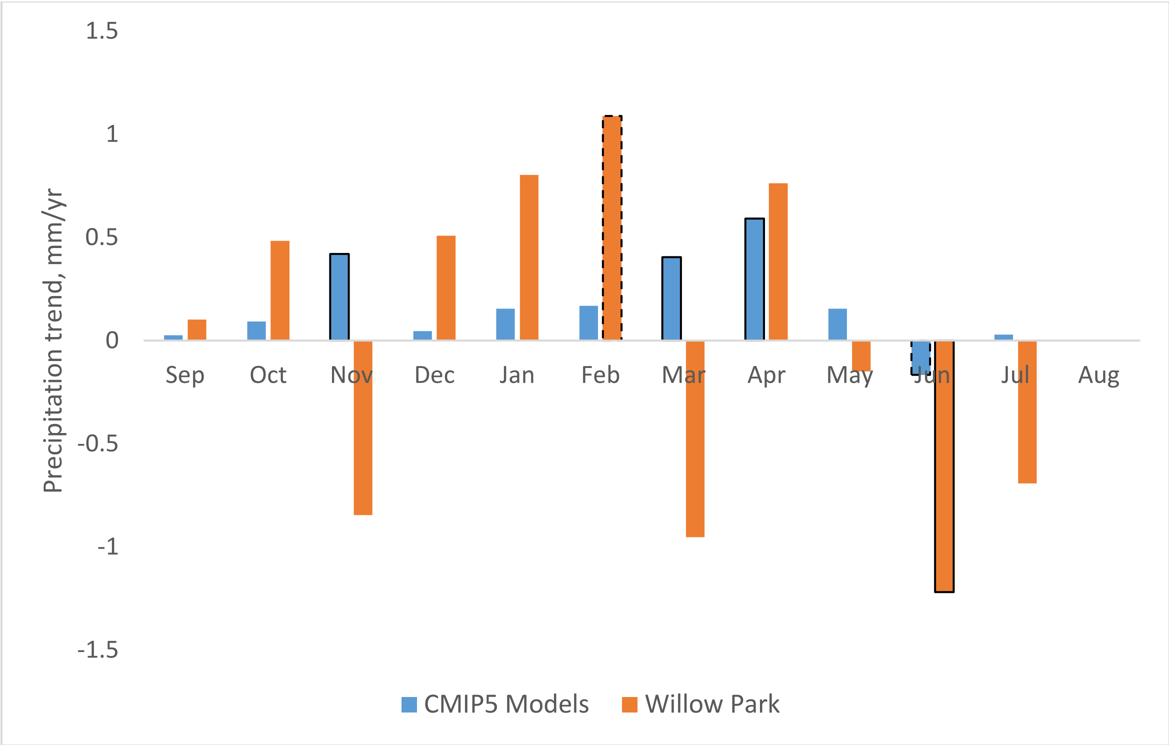

Figure 4.27. Trends in monthly precipitation, 1981-2015, simulated by the CMIP5 model ensemble and observed at Willow Park.

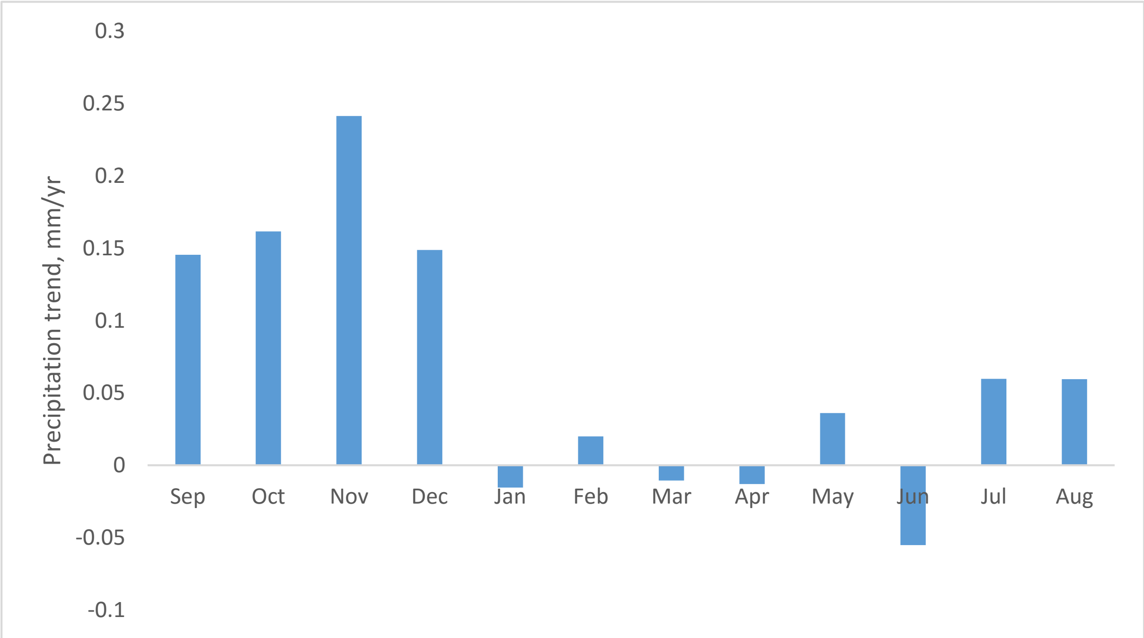

Figure 4.28. Trends in monthly precipitation, 2015-2099, simulated by the CMIP5 model ensemble.

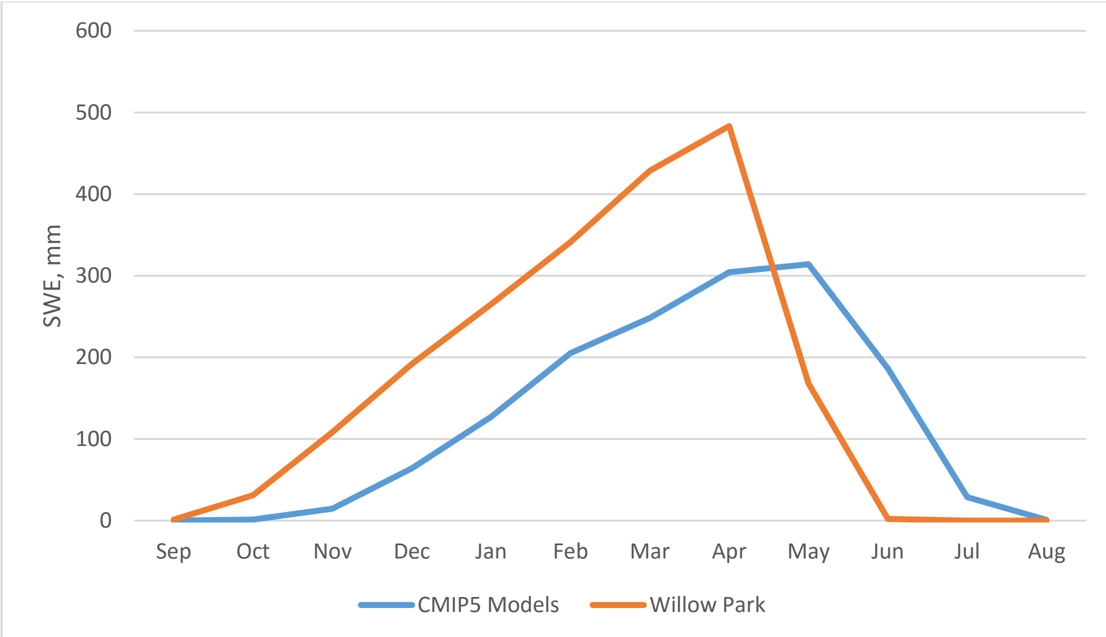

Figure 4.29. Average monthly SWE, 1981-2015, simulated by the CMIP5 model ensemble and observed at Willow Park.

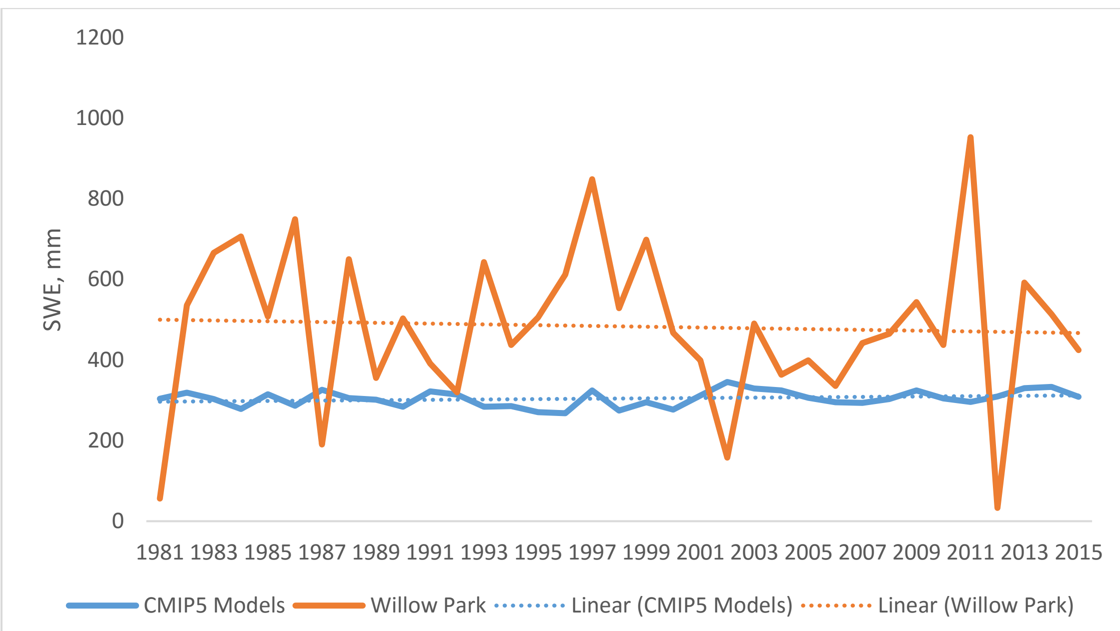

Figure 4.30. April 1 SWE, 1981-2015, simulated by the CMIP5 model ensemble and observed at Willow Park.

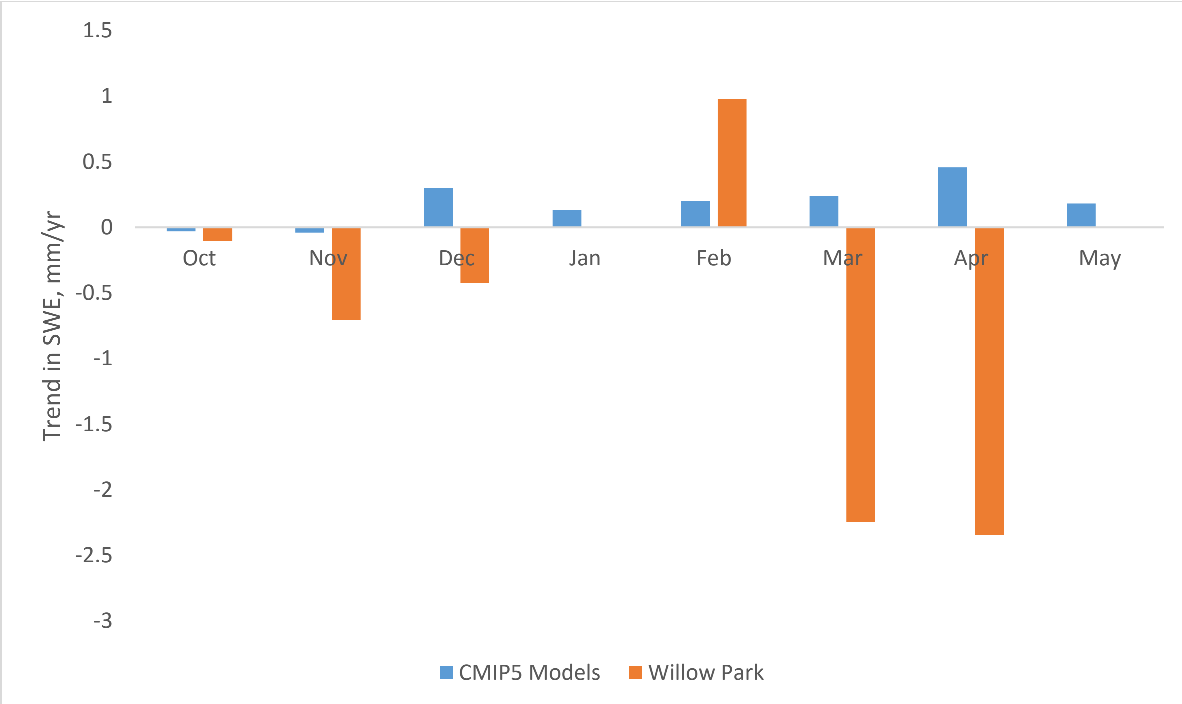

Figure 4.31. Trends in monthly SWE, 1981-2015, simulated by CMIP5 model ensemble and observed at Willow Park.

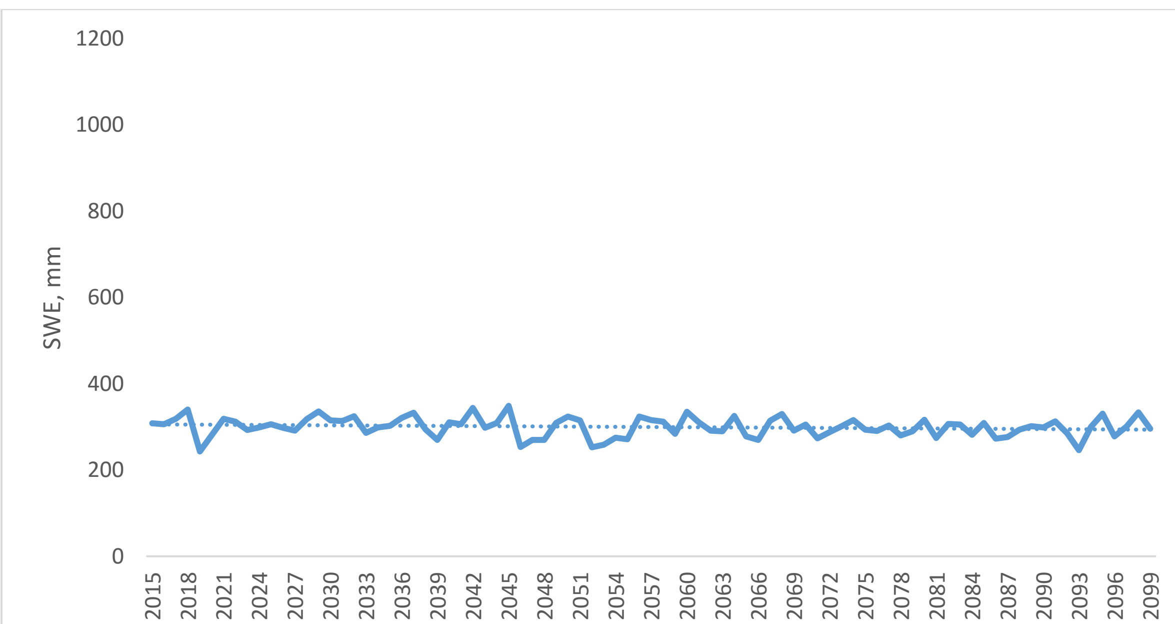

Figure 4.32. Projected April 1 SWE, 2015-2099, simulated by CMIP5 model ensemble. Y-axis has same scale as in Figure 4.30.

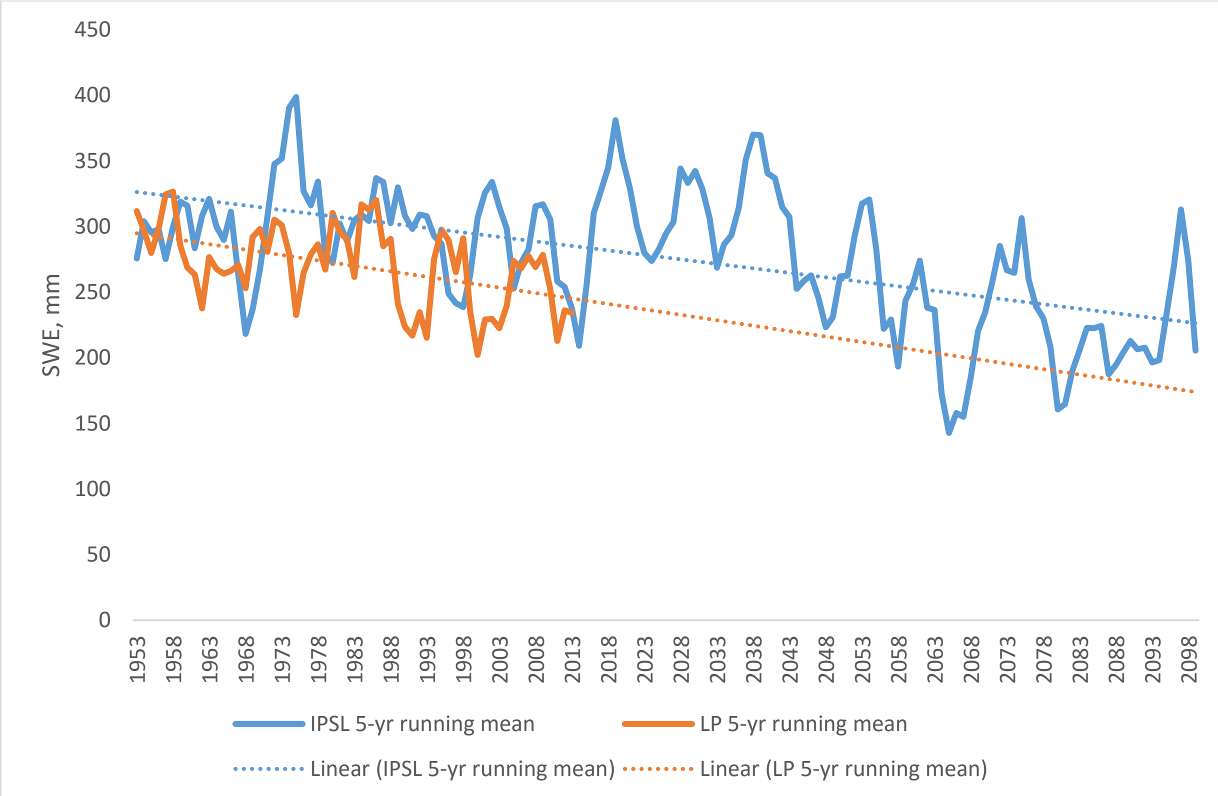

Figure 4.33. Comparison of smoothed trends in April 1 SWE between observed Longs Peak snow course data and IPSL model.

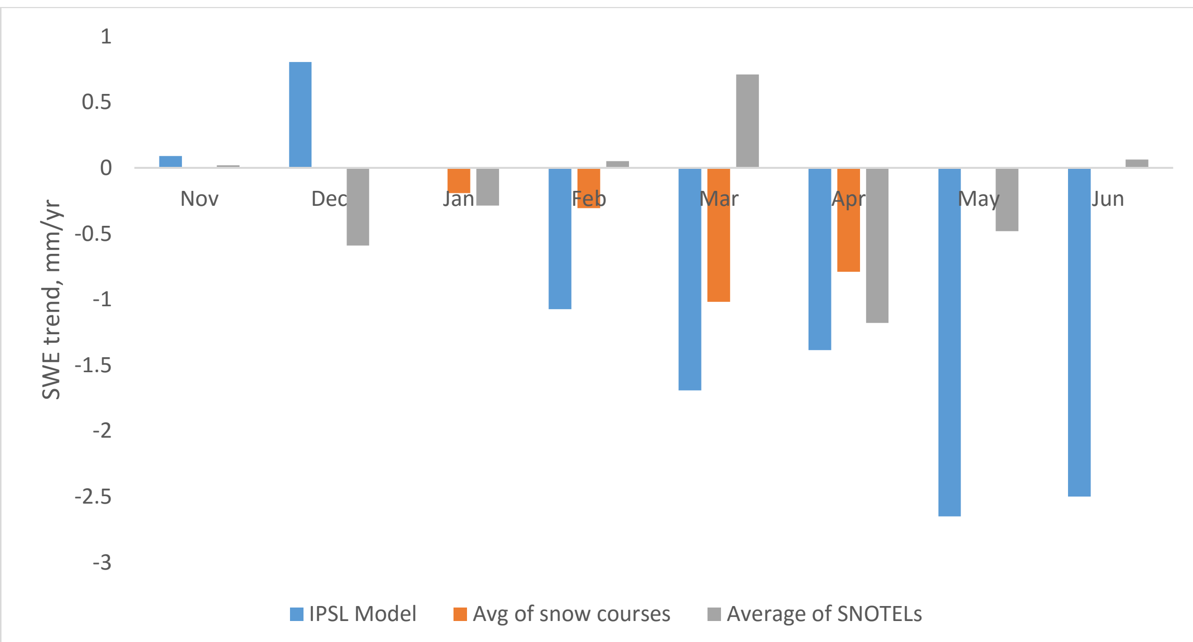

Figure 4.34. Trend in monthly SWE, 1981-2015, simulated by IPSL model and observed averages at snow courses and SNOTELs.

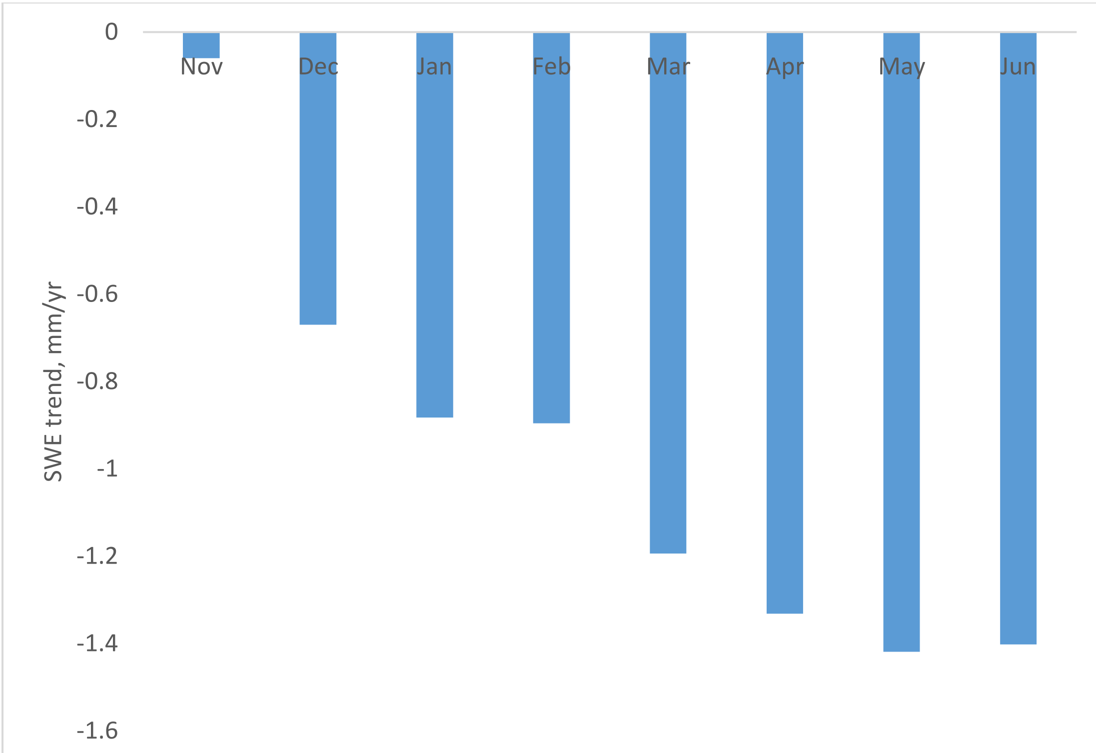

Figure 4.35. IPSL model projections for monthly SWE, 2015-2099.

CHAPTER 5. DISCUSSION

5.1 Trends in Snow Water Equivalent

5.1.1 Variability in Annual Measures of SWE

Evaluating trends in an annual time series of data involves making a distinction between inter-annual variability and long-term trends (Bradley *et al.*, 2007; Venable *et al.*, 2012). In general, longer time series produce more accurate estimation of trends than shorter ones. Time series shorter than 15 years were found to produce inaccurate estimates of temperature trends (IPCC, 2013). Precipitation in the western United States is often said to vary according to the double sunspot cycle, which has a period of about 22 years (Vines, 1982; Fu *et al.*, 2012), suggesting that the period for trend analysis of precipitation should exceed 22 years. As temperature and precipitation are the primary factors that influence accumulation and melt of the seasonal snowpack in the western United States (Hamlet *et al.*, 2005), trends in measures of snowpack dynamics should also be based on time series that exceed these minimum durations. The availability of SWE data dictates the maximum length of record that can be considered. For example, there are 80 years for the earliest snow courses, such as Wild Basin established in 1936, and 36 years for the first group of SNOTEL stations, such as Lake Irene established in 1980. Fortunately the latter periods provide annual series of SWE values sufficiently long for meaningful analysis of long-term trends despite the strong inter-annual variability.

In this investigation, the short-term, inter-annual variability in April 1 SWE at a typical SNOTEL station, such as Willow Park, was more closely related to precipitation than to temperature (Figure 4.3). Regression analysis (Figure 5.1) showed this greater correlation of April 1 SWE with precipitation than with temperature, indicating that short-term SWE variability in the western United States, and particularly in the colder inland mountains, is primarily related to variability in precipitation, while long-term SWE trends are primarily related to trends in temperature (Hamlet *et al.*, 2005; Serreze *et al.*, 1999).

5.1.2 Monthly Trends in SWE

The generally declining trends in snow water equivalent at snow courses and SNOTEL sites in and near Rocky Mountain National Park are consistent with previous studies (see below) that have found trends toward rising temperatures and decreasing snowpack in the region. This study, however, goes further to more closely examine patterns of variability in SWE trends throughout the snow season, and examining those patterns in relation to trends in near-freezing temperatures during the shoulder months of November and March, and to explore SWE trends in the context of elevation-dependent warming.

Monthly trends in April 1 SWE at snow courses and SNOTEL sites in this investigation ranged from an increase of 2 mm/yr to a decrease of 5 mm/yr, but most trends were negative. This is consistent with trends for various recent periods found for the northern Front Range of Colorado (Hamlet *et al.*, 2005, about -2.5 mm/yr; Regonda *et al.*, 2005, -2 to -4 mm/yr; Clow, 2010, -1.2 to -2.7 mm/yr), the upper Colorado River Basin (Harpold *et al.*, 2012, -1 to -5 mm/yr for annual peak SWE), and the Rocky Mountains from Colorado to British Columbia (Mote *et al.*, 2005, about -0.1 to -0.5 mm/yr).

While the average rate of decreasing trend with elevation was similar for most months, except January, as seen by the slopes in Figure 4.6, the y-intercept was the lowest (~2650m, similar for January). Conversely, the y-intercept for March 1st is the highest (~3075m). The change from March 1st to April 1st is seen since March is the month with the greatest negative change in SWE trends (Figure 4.5b). The change in SWE over the month of November also decreases at most stations (Figure 4.5b). Thus the months near the beginning (November) and end (March) of the snow accumulation season are times that have seen the greatest shift toward less accumulation (November) and/or more melt (March) (Figure 4.13). The decreasing trends in March change in SWE are observed over both long-term and short-term time scales, as illustrated in the lower left quadrant of Figure 4.8. All but five of the trends for March change in SWE occur to the left of the 1:1 line in Figure 4.8. This indicates that the pattern of negative trends in March change in SWE was stronger during the period 1981-2015 than during the longer term period. Finally, the stronger clustering of trends for March change in SWE in the lower left quadrant in Figure 4.8, compared with the more scattered pattern of April 1 SWE in Figure 4.7, indicates that the negative shift in March change in SWE (less accumulation or more melt or both) is more pronounced than the trend in April 1 SWE (Figure 4.8). Since more SWE accumulates during February (Figure 4.5b) and less is still present on April 1, the change in SWE between March 1 and April 1 is accentuated.

Although the negative trend in March change in SWE was stronger during the period 1981-2015 than during the longer period of snow course records, the general pattern of declining trends in April 1 SWE was slightly stronger over the longer period (Figure 4.7). This suggests that the trend toward increased SWE accumulation during February (Figure 4.5b) may have occurred mostly during the recent 35 years. It also indicates that the general trend toward declining April 1 SWE is consistent over at least 8 decades. This is consistent with an earlier conclusion that the cause of declining trends in April 1 SWE in the western United States is widespread long-term warming, and not decadal climate variability such as would be caused by the cyclical patterns of the Pacific Decadal Oscillation (Hamlet *et al.*, 2005).

At some of the northerly stations, for the longer period of record, there were increasing trends in SWE from the beginning (~1936) to the mid-1970s, followed by a decreasing trend afterwards (Fassnacht and Hultstrand, 2015). This is similar to what Chen and Grasby (2009) hypothesized with synthetic data, but contrary to what Venable *et al.* (2012) found with temperature data.

5.1.3 Other Measures of SWE Derived from Daily Data

The daily time step of SNOTEL stations provides opportunities for examining a variety of measures defining snowpack dynamics (e.g., Fassnacht *et al.*, 2014), in addition to the traditional measure of SWE on the first of the month. The most obvious measure is annual peak SWE (Figure 3.1). As in an earlier study (Clow, 2010), this investigation did not find a clear pattern of trends in annual peak SWE in this part of Colorado. However, Clow (2010) did find clear declining trends in annual peak SWE in southern and western Colorado (Derry and Fassnacht (2015) found declines in the SWE south of 38.75 degrees North latitude), suggesting that weather patterns influencing snowpack development in the north-central mountains of Colorado are probably different from those affecting other parts of the state. Also, Clow (2010) used a shorter time period (~1987 to 2007) and used bulk trends from the Regional Kendall Test, that Fassnacht *et al.* (2016) illustrated masked local climate trends. One pattern that did emerge in the trends in annual peak SWE is a tendency for high-elevation sites to have declining trends, while low-elevation sites had increasing trends (Figure 4.9). This is consistent with studies that have found a tendency for more rapid warming at higher elevation (Diaz and Eischeid, 2007; McGuire *et al.*, 2012). It is also consistent with trends toward greater April and May precipitation at SNOTEL sites in the study area (Figure 4.17).

The combination of increasing and decreasing trends for date of annual peak SWE (Figure 4.9) found in this investigation are consistent with the finding of little change in date of peak SWE (Harpold *et al.*, 2012). They are also mostly consistent with the finding of slight trends toward earlier peak SWE on the western side of the Front Range and later peak SWE on the eastern side, as well as towards an earlier peak SWE at warmer (lower) sites (Hamlet *et al.*, 2005).

Comparison of annual peak SWE with April 1 SWE (Table 4.1) confirms the earlier finding that April 1 SWE can significantly underestimate annual peak SWE (Bohr and Aguado, 2001). In this investigation the difference averaged 22 percent, a much larger difference than the 12 percent reported in the earlier study. This reflects the fact that SNOTEL stations in the study area include many of the higher, colder, and snowier stations in the state. Using May 1 SWE to estimate annual peak SWE would come closer to an accurate approximation for sites in this study area, but would still average 7 percent less than actual peak SWE.

5.1.4 Patterns in SWE Trends

The day-by-day analysis of SWE trends shows both similarities and differences among SNOTEL stations (Figures 4.10 and 4.11). At most stations, the trend on most days is toward less SWE, but at some stations, such as Bear Lake and Copeland Lake, there are many or all days with positive trends in SWE. The climatic influences on these daily SWE trends involve monthly patterns of trends in temperature and precipitation (Figure 5.2). This pattern of different climatic forcing mechanisms controlling snowpack dynamics at different times of the snow season has been noted in previous studies (Knowles *et al.*, 2006). In some months such as November and

March, decreasing precipitation and increasing temperatures combine to reduce SWE accumulation. In others, such as December, February, April, and May, the reverse is true, resulting in positive trends in SWE change. During January, precipitation and temperature trends are both increasing, and apparently the increasing precipitation trend outweighs the temperature trend. The apparent discrepancies between SWE and change in SWE in December and April are explained by the strong negative trends in SWE during the preceding months, November and March, respectively. These strong negative antecedent trends mean that the SWE trend in the following month can still be negative, even while the trend in change in SWE can be positive, since the SWE change is measured from an increasingly lower base.

The oscillating pattern of greater decrease or less increase in SWE during November; less decrease or greater increase during December to early March; greater decrease or less increase during mid-March to early April; less decrease or greater increase during mid-April to early May; and finally greater decrease or less increase during mid-May to June, is consistent at the four SNOTEL stations analyzed in detail (Figure 4.11). This suggests that the pattern of climatic forcing mechanisms described above is consistent across the study area.

5.1.5 Trends in Number of Days per Year with Threshold Accumulation and Melt Values

Five of the SNOTEL stations have trends toward slightly fewer days per year with SWE accumulation exceeding 10 mm (Table 4.2). The two that had positive trends, which were the only statistically significant trends, were Deadman Hill and Copeland Lake, both of which had positive trends. These stations are both on the eastern side of the Front Range (in Clow's purple grouping), and frequently receive snow from upslope storms that bring moisture from the east up into the Front Range (Figure 4.17d). This further explains the positive SWE trends at Copeland Lake.

Trends in days with accumulation exceeding 5 mm were mostly positive, but none were statistically significant. Trends in days with accumulation exceeding 0 mm were all positive, and almost all statistically significant. The daily SWE increases during those times are mostly less than 5 mm, as seen by the difference in significant trends for the 0 and 5 mm threshold. This consistent pattern of increasing trends in small increases in days with accumulation is likely a reflection of the trends toward increased precipitation and increased SWE accumulation during December, January, February, and April (Figure 4.17). An alternate explanation, however, could be increased variability, or noise, in sensor readings (Avanzi *et al.*, 2014.) Trends toward melt also increased. Nearly all of the trends in days with melt exceeding these three thresholds are positive (meaning more melt). The lowest threshold (0 mm) has the most significant trends, while the highest threshold (-10 mm) has the least significant trends. This may indicate that more melt is occurring in small daily amounts, but can also include increased sensor noise.

5.2 Trends in Precipitation

While the pattern of monthly trends in precipitation helps to explain the observed trends in SWE, the magnitudes of the precipitation trends observed in this study, on the order of 1 mm/yr, are small compared with trends observed in other parts of the western United States, which exceeding 3 mm/yr (Mote, 2003; Regonda *et al.*, 2005). This is consistent with other studies that have found little change in precipitation in this part of Colorado (Hamlet *et al.*, 2005; Ray *et al.*, 2008; Lukas *et al.*, 2014). The time series of total cold-season (October to June) precipitation (Figure 5.3) in this investigation does not show a cyclical pattern that would indicate strong correlation with cyclical atmospheric or sea-surface temperature phenomena, as has been observed closer to the ocean in the western United States (Hamlet, 2005).

In the Sierra and Cascade mountain ranges, the fraction of winter precipitation falling as snow has been decreasing by as much as 60 percent in 54 years, or about 4 mm/yr (Knowles *et al.*, 2006), but in this study area, the ratio of SFE to cold-season precipitation has been more consistent (Figure 4.17). The average trend in the ratio of total annual SFE/total Oct-June precipitation varied from zero to about +20 percent over 35 years, or about 0 to +1 mm/yr. The largest single-month decreasing trend was October at Joe Wright (Figure 4.17a), which had rate of about -20 percent over 35 years (1981-2015), or about -1 mm/yr. This is consistent with findings that the Colorado Rockies are sufficiently cold that the fraction is minimally affected by recent warming trends (Knowles *et al.*, 2006).

Many studies of precipitation trends depend on data from weather stations that do not measure snow, or do not use snow data. Also, most of these weather stations are at elevations below the zone of persistent seasonal snow accumulation (Richer *et al.*, 2013). Accordingly, it can be difficult to determine the phase of precipitation, i.e., rain or snow, on a particular day in the snow zone (Fassnacht and Soulis, 2002). Temperature records are used to make this determination, but the temperature threshold below which precipitation falls in frozen form is not always constant (Fassnacht *et al.*, 2013; Harder and Pomeroy, 2014). This underscores an advantage of calculating precipitation trends using records from SNOTEL stations, as the SWE data are available to help determine precipitation phase.

5.3 Trends in Temperature

The general trends during 1981-2015 toward warmer temperatures in the study area are similar to observations found in other recent reports (Hamlet *et al.*, 2005; Knowles *et al.*, 2006; Diaz and Eischeid, 2007). The Loch Vale temperature trend in this investigation represents an update to the trend reported for 1983-2007 in Clow (2010), including new and revised data (Colorado State University, 2016). The months with the strongest warming trends were November, March, June, and July, when trends exceeded 0.4˚C/decade (Figure 4.21). By comparison, average annual statewide temperatures in Colorado during 1980-2010 have been increasing at about 0.37˚C/decade (Lukas *et al.*, 2014). The northern Colorado Rockies appear to be warming more quickly than other parts of the state (Ray *et al.*, 2008; Lukas *et al.*, 2014). A comparison of seasonal temperature trends from 1957 to 2006 in four mountainous areas of the globe found that the Colorado Rocky Mountains and the Swiss Alps had the strongest overall warming trends, and that the warming trend in the Colorado Rockies (about 0.5˚C/decade) was particularly strong in the spring (Rangwala and Miller, 2012). Wet-day (days with precipitation) daily minimum temperatures, too, showed strong warming trends, especially during March, in the Colorado Rockies and elsewhere, with increasing trends of 0.25˚C/decade during 1949-2004 (Knowles *et al.*, 2006). Based on temperature records recorded at SNOTEL stations from 1991 to 2012, homogenized to adjust for effects of sensor upgrades, Oyler *et al.* (2015) found that warming trends in maximum temperatures in the mountains of Colorado, like those of other interior mountain ranges in the western U.S., averaged in the range of 0.25 to 0.75˚C/decade.

Temperature records from SNOTEL stations have been shown to contain a positive bias related to changes in sensors, sensor locations, and operational protocols (Oyler *et al.*, 2015). Nevertheless, temperature records from SNOTEL stations can be useful in making relative comparisons among stations. Using the temperature records from the SNOTEL stations preserves the association with the other data collected at the same stations, notably SWE and precipitation. These were used to analyze trends in April 1 SWE in relation to trends in temperature and precipitation. Based on the technique used by Mote *et al.* (2003), trends in April 1 SWE at the SNOTELS were plotted as circles with diameters proportional to the SWE trends, along a pair of axes representing trends in total October-June precipitation and average October-June temperature at the SNOTEL stations (Figure 5.4). The results for the RMNP area are similar to those in the Pacific Northwest (Mote, 2003), in that most of the April 1 SWE is decreasing, most are associated with slightly decreasing October-June precipitation at SNOTEL stations, and all are associated with increasing October-June temperatures. However, these temperature trends (Figure 5.4) are artificially amplified (Oyler *et al.* 2015). If the data were properly homogenized, the temperature trends could be in the range of zero to 0.3 degrees C/year, similar to those found at Allenspark, Grand Lake, and Loch Vale. One station (Lake Eldora) has increasing temperature, increasing precipitation, and significantly increasing SWE, suggesting that the greater precipitation was sufficient to counteract the effects of warming and resulted in increased SWE. Another station (Deadman Hill) has increasing temperature and precipitation, but decreasing SWE, suggesting that at this station the greater precipitation was not sufficient to counteract the effects of warming. A further station (Willow Creek Pass) has increasing temperature, decreasing precipitation, and increasing SWE, suggesting that other factors may also be involved. The overall range of the three variables in this study area is less than the variability reported for the same variables in the Pacific Northwest (Mote, 2003). This is consistent with Losleben and Pepin (2003) who showed the northern Colorado Rockies are less susceptible to temperature-induced negative SWE trends and variability is less than other mountainous areas in the western U.S.

The warming trends observed in this investigation do not appear to follow a cyclical pattern, but instead are relatively consistent in their rate of warming (Figure 4.19). The longer period of record (1949-2015) at the Grand Lake cooperative weather station provides an indication of when this recent warming trend may have begun (Figure 4.20). The time series of average annual temperature at Grand Lake shows a cooling trend from 1949 to 1973, followed by the warming trend that continues to the present. This is consistent with other studies that have identified the early to mid-1970s as the beginning of the current warming trend (Tebaldi *et al.*, 2012), and the 1990s to the present as the period with the strongest warming signal (Ray *et al.*, 2008; Santer *et al.*, 2011).

5.4 Trends in Freezing Level and Maximum Daily Temperatures above Zero

Results of the analysis of simulated free-atmosphere freezing level showed that only during November and March, the average level of the zero-degree Celsius isotherm was about 400 m below the elevation of the lowest SNOTEL stations (Figure 4.22). In addition, trends in average freezing level were rising more rapidly in November and March than in other cold-season months in between. The trend in freezing level for February was slightly negative. This suggests that during November and March the daily temperature cycle brings warmer-than-freezing temperatures to the elevation zone of the SNOTEL stations for some portion of each of those months. It would then follow that warming trends during these two months are likely to increase the portion of the month with warmer-than-freezing weather, resulting in increased snowmelt. The monthly patterns and trends in daily maximum temperatures warmer than zero at the highest (Lake Irene) and lowest (Copeland Lake) SNOTEL stations provided a test of this hypothesis. Results show that November and March had similar values (Figure 4.23). During these two months the summation of daily maximum temperatures warmer than zero at these two stations ranged from 75 to 250 total degrees, implying a monthly average maximum temperature of about 2.5 to 8.3 degrees. Therefore, both November and March have temperature ranges based on this measure such that a relatively small increase in temperature would cause a relatively large proportional increase in the amount of warmer-than-freezing weather available to help melt snow.

5.5 Climatic Influence on SWE Trends

Trends in both precipitation and temperature influence trends in SWE in and near Rocky Mountain National Park. While year-to-year variability in SWE is influenced primarily by variability in precipitation, the primary reason for the long-term SWE declines in and near Rocky Mountain National Park appears to be increased melt during periods of the snow season formerly characterized by little melt. This increased melt is associated with long-term trends toward warming temperatures, especially during November and March. In their assessment of the Rocky Mountains from Colorado to British Columbia, Mote *et al.* (2005) concluded that warming produces lower spring SWE largely by increasing the frequency of melt events. Hamlet *et al.* (2005) concluded that decreased April 1 SWE in the Rockies from 1947-2003 is due primarily to widespread warming. In this study, we looked at total cold-season precipitation and snowfall at SNOTEL sites, and found no trend toward decreased precipitation or snowfall. Also, snowfall as a proportion of total cold-season precipitation showed no trend, except at the lowest SNOTEL station (Copeland Lake), where there was a trend of increasing snowfall. This agrees with Clow (2010), Mote *et al.* (2005), and Harpold *et al.* (2012), indicating that the SWE declines are not related to an overall decrease in winter precipitation, or a change from snow to rain. While there was a shift from snow to rain widespread over much of western North America and even for parts of Colorado, no such shift was observed for the northern Colorado Front Range (Knowles *et al.*, 2006).

As shown in Figure 4.21, March and November are also the two months during the snow accumulation season with the strongest warming trends. This finding reinforces an earlier finding (Mote *et al.*, 2005) that warming produces lower spring SWE largely by increasing the frequency of melt events, not by simply enhancing the likelihood of rain instead of snow. Indeed, Mote *et al.* (2005) found significant correlation between April 1 SWE and total daily melt events. In this study there was no significant correlation between April 1 SWE and number of melt events, but trends in number of days with melt (>5 mm loss in SWE) were significantly increasing at five SNOTEL stations and not significantly increasing at six other, closely matching the downward trends in April 1st SWE.

5.6 Trends in Magnitude and Timing of Peak SWE versus SWE Change

Annual peak SWE has decreased at 7 stations (increased at 5) and timing of the annual peak has shifted earlier at four and later at eight SNOTEL stations. The timing of peak SWE, in part, explains why some stations have had a shift in date of peak SWE while others have not. The period during mid-March to early April is characterized by warming and drying trends, when SWE trends become less positive or more negative (Figures 4.11, 4.13, 5.2). For simplicity, this period will be called the “March thaw”. At the higher and colder sites such as Joe Wright and

Willow Park (Figure 5.5), the March thaw occurs prior to the peak, during a period of net SWE accumulation. The March thaw inhibits the accumulation, but as of 2015 has not caused the SWE trends to shift from net accumulation to net loss. At the lower and warmer sites, such as Copeland Lake (Figure 5.6), the March thaw occurs after the peak, during a period of net SWE loss, when further acceleration of the SWE loss trend has no impact on the earlier annual peak. At both of these types of sites, peak SWE therefore occurs during a period with a trend toward cooler and wetter conditions. Therefore, the warming trends during the March thaw do not affect the timing or magnitude of the annual peak, which is actually likely to have a trend toward increasing magnitude and later timing, in response to the trends toward cooler and wetter conditions. However, at some sites, such as Phantom Valley (Figure 5.7), which is intermediate in elevation, annual peak SWE occurs during the March thaw. Moreover, at this site, the temperatures during the March thaw are such that the trend in 15-day SWE change has shifted from net accumulation to net loss of SWE. This is the requisite for a shift toward an earlier peak SWE, and as a result, the timing of peak SWE has shifted 10 days earlier over 35 years. The relation between trend in date of peak SWE and peak SWE, therefore, is a curvilinear pattern, in which early and late dates of peak SWE show trends toward later peak SWE, while dates of peak SWE during the March thaw show trends toward earlier peak SWE (figure 5.8). In the future, if the March thaw trend continues, SNOTEL stations such as Willow Park are likely to see SWE trends during the March thaw shift from net accumulation to net loss, and at that time the magnitude and timing of annual peak SWE are likely to shift toward smaller and earlier values.

5.7 Elevation Dependent SWE Loss

Negative trends in several measures of SWE in this investigation are more prevalent at higher elevations in the zone that includes the SNOTEL stations, than at lower ones. All of the monthly SWE trends from snow courses and SNOTEL stations have a negative correlation with elevation (Figure 4.6). Similarly, the trends in annual peak SWE (Figure 4.9), and in days per year with SWE over 100 mm (Figure 4.14) show more negative trends at higher elevations than lower ones. Trends in some other measures, however, suggest an opposite pattern. Trends in date of annual peak SWE show more shifts to later dates at higher-elevation sites, and more shifts to earlier dates at lower-elevation sites (Figure 4.8.2). Warming of annual average temperature is limited at the higher-elevation site (Loch Vale), and greater at the lower-elevation sites (Allenspark and Grand Lake) (Figure 4.21). Similarly, trends in average cold-season temperature as measured at the SNOTEL stations show no correlation with elevation (not shown). A factor other than temperature is likely influencing the elevation dependence of negative SWE trends. Trends in October-June snowfall equivalent also show a negative correlation with elevation (Figure 5.9). This suggests that the predominance of negative trends in SWE at higher elevation sites may be related to less accumulation, rather than more temperature-induced melt. The elevation dependence of negative SWE trends in the study area contrasts with results of a study in the Pacific Northwest, which found a positive correlation between elevation and April 1 SWE (Mote, 2003). The negative correlation between SWE trends and elevation found in this investigation contrasts with the earlier conclusion that higher-elevation, colder sites in the Colorado Rockies tend to be more immune to effects of climate change than lower sites (Mote *et al.*, 2005).

Some researchers have found evidence for elevation-dependent warming in and near the study area. Along an elevation transect in the Front Range of the Colorado Rockies just south of Rocky Mountain National Park, the strongest warming trend (0.42-0.44˚C/decade) during 1953-2008 was in maximum temperatures at the sites with elevations similar to the SNOTEL stations (2591–3048 m) (McGuire *et al.*, 2012). The strongest warming trends in that study during 1989-2008 were in maximum temperatures at 2591m (0.85˚C/decade), and in minimum temperatures at 3048 m (0.5˚C/decade). Using data for 1979-2006 from the PRISM (Parameter-elevation Regressions on Independent Slopes Model), Diaz and Eischeid (2007) found mean annual warming of 0.3˚C/decade at 1500 m, compared to 0.7 ˚C/decade at 3500 m, and 0.96 ˚C/decade at 4000 m. Models of atmospheric processes over mountains, as well as numerous other observational studies, support the conclusion that several atmospheric warming mechanisms are stronger at higher elevations, or at least in the subalpine zone, than at lower elevations (Rangwala and Miller, 2012; Pepin *et al.*, 2015). Apparently the complex interaction of temperature, precipitation, snowfall, topography, and weather patterns in these mountainous areas creates some opportunities for additional research into elevation dependence of trends.

5.8 Albedo-Feedback and Humidity Effects

One of the more likely mechanisms that would account for the warming and SWE loss during March and November is albedo-feedback warming (Pepin *et al.*, 2015; Rangwala and Miller, 2012). This process tends to occur near the zero isotherm, which is in the vicinity of the SNOTEL sites during November and March. Loss of snow cover reduces the albedo, which enhances absorption of solar radiation, leading to more warming. The effect is to shorten the snow accumulation season at both ends. Thus, while the high, cold mountains of the northern

Front Range are still cold enough to support a strong snow accumulation season during December-February, they are showing the typical signs of warming and SWE loss during November and March. Another elevation-dependent warming mechanisms is positive feedback related to increased humidity that yields more downward longwave radiation; this has been found to be significant in Colorado (Naud *et al.*, 2013). The occurrence of dust on snow has been mentioned as a possible factor enhancing melt through albedo feedback at higher elevations (Lukas *et al.*, 2014; Rangwala and Miller, 2012; Painter *et al.*, 2010). However, the influence of dust on elevation dependent warming in the study area is likely to be limited as the Front Range is distant from the Four Corners area, the source of most dust affecting snow in Colorado, and there appears to be little or no relation between dust deposition and elevation (Painter, T.H., personal communication, 2015). Another possible factor could be additional albedo feedback caused by deposition of black carbon and other organic-matter particles on snow (Flanner *et al.*, 2009).

5.9 Niveograph Interpolation from Snow Course Data

The technique of model adjustment using time pro-rated multipliers provides a simple method that allows the model niveographs to be bent and shaped to closely match the observed monthly data. As there is also significant interannual variability in the specific shape of the niveograph, this flexibility to bend the model to fit each year's data helps to improve overall accuracy. The close fit between adjusted niveograph models based on daily niveographs at four index SNOTEL stations and observed daily niveographs at Willow Park suggest that SNOTEL records can be used to estimate daily niveographs at nearby snow courses. This approach represents a significant improvement over the assumption that April 1 SWE represents the annual peak SWE at snow courses.

5.10 Reconstructions of Past SWE Trends

According to the Gunnison SWE reconstruction (Woodhouse, 2003), in the past a site near the study area experienced a steeper declining trend in SWE than was found in this investigation, but the duration of the trend was much shorter (13 years compared with 80 years in this study). Similarly, estimates of declining SWE trends inferred from the hydrologic reconstructions in the Front Range (Woodhouse and Lukas, 2006) suggest the occurrence of past declining SWE trends with similar slopes and durations similar to the period of SNOTEL record, but still shorter than the consistent declining SWE trends of the snow course records. However, the temperature reconstruction from Almagre Mountain (Berkelhammer, 2010, LaMarche and Stockton, 1974) suggests that past trends in temperature near the study area have included warming trends with durations similar to or greater than the recent 40-year warming trend noted for Grand Lake. This suggests that the past few centuries are not entirely without precedent for the warming trend that has been observed in the study area during 1981-2015.

5.11 Future Projections of SWE Trends

The climate and hydrologic model projections discussed in the results section consider the atmospheric greenhouse gas loading that distinguishes the current period of warming trends from past warming trends (Maurer *et al.*, 2007). Therefore, the projections for continued warming throughout the 21st century, common to all the models discussed, suggest that climatic conditions driving snowpack dynamics for the remainder of this century will be different from the cyclical patterns of the past. The general agreement between the high-resolution WRF-Hydro model and observed monthly trends in SWE lend some additional credence to this model (Rasmussen *et al.*, 2014). The primary discrepancies are for the months of November and March, during which the modeled trends projected more precipitation while the trends at Willow Park were toward less precipitation (Figure 4.27). The general pattern of future precipitation and SWE, however, appears to reflect current trends. Monthly trends in precipitation and SWE for the next 3.5 decades for the Colorado mountains, simulated using the WRF-Hydro model, were quite similar to trends observed at snow courses and SNOTEL stations in this investigation. Specifically, the simulated trends for precipitation showed increasing precipitation early in the snow season, especially December through April, and decreasing precipitation later in the spring. The simulated trends for SWE showed little or no loss in SWE during October through February, but greater SWE loss during March through June. Therefore, enhanced precipitation during the early part of the snow season (October to February), especially higher than 3000 m, is likely to preserve and, in some places, even raise the average SWE during those months. This enhancement of precipitation and SWE is likely to maintain or raise current SWE conditions as late as April 1 in parts of the study area. As spring progresses, enhanced warming and decreased precipitation are likely to accelerate rates of SWE loss, leading to earlier and lower peak SWE, more rapid melt, and earlier runoff By 2050 April 1 SWE in much of Colorado is projected to decrease by 25% compared with 2014 (Rasmussen *et al.*, 2014), with probably lesser declines in this study area.

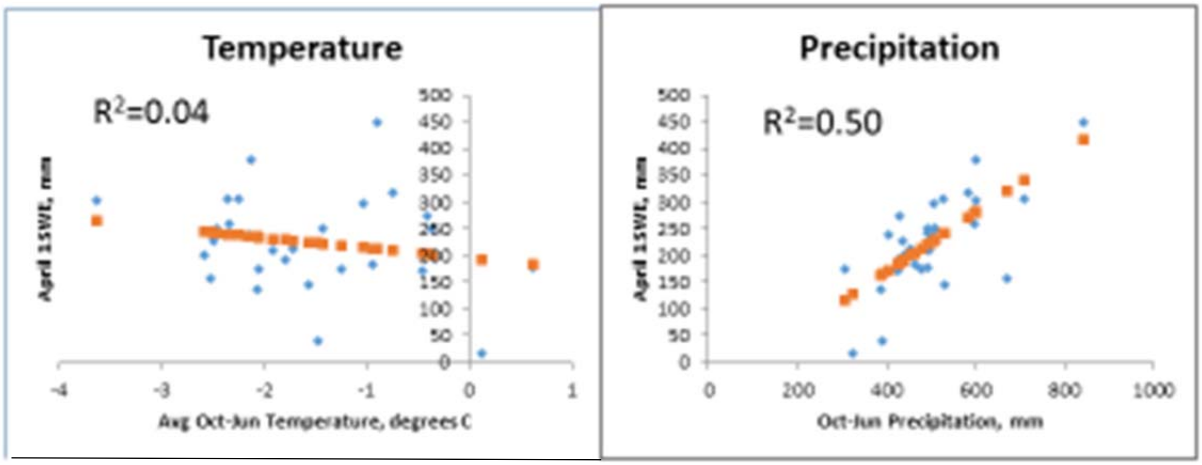

Figure 5.1. Regression analyses of April 1 SWE on October-June temperature and precipitation at Phantom Valley SNOTEL.

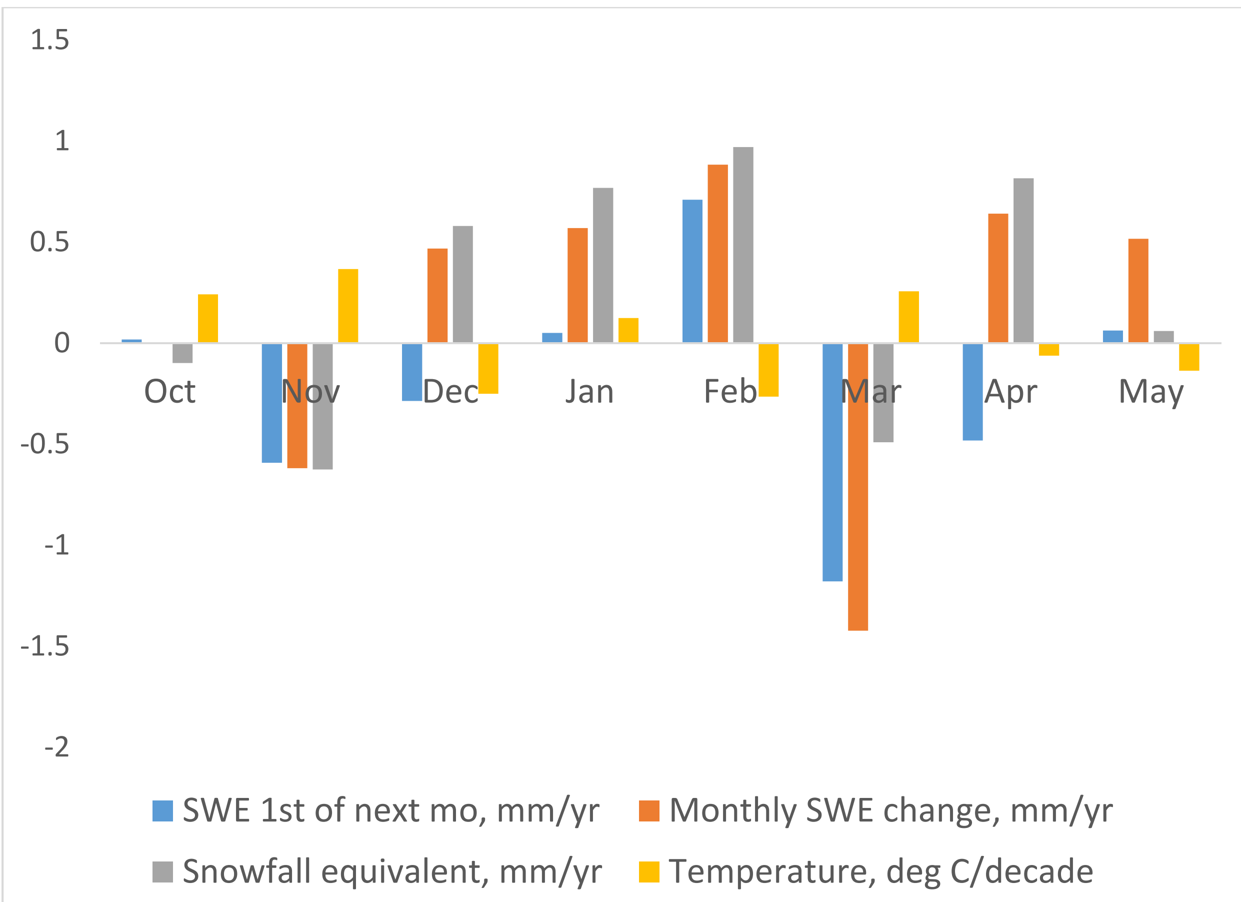

Figure 5.2. Average trends in SWE, monthly change in SWE, SFE, and temperature, by month.

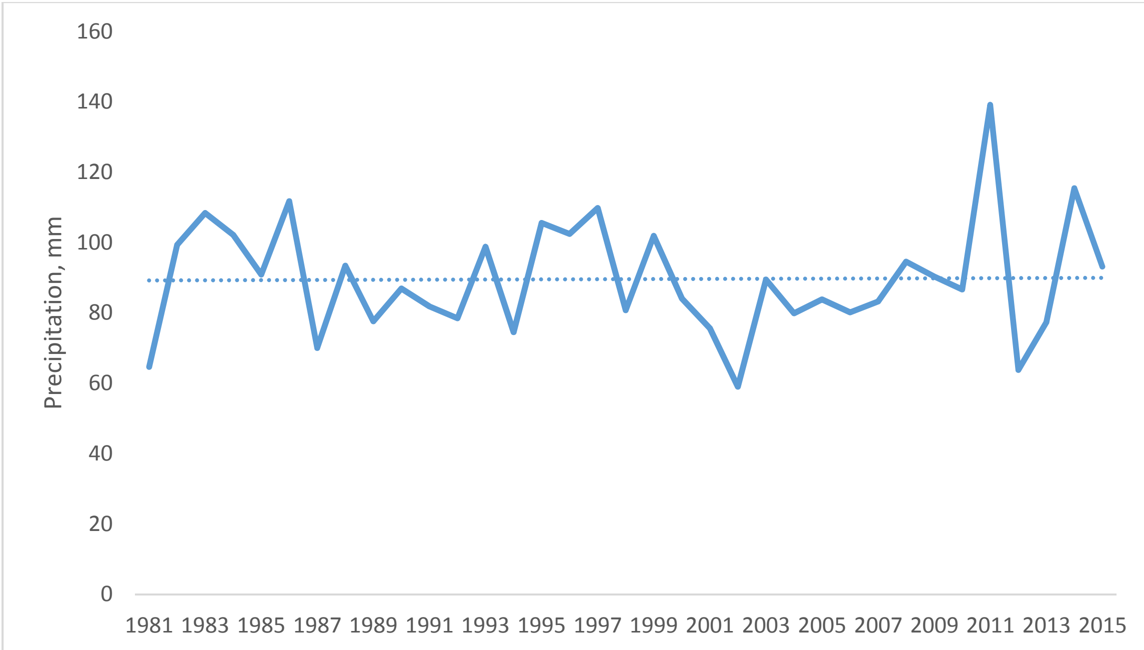

Figure 5.3 Total Oct-June precipitation at Willow Park SNOTEL, 1981-2015.

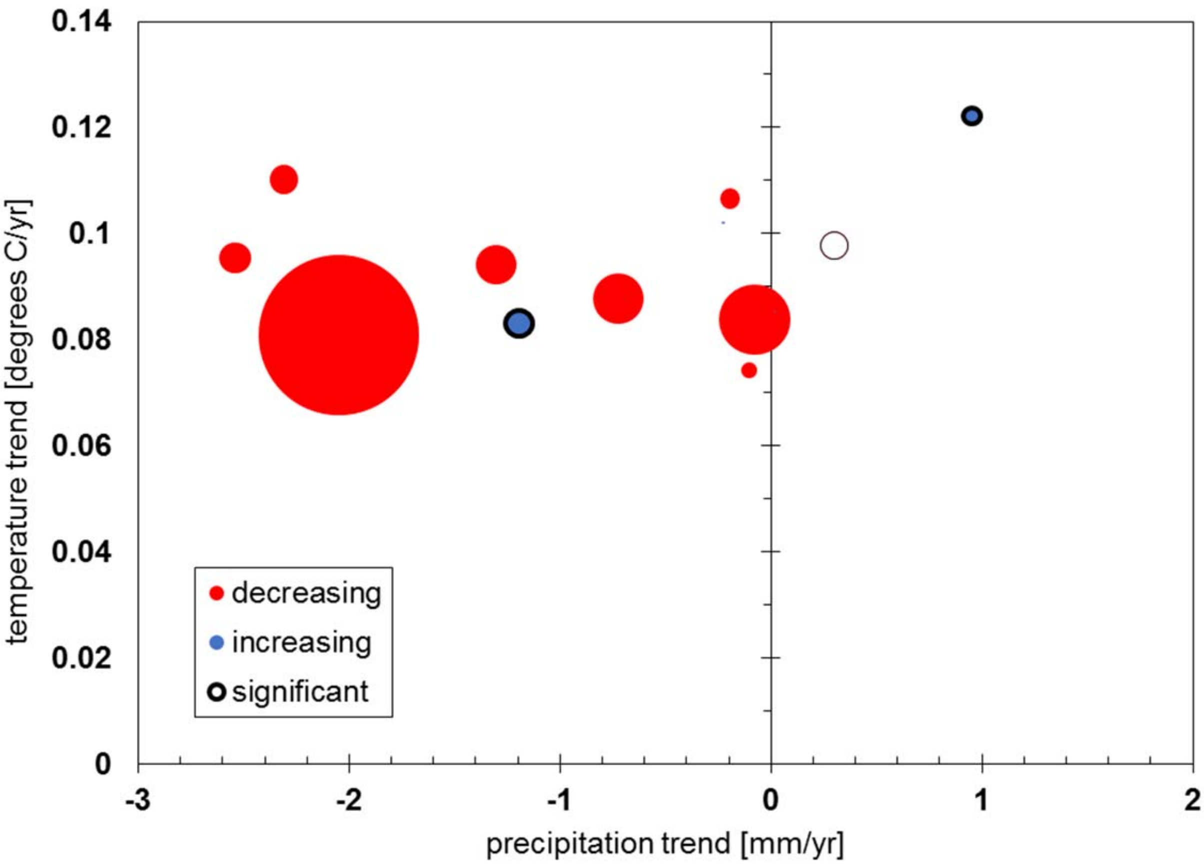

Figure 5.4. Trends in April 1 SWE at SNOTELs in relation to trends in precipitation and temperature at the SNOTELs.

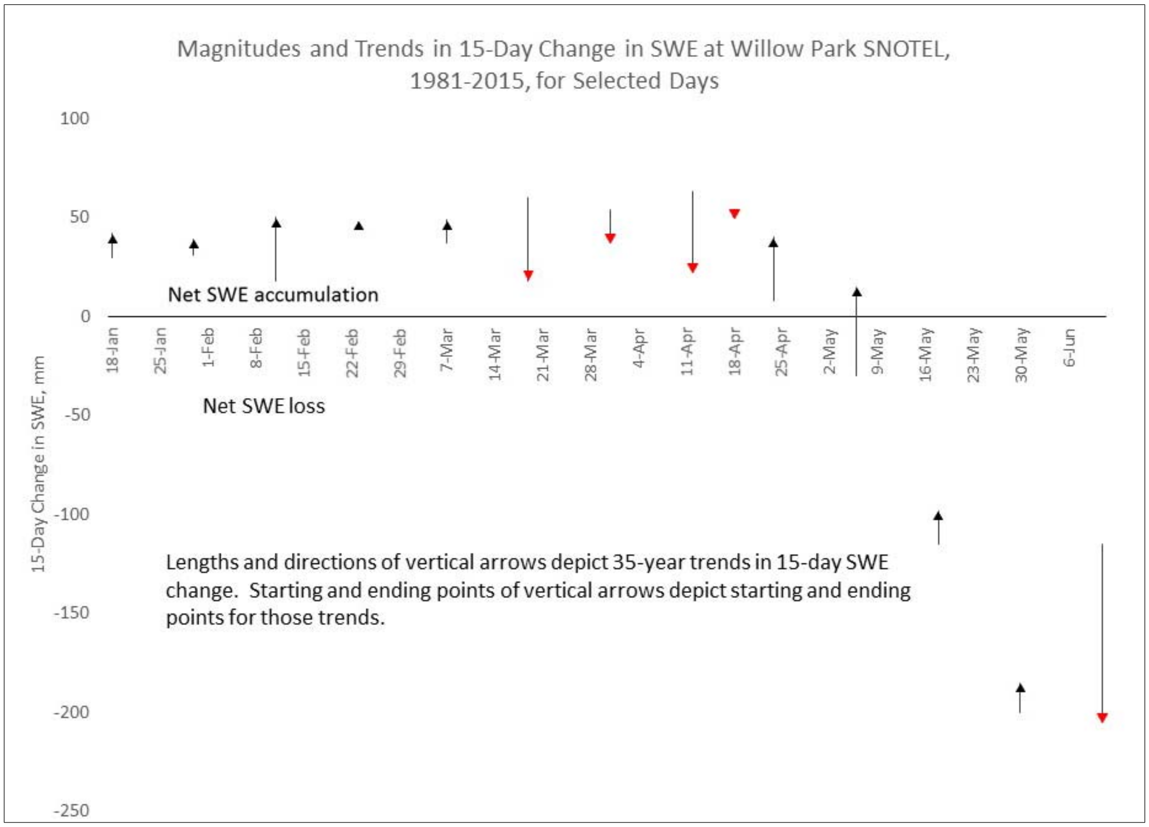

Figure 5.5. Magnitudes and trends in 15-day change in SWE at Willow Park SNOTEL, 1981-2015, for selected days

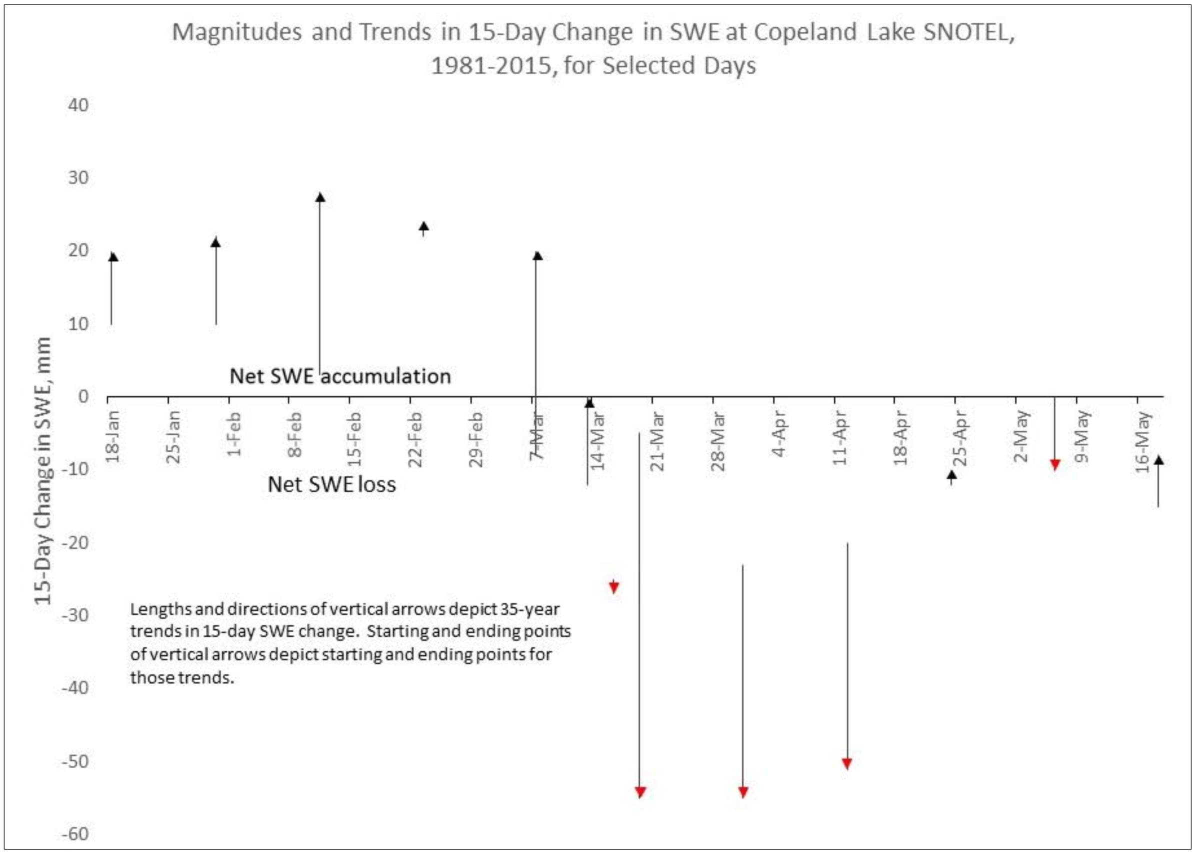

Figure 5.6. Magnitudes and trends in 15-day change in SWE at Willow Park SNOTEL, 1981-2015, for selected days.

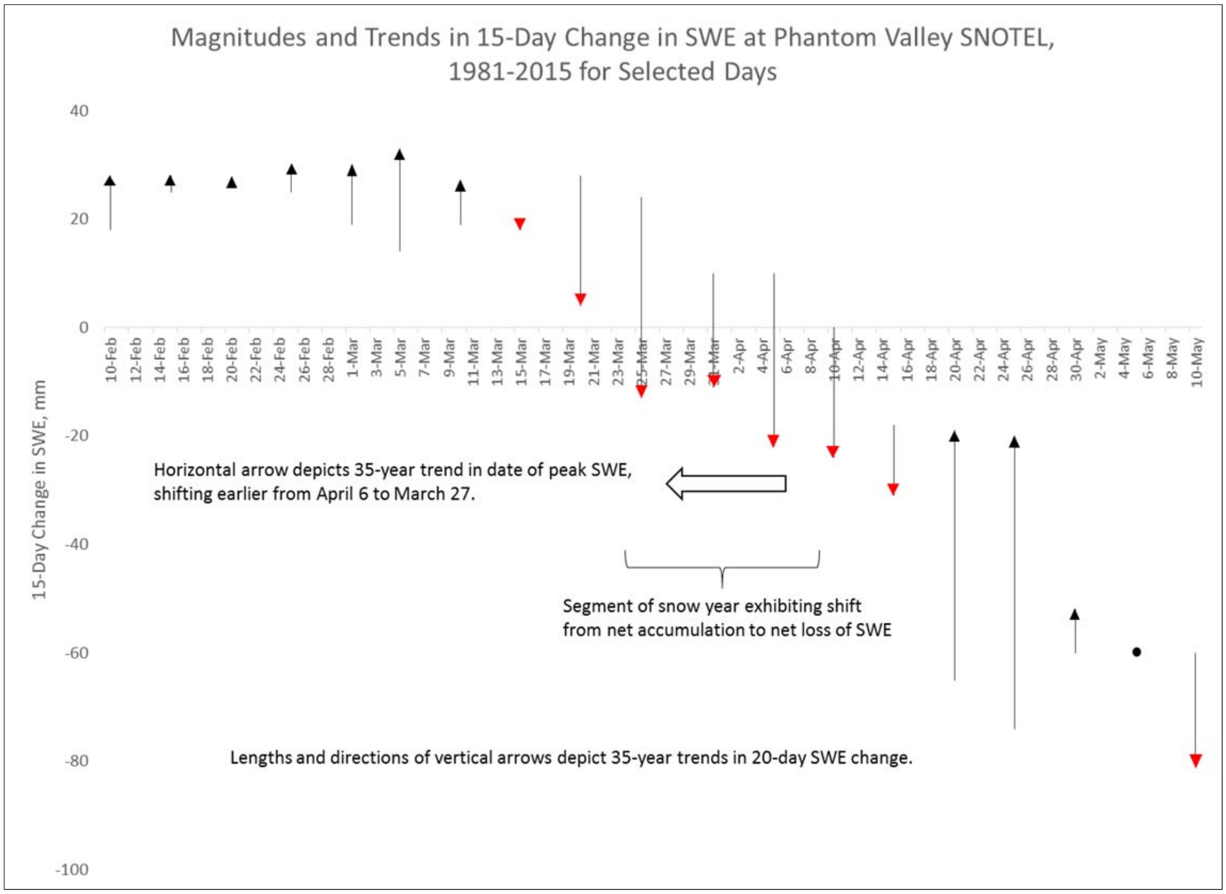

Figure 5.7. Magnitudes and trends in 15-day change in SWE at Phantom Valley SNOTEL, 1981-2015, for selected days.

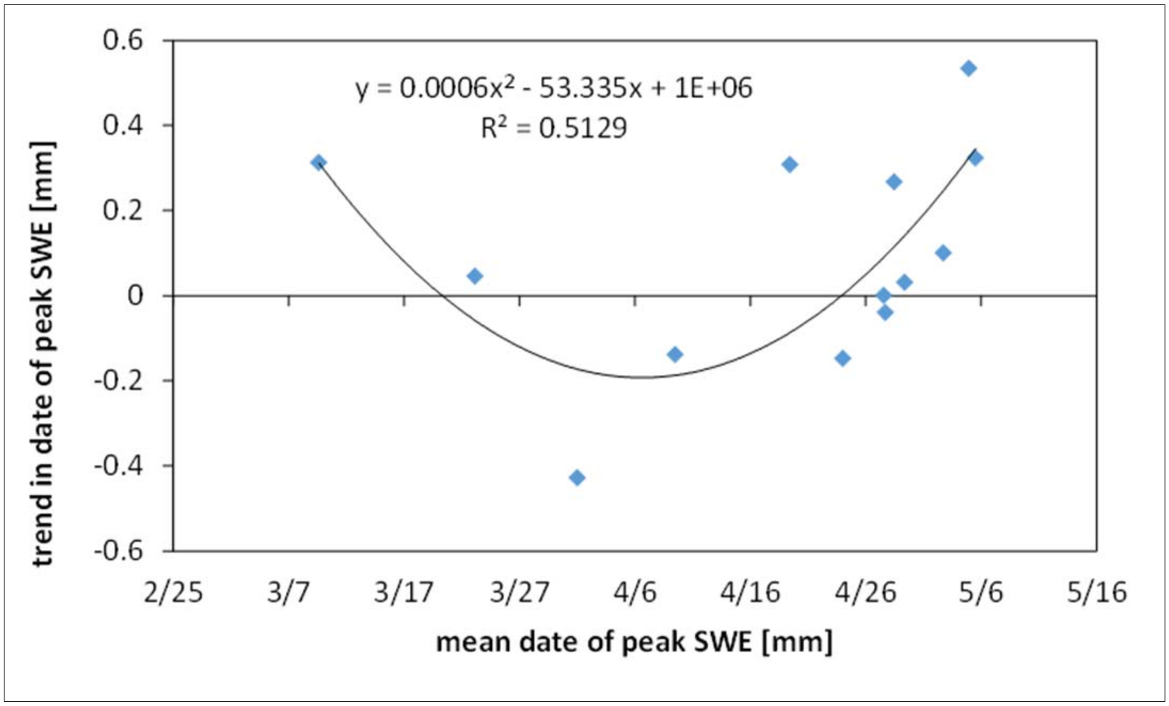

Figure 5.8. Relation between trend in date of peak SWE and date of peak SWE at SNOTEL stations.

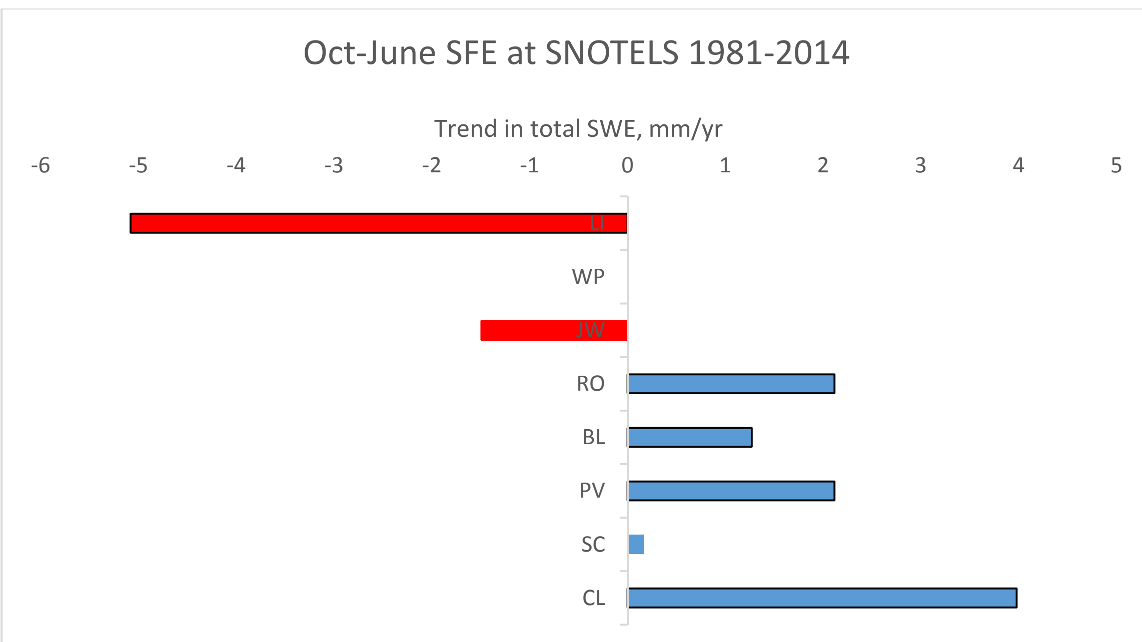

Figure 5.9. Trends in total October-June SFE at SNOTELS, listed by elevation. (NOTE: This graph will be updated to include all 13 SNOTELs and the year 2015.)

CHAPTER 6. RAMIFICATIONS FOR NATURAL RESOURCE MANAGEMENT

The findings of this investigation suggest some points that are likely to be of interest to natural resource managers working in the study area, as well as beyond. Like other areas of the western United States that depend on the seasonal snowpack for water supply and other natural resources benefits, Rocky Mountain National Park and vicinity are undergoing climatic changes that are altering patterns of accumulation and melt of the snowpack. In comparison to areas such as the Sierra and Cascade mountain ranges in California and the Pacific Northwest, as well as mountains farther south and west in Colorado, the north-central Colorado mountains in the study area are experiencing less variability in temperature, precipitation, and snow water equivalent (SWE), and weaker long-term trends. However, the observed and projected trends are still noteworthy. December, January, and February are still dependable months for snow accumulation, and are likely to continue as such for much of the rest of this century. In November and March, however, warming and drying trends are reducing SWE accumulation at the beginning and end of the winter. March is shifting from a month of dependable SWE accumulation to a month with less accumulation and, in some years and some places, net loss in SWE. Spring storms often bring additional SWE accumulation, but rising temperatures create variable conditions and lead to rapid melt.

6.1 Ramifications for Water Supply

Overall precipitation is not decreasing in the study area. In December, January, and February, April, and May, precipitation trends are positive. However, trends toward less accumulation or greater loss of SWE during March result in less SWE in the snowpack on April 1, and variable warming trends later in the spring are likely to enhance snowmelt. Water managers will have to rely less on the snowpack for water storage, and more on artificial storage projects, whether above or underground. Warming trends during the growing season are projected to enhance evapotranspiration. This will reduce soil moisture and streamflow, stress vegetation, and will probably result in increased demand for water for irrigation.

6.2 Ramifications for Managers of Other Natural Resources

Reduced soil moisture and streamflow, and increased stress on vegetation, are likely to lead to increased risk of wildfires during longer, drier summers in the study area (Westerling *et al.*, 2006). This will impact risk to structures and people, put more aerosols into the atmosphere, and, together with the reduced streamflow, alter aquatic and riparian ecosystems. Terrestrial ecosystems that rely on snow for habitat or protection will also be altered as spring SWE is reduced, and will have drier soil and vegetation in the summer.

6.4 Ramifications for Winter Recreation

Snow-based recreation is likely to become more concentrated during the months with more dependable SWE accumulation, January and February. It is also likely to become more concentrated in higher-elevation parts of the park and vicinity, where snow will be more dependable. Variable conditions during the spring may result in increased danger of avalanches. As warming continues there will be fewer days with snow suitable for skiing or snowshoeing.

CHAPTER 7. CONCLUSIONS

The seasonal snowpack in and near Rocky Mountain National Park is undergoing changes that will pose challenges for water providers, natural resource managers, and winter recreation enthusiasts. Assessing long-term trends in measures of the seasonal snowpack, and in the climatic factors that influence its accumulation and melt, helps to characterize those challenges. In particular, evaluating the patterns of variation in those trends during the snow season provides new understanding as to the causes, specific ramifications, and likely future course of the trends. In addition, placing the current 35-year trends in the longer context of longer-term observational records, and paleoclimate tree-ring reconstructions provides useful comparisons of current and past trends. Finally, projections of future trends provided by linked climate and hydrologic models offer a sense of how these trends are likely to affect the snowpack of the future.

Several factors are working to limit snowpack changes in the study area, in comparison to snowpack changes elsewhere in the western United States, at least for the next few decades. One mitigating factor is the relatively limited variability in temperature and precipitation, and hence snow water equivalent (SWE), in the study area, compared with other areas such as the Cascades, Sierras, and mountains of western and southern Colorado. The relatively limited variability applies mostly to time scales of one to a few years; over longer time scales, the study areas appears to have warming trends as rapid or more rapid than those found in other parts of the western U.S. Short-term variability in precipitation is correlated with short-term variability in SWE, while long-term trends in both precipitation and temperature appear to influence long-term trends in SWE. A second mitigating factor is the inland location and continental climate of the study area. This tends to isolate the study area from influences of the ocean, including moderate temperatures, and the cyclical oscillations that bring stronger variability to mountains closer to the coast. The continental climate also helps to keep winter temperatures sufficiently cool to support snowpack development and retention. A third mitigating factor is the relatively high elevation of the snow zone in the study area, which reinforces the cool winter temperatures. A fourth mitigating factor is an observed trend toward greater precipitation in the core winter months of December, January, and February. This trend is crucial to maintaining and, in some cases, enhancing the snowpack during these months. All four of these mitigating factors are likely to continue influencing the snowpack in the study area throughout the present century. A fifth mitigating factor is an observed trend toward enhanced precipitation and cooler temperatures during mid-April to early May, a crucial period for retaining SWE into the beginning of the drier summer months. This trend may be related to the occurrence of spring upslope conditions that bring moisture to the Front Range of the Rockies from the east. This factor is not as wide-spread geographically as the other factors mentioned. In addition, linked climate and hydrologic models that are capable of simulating the effects of the first four factors do not project this fifth factor as a significant influence on snowpack dynamics in the future. Therefore, there is some uncertainty about the future persistence of this trend in the study area. Model projections suggest that spring warming will reduce SWE accumulation and enhance melt during April and May. These mitigating factors help to explain why, at most of the 13 SNOTEL stations in the study, the timing of annual peak SWE has not shifted to an earlier date, and monthly SWE trends during February are positive.

In spite of these mitigating factors, declining trends in several measures of SWE in the study area are apparent. In every month, some of the trends in monthly SWE at the 23 snow courses and 13 SNOTELs included in the study are decreasing. During November, December, March, and April most of the trends in monthly SWE are decreasing. Annual peak SWE is decreasing at most of the 13 SNOTEL stations, number of days per year with measurable SWE loss is increasing at all of the SNOTELs, and number of days per year with over 100 mm of SWE is decreasing at most of the SNOTELs.

The strongest negative trends in SWE occur during November and March. These trends are associated with both increasing temperatures and decreasing precipitation in those months. The increasing temperatures bring more above-freezing weather, and hence more snowmelt, to SNOTEL stations in the study area. At SNOTEL stations such as Phantom Valley, where the "March thaw" coincides with the timing of annual peak SWE, annual peak SWE is reduced and the timing of the peak is shifted earlier. At other stations, where annual peak SWE occurs prior to or after the March thaw, trends toward cooler and wetter conditions tend to preserve or enhance the magnitude and timing of annual peak SWE.

While most of the trends discussed in this investigation pertain to the 35-year period 1981-2015, evaluation of SWE data from snow courses starting in the 1930s shows that declining SWE trends have been consistent for the full 80-year record of these snow courses. In contrast, evaluation of temperature data starting in 1949 at Grand Lake shows that, consistent with trends in many other areas, the current warming trend in the study area began about 1973, following a period of cooling.

As found by Clow (2010), there are some differences in SWE trends at SNOTEL stations on the western (orange cluster) and eastern (purple cluster) sides of the Front Range. The sites on the western side have more negative monthly trends in SWE, while the sites on the eastern side have fewer such trends. This difference is especially noteworthy during February, April, and May.

Trends in October-June precipitation are relatively small in comparison with trends seen in other parts of the western United States, and there is as yet no trend toward decreased fraction of October-June precipitation falling as snow. The precipitation trends that do exist, along with the temperature trends, help to explain the pattern of SWE trends that tend to decrease in November and March and increase in other months.

In contrast to results from the Pacific Northwest, which found stronger decreasing trends in SWE at lower elevations in the mountains, in this investigation several of the measures of SWE had trends indicating greater SWE loss at higher elevations. This pattern was not consistent with all measures of SWE, however, and the temperature trends examined in this investigation did not show a consistent pattern of more rapid warming at higher elevation.

Tree-ring reconstructions that provide indications of past multi-decade trends in SWE near the study area suggest that past multi-decade trends may have been steeper than the present trend, but did not last as long. A reconstruction of past temperature near the study area suggests that several multi-decade periods in the past may have had warming trends with longer duration than the approximately 40 years, so far, of the current trend.

Linked climate and hydrologic models, especially the high-resolution WRF-Hydro model designed for accurate simulation of atmospheric processes over the complex topography of the study area, project that the current warming trends will continue, and will overcome the effects of increased precipitation to enhance spring snowmelt in most of the study area. At some of the highest elevation sites the models project that, even with warming, the temperatures will remain below freezing, and continued enhanced precipitation may preserve the snowpack.

These findings suggest that, in Rocky Mountain National Park and vicinity, the core winter months of December, January, and February are likely to continue with current or perhaps even enhanced trends in snow water equivalent. However, the snow season is likely to begin later, due to continued warming and drying trends in November. March is likely to shift from being a key month for snow accumulation, to being a period with less accumulation and some loss of SWE. Conditions during April and May are likely to variable, with strong SWE accumulation from upslope storms in some years, but increased warming causing more rapid melt. Spring runoff is likely to begin earlier. During the longer, warmer, summers, soil moisture is likely to decline, contributing to decreased streamflow, stress for vegetation, and increased fire danger. Winter recreation will become more concentrated during January and February. The variable conditions during the spring will create increased risk for avalanches. Downstream water users will need to rely less on snowmelt runoff and more on stored water.

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