

THESIS

A CLIMATOLOGICAL STUDY OF SNOW COVERED AREAS  
IN THE WESTERN UNITED STATES

Submitted by

Cara Moore

Department of Ecosystem Science and Sustainability

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Master's Committee:

Advisor: Stephanie Kampf

Steven Fassnacht

Jason Sibold

## ABSTRACT

### A CLIMATOLOGICAL STUDY OF SNOW COVERED AREAS IN THE WESTERN UNITED STATES

Snow accumulation and timing of melt affect the availability of water resources for the Western United States. Climate warming can significantly impact the hydrology of this region by decreasing the amount of precipitation falling as snow and altering the timing of snowmelt and associated runoff. Therefore, it is essential to characterize how regional climatology affects snow accumulation and ablation and to identify areas that may be especially sensitive to climate warming. This can help resource managers plan appropriately for hydrologic changes. This study utilizes 11-year average (2000 – 2010) MODIS Snow Cover Area (SCA) and Land Surface Temperature (LST) data and annual PRISM precipitation to determine how elevation, slope orientation, latitude, and continentality influence regional characteristics of SCA and LST for early April, early May, early June, and early July in four focus regions: the Colorado Rockies, the Sierra Nevada, the Washington Cascades, and the Montana Rockies. Then, using monthly averages of the 11-year MODIS SCA for January to June, we examine the spatiotemporal evolution of the snowpack and LST throughout the Western U.S. We use threshold values of January to July 11-year average SCA to determine the duration of snow persistence and delineate zones of intermittent, transitional, persistent and seasonal snow. Within the transitional and persistent snow zones, we use 11-year average LST data for January-February-March ( $LST_{JFM}$ ) to categorize five different snow sensitivity zones. Areas with the highest winter average land surface temperatures are assumed to be most sensitive to climate warming, whereas areas with the lowest land surface temperature are assumed to be least sensitive.

Results show that snow cover tends to increase with increasing elevation, and the elevation of snow cover is lower in higher latitudes, maritime environments, and most western slopes. Land surface temperature tends to decrease with increasing elevation, increasing latitude, and tends to be colder on most western slope sites. The largest divergence between eastern and western slope SCA and LST characteristics is observed in the Sierra Nevada, while little divergence is observed in the Colorado Rockies. Snow cover in the Western U.S. is observed predominantly along two main axes: from north to south along the Cascades and the Sierra Nevada, and from northwest to southeast along the axis of the Rocky Mountain Cordillera. The snow line is lowest in the Washington Cascades and highest in the Colorado Rockies; between these two areas a northwest/southeast elevation gradient is observed. The warmest snow zones (warmest JFM<sub>LST</sub>) are at lower elevations of the Cascades/Sierra Nevada and in the southwest, whereas coolest snow zones (coldest JFM<sub>LST</sub>) are in the interior northern Rockies, mid to higher elevations of the Cascades, and the higher elevations of the Colorado Rockies and the Sierra Nevada. The warmest snow zones are likely to be most sensitive to climate warming, as these locations are vulnerable to shifting toward intermittent winter snow cover.

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## LIST OF ABBREVIATIONS

**AVHRR:** Advanced Very High Resolution Radiometer

**CA:** California

**ENSO:** El Niño Southern Oscillation

**GOES:** Geostationary Operational Environmental Satellite

**huc:** Hydrologic Unit Code

**huc100:** 100m Elevation Band Division of a huc

**ID:** Idaho

**ISZ:** Intermittent Snow Zone

**LST:** Land Surface Temperature

$\overline{\text{LST}}$ : Focus Region Land Surface Temperature

$\overline{\text{LST}}_{0^{\circ}\text{C}}$ : Elevation Where Focus Region Land Surface Temperature is Less Than 0°C

$\text{LST}_{\text{JFM}}$ : January-February-March Average Land Surface Temperature

**MODIS:** Moderate Resolution Imaging Spectroradiometer

**msl:** Mean Sea Level

**MT:** Montana

**n:** Length of Date Series

**NASA:** National Aeronautics and Space Administration

**NOHRSC:** National Weather Service National Operational Hydrologic Remote Sensing Center

**NS:** Number of Pixels Classified as No Snow

**NSDI:** Normalized Snow Difference Index

**OR:** Oregon

**p:** Number of Fitted Parameters in the SCoDMod Equation

**P:** Precipitation

**$\bar{P}$ :** Linear Fit to Focus Region Precipitation

**PDO:** Pacific Decadal Oscillation

**PRISM:** Parameter-Regressions on Independent Slopes Model

**PSZ:** Persistent Snow Zone

**Q50:** Day of Year Depletion to 50% SCA

**$\overline{Q50}$ :** Day of Year Depletion to 50%  $\overline{SCA}$

**RMSE:** Root Mean Squared Error

**$R^2$ :** Square of the Correlation Coefficient

**S:** Number of Pixels Classified as Snow

**SCA:** Snow Cover Area

**$SCA_i$ :**  $i^{\text{th}}$  SCA in a Date Series

**$\overline{SCA}$ :** Focus Region Average Snow Cover Area

**$\overline{SCA}_{50\%}$ :** Elevation Where the Focus Region  $\overline{SCA}$  is greater than 50%

**$\overline{SCA}_{90\%}$ :** Elevation Where the Focus Region  $\overline{SCA}$  is greater than 90%

**$SCA_{\text{APR}}$ :** April Snow Cover

**$SCA_{\text{FEB}}$ :** February Snow Cover

**$SCA_{\text{JAN}}$ :** January Snow Cover

**$SCA_{\text{JUN}}$ :** June Snow Cover

**$SCA_{\text{MAR}}$ :** March Snow Cover

**$SCA_{\text{MAY}}$ :** May Snow Cover

**$\overline{SCA}_{\text{max}}$ :** Percentage of Maximum Snow Cover

**$SCA_{\text{SCoDMod}}$ :** SCodMod Modeled Snow Cover

**SCA<sub>YR\_DY</sub>**: Snow Cover Percentage on a Given Day in a Given Year

**SCoDMod**: Snow Cover Depletion Model

**SP**: Snow Persistence

**SSL<sub>1</sub>**: 1<sup>st</sup> Least Sensitive Snow Zone

**SSL<sub>2</sub>**: 2<sup>nd</sup> Least Sensitive Snow Zone

**SSM<sub>1</sub>**: 1<sup>st</sup> Most Sensitive Snow Zone

**SSM<sub>2</sub>**: 2<sup>nd</sup> Most Sensitive Snow Zone

**SSM<sub>3</sub>**: 3<sup>rd</sup> Most Sensitive Snow Zone

**SSM/I**: Special Sensor Microwave/Imager

**SSZ**: Seasonal Snow Zone

**SWE**: Snow Water Equivalent

**TSZ**: Transitional Snow Zone

**USGS**: United States Geological Survey

**UT**: Utah

**WA**: Washington

**Z<sub>SCA\_max</sub>**: Elevation of Maximum  $\overline{SCA}$

**$\sigma_{LST}$** : Land Surface Temperature Standard Deviation

**$\overline{\sigma}_{LST}$** : Focus Region Land Surface Temperature Standard Deviation

**$\sigma_{SCA}$** : Snow Cover Area Standard Deviation

**$\overline{\sigma}_{SCA}$** : Focus Region Snow Cover Area Standard Deviation

## **CHAPTER 1: INTRODUCTION, BACKGROUND AND OBJECTIVES**

### **1.1 INTRODUCTION**

The snowpack is the largest reservoir of water in the western region of the United States, contributing 50-80% of the total water supply (Doesken and Judson, 1997). The Western U.S. depends on this snowmelt runoff to provide water for municipal, industrial, recreational and agricultural needs (Barnett et al., 2008). The location and duration of snow cover affects the local and global energy balance due to the high albedo of snow compared to land surfaces (DeBeer and Pomeroy, 2009; Déry et al., 2005) along with the insulating properties of snow (Déry et al., 2005). In mountain environments, the persistence of snow affects the plant growing season and the amount of water available for soil moisture (Billings and Bliss, 1959). The timing and magnitude of available surface runoff in the Western U.S. depends on the amount of snowpack accumulated in a season and the timing and rate of melt (Barnett et al., 2008; Nolin and Daly, 2006; Stewart, 2009). Therefore, the Western U.S. could be adversely affected by climate warming, which can significantly alter snow environments (CIRMOUNT, 2006).

Already, the Western U.S. has experienced some of the greatest warming in the country, and changes in snowpack and streamflow characteristics have been observed (Stewart, 2009). Many areas of the Western U.S. have documented reductions in April 1<sup>st</sup> snow water equivalent (SWE) (Hamlet et al., 2005; Mote et al., 2005; Stewart, 2009) and an increase in the fraction of precipitation falling as rain rather than snow (Knowles et al., 2006; Pederson et al., 2010; Stewart, 2009), which decreases the amount of water available for spring and summer runoff.

These trends have been particularly apparent in areas characterized by warmer winter and/or spring time air temperatures, and future warming is predicted to further exacerbate these problems (Hamlet et al., 2005; Knowles et al., 2006; Stewart, 2009). Several studies have documented earlier dates of snowmelt onset and earlier hydrograph rise (Cayan et al., 2001; Clow, 2010; Stewart, 2009). Decreases in the duration of the snowcover causes a negative feedback loop due to the corresponding decrease in albedo; thus as snowpacks decrease in response to warmer temperatures, increased absorption of solar radiation causes an intensification of warming trends and further reductions in snowpack (Déry et al., 2005; Stewart, 2009). As such, water resources in the Western United States may see significant changes in response to continued climate warming (Barnett et al., 2008; Huber et al., 2005; Mote et al., 2005; Nolin and Daly, 2006; Stewart, 2009). Such changes will have significant impacts on the western water budget, which will in turn affect environmental, economic, social, and political realms.

## 1.2 BACKGROUND

Measurements of the snowpack are essential for monitoring and modeling snowpack characteristics and trends. Measurements of snowpack occur at a point through manual or automated procedures, or with remote sensing, both of which have their respective advantages and disadvantages. Snow water equivalent (SWE), a measurement common at ground-based point stations, is the amount of water held in a snowpack. From a resource management perspective, the spatial distribution of SWE is a top priority to assess the potential runoff during the melt season. However, across the Western U.S. SWE has traditionally been measured at sparse point locations (Martinec and Rango, 1981; Stewart, 2009), which have to be interpolated

in space in order to estimate the volume of water held in a snowpack (Cline et al., 1998). Since SWE can exhibit great variability across space, especially in mountain regions (DeBeer and Pomeroy, 2009; Déry et al., 2005; Elder et al., 1991), this can lead to large errors in estimates of SWE. Furthermore, snowfall measured with rain gauges may significantly under-represent the amount of actual snowfall due to undercatch (Fassnacht, 2004). Additionally, there is a lack of snowpack measurements above the tree line (Mizukami and Perica, 2008), an area that can receive high amounts of snow accumulation. Therefore, snowpack in the mountainous regions of the Western United States may be substantially undersampled through point measurement techniques (CIRMOUNT, 2006; Stewart, 2009).

Temperature measurements are essential for many modeling and modeling applications, especially those concerning climate change. Point measurements of temperature demonstrate many of the same disadvantages as point snowpack measurements. Temperature is highly variable over space, yet climate stations measuring temperature in the Western U.S. are relatively sparse and are often situated in valley bottoms (Barry, 2008; Cayan, 1996). Most weather stations measure air temperature; there are very few ground based stations which measure surface temperature (Coll et al., 2005). Despite the relatively sparse ground based network of snow and temperature measurements, these measurement are often employed in studies describing regional climatology and climate change trends (Armstrong and Armstrong, 1987; Cayan, 1996; Clow, 2010; Knowles et al., 2006; Mizukami and Perica, 2008; Mock, 1996; Mote et al., 2005; Serreze et al., 1999).

Remote sensing offers exciting opportunities to expand and improve upon snowpack monitoring on a global scale, but it is not without its drawbacks. Remote sensing is useful to monitor the seasonal (Martinec and Rango, 1986) and long term trends (Robinson and Frei,



2000) in snow cover. Although remote sensing can provide continuous measurements, tradeoffs must be made between spatial and temporal resolution. Higher temporal resolution is typically associated with reduced spatial resolution and vice versa. Passive microwave remote sensing has been used to measure SWE, but it has a very coarse spatial resolution (e.g. ~25km grid cell for the SSM/I product) and presents difficulties in measurements of wet snowpacks or snow in mountainous terrain (Blyth, 1993; Grody and Basist, 1996). Snow covered area (SCA) provides a useful measurement for several snowpack monitoring and modeling applications. SCA is easily observable because snow has high reflectance in the visible, thermal infrared, and microwave wavelengths of the electromagnetic spectrum (Hall and Martinec, 1985). Although SCA measurements do not supply information on the volume of water held in a snowpack, they do provide information on the spatial distribution of snow, which is important for distributed and semi-distributed hydrologic modeling (Cline et al., 1998; DeBeer and Pomeroy, 2009; DeWalle and Rango, 2008; Déry et al., 2005; Martinec and Rango, 1981) and global atmospheric modeling (Robinson and Frei, 2000).

Several snow cover products are available; two commonly used snow cover products are from the National Weather Service National Operational Hydrologic Remote Sensing Center (NOHRSC) and NASA's Moderate Resolution Imaging Spectroradiometer (MODIS). NOHRSC is a modeled snow cover product produced daily at a 1km resolution using imagery from the Advanced Very High Resolution Radiometer (AVHRR) and the Geostationary Operational Environmental Satellite (GOES) (Maurer et al., 2003). MODIS provides snow cover using the Normalized Snow Difference Index (NSDI), which is calculated as the differences between surface reflectance observed in visible (0.545 – 0.565 $\mu$ m) and near infrared (1.628-1.652 $\mu$ m) wavelengths divided by the sum of these reflectances. MODIS data are provided at a 500 m grid

resolution, daily and as an 8-day composite, from both the Terra (available since 2000) and Aqua (available since 2002) satellite platforms (Hall et al., 2002). Maurer (2003) found the daily MODIS data product to be superior to the NOHRSC product, classifying fewer pixels as cloud and more accurately classifying snow/no-snow pixels based on comparison to ground-based station observations of SWE.

Cloud cover obscuration is one of the main disadvantage of remotely sensed SCA (Parajka and Blöschl, 2008; Parajka et al., 2010; Robinson and Frei, 2000). For MODIS imagery cloud cover can be greatly reduced by using imagery from the Terra and Aqua satellite platforms, which pass over the same geolocation approximately 12 hours apart, to create a single image composite (Parajka and Blöschl, 2008) or by using the 8-day composite MODIS SCA product (Hall et al., 2002).

In addition to the MODIS SCA products, MODIS also provides daily measurements of land surface temperature (LST), which could be useful for energy balance modeling and monitoring global climate change (Justice et al., 1998). The product is sensitive to cloud temperature, and therefore cloud covered pixels are excluded from the product. The MODIS land surface temperature product is based on measurements of thermal radiation (within the 3.5 - 4.2  $\mu\text{m}$  range) emitted from the land surface. The land surface temperature data product has been available since 2000. Data are provided at a 1km gridded resolution for daily or 8-day time intervals. The accuracy of the product has been tested by comparing remotely sensed MODIS LST measurements to in-situ LST measurements and was found to have a high degree of accuracy ( $\pm 1^\circ\text{K}$ ) (Wan, 2008; Wan et al., 2004). Thus the product can provide an overview of the spatiotemporal characteristics of land surface temperature. Since a decade-long MODIS record exists, which has continuous spatial coverage, this study utilizes MODIS derived snow

cover area and land surface temperature, rather than point station data, to analyze the spatiotemporal characteristics of snow and land surface temperature and to assess the sensitivity of seasonally snow covered areas to climate change in the Western United States.

### 1.3 OBJECTIVES

The primary objectives of this research are to (1) determine the spatiotemporal variability of snow cover and land surface temperature across the Western United States, (2) identify and map snow zones with similar characteristics, and (3) identify snow covered areas that may be most sensitive to climate change. To meet these objectives, we investigate the spatial and temporal patterns of MODIS snow covered area and land surface temperature as they relate to physiographic characteristics across the Western United States. Within smaller focus regions, we examine how snow covered area, land surface temperature, and precipitation vary with elevation, maritime versus continental climates, higher-mid latitude versus mid-latitude environments, and eastern versus western slopes. Then, for the entire Western U.S., we use average winter and springtime snow cover to identify zones of persistent, transitional, intermittent, and seasonal snowpacks, determine the spatiotemporal distribution of land surface temperature within seasonally snow covered areas, and use average wintertime land surface temperature to identify which snow zones may be sensitive to climate warming.

## CHAPTER 2: DATA AND METHODS

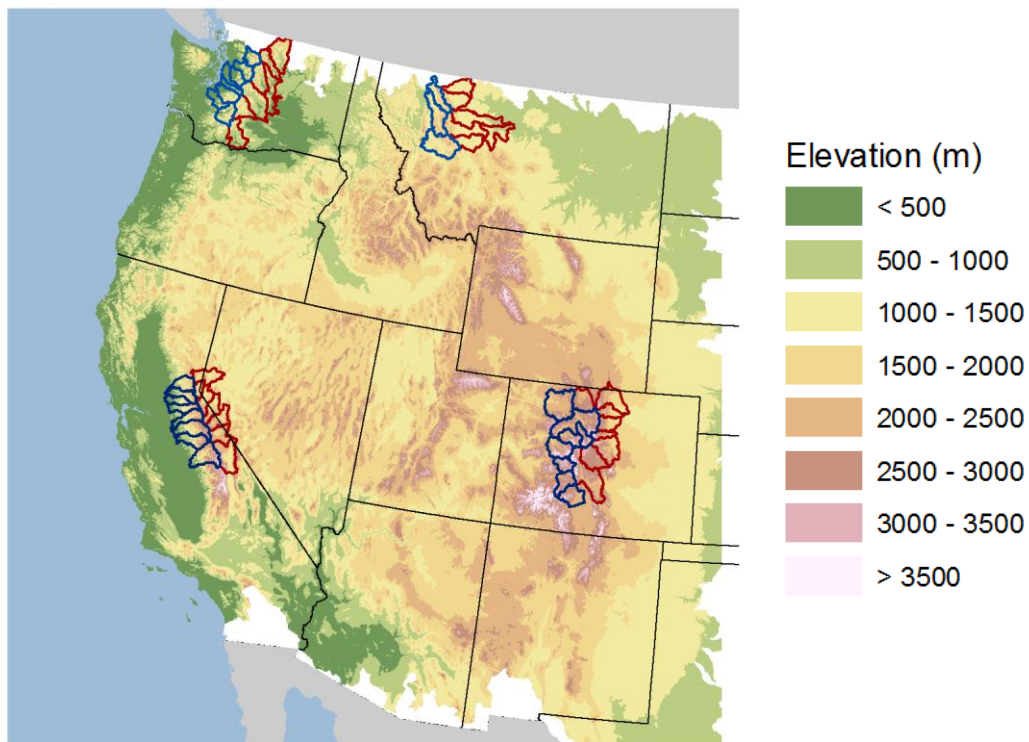
### 2.1 STUDY AREA

The study area includes basins located within Western United States west of the 100<sup>th</sup> meridian (Figure 2.1). The area is divided into hydrologic basins defined by the 8-digit USGS hydrologic unit code (huc); only basins that lie entirely within the United States border are included in analyses. Each basin is divided into 100 m elevation contours; these elevation zones are the spatial unit used for analyses in this study, and they are referred to as huc100. The huc100 spatial unit is chosen to analyze the influence of elevation on snow cover, land surface temperature, and precipitation in the Western U.S. Elevation of the Western United States study area ranges from 86 m below mean sea level (msl) in Death Valley, California, to 4,421 m above msl, a mere 120 km away at the peak of Mt. Whitney (Figure 2.1).

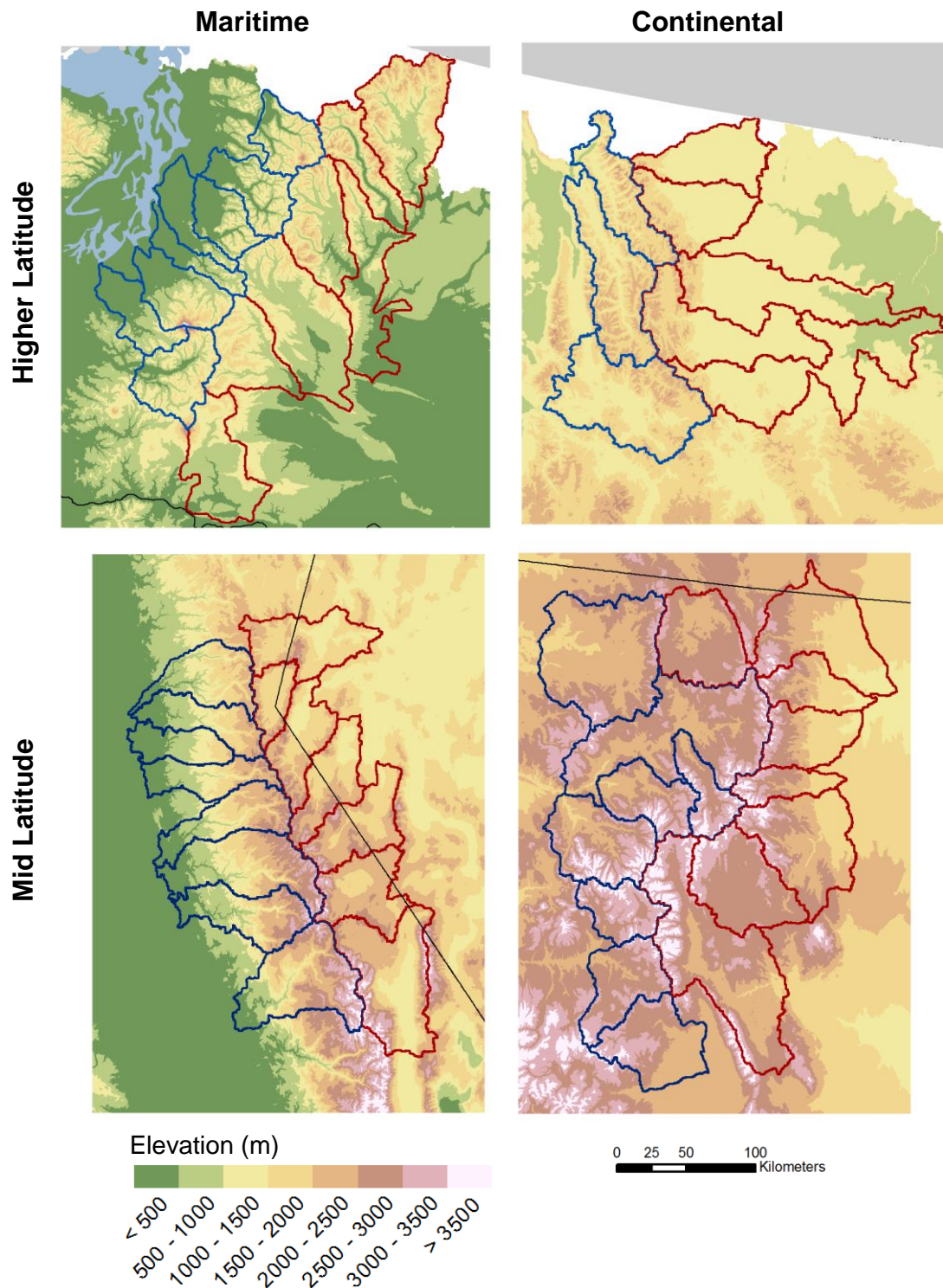
Throughout the Western U.S. precipitation and temperature are highly variable; Figure 2.3 shows the average annual precipitation and average annual temperature for 1971 to 2000 from PRISM. As a whole, the Western U.S. boasts a wide variety of climates, including coastal environments, deserts, montane regions, and rainforest. Continental snow environments tend to have cold, dry winters with shallower, less dense snowpacks while maritime snow environments are typically characterized by milder winters and deeper, denser snowpacks (Armstrong and Armstrong, 1987; Mizukami and Perica, 2008; Serreze et al., 1999).

Within the Western U.S. study region, four areas are defined for in-depth analyses: the Colorado Rockies, Sierra Nevada, Washington Cascades, and Montana Rockies (Figure 2.1;

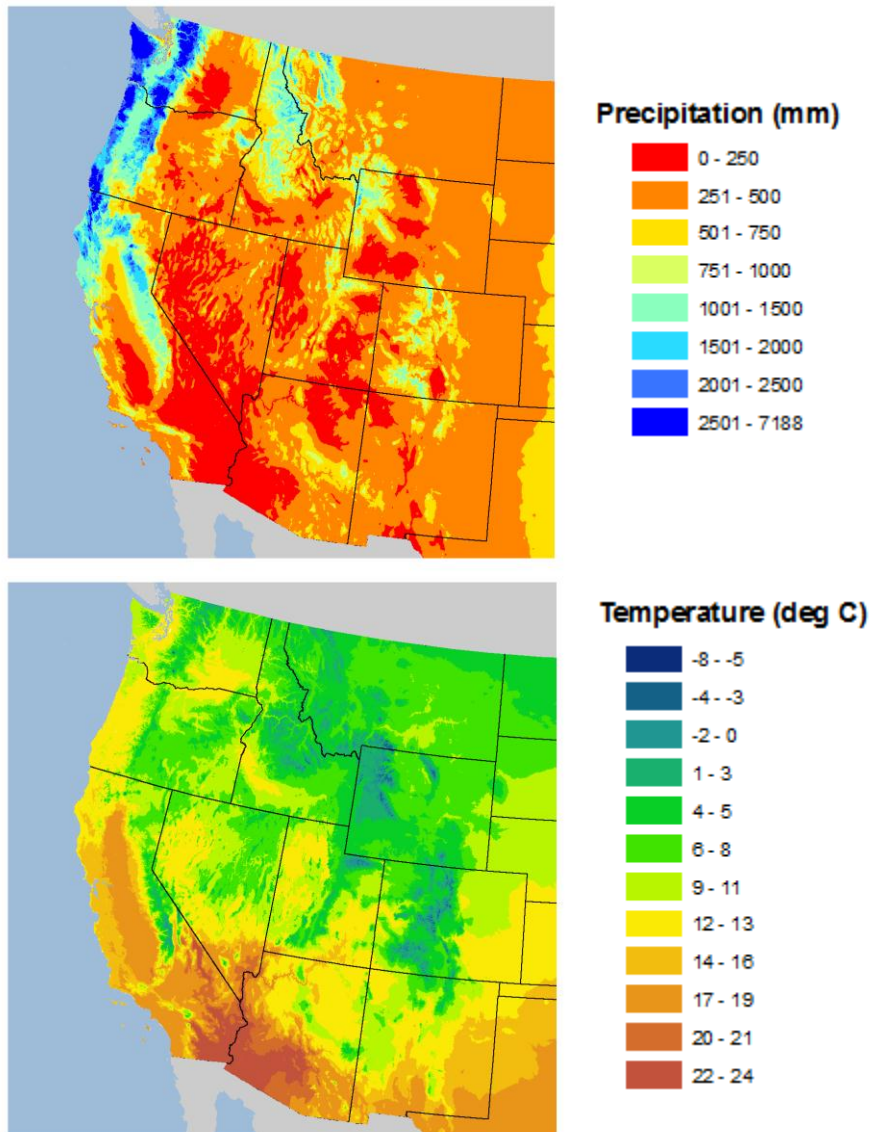
Figure 2.2). These in-depth areas are chosen to represent examples of four different climate zones, selected to illustrate the influence of latitude and continentality on snow cover, land surface temperature, and precipitation in the Western U.S. These in-depth areas are further subdivided into western and eastern slope regions, referred to in this paper as focus regions, to examine the effects of primary slope orientation in each area. Focus region huc statistics are included in Appendix A.



**Figure 2.1** Elevation of Western United States study areas. Eastern slope focus region hucs are outlined in red, and western slope focus region hucs are outlined in blue.



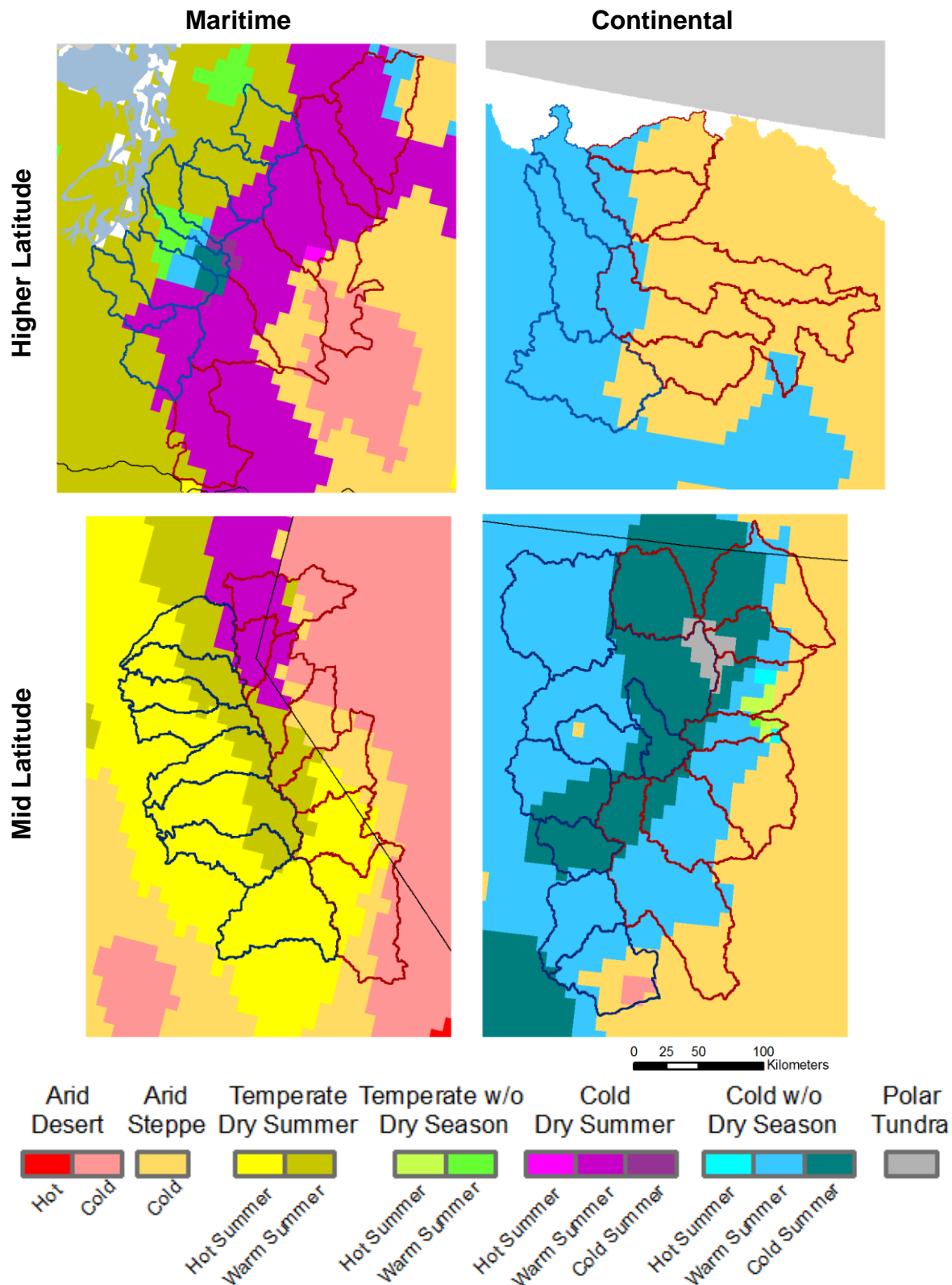
**Figure 2.2** Elevation of the (a) Washington Cascades (b) Montana Rockies (c) Sierra Nevada and (d) Colorado Rockies focus study regions. Western slope focus regions are shown in blue, and eastern slope focus regions are shown in red.



**Figure 2.3** Average annual precipitation and average annual temperature (1971-2000) in the Western United States. Source: Oregon State University PRISM model (<http://www.prism.oregonstate.edu/>).

The Köppen-Geiger climate classifications for the focus regions are displayed in Figure 2.4 (Peel et al., 2007). The Sierra Nevada is largely classified as a temperate climate with a hot or warm dry summer, especially along the western portions, but also includes cold climates with a warm dry summer, as well as cold arid steppe/desert. The western portion of the Washington Cascades is classified primarily as temperate with a dry/hot summer, while the eastern portion is cold with a dry warm summer and arid steppe. The Montana Rockies are located at a relatively high latitude and exhibit both maritime and continental climate characteristics, with the western slope being a predominantly maritime climate while the eastern slope is predominantly continental (Finklin, 1986). The Köppen-Geiger climate classification for the Montana Rockies focus regions is a cold climate without a dry season and a warm summer on the western portion and on the eastern portion the climate is characterized as cold with a dry winter and a hot summer (Peel et al., 2007). The Colorado Rockies are a predominantly cold climate without a dry season and with a warm or cold summer, with highest areas classified as polar tundra. Near Denver, Colorado the climate is classified as temperate without a dry season and a hot or warm summer, and cold without a dry season with a hot summer. Further east along the great plains region of Colorado is classified as cold arid steppe with a small area of cold arid desert.





**Figure 2.4** Köppen-Geiger climate classification (Peel et al., 2007) for the (a) Washington Cascades (b) Montana Rockies (c) Sierra Nevada and (d) Colorado Rockies focus study regions. Western slope focus regions are shown in blue, and eastern slope focus regions are shown in red.

## 2.2 DATA

Within the study region, we examined spatiotemporal differences of MODIS snow covered area, MODIS land surface temperature, and the spatial differences in precipitation as estimated by the PRISM model. Focus regions are utilized to facilitate in-depth understanding of regional patterns and the Western U.S. region is analyzed to assess wider-scale patterns.

### 2.2.1 *Snow Covered Area*

The SCA data used in this study are the NASA MODIS/Terra Snow Cover 8-Day L3 Global 500m Grid, Version 5 (Product code MOD10A2) available from the National Snow and Ice Data Center (NSIDC <<http://nsidc.org/data/>>) covering the time period of January 1<sup>st</sup> to August 12<sup>th</sup> for 2000 to 2010. The Terra MODIS satellite did not start collecting data until February 26<sup>th</sup> of 2000, so there is no snow cover data prior to that time. The MODIS SCA product is derived with the normalized snow difference index (NSDI), which divides the difference between the surface reflectance measured by band 4 (0.545 – 0.565 $\mu$ m) and band 6 (1.628-1.652 $\mu$ m) by the sum of band 4 and band 6:

$$\text{NSDI} = \frac{(\text{Band 4} - \text{Band 6})}{(\text{Band 4} + \text{Band 6})} \quad (2.1)$$

A pixel is considered snow if reflectance of band 2 (0.841 – 0.876 $\mu$ m) is greater than 11% and the NSDI is greater than 0.4 (Hall et al., 2002). The algorithm classifies each pixel as snow, no snow, cloud, lake ice, ocean, detector saturated, fill, night, no decision, and data missing. The 8-day data product classifies a pixel as snow covered if any day within the eight day period is covered in snow. The data are available in a gridded sinusoidal projection at a 500m gridded spatial resolution. Several quality assessments studies of the MODIS SCA product have found it to be an accurate and reliable data product (Hall and Riggs, 2007; Hall et al., 2002). Although a

daily MODIS SCA data product is available, the 8-day composite is used in this study due to the marked decrease in cloud cover, which is the primary limitation of the MODIS snow cover products (Parajka and Blöschl, 2008; Parajka et al., 2010; Robinson and Frei, 2000).

### 2.2.2 *Land Surface Temperature*

The land surface temperature data product used in this study is the MODIS Land Surface Temperature and Emissivity 8-Day L3 Global 1km product (Product code MOD11A2) provided by NASA for the time period of January 1<sup>st</sup> to August 12<sup>th</sup> for 2000 to 2010. As mentioned previously, there are no land surface temperature data prior to February 26<sup>th</sup> 2000 from the Terra satellite sensor. MODIS Land Surface Temperature is determined by measuring the thermal radiation emitted by the land surface in clear sky conditions with bands 31 (10.780 – 11.280  $\mu\text{m}$ ) and 32 (11.770 – 12.270  $\mu\text{m}$ ) and employing a generalized split window algorithm, described in detail in Wan (2004). These data have been validated as reliable for scientific use and are considered accurate within 1°C (Wan, 2008; Wan et al., 2004). These data provide 8-day average day time land surface temperatures at a 1km gridded resolution.

### 2.2.3 *Precipitation*

The precipitation values used in this study were derived by the Parameter-elevation Regression on Independent Slopes Model (PRISM) developed by Oregon State University (<<http://www.prism.oregonstate.edu/>>). PRISM provides annual average precipitation from 1971 to 2000 at an 800 m gridded resolution. The PRISM model uses point precipitation data from weather stations and interpolates precipitation over space by taking into consideration the influences of topography (Neilson and Phillips, 1994). It is considered a high quality product

and has been implemented in many climatological and hydrological studies (Nolin and Daly, 2006).

### 2.3 DATA PROCESSING AND PREPARATION

MODIS snow cover and land surface temperature data for the Western U.S. consists of 9 separate tiles, which are mosaicked and reprojected into the USGS version of the USA Contiguous Albers' equal area conic projection using IDL. This process is repeated for each 8-day image in the 2000-2010 dataset. Then, for each 8-day image, the SCA percentage is calculated for every huc100, (the 100 m elevation zones that are the spatial unit of analysis for this study), by Equation 2.2:

$$SCA_{YR\_DY} = \frac{S}{(S+NS)} * 100\% \quad (2.2)$$

where  $SCA_{YR\_DY}$  is the snow cover percentage on a certain day for a certain year,  $S$  is the number of pixels classified as snow, and  $NS$  is the number of pixels classified as no snow. If a pixel is classified as cloud, lake ice, ocean, detector saturated, fill, night, no decision, or data missing it is not included in analyses. If a huc100 contains greater than 20% cloud cover, it is excluded from the results.

Land surface temperature for a certain day of year for the period from 2000 to 2010 is calculated as the arithmetic mean of all pixel values within each huc100. Cloud cover is not taken into consideration due to the fact that MODIS does not calculate LST for cloud covered pixels.

Data analyses in this study consider both the 11-year average characteristics and annual variability in SCA. For both land surface temperature and snow covered area, four 8-day time periods are used for temporal comparison of the ablation period across the focus regions: March

30<sup>th</sup> to April 6<sup>th</sup>, referred to as early April; May 1<sup>st</sup> to May 8<sup>th</sup>, referred to as early May; June 2<sup>nd</sup> to the June 9<sup>th</sup>, referred to as early June, and July 4<sup>th</sup> to July 11<sup>th</sup>, referred to as early July.

These analyses use both the average and standard deviation of snow covered area for each of the selected 8 day periods (early April, early May, early June, and early July) from 2000 to 2010, which are determined for every huc100, and are referred to as  $SCA$  and  $\sigma_{SCA}$ , respectively. Focus region comparisons incorporate an additional processing step to facilitate regional comparisons. Each focus region contains several different hucs; to estimate the aggregated characteristics over the entire focus region,  $SCA$  and  $\sigma_{SCA}$  for all hucs are binned and averaged by 100 m elevation zone, referred to as  $\overline{SCA}$  and  $\overline{\sigma}_{SCA}$ . To compare  $\overline{SCA}$  across focus regions, the elevation zone where  $\overline{SCA}$  is greater than 50% and 90% are derived, referred to as  $\overline{SCA}_{50\%}$  and  $\overline{SCA}_{90\%}$ , respectively. On average  $\overline{SCA}_{50\%}$  is chosen to represent elevations that have a 50% probability of being snow covered on a specific date, whereas  $\overline{SCA}_{90\%}$  represents areas that maintain close to full snow cover on a specific date. Additionally, the elevation and percentage where  $\overline{SCA}$  is at a maximum is determined ( $Z_{\overline{SCA}_{max}}$  and  $\overline{SCA}_{max}$ , respectively). The variable  $\overline{\sigma}_{SCA}$  is used to evaluate when and where  $\overline{SCA}$  exhibits high interannual variability across and within focus regions.

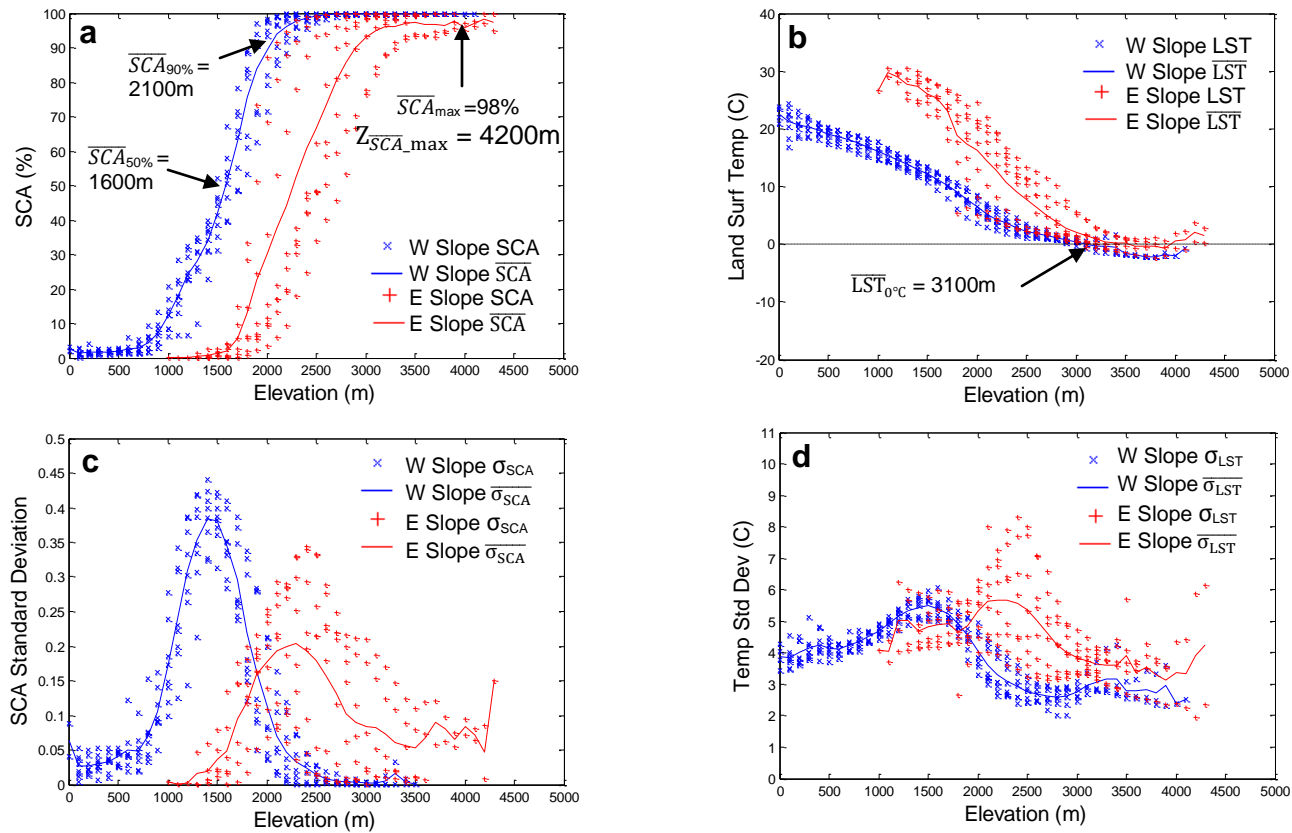
A similar procedure is employed to process land surface temperature regionally and within focus regions. Like snow covered area, the average and standard deviation for land surface temperature for each 8 day period from 2000 to 2010 is determined for every huc100, referred to as  $LST$  and  $\sigma_{LST}$ , respectively. Within each focus region,  $LST$  and  $\sigma_{LST}$  are binned and averaged across 100 m elevation zones, referred to as  $\overline{LST}$  and  $\overline{\sigma}_{LST}$ . From these averaged values, a comparative metric for the land surface temperature is derived by taking the elevation zone where  $\overline{LST}$  is less than 0°C ( $\overline{LST}_{0^{\circ}C}$ ). Similar to  $\overline{\sigma}_{SCA}$ , the variable  $\overline{\sigma}_{LST}$  is useful in showing

spatiotemporal patterns in the focus region land surface temperature standard deviation. Figure 2.5 depicts an example of  $SCA$ ,  $\overline{SCA}$ ,  $\overline{SCA}_{50\%}$ ,  $\overline{SCA}_{90\%}$ ,  $Z_{\overline{SCA}_{max}}$ ,  $\sigma_{SCA}$ ,  $\overline{\sigma}_{SCA}$ ,  $LST$ ,  $\overline{LST}$ ,  $\overline{LST}_{0^{\circ}C}$ ,  $\sigma_{LST}$  and  $\overline{\sigma}_{LST}$  versus elevation for early April in the Sierra Nevada, and variables are summarized in Table 2.1.

**Table 2.1** Snow cover and land surface temperature variables for the study areas.

<b>Variable</b>	<b>Definition</b>
$LST$	11 year-average land surface temperature for each huc100
$\overline{LST}$	11 year-average land surface temperature binned by 100 m elevation band for each focus region for each 8-day period
$\overline{LST}_{0^{\circ}C}$	Elevation where $\overline{LST}$ is less than $0^{\circ}C$
$SCA$	11 year-average snow cover area for each huc100
$\overline{SCA}$	11 year-average snow cover area binned by 100 m elevation band for each focus region for each 8-day period
$\overline{SCA}_{50\%}$	Elevation where $\overline{SCA}$ is greater than 50%
$\overline{SCA}_{90\%}$	Elevation where $\overline{SCA}$ is greater than 90%
$\overline{SCA}_{max}$	Percentage of maximum $\overline{SCA}$
$Z_{\overline{SCA}_{max}}$	Elevation where $\overline{SCA}$ at a maximum
$\sigma_{LST}$	11 year land surface temperature standard deviation for each huc100
$\overline{\sigma}_{LST}$	11 year land surface temperature standard deviation binned by 100 m elevation band for each focus region for each 8-day period
$\sigma_{SCA}$	11 year snow cover area standard deviation for each huc100
$\overline{\sigma}_{SCA}$	11 year snow cover area standard deviation binned by 100 m elevation band for each focus region for each 8-day period

The PRISM model is used to determine precipitation patterns across the Western U.S. and to compare precipitation versus elevation characteristics between focus regions. The PRISM 1971 – 2000 average annual precipitation ( $P$ ) are binned by 100 m elevation bands within the focus regions, and a linear trend line is fit to each focus region (referred to as  $\overline{P}$ ) to describe the relationship between precipitation versus elevation.



**Figure 2.5** Examples of SCA and LST metrics for the western and eastern slope Sierra Nevada focus region in early April, 2000-2010. (a) Average snow covered area (SCA), focus region average snow covered area ( $\overline{SCA}$ ), elevation where  $\overline{SCA} > 50\%$  ( $\overline{SCA}_{50\%}$ ), elevation where  $\overline{SCA} > 90\%$  ( $\overline{SCA}_{90\%}$ ), and elevation ( $Z_{\overline{SCA}_{\max}}$ ) and percentage ( $\overline{SCA}_{\max}$ ) of focus region maximum snow covered area, (b) Average land surface temperature (LST), focus region average land surface temperature ( $\overline{LST}$ ), and the elevation where is  $\overline{LST}$  crosses the  $0^{\circ}\text{C}$  threshold ( $\overline{LST}_{0^{\circ}\text{C}}$ ), (c) Snow covered area standard deviation  $\sigma_{SCA}$  and focus region snow covered area standard deviation  $\overline{\sigma}_{SCA}$ , (d) Land surface temperature standard deviation ( $\sigma_{LST}$ ) and the focus region standard deviation ( $\overline{\sigma}_{LST}$ ).

## 2.4 SNOW COVER DEPLETION MODEL

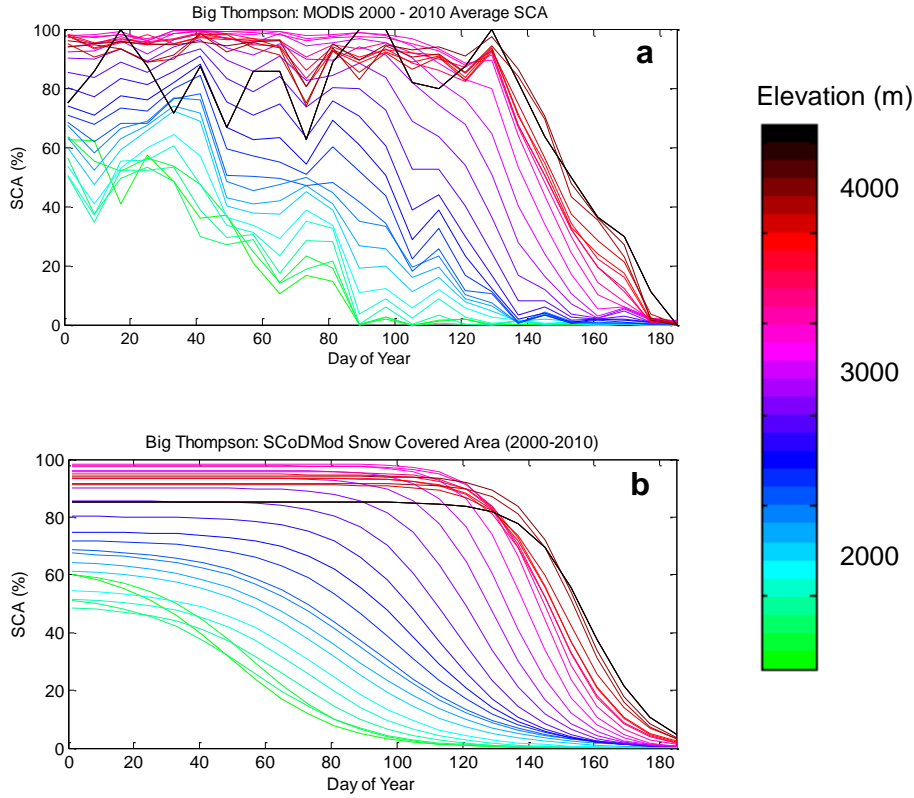
The data analyses described in Section 2.3 focus on particular dates during the ablation period, and this study is also concerned with continuous snow cover depletion characteristics. Figure 2.6a shows an example of SCA calculated for huc100's in the Big Thompson huc in Colorado for each 8-day period. The nature of the 8-day maximum snow extent can make annual data noisy, especially in lower elevation zones with intermittent snow cover, where spring snow storms can create abrupt increases in snow cover with limited SWE that usually melts quickly after storm events. This noise can be propagated into the eleven year average, causing spikes in the average and making it difficult to determine melt timing metrics, specifically the day of year for 50% snow cover loss.

We therefore developed a Snow Cover Depletion Model, termed the SCoDMod, and applied it to the MODIS 8-day SCA data from 2000 to 2010 to address these issues. The SCoDMod is a modified sigmoid function fit to 2000 -2010 average snow cover from January 1<sup>st</sup> through August 12<sup>th</sup> for each huc100 and is represented by Equation 2.3:

$$SCA\% = p1 / (1 + \exp\left(\frac{t-p2}{p3}\right)) \quad (2.3)$$

where SCA% is the percent snow covered area, t is the day of year starting with Jan 1<sup>st</sup> as day 1, and p1, p2, and p3 are fitted parameters. The SCoDMod smooths the average SCA decrease curve, thus highlighting the ablation period snow cover loss characteristics each huc100. Every SCoDMod function for each huc100 is calculated by running a MATLAB code, which executes a nonlinear regression on the 11 year average data and determines parameters based on a nonlinear regression least squares curve fitting algorithm. Using SCoDMod, we can derive objective indices of snow cover loss timing that are less affected by the intermittent spring snow





**Figure 2.6** (a) MODIS 2000 – 2010 average snow covered area versus day of year, and (b) SCoDMod derived average snow covered area versus day of year for the Big Thompson basin, Colorado.

events that cause spikes in the loss curve. The climatic smoothing of the raw MODIS data are apparent in Figure 2.6b, which shows the results of the SCoDMod derived SCA fit to the Big Thompson huc in Colorado. The goodness of the SCoDMod fit is determined by the root mean square error (RMSE), which is calculated as (Equation 2.4):

$$RMSE = \sqrt{\frac{\sum_{i=1}^n (SCA_i - SCA_{SCoDMod})^2}{(n-p)}} \quad (2.4)$$

where  $SCA_i$  is the  $i^{\text{th}}$  SCA in the date series,  $SCA_{SCoDMod}$  is the SCA modeled by the fitted SCoDMod equation,  $n$  is the length of the date series, and  $p$  is equal to 3, the number of fitted parameters. The SCoDMod RMSE for the focus regions is shown in Figure 2.7. Once

SCoDMod is fit to 2000 to 2010 average snow cover data for each huc100, we determine quantile metrics indicating the day of year that the SCA reaches a value of 50% by rearranging the SCoDMod equation as (Equation 2.5):

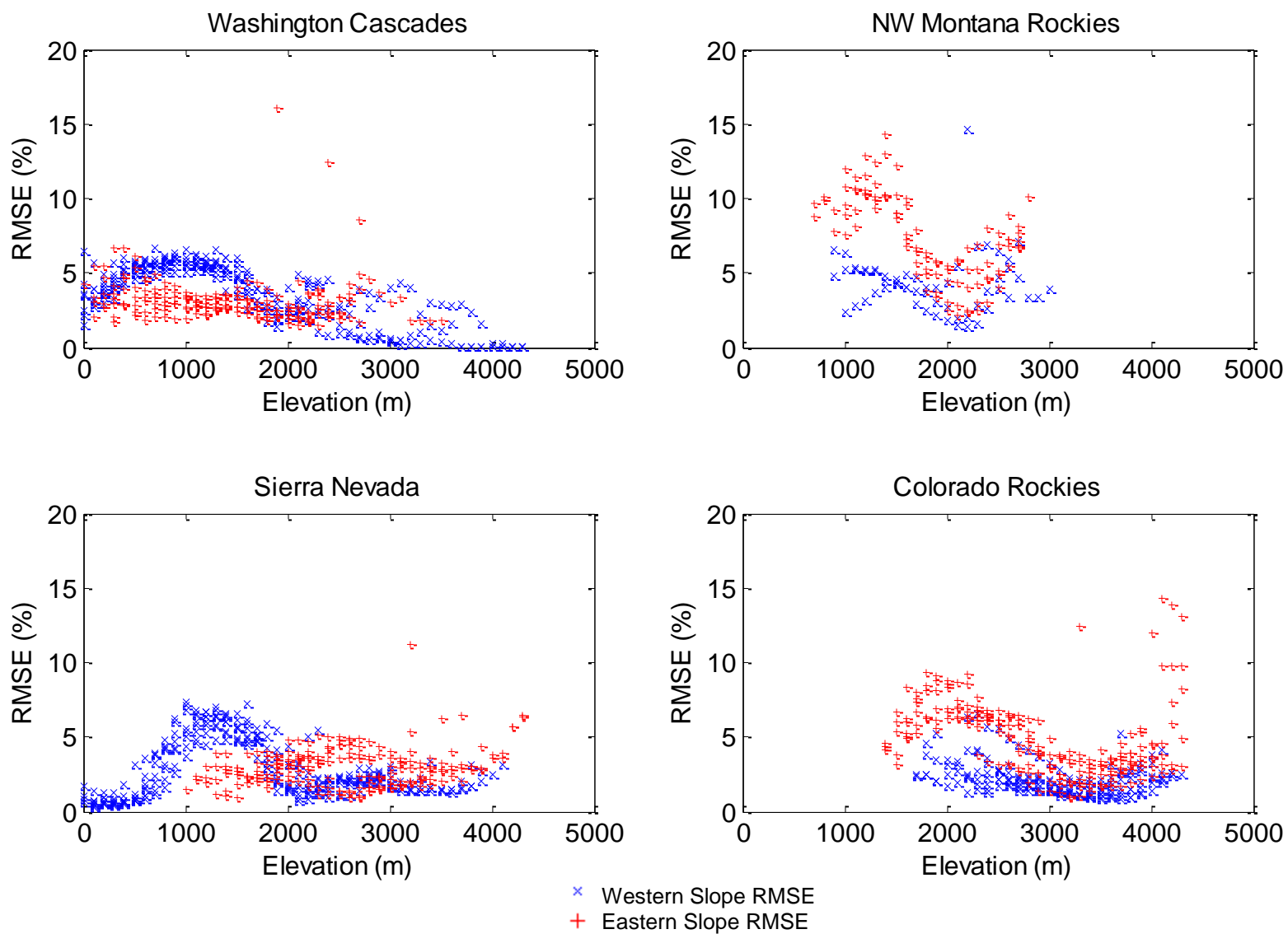
$$Q50 = p3 * \ln\left(\frac{p1}{SCA} - 1\right) + p2 \quad (2.5)$$

where Q50 is day of year that the snow cover in a huc100 reaches 50%, on average. The Q50 variable is binned and averaged across 100 m elevation zones for each focus region (referred to as  $\overline{Q50}$ ) and used to analyze depletion patterns between and within each focus region.

## 2.5 SNOW AND TEMPERATURE ZONES

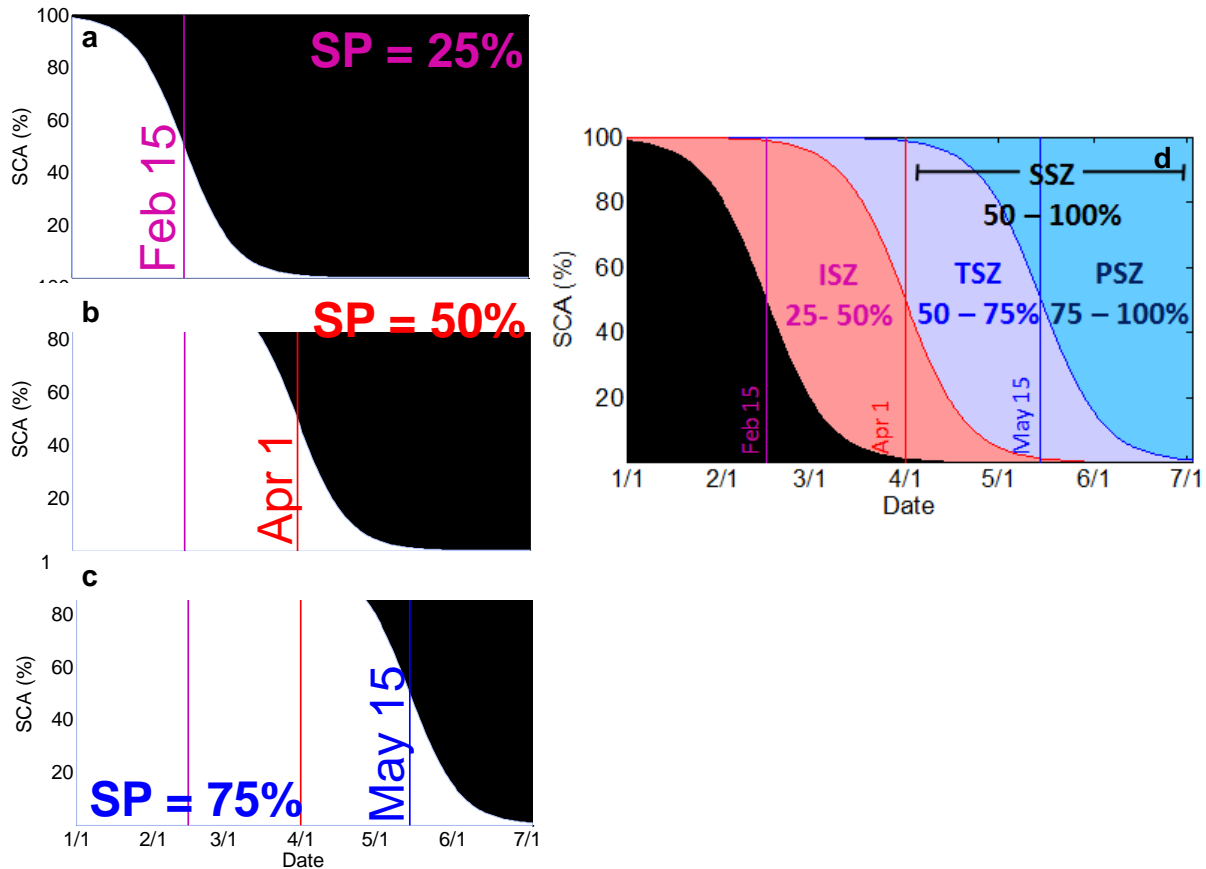
### 2.5.1 *Western U.S. Analyses of Snow Cover and Snow Zones*

Monthly snow cover maps are determined by averaging the 2000-2010 average SCA over a month, which effectively means averaging four 8-day consecutive time periods for each month from January to July for each huc100. These times periods are January 1<sup>st</sup> to February 1<sup>st</sup>, referred to as January snow cover area (SCA<sub>JAN</sub>), February 2<sup>nd</sup> to March 5<sup>th</sup>, referred to as February snow cover area (SCA<sub>FEB</sub>), March 6<sup>th</sup> to April 6<sup>th</sup>, referred to as March snow cover area (SCA<sub>MAR</sub>), April 7<sup>th</sup> to May 8<sup>th</sup>, referred to as April snow cover area (SCA<sub>APR</sub>), May 9<sup>th</sup> to June 9<sup>th</sup>, referred to as May snow cover area (SCA<sub>MAY</sub>), and June 10<sup>th</sup> to July 11<sup>th</sup>, referred to as June snow cover area (SCA<sub>JUN</sub>). Irregular start and stop dates reflect the timing of the 8-day MODIS product, which does not always provide data that correspond with the first day in each month (each time period is 32 days long).



**Figure 2.7** Root mean squared error (RMSE) of SCoDMod fit to SCA in focus regions in the Western United States. Western slope is shown in blue, and eastern slope is shown in red.

In this study, we calculate the average SCA for the 11 years from Jan 1st to July 3<sup>rd</sup> (23 8-day periods) and call this snow cover persistence (SP) for each huc100. By using this calculation, areas with SP greater than 50% are typically snow covered until April 1<sup>st</sup> on average, which is the mid-point of the time period. Additionally, April 1<sup>st</sup> is typically considered to be close to the date of maximum snow accumulation (Mote et al., 2005; Serreze et al., 1999). We then define four snow zones based on SP, summarized in Table 2.2: the intermittent snow zone (ISZ), the persistent snow zone (PSZ), the transitional snow zone (TSZ), and the seasonal snow zone (SSZ). If we assume an idealized snow loss curve, the intermittent snow zone represents areas that ablate to 50% snow cover between February 15<sup>th</sup> and April 1<sup>st</sup> (Figure 2.8a,d). The ISZ can also include areas subject to with frequent short lasting snow events. The persistent snow zone includes areas that, on an idealized snow loss curve, melt to 50% snow cover on or after May 15<sup>th</sup>, as well as areas that may have snow persistence year round (Figure 2.8c,d). The transitional snow zone lies between the PSZ and the ISZ corresponds to areas would melt to 50% snow cover on the idealized curve between April 1<sup>st</sup> and May 15<sup>th</sup> (Figure 2.8b,d). The seasonal snow zone (SSZ) is a combination of the latter two snow zones and represents areas which melt on the idealized curve to 50% snow cover on or after April 1<sup>st</sup> (Figure 2.8b,d). Elevation of the snow zones are used to illustrate differences in snow characteristics across the Western U.S. Areas within the SSZ are those with high potential to generate snowmelt runoff.



**Figure 2.8** Conceptual model illustrating snow persistence and snow zone definitions. Assuming an idealized snow cover depletion curve, snow persistence (SP) of 25% corresponds to areas that melt to 50% snow cover on February 15<sup>th</sup> (a), snow persistence (SP) of 50% corresponds to areas that melt to 50% snow cover on April 1<sup>st</sup> (b), snow persistence (SP) of 75% corresponds to areas that melt to 50% snow cover on March 15<sup>th</sup> (c). Right image shows idealized snow loss curves for the intermittent snow zone (ISZ), transitional snow zone (TSZ), persistent snow zone (PSZ) and seasonal snow zone (SSZ) (d).

**Table 2.2** Snow zone definitions based on January 1<sup>st</sup> – July 3<sup>rd</sup> snow persistence (SP) for the Western United States.

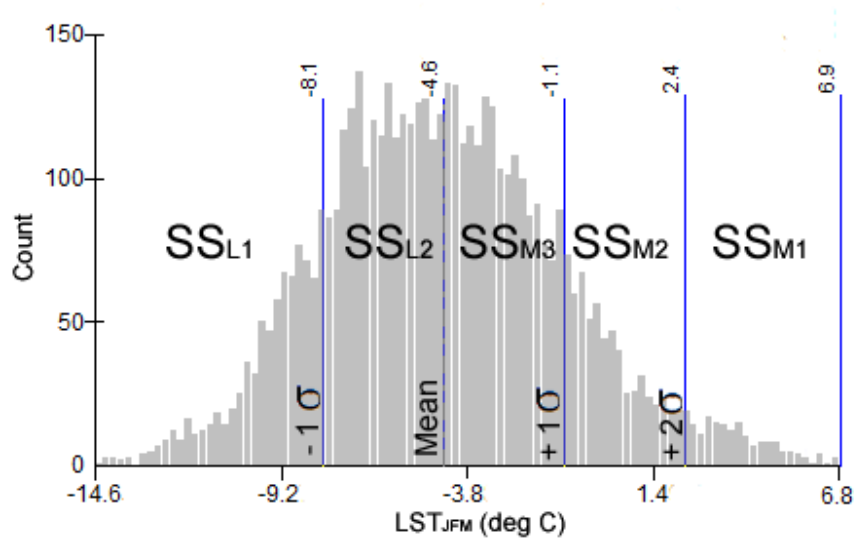
	Greater than SP (%)	Less than SP (%)
<b>Intermittent Snow Zone (ISZ)</b>	25	50
<b>Transitional Snow Zone (TSZ)</b>	50	75
<b>Persistent Snow Zone (PSZ)</b>	75	100
<b>Seasonal Snow Zone (SSZ)</b>	50	100

### 2.5.2 Sensitive Snow Zones

The 2000 – 2010 average land surface temperature within the SSZ is averaged from January 1<sup>st</sup> to March 29<sup>th</sup> (referred to as  $LST_{JFM}$ ) and is used as a proxy to identify seasonally snow covered environments that may be sensitive to climate warming. In this study we use temperature thresholds based on the mean and standard deviation of  $LST_{JFM}$  to define the sensitive snow zones. Although the data are slightly skewed toward higher temperatures (Figure 2.9), the spread of  $LST_{JFM}$  in the SSZ is nearly normal and has a mean of  $-4.6^{\circ}\text{C}$  and a standard deviation of  $3.5^{\circ}\text{C}$ . We define 5 sensitive snow zones: the 1<sup>st</sup> most sensitive snow zone ( $SS_{M1}$ ), the 2<sup>nd</sup> most sensitive snow zone ( $SS_{M2}$ ), the 3<sup>rd</sup> most sensitive snow zone ( $SS_{M3}$ ), the 2<sup>nd</sup> least sensitive snow zone ( $SS_{L2}$ ), and the 1<sup>st</sup> least sensitive snow zone ( $SS_{L1}$ ) (Figure 2.9; Table 2.3). The areas that have a  $LST_{JFM}$  above the mean are considered the most sensitive snow zones, and the areas that have a  $LST_{JFM}$  below the mean are considered the least sensitive snow zones. The area with  $LST_{JFM}$  greater than the 3<sup>rd</sup>  $LST_{JFM}$  standard deviation above the mean is very small, so it is combined with the area within the 3<sup>rd</sup> standard deviation above the mean to define the 1<sup>st</sup> most sensitive snow zone ( $SS_{M1}$ ) ( $LST_{JFM} > 2.4^{\circ}\text{C}$ ). Similarly, the area included within the 3<sup>rd</sup> standard deviation below the mean is very small, and there are no areas that fall outside the 3<sup>rd</sup> standard deviation below the mean (due to the positive skew in the data), so the 1<sup>st</sup> least sensitive snow zone ( $SS_{L1}$ ) includes all areas within the second and third  $LST_{JFM}$  standard deviation below the mean ( $LST_{JFM} < -8.1^{\circ}\text{C}$ ). In addition to the three-month averaged  $LST_{JFM}$ , the temporal evolution of each 8-day 2000-2010 average LST for the period from January to May is considered for qualitative analyses of areas that are potentially sensitive to climate change.

**Table 2.3** Sensitive snow zone definitions, based on the mean and standard deviation of the January-March 11-year average land surface temperature ( $LST_{JFM}$ ) within the seasonal snow zone (SSZ), for the Western United States.

	Cooler than (°C)	Warmer than (°C)
<b>1<sup>st</sup> Most Sensitive Snow Zone (<math>SS_{M1}</math>)</b>	NA	2.4
<b>2<sup>nd</sup> Most Sensitive Snow Zone (<math>SS_{M2}</math>)</b>	-2.4	-1.1
<b>3<sup>rd</sup> Most Sensitive Snow Zone (<math>SS_{M3}</math>)</b>	-1.1	-4.6
<b>2<sup>nd</sup> Least Sensitive Snow Zone (<math>SS_{L2}</math>)</b>	-4.6	-8.1
<b>1<sup>st</sup> Least Sensitive Snow Zone (<math>SS_{L1}</math>)</b>	-8.1	NA



**Figure 2.9** Histogram of the 2000-2010 average LST from January to March ( $LST_{JFM}$ ) for the seasonal snow zone (SSZ) in the Western United States. Blue lines show sensitive snow zone divisions, and the blue dashed line shows the mean of the data.

## CHAPTER 3: FOCUS REGION ANALYSES

### 3.1 INTRODUCTION

The Western United States is a topographically diverse landscape spanning the mid-latitudes south of the 49<sup>th</sup> parallel. Climatic conditions are influenced by a variety of factors that can be described by three major process scales: broad-scale, meso-scale, and local-scale (Mock, 1996; Shinker, 2010). Broad-scale (>100 km) processes include synoptic patterns, which are largely influenced by latitude and global circulation. Meso-scale (10-100 km) climatic patterns in the Western U.S. are the result of proximity to the coast (continentality). Finally, terrain physiography impacts climatic conditions on a local scale (1-10 km) (Mock, 1996; Shinker, 2010). In this study, four in-depth areas, divided into western and eastern slope focus regions, are used to assess the influence of broad-scale, meso-scale, and local-scale processes on precipitation (P), land surface temperature ( $\overline{LST}$ ), and snow cover area ( $\overline{SCA}$ ) in the Western United States. This focus region analysis uses snapshots in time and space to improve the understanding of the spatial variations in P and spatial and temporal variations in  $\overline{SCA}$  and  $\overline{LST}$  in the Western United States. The four in-depth areas are chosen to investigate maritime-continental and latitudinal effects. Since the majority of air masses providing moisture to the Western U.S. are of a westerly origin, dividing each in-depth area into western and eastern slope focus regions provides insight on the characteristics of leeward versus windward slopes on precipitation, land surface temperature, and snow cover. Using these focus regions, the specific

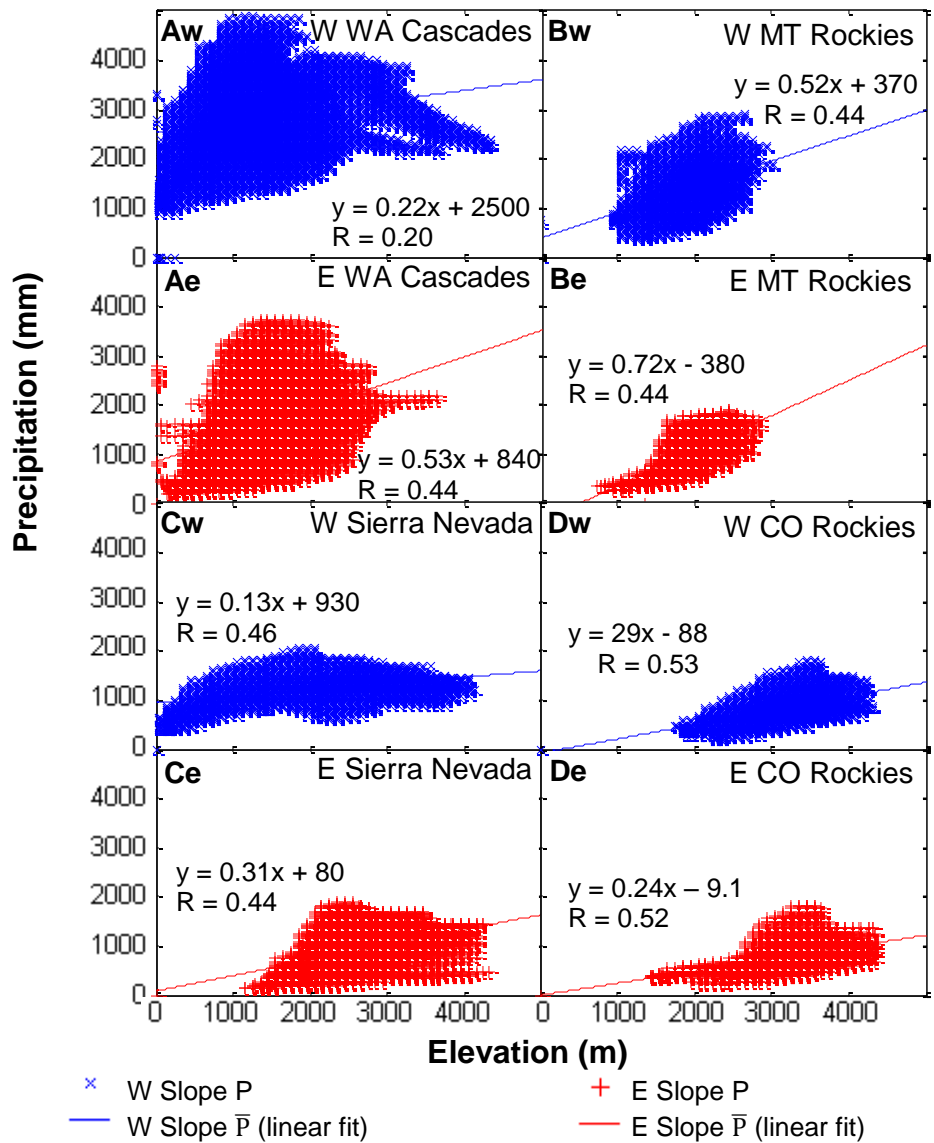


objectives of this chapter are to evaluate the influence of (1) elevation, (2) latitude, (3) continentality, and (4) primary orientation of slope (western versus eastern) on the spatial variability of precipitation and the spatiotemporal variability of snow cover and land surface temperature in the Western United States.

## 3.2 RESULTS

### 3.2.1 *Precipitation*

Annual western and eastern slope precipitation ( $P$ ) and a linear fit of precipitation versus elevation ( $\bar{P}$ ) are shown in Figure 3.1 for the Washington Cascades, the Montana Rockies, Sierra Nevada and the Colorado Rockies focus regions. For all focus regions,  $\bar{P}$  increases with elevation; however, this rate varies between regions. Because  $P$  exhibits large variability across elevation zones, the linear fit  $\bar{P}$  is poor ( $R^2$  between 0.04 and 0.28) and therefore only loosely describes the general trends of precipitation versus elevation in the focus regions. The rate of increase of  $\bar{P}$  with elevation for the focus regions is summarized in Table 3.1. The western Sierra Nevada has the lowest  $\bar{P}$  rate increase with elevation ( $13 \text{ mm } 100\text{m}^{-1}$ ; Figure 3.1Cw) while the highest rate of increase is observed on the eastern slope of the Montana Rockies ( $72 \text{ mm } 100\text{m}^{-1}$ ; Figure 3.1Be). The greatest y-intercept is observed on the western slope of the Washington Cascades ( $2500\text{mm}$ ; Figure 3.1Aw).



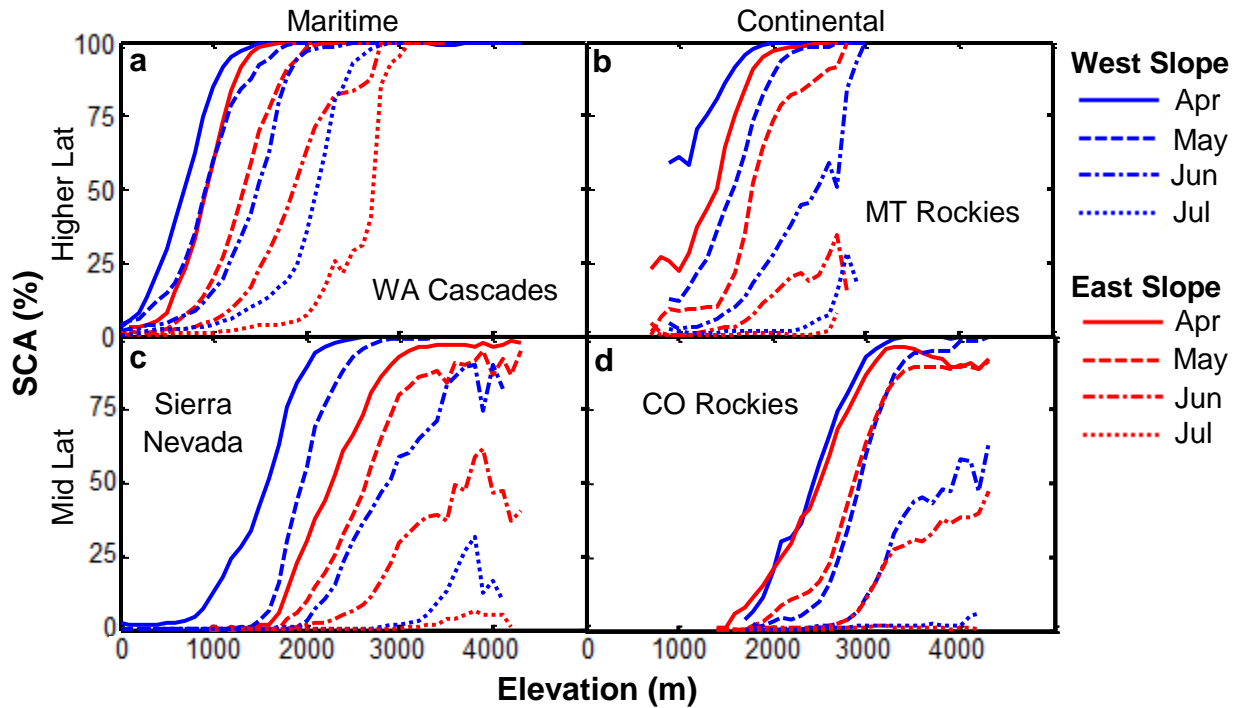
**Figure 3.1** Precipitation versus elevation for western (blue) and eastern (red) slopes in the Washington Cascades, Montana Rockies, Sierra Nevada, and Colorado Rockies. Data are from the PRISM model (<http://www.prism.oregonstate.edu/>).

**Table 3.1** Average precipitation rate of increase per 100 m increase in elevation, y-intercept and  $R^2$  of the linear fit of PRISM (<http://www.prism.oregonstate.edu/>) modeled precipitation versus elevation for the Washington Cascades, Montana Rockies, Sierra Nevada, and Colorado Rockies focus regions.

Precipitation ( $\bar{P}$ - linear fit) Increase with Elevation and Y Intercept						
Focus Study Site	Precipitation Increase with Elevation		Y Intercept		$R^2$	
	W Slope (mm 100m <sup>-1</sup> )	E Slope (mm 100m <sup>-1</sup> )	W Slope (mm)	E Slope (mm)	W Slope	E Slope
Washington Cascades	22	53	2500	840	0.04	0.19
Montana Rockies	52	72	370	-380	0.19	0.19
Sierra Nevada	13	31	930	80	0.21	0.19
Colorado Rockies	29	24	-88	-9.1	0.28	0.27

### 3.2.2 Snow Covered Area

Focus region snow cover area versus elevation for early April, early May, early June, and early July are shown in Figure 3.2 for the focus regions, and Table 3.2 summarizes the elevation of  $\overline{SCA}_{50\%}$ ,  $\overline{SCA}_{90\%}$ , and  $Z_{\overline{SCA}_{max}}$  and the percentage of  $\overline{SCA}_{max}$ . Figure 3.2a illustrates snow cover versus elevation through the ablation period for the Washington Cascades. In early April in the Washington Cascades  $\overline{SCA}_{50\%}$  is observed at 700 m on the western slope and 1000 m on the eastern slope, while the western versus eastern slope  $\overline{SCA}_{90\%}$  is at 1100 m versus 1300 m, respectively.  $\overline{SCA}_{max}$  reaches 100% for both western (1600 m) and eastern (1700 m) slopes. As the ablation season progresses, an increase in elevation of  $\overline{SCA}_{50\%}$ ,  $\overline{SCA}_{90\%}$  and  $Z_{\overline{SCA}_{max}}$  is observed. In early July  $\overline{SCA}_{50\%}$  and  $\overline{SCA}_{90\%}$  are at 2200 m and 2500 m, respectively for western slope and at 2800 m and 2900 m, respectively, for the eastern slope. For both eastern and western slope Washington Cascades a  $\overline{SCA}_{max}$  of 100% is still observed in early July at higher elevation (3000 m, both slopes). At the highest elevations, year round SCA is observed (not shown).



**Figure 3.2** Focus region average snow covered area ( $\overline{SCA}$ ) versus elevation for early April, early May, early June, and early July in the Washington Cascades, Montana Rockies, Sierra Nevada, and Colorado Rockies. Eastern slopes are shown in red, and western slopes are shown in blue.

During early April in the Montana Rockies, western slope  $\overline{SCA}_{50\%}$  and  $\overline{SCA}_{90\%}$  are at 900 m and 1600 m, respectively, while eastern slope  $\overline{SCA}_{50\%}$  and  $\overline{SCA}_{90\%}$  are at 1400 m and 1800 m (Figure 3.2b; Table 3.2). The elevations of  $\overline{SCA}_{50\%}$  and  $\overline{SCA}_{90\%}$  rise during early May and early June on the western slope, and neither  $\overline{SCA}_{50\%}$  and  $\overline{SCA}_{90\%}$  are observed on the western slope in early July. On the eastern slope the elevations of  $\overline{SCA}_{50\%}$  and  $\overline{SCA}_{90\%}$  rise during early May, but  $\overline{SCA}_{50\%}$  and  $\overline{SCA}_{90\%}$  are not observed in early July. Early April  $\overline{SCA}_{\max}$  of 100% is observed on both western (3100 m) and eastern (2600 m) slopes. By early June the west slope still retains a  $\overline{SCA}_{\max}$  of 100% at 3000 m. However, the  $\overline{SCA}_{\max}$  of the eastern slope has depleted to 34% at 2700 m. By early July most snow has depleted on both western and eastern slopes.

**Table 3.2** Elevation (m) of  $\overline{SCA}$  depletion to 50% ( $\overline{SCA}_{50\%}$ ), 90% ( $\overline{SCA}_{90\%}$ ), and maximum  $\overline{SCA}$  ( $Z_{\overline{SCA\_max}}$ ), and the percentage of maximum  $\overline{SCA}$  ( $\overline{SCA}_{max}$ ) for the Washington Cascades, the Montana Rockies, the Sierra Nevada, and the Colorado Rockies in early April, early May, early June, and early July.

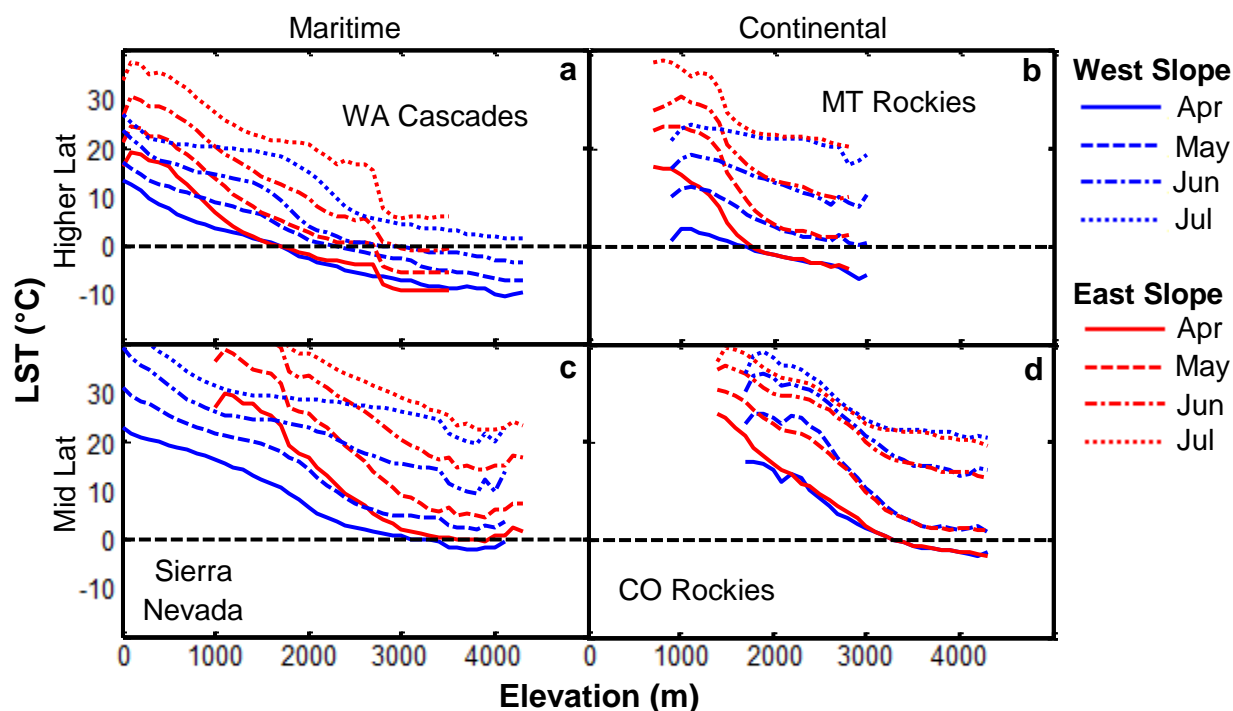
	Early April	Early May	Early June	Early July
<b>Washington Cascades</b>				
W Slope $\overline{SCA}_{50\%}$	700	1000	1500	2200
E Slope $\overline{SCA}_{50\%}$	1000	1400	1900	2800
W Slope $\overline{SCA}_{90\%}$	1100	1500	1900	2500
E Slope $\overline{SCA}_{90\%}$	1300	1800	2800	2900
W Slope $\overline{SCA}_{max}$	100%	100%	100%	100%
W Slope $Z_{\overline{SCA\_max}}$	1600	2000	2600	3000
E Slope $\overline{SCA}_{max}$	100%	100%	100%	100%
E Slope $Z_{\overline{SCA\_max}}$	1700	2000	2800	3000
<b>Montana Rockies</b>				
W Slope $\overline{SCA}_{50\%}$	900	1600	2500	-
E Slope $\overline{SCA}_{50\%}$	1400	1800	-	-
W Slope $\overline{SCA}_{90\%}$	1600	2100	2900	-
E Slope $\overline{SCA}_{90\%}$	1800	2600	-	-
W Slope $\overline{SCA}_{max}$	100%	100%	100%	28%
W Slope $Z_{\overline{SCA\_max}}$	3100	2800	3000	2800
E Slope $\overline{SCA}_{max}$	100%	100%	34%	8%
E Slope $Z_{\overline{SCA\_max}}$	2600	2800	2700	2700
<b>Sierra Nevada</b>				
W Slope $\overline{SCA}_{50\%}$	1600	2000	2900	-
E Slope $\overline{SCA}_{50\%}$	2300	2700	3800	-
W Slope $\overline{SCA}_{90\%}$	2100	2500	3800	-
E Slope $\overline{SCA}_{90\%}$	3000	3600	-	-
W Slope $\overline{SCA}_{max}$	100%	100%	90%	32%
W Slope $Z_{\overline{SCA\_max}}$	2600	3400	3800	3800
E Slope $\overline{SCA}_{max}$	98%	95%	61%	6%
E Slope $Z_{\overline{SCA\_max}}$	4200	3900	3900	3800
<b>Colorado Rockies</b>				
W Slope $\overline{SCA}_{50\%}$	2500	3000	4000	-
E Slope $\overline{SCA}_{50\%}$	2500	2900	-	-
W Slope $\overline{SCA}_{90\%}$	3000	3400	-	-
E Slope $\overline{SCA}_{90\%}$	3100	3700	-	-
W Slope $\overline{SCA}_{max}$	100%	100%	64%	6%
W Slope $Z_{\overline{SCA\_max}}$	3500	4300	4300	4300
E Slope $\overline{SCA}_{max}$	97%	92%	47%	1%
E Slope $Z_{\overline{SCA\_max}}$	3300	4300	4300	3200

Sierra Nevada focus region  $\overline{SCA}$  versus elevation for the spring ablation period is shown in Figure 3.2c. In early April in the Sierra Nevada,  $\overline{SCA}_{50\%}$  is at 1600 m on the western slope and at 2300 m on the eastern slope (Figure 3.2c; Table 3.2). The elevation of  $\overline{SCA}_{50\%}$  progressively increases for both slopes through June. Early April  $\overline{SCA}_{90\%}$  is at 2100 m and 3000 m on the western and eastern slopes, respectively, and increases in elevation through June on the western slope and through early May on the eastern slope. In early April on the western slope  $\overline{SCA}_{\max}$  reaches 100% at 2600 m and on the eastern slope it reaches 98% at 4200 m. By early June western slope  $\overline{SCA}_{\max}$  has depleted to 90% at 3800 m, and eastern slope has depleted to 61% at 3900 m. In early July nearly all snow cover has melted on the eastern slope ( $\overline{SCA}_{\max} = 6\%$  at 3800 m), and most snow cover has melted on the western slope ( $\overline{SCA}_{\max} = 32\%$  at 3800 m).

In early April in the Colorado Rockies, both the western and eastern slopes have  $\overline{SCA}_{50\%}$  at 2500 m (Figure 3.2d; Table 3.2). Western slope  $\overline{SCA}_{90\%}$  is at 3000 m, and eastern slope is at 3100 m. In early May  $\overline{SCA}_{50\%}$  and  $\overline{SCA}_{90\%}$  rise in elevation for both slopes, and by early June  $\overline{SCA}_{50\%}$  is observed only on the western slope (4000 m) and neither slope reaches  $\overline{SCA}_{90\%}$ . Early April  $\overline{SCA}_{\max}$  is 100% for the western slope at 3500 m and 97% at 3300 m on the eastern slope. Both slopes reach  $Z_{\overline{SCA}_{\max}}$  of 4300 m in early June, with western slope  $\overline{SCA}_{\max}$  at 64% and eastern slope at 47%. Nearly all snow melts by early July, with both western and eastern slopes having  $\overline{SCA}$  less than 6% at all elevations.

### 3.2.3 Average Land Surface Temperature

Early April, early May, early June, and early July land surface temperatures versus elevation for all study regions are illustrated in Figure 3.3, and  $\overline{LST}_{0^{\circ}\text{C}}$  is summarized in Table 3.3. For all time periods in all focus regions, land surface temperature tends to decrease with increasing elevation; however this rate of change slows at higher elevations. Early April  $\overline{LST}_{0^{\circ}\text{C}}$  is at 1700m for both eastern and western slopes of the Washington Cascades (Figure 3.3a; Table 3.3). During early May and early June  $\overline{LST}_{0^{\circ}\text{C}}$  rises in elevation for both slopes, but occurs at a higher elevation on the eastern slope than that observed on the western slope. Neither eastern nor western slope Washington Cascades has focus region average land surface temperatures that cross the zero degree threshold in early July. In the Montana Rockies, western slope  $\overline{LST}_{0^{\circ}\text{C}}$  is at 1700 m in early April, and eastern slope  $\overline{LST}_{0^{\circ}\text{C}}$  is at 1800 m (Figure 3.3b; Table 3.3). For both western and eastern slopes  $\overline{LST}_{0^{\circ}\text{C}}$  is not observed on or after early May in the Montana Rockies. In the Sierra Nevada Mountains in early April,  $\overline{LST}_{0^{\circ}\text{C}}$  is at 3100 m on the western slope and 3600 m on the eastern slope (Figure 3.3c; Table 3.3). On both western and eastern slope Sierra Nevada  $\overline{LST}$  is above the  $0^{\circ}\text{C}$  threshold by early May. In early April in the Colorado Rockies, both western and eastern slope  $\overline{LST}$  crosses the  $0^{\circ}\text{C}$  threshold ( $\overline{LST}_{0^{\circ}\text{C}}$ ) at 3300 m (Figure 3.3d; Table 3.3). By early May neither eastern nor western slope focus regions have average  $\overline{LST}$  less than  $0^{\circ}\text{C}$ .



**Figure 3.3** Focus region average land surface temperature ( $\overline{LST}$ ) versus elevation for early April, early May, early June, and early July in the Washington Cascades, Montana Rockies, Sierra Nevada, and Colorado Rockies. Eastern slope is shown in red, and western slope is shown in blue.

**Table 3.3** Elevation at which  $\overline{LST}$  is less than  $0^{\circ}\text{C}$  ( $\overline{LST}_{0^{\circ}\text{C}}$ ) for the Washington Cascades, the Montana Rockies, the Sierra Nevada, and the Colorado Rockies in early April, early May, early June, and early July.

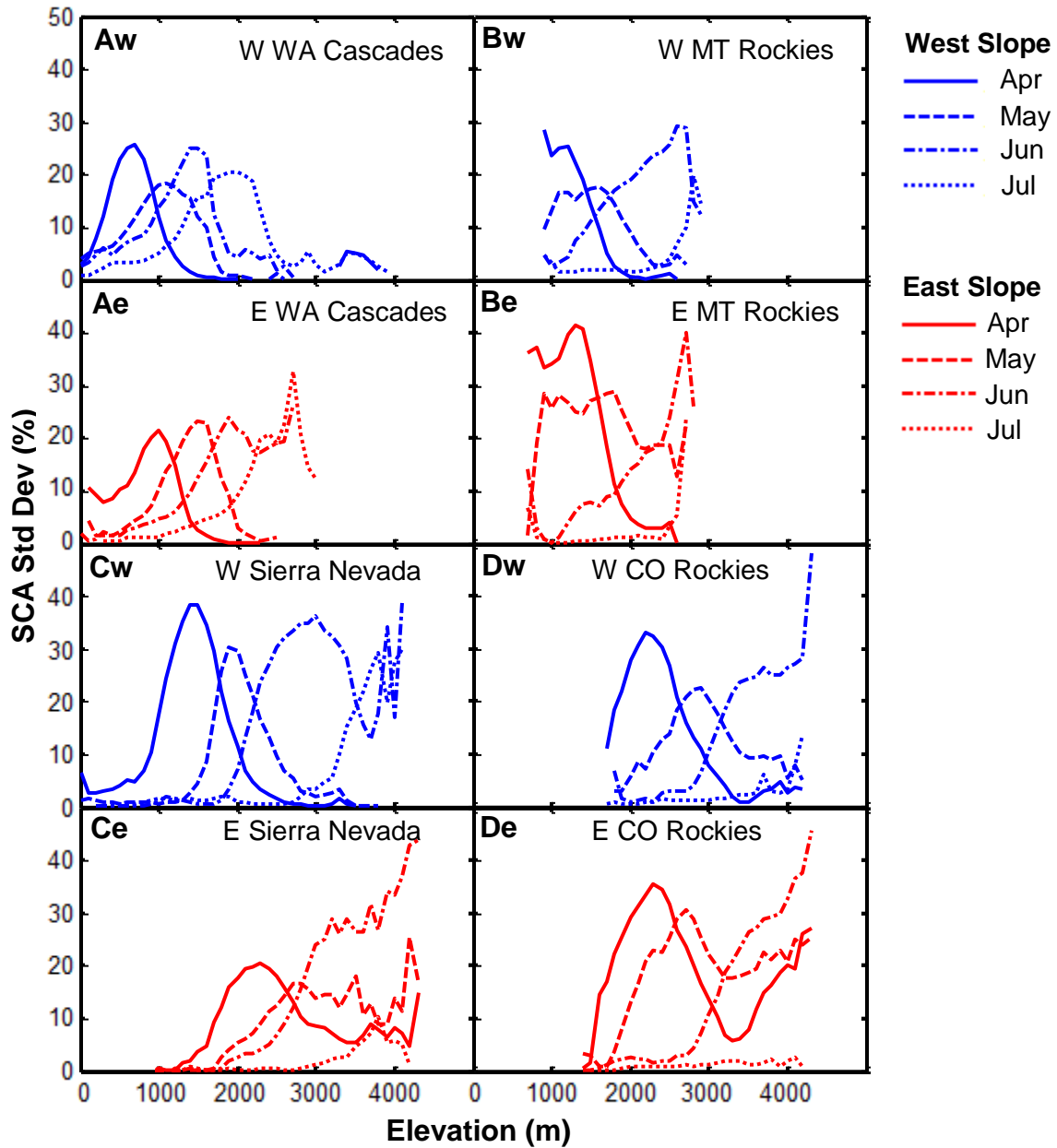
	Early April	Early May	Early June	Early July
<b>Washington Cascades</b>				
W Slope $\overline{LST}_{0^{\circ}\text{C}}$	1700	2300	2800	-
E Slope $\overline{LST}_{0^{\circ}\text{C}}$	1700	2800	3000	-
<b>Montana Rockies</b>				
W Slope $\overline{LST}_{0^{\circ}\text{C}}$	1700	-	-	-
E Slope $\overline{LST}_{0^{\circ}\text{C}}$	1800	-	-	-
<b>Sierra Nevada</b>				
W Slope $\overline{LST}_{0^{\circ}\text{C}}$	3100	-	-	-
E Slope $\overline{LST}_{0^{\circ}\text{C}}$	3600	-	-	-
<b>Colorado Rockies</b>				
W Slope $\overline{LST}_{0^{\circ}\text{C}}$	3300	-	-	-
E Slope $\overline{LST}_{0^{\circ}\text{C}}$	3300	-	-	-



### 3.2.4 Snow Covered Area Standard Deviation

The focus region snow covered area standard deviations are shown in Figure 3.4. The western slope Washington Cascades  $\bar{\sigma}_{SCA}$  peaks at 700 m (26%) in early April (Figure 3.4Aw). The peak in  $\bar{\sigma}_{SCA}$  progressively moves up in elevation through the ablation period, reaching 1100 m (18%) in early May, 1400-1500 m in early June (25%), and 2000 m (21%) in early July. On the eastern slope,  $\bar{\sigma}_{SCA}$  peaks at 1000 m (22%) in early April, and the peak increases in elevation to 1500 m (23%) in early May (Figure 3.4Ae). In early June, eastern slope Washington Cascades  $\bar{\sigma}_{SCA}$  increases linearly from 1200 m (6%) to 1900 m (24%) and ranges between 17% to 26% from 1900 m to 2700 m. By early July  $\bar{\sigma}_{SCA}$  tends to increase linearly between 1800 m (6%) and 2700 m (33%).

In early April on the western slope of the Montana Rockies, the largest standard deviations in focus region snow cover are observed at the lowest elevations (below 1700 m; maximum 29% at 900 m (Figure 3.4Bw). The peak in early May western slope  $\bar{\sigma}_{SCA}$  is lower in value (18%) and higher in elevation (1600 m) than in early April. In early June  $\bar{\sigma}_{SCA}$  increases linearly with elevation from 1000 m (3%) to 2600 m (29%). By early June, lower elevations (less than 2500 m) have low  $\bar{\sigma}_{SCA}$  (less than 5%), and the highest  $\bar{\sigma}_{SCA}$  is observed at 2800 m (20%). The greatest early April  $\bar{\sigma}_{SCA}$  on the eastern slope of the Montana Rockies is at lower elevations (less than 1900 m) with a peak at 1300 m (41%) (Figure 3.4Be). High  $\bar{\sigma}_{SCA}$  (ranging from 25% to 29%) is observed between 900 m and 1800 m in early May. In early June eastern slope  $\bar{\sigma}_{SCA}$  generally increases with elevation from 1100 m (1%) to 2700 m (40%). By early July all elevation zones below 2700 m have less than 7%  $\bar{\sigma}_{SCA}$ , and at 2700 m a  $\bar{\sigma}_{SCA}$  of 24% is observed.



**Figure 3.4** Focus region standard deviation of snow covered area ( $\bar{\sigma}_{SCA}$ ) for early April, early May, early June, and early July in the Washington Cascades, Montana Rockies, Sierra Nevada, and Colorado Rockies. Eastern slope is shown in red, and western slope is shown in blue.

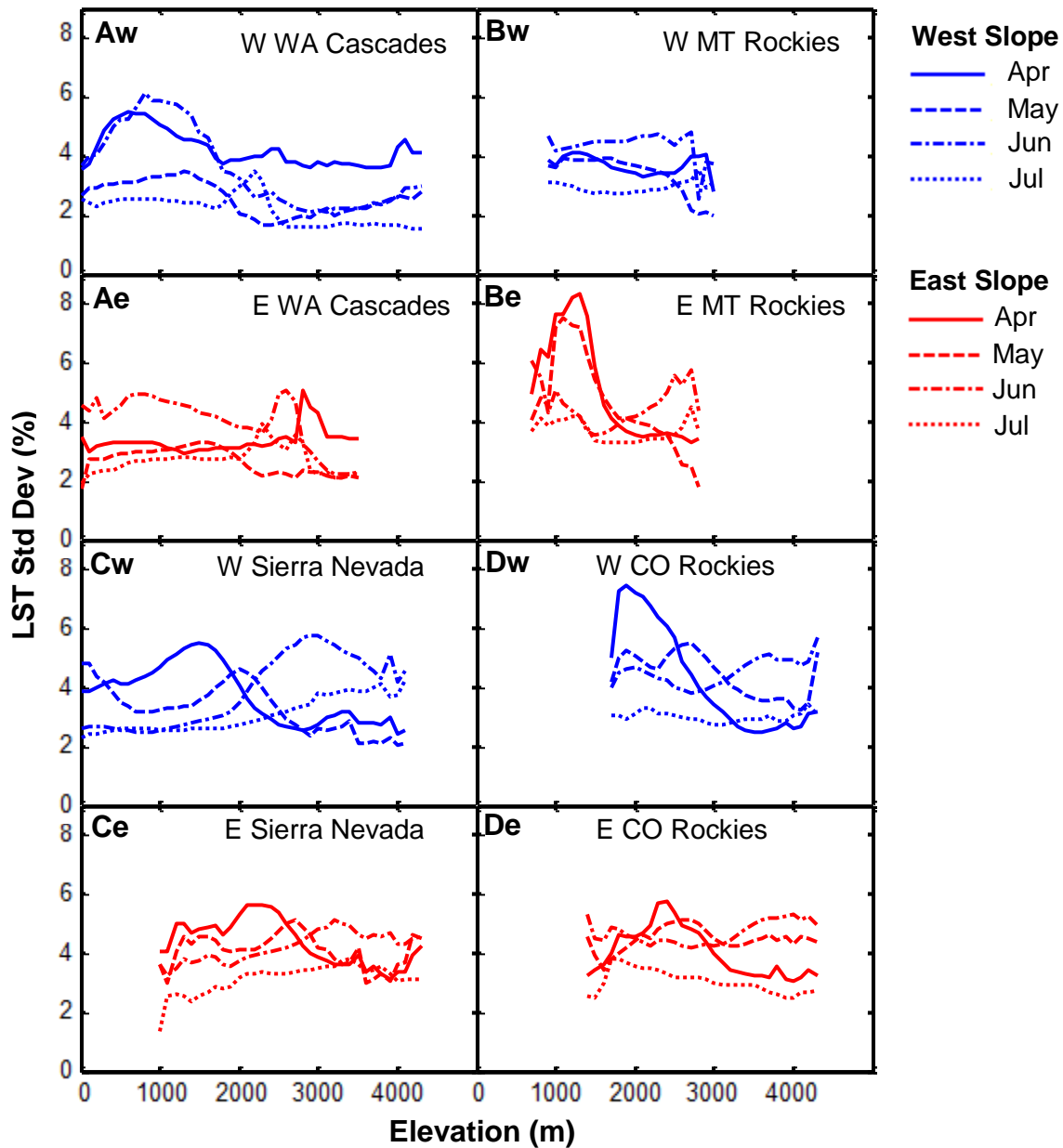
In early April on the western slope in the Sierra Nevada,  $\bar{\sigma}_{SCA}$  peaks at 1400 m at 38% (Figure 3.4Cw). By early May the peak has risen to 1900 m and is lower in value (30%). In early June  $\bar{\sigma}_{SCA}$  peaks at 3000 m (36%), and the peak is wider than observed in preceding months. By early July low snow covered area standard deviations are observed below 3100 m (< 5%), and standard deviation increases linearly above this elevation, with maximum  $\bar{\sigma}_{SCA}$  at 4100 m (30%). Eastern slope Sierra Nevada snow cover area standard deviation peaks at 2300 m (20%) in early April (Figure 3.4Ce). In early May  $\bar{\sigma}_{SCA}$  increases linearly from 1700 m (0%) to 2700 m (17%) and varies randomly from 2700 m and 4300 m between 9 and 25%. Early June eastern slope Sierra Nevada  $\bar{\sigma}_{SCA}$  generally increases linearly above 1800 m (1%) to a peak of 44% at 4300 m. Early July  $\bar{\sigma}_{SCA}$  is minimal on the eastern slope, with  $\bar{\sigma}_{SCA}$  less than 6% at most elevation zones and less than 11% at all elevation zones.

In early April the standard deviation of focus region snow covered area ( $\bar{\sigma}_{SCA}$ ) in western Colorado peaks at 33% at 2200m (Figure 3.4Dw). As the ablation season progresses, this peak rises in elevation to 2900 m but is lower in amount (23%) by early May, and in early June the highest  $\bar{\sigma}_{SCA}$  is observed at the highest elevations (maximum of 48% at 4300 m). By early July  $\bar{\sigma}_{SCA}$  is small, with all but the highest elevation zone exhibiting  $\bar{\sigma}_{SCA}$  of less than 6%. Eastern slope Colorado Rockies  $\bar{\sigma}_{SCA}$  follows a similar trend (Figure 3.4De). Early April  $\bar{\sigma}_{SCA}$  peaks at 2300 m (35%) and increases in elevation (2700 m) but decreases in value (31%) by early May. By early June  $\bar{\sigma}_{SCA}$  increases linearly starting at 2700 m (3%) and peaks at the highest elevation at 45% (4300 m). By early July the variance in snow cover is minimal, and all elevations have less than 3%  $\bar{\sigma}_{SCA}$ .

### 3.2.5 *Land Surface Temperature Standard Deviation*

The focus region standard deviations for land surface temperature are shown in Figure 3.5. In early April on the western slope of the Washington Cascades,  $\bar{\sigma}_{LST}$  shows a slight peak at 600 m (5.5°C) (Figure 3.5Aw). Early May  $\bar{\sigma}_{LST}$  ranges from 1.7°C to a small peak of 3.5°C at 1300 m. The largest peak in  $\bar{\sigma}_{LST}$  on the western slope of the Washington Cascades is in early June at 800 m (6.2°C). Early July  $\bar{\sigma}_{LST}$  ranges between 1.5 and 3.5°C, with a slight peak at 2200 m. On the eastern slope of the Washington Cascades, no strong trend is present in  $\bar{\sigma}_{LST}$  from early April to early July (Figure 3.5Ae). Early April  $\bar{\sigma}_{LST}$  ranges from 3.0 to 5.0°C, with a small peak at 2800 m. In early May  $\bar{\sigma}_{LST}$  ranges between 1.7 and 3.3°C, from 2.2 to 5.1°C in early June, and from 1.9 to 4.0°C in early July.

Western slope Montana Rockies do not show a distinct relationship between  $\bar{\sigma}_{LST}$  and elevation (Figure 3.5Bw). Early April  $\bar{\sigma}_{LST}$  ranges from 2.8 to 4.1°C, early May  $\bar{\sigma}_{LST}$  ranges from 2.0 to 4.0°C, early June  $\bar{\sigma}_{LST}$  ranges from 2.6 to 4.8°C, and early July  $\bar{\sigma}_{LST}$  ranges from 2.8 to 3.5°C. On the eastern slope a strong peak in  $\bar{\sigma}_{LST}$  is observed in early April at 1300 m (8.4°C) and in early May at 1100 m (7.5°C) (Figure 3.5Be). In early June  $\bar{\sigma}_{LST}$  ranges between 3.5 and 6.1°C, and in early July  $\bar{\sigma}_{LST}$  ranges between 3.3 and 4.5°C.



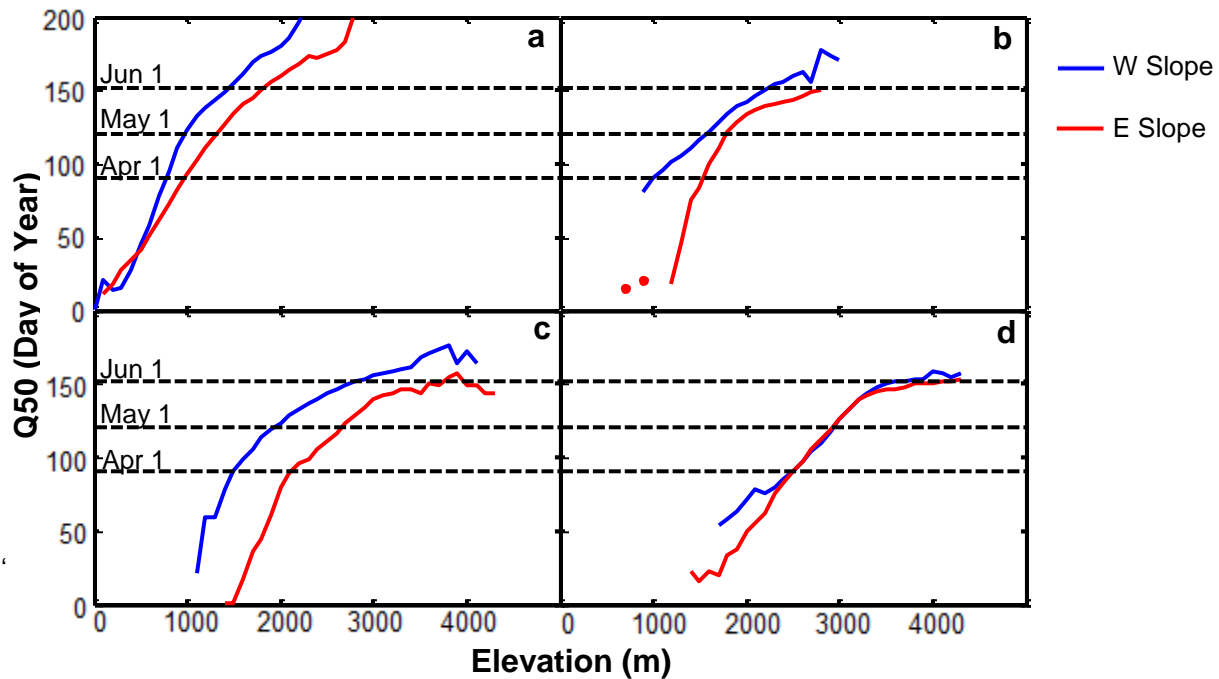
**Figure 3.5** Focus region standard deviation of land surface temperature ( $\bar{\sigma}_{LST}$ ) for early April, early May, early June, and early July in the Washington Cascades, Montana Rockies, Sierra Nevada, and Colorado Rockies. Eastern slope is shown in red, and western slope is shown in blue.

In early April on the western slope of the Sierra Nevada,  $\bar{\sigma}_{LST}$  peaks at 5.5°C at 1500 m (Figure 3.5Cw). Early May shows bimodal peaks at 0 m (4.8°C) and 2000 m (4.6°C). In early June,  $\bar{\sigma}_{LST}$  peaks at 5.8°C at 2900 m. Early July  $\bar{\sigma}_{LST}$  increases slightly with increasing elevation and ranges from 2.3 to 4.2°C. The land surface temperature standard deviation on the eastern slope of the Sierra Nevada peaks at 5.7°C at 2200-2300 m in early April (Figure 3.5Ce). In early May  $\bar{\sigma}_{LST}$  ranges from 3.0 to 5.1°C, from 3.2 to 5.1°C in early June and from 1.3 to 3.8°C in early July.

In early April,  $\bar{\sigma}_{LST}$  on the western slope of the Colorado Rockies ranges from 2.5°C to a peak of 7.4°C at 1900 m (Figure 3.5Dw). For the rest of the ablation period, there is no discernible trend in  $\bar{\sigma}_{LST}$  versus elevation, with  $\bar{\sigma}_{LST}$  ranging between 3.2 to 5.5°C in early May, 3.8 to 5.7°C in early June, and 2.8 to 3.5°C in early July. On the eastern slope of the Colorado Rockies,  $\bar{\sigma}_{LST}$  peaks from 2300-2400 m at 5.7°C in early April (Figure 3.5De). Early May to early July  $\bar{\sigma}_{LST}$  does not reveal strong trends versus elevation and ranges from 3.4 to 5.1°C in early May, 4.2 to 5.3°C in early June, and 2.5 to 3.8°C in early July.

### 3.2.6 Day of Year Depletion to 50% $\overline{SCA}$ ( $\overline{Q50}$ )

The focus region day of year depletion to 50%  $\overline{SCA}$  ( $\overline{Q50}$ ) is shown in Figure 3.6. On the western slope of the Washington Cascades the  $\overline{Q50}$  versus elevation curve is very steep (15.5 days 100 m<sup>-1</sup>) between 300 and 1000 m (Figure 3.6a). This rate of change slows to 6.5 days 100 m<sup>-1</sup> from 1000 to 1700 m, and there is a slight decrease (5.4 days 100 m<sup>-1</sup>) from 1700 to 2200 m. On the eastern slope of the Washington Cascades the rate of change from 100 to 1500 m<sup>-1</sup> is 8.8 days 100 m<sup>-1</sup>, and it slows to 4.9 days 100 m<sup>-1</sup> from 1500 to 2300 m. The lowest elevation  $\overline{Q50}$  on the western slope happens on January 1<sup>st</sup> at 0 m and lasts until after July



**Figure 3.6** Focus region day of depletion to 50%  $\overline{SCA}$  ( $\overline{Q50}$ ) for early April, early May, early June, and early July in the Washington Cascades, Montana Rockies, Sierra Nevada, and Colorado Rockies. Eastern slope is shown in red, and western slope is shown in blue.

19<sup>th</sup> above 2200 m. The lowest elevation  $\overline{Q50}$  on the eastern slope happens on January 11<sup>th</sup> at 100 m and lasts until after July 19<sup>th</sup> above 2700 m.

The western slope of the Montana Rockies maintains a similar  $\overline{Q50}$  rate of change for all elevation zones up to 2600 m (Figure 3.6b). The rate of change is 5.8 days 100 m<sup>-1</sup> from 900 to 1900 m and is slightly less (3.5 days 100 m<sup>-1</sup>) from 1900 to 2600 m. The eastern slope rate of change is very steep (28.2 days 100 m<sup>-1</sup>) from 1200 to 1400 m. From 1400 to 1900 m it decreases to 10.7 days 100 m<sup>-1</sup>, and above 1900 m the rate of change dramatically decreases to 3.7 days 100 m<sup>-1</sup> (1900 – 2200 m) and 1.8 days 100 m<sup>-1</sup> (2200 – 2800 m). Western slope  $\overline{Q50}$  ranges from March 22<sup>nd</sup> at 900 m to June 27<sup>th</sup> at 2800 m while eastern slope  $\overline{Q50}$  ranges from Jan 19<sup>th</sup> at 1200 m to May 30<sup>th</sup> at 2800 m.

On the western slope in the Sierra Nevada,  $\overline{Q50}$  versus elevation increases rapidly from 1300 m to 1800 m at a rate of 10.8 days 100 m<sup>-1</sup> (Figure 3.6c). Above 1800 m the rate of change

decreases to 3.3 days  $100 \text{ m}^{-1}$  between 1800 m and 3200 m and further decreases to 0.6 days  $100 \text{ m}^{-1}$  from 3200 to 4100 m. On the eastern slope there is a steep rate of change (13.4 days  $100 \text{ m}^{-1}$ ) from 1500 m to 2200 m. This rate of change slows to 4.6 days  $100 \text{ m}^{-1}$  from 2200 m to 3300 m and to 0.4 days  $100 \text{ m}^{-1}$  from 3300 m to 4100 m. In higher elevations (2700 to 4100 m) on the western slope  $\overline{Q50}$  ranges between May 29<sup>th</sup> at 2700 m and June 25<sup>th</sup> at 3800 m, and in higher elevations (3000 to 4300 m) on the eastern slope  $\overline{Q50}$  ranges between May 20<sup>th</sup> at 3000 m and June 6<sup>th</sup> at 3900 m.

Areas above 2500 m in the Colorado Rockies ablate to 50% snow cover ( $\overline{Q50}$ ) on or after March 30<sup>th</sup> (Figure 3.6d). In the Colorado Rockies, both western and eastern slope exhibit a similar  $\overline{Q50}$  versus elevation relationship from 2500 m to 3200 m. Between these middle-elevation zones the rate of change of  $\overline{Q50}$  versus elevation is 6.9 days  $100 \text{ m}^{-1}$  on the western slope and 7.0 days  $100 \text{ m}^{-1}$  on the eastern slope. Above 3300 m, a stark break in the rate of change appears on both the western and eastern slopes, such that increasing elevation above 3300 m produces little delay in the date of  $\overline{Q50}$ . From 3500 m to 4300 m the rate of change is 0.9 days  $100 \text{ m}^{-1}$  on the western slope and 1.0 days  $100 \text{ m}^{-1}$  on the eastern slope. In high elevations (3400 – 4300 m) in the Colorado Rockies, the western slope  $\overline{Q50}$  occurs between May 28<sup>th</sup> and June 6<sup>th</sup>, and the eastern slope  $\overline{Q50}$  occurs between May 25<sup>th</sup> and June 3<sup>rd</sup>.

### 3.3 DISCUSSION

Focus region analyses highlight regional spatial differences in precipitation and spatiotemporal differences in snow cover and land surface temperature in the Western United States. By considering these patterns as the result of a combination of processes at multiple scales, insights into the controls on precipitation, snow cover, and land surface temperature are



gained. Differences in land surface temperature and snow cover observed between the focus regions illustrate the impacts of latitude, which operates at a broad-scale ( $> 100$  km) and influences synoptic atmospheric circulation patterns and the availability of solar energy. Maritime/continental divisions demonstrate mesoscale (10-100 km) impacts of continental location, and physiographic elements, such as the elevation and western-eastern slopes, help explain differences observed at the local-scale (1-10 km). In some cases the combined effects of these different scales may highlight certain climatic characteristics, while in others the impacts may be masked or diminished. In considering the influences of all three scales on climate, we arrive at a more complete understanding of the causes for spatiotemporal variations in snow cover and land surface temperature and the spatial variations in precipitation in the Western U.S.

### 3.3.1 *Precipitation*

In mid-latitudes, precipitation tends to increase with increasing elevation (Barry, 2008), as is observed for all focus regions in this study; however, precipitation can be highly variable over space, illustrated by the very low  $R^2$  for the linear fits of  $\bar{P}$  versus elevation in the focus regions (Table 3.1). Increasing precipitation with elevation is a result of orographic influence caused by local-scale physiography (Barry, 2008; Mock, 1996; Shinker, 2010). The higher latitude sites have a greater slope for the linear fit of  $\bar{P}$  versus elevation than the mid latitude sites (Table 3.1) and tend to have greater overall precipitation than the mid-latitude sites. More precipitation at higher latitudes is in part caused by broad-scale synoptic weather patterns that tend to bring more frequent wintertime storms to the areas above the 41<sup>st</sup> parallel in the Western U.S. (Mitchell, 1976). The rate of increase of  $\bar{P}$  versus elevation is greater on the eastern slope than the western slope sites for all focus regions except the Colorado Rockies (Table 3.1). As

will be seen in the following discussion of snow cover and land surface temperature, the Colorado Rockies are different from the other study regions in that western/eastern slope differences are not readily apparent. Higher precipitation amounts on the western slope of the Sierra Nevada, the Washington Cascades (Price, 1986; Shinker, 2010), and the Montana Rockies (Figure 3.1Aw,Ae,Bw,Be,Cw,Ce) are common (Finklin, 1986) because these western slopes are the windward facing and are the first to receive the westerly moist air masses coming off the Pacific ocean (Price, 1986). The Washington Cascades have the highest overall annual precipitation for all the focus regions (Figure 3.1Aw,Ae), with greater amounts of precipitation observed on the western slope. The Pacific Northwest, which includes the Washington Cascades, receives some of the highest amounts of wintertime precipitation in the Western U.S. (Serreze et al., 1999; Shinker, 2010).

### 3.3.2 *Snow Covered Area: Elevation*

Elevation is a local-scale physiographic characteristic that plays a strong role in explaining snow cover within the focus regions. Both western and eastern slope focus regions demonstrate increasing snow cover with increasing elevation for time periods when snow exists (Figure 3.2). Although snow cover can vary greatly dependent on other variables (e.g. land use, aspect, and vegetation), for the spatial unit used in this study (huc100) elevation appears to play the predominant role within each region. Considering that snow depth (Fassnacht et al., 2003) and snow cover duration increase with increasing elevation (Barry, 2008) it is intuitive that snow cover should also increase with increasing elevation. In mid-latitudes at the date of maximum snow accumulation for a region, higher elevations will tend to have higher amounts of snow due to the combined influence of lower temperatures and higher precipitation with increasing

elevation (Barry, 2008). Additionally, it has been found that higher elevations tend to melt later than lower elevations in the Western U.S. (Clow, 2010). In mountain environments higher elevations generally correspond with increased topographic complexity; therefore, although some mountainous areas will receive very high solar radiation inputs with increasing elevation due to decreased atmosphere, the topographic diversity causes shading in many areas and therefore mountains have lower overall net radiation, which allows snowpacks to persist longer (Barry, 2008; Price, 1986). Therefore, higher elevations tend to melt later due to a combination of these lower temperatures and lower overall net radiation (Barry, 2008).

Although the snow cover versus elevation curve for different focus regions tends to be similar in shape in early April, the elevation at which a certain snow cover percentage is observed differs markedly between regions (see Table 3.2; Table 3.4; Figure 3.2). The lowest elevation and longest lasting snow cover is observed in the Washington Cascades (Figure 3.2a), likely due to cold temperatures and high SWE input (Serreze et al., 1999). The highest elevation snow cover is observed in the Colorado Rockies (Figure 3.2d), which receives the lowest wintertime SWE inputs (Serreze et al., 1999) of the study regions. This high-elevation, low SWE observed in the Colorado Rockies is probably due to the combined effects of the mid-latitude location and continental climate.

**Table 3.4** Early April elevation of  $\overline{SCA}_{50\%}$  for the western and eastern slopes of the Colorado Rockies, Sierra Nevada, Washington Cascades, and Montana Rockies.

<b>Elevation Where <math>\overline{SCA} &gt; 50\%</math></b>		
<b>Focus Study Site</b>	<b>W Slope (m)</b>	<b>E Slope (m)</b>
Washington Cascades	700	1000
Montana Rockies	900	1400
Sierra Nevada	1600	2300
Colorado Rockies	2500	2500

### 3.3.3 *Snow Covered Area: Western Versus Eastern Slope*

In the Western U.S., the location of focus regions on western or eastern slopes can result in large differences in the amounts of snow received (Figure 3.2). Western and eastern slope differences in snow cover illustrate the impacts of local-scale physiography on precipitation patterns. The largest divergence in western versus eastern slope snow cover characteristics are observed in the Sierra Nevada (Table 3.2; Figure 3.2c). Conversely, the Colorado Rockies show little to no divergence between western and eastern slope snow cover characteristics during the ablation period. Both the Washington Cascades and the Montana Rockies show an intermediate divergence between western and eastern slope snow cover characteristics, and although not as pronounced as that observed in the Sierra Nevada, a western-eastern slope difference is readily apparent (Table 3.2; Table 3.4; Figure 3.2a-c).

It has been long established that there is a strong rain shadow effect in the Sierra Nevada and the Washington Cascades, with much greater amounts of precipitation observed on the western slope (Price, 1986). The Montana Rockies also act as a barrier that causes the western slope to exhibit characteristics of a maritime environment while the eastern slope is more like a continental environment (Finklin, 1986). The divergence in west and east slope snow cover in Sierra Nevada, the Washington Cascades, and the Montana Rockies focus regions illustrates the impacts of the rain-shadow effect on snow cover. The smaller differences in snow cover for west and east slopes in the Washington Cascades and the Montana Rockies are possibly caused by the generally lower elevations of these mountain ranges compared to the Sierra Nevada (see Figure 2.2a-c), thus creating less of a barrier to air flow. Since snowfall is largely a function of temperature (Auer, 1974; U.S. Army Corps of Engineers, 1956; Fassnacht et al., 2001; Fassnacht and Soulis, 2002), these smaller differences may also be the result of cooler temperatures overall

at the higher latitude focus regions and warmer temperatures overall in the lower latitude focus regions, which would cause more precipitation to fall as rain rather than snow on the eastern slopes in the Sierra Nevada. Although the Colorado Rockies will act as a barrier for westerly air masses, the eastern slope also receives frequent spring upslope events (Barry, 2008; Changnon et al., 1993) as well as southerly storms from the Gulf of Mexico (Barry, 2008), which can contribute large amounts of precipitation. These upslope events and southerly storms may reduce the influence of western and eastern slope on snow cover in the Colorado Rockies relative to the other focus regions.

#### 3.3.4 *Snow Covered Area: Maritime versus Continental*

Continentality, or distance from water sources, influences snow cover characteristics in the Western United States at the mesoscale. As observed in this study, the theoretical elevation of the snowline tends to increase with increasing continentality worldwide (Ives et al., 1974). For each latitudinal and slope group (mid-latitude versus higher-latitude and west slope versus east slope), maritime sites have snow cover at lower elevation than continental sites (Figure 3.2; Table 3.2; Table 3.4). One noteworthy observation is that the Sierra Nevada rain shadow effect causes the eastern slope of the Sierra Nevada to behave more similarly to the continental Colorado sites, over 1,000 km away, than to the adjacent western slope Sierra Nevada. Differences in snowpack properties based on continental location have been noted by other studies (Armstrong and Armstrong, 1987; Serreze et al., 1999). Snowpacks are generally denser in maritime regions of the Western U.S. (Armstrong and Armstrong, 1987). Serreze (1999) found the more maritime Pacific Northwest to have the lowest snow to precipitation ratios in the Western U.S because these areas are close the freezing threshold. Fassnacht et al. (2001)

summarized studies of the probability of snow based on temperature using work from the U.S. Army Corps of Engineers (1956) and Auer (1974), which show that for a given temperature, precipitation is less likely to fall as snow in the Sierra Nevada than compared to data from the entire U.S. Although these studies highlight that maritime snow may be more sensitive to temperature, the lower snowlines observed in maritime regions are probably most strongly influenced by the decreased moisture content of continental air masses (Price, 1986), which causes an increase in the lifting condensation level required for precipitation to occur (Price, 1986).

### 3.3.5 *Snow Covered Area: Latitude*

Latitude plays a broad scale role in influencing snow cover versus elevation relationships in the Western U.S. It is well established that the snow line tends to occur at lower elevation with increasing latitude (Barry, 2008; Ives et al., 1974), a phenomena observed in this study: for each respective western and eastern slope focus region, higher latitude maritime (continental) sites exhibit lower elevation snow cover than mid-latitude maritime (continental) sites (Figure 3.2; Table 3.2; Table 3.4). Latitude, and the corresponding differences in insolation, has a large influence on the elevation of snow cover. The mid-latitude sites receive greater amounts of winter insolation than the higher-latitude sites (Shinker, 2010) thus causing warmer temperatures at mid-latitude sites. Since temperature will largely determine whether precipitation falls as rain or snow (Auer, 1974; U.S. Army Corps of Engineers, 1956; Fassnacht et al., 2001), and higher latitudes tend to have lower winter time temperatures, it follows that the rain/snow threshold will tend to be higher in elevation at the mid-latitude sites.

### 3.3.6 *Snow Covered Area: Standard Deviation*

We postulate that the peaks in snow cover area standard deviation are indicative of the interannual variability of the snow line. All focus regions follow a similar progression through time: in early April,  $\bar{\sigma}_{SCA}$  is greatest at lower to mid elevations, and as the season progresses this greater variation occurs at higher elevations (Figure 3.4). By June, the highest  $\bar{\sigma}_{SCA}$  values are generally observed at the highest elevation zones. By early July for all focus regions except the Washington Cascades  $\bar{\sigma}_{SCA}$  is minimal because  $\overline{SCA}$  has almost completely depleted; higher  $\bar{\sigma}_{SCA}$  is observed in the highest elevations of the Washington Cascades because snow cover often still exists at these locations in July. The largest snow cover standard deviation is observed in early April at the lower elevations on the eastern slope of the Montana Rockies (Figure 3.4Be), probably due to frequent spring storm events on the Great Plains. Overall, for a given time period SCA variability tends to be lowest in areas that have close to 100% snow cover (e.g. high elevation areas), or in areas with no snow cover (e.g. low elevation areas). In the San Juan Mountains of Colorado interannual variability of maximum snow pack has been found to decrease with increasing elevation, while areas closer to the snowline showed the greatest variability (Caine, 1975). Variability is highest in areas that are between higher elevation persistent snow and lower elevation intermittent snow, such as the mid elevation Sierra Nevada, as well as areas which are subject to short lasting spring storm cover events, such as low elevation eastern slope Montana Rockies. The peaks in  $\overline{SCA}$  standard deviation correspond to areas that exhibit the most sensitivity to weather conditions from year to year.

### 3.3.7 $\overline{Q50}$ : Day of Year Depletion to 50% Snow Cover

The day of depletion to 50% snow cover ( $\overline{Q50}$ ) versus elevation offers insights into the progression of snowmelt through time. The rate of change of  $\overline{Q50}$  versus elevation tends to decrease with increasing elevation for all study regions (Figure 3.6). In some regions, the rate of change in ( $\overline{Q50}$ ) versus elevation is relatively consistent, as observed in the Washington Cascades (Figure 3.6a); in others, the rate of change gradually decreases with increasing elevation, as observed in the Sierra Nevada (Figure 3.6c). In the Colorado Rockies (Figure 3.6d) and the eastern slope of the Montana Rockies (Figure 3.6b), there is an abrupt break in the rate of change at high elevation (above 3000 m), indicating that all snow cover at those high elevations is depleting to 50% in a very short time period (10 days for the Colorado Rockies). Our field experiences in the Colorado Rockies lead us to suggest that at the beginning of the ablation season these higher elevation sites have little to no melt because temperatures remain very cold. The lower and middle elevations, on the other hand, are progressively warming and melting as the ablation season marches through time. Towards the end of the ablation period temperatures eventually warm enough to induce rapid melting everywhere. Since the Colorado Rockies and the Montana Rockies receive lower snowfall amounts (Armstrong and Armstrong, 1987; Serreze et al., 1999), less time is required to reduce snow cover. In the Washington Cascades, temperatures at higher elevations remain cold enough to allow snow cover to persist year round, thus this break in rate of snow cover loss with elevation is not observed.

### 3.3.8 Land Surface Temperature: Elevation

Western and eastern slope land surface temperatures of the focus regions demonstrate a strong local-scale dependence on elevation, with increasing elevation corresponding to



decreasing temperatures (Figure 3.3). This relationship is most pronounced at lower and mid-elevation zones; at the highest elevation zones,  $\overline{\text{LST}}$  does not change as rapidly with increasing elevation, creating a backwards “J” shaped response of  $\overline{\text{LST}}$  versus elevation (Figure 3.3). This may be the result of the fact that high elevations in the Western U.S. tend to warm slower than middle and lower elevations (Shinker, 2010). Similarly, lower elevations tend to have wider daily temperature ranges than higher elevation sites due more frequent turbulent mixing of air in higher elevation mountains (Barry, 2008; Price, 1986). Additionally, the high albedo of snow will cause less absorption of solar energy than bare ground and therefore less warming. Both the high heat capacity and frozen state of snow will cause less variable LST than would be observed on bare ground (Barry, 2008). While the land surface temperature versus elevation curve is generally similar in shape for the study regions, land surface temperatures can vary greatly between study regions. Comparing LST in similar elevation zones across focus regions, the coldest land surface temperatures are observed in the Washington Cascades (Figure 3.3a; Table 3.3). Additionally, these lower temperatures last much longer in the Cascades, where  $\overline{\text{LST}}_{0^{\circ}\text{C}}$  lasts until June, whereas the  $\overline{\text{LST}}_{0^{\circ}\text{C}}$  at all other focus sites is not observed after early April (Figure 3.3a-d; Table 3.3).

### 3.3.9 Land Surface Temperature: Western Versus Eastern Slope

Location on the western or eastern slope can cause local-scale differences in land surface temperature (Figure 3.3). The largest divergence between western and eastern slope land surface temperatures is in the Sierra Nevada at lower and middle elevation zones (less than 3000 m) (Figure 3.3c). In contrast, the Colorado Rockies land surface temperature versus elevation relationship is similar for western and eastern slopes (Figure 3.3d). In the Washington Cascades

and the Montana Rockies the widest deviations between western and eastern slope land surface temperature is observed at the lowest elevations (Figure 3.3a,b). Since the eastern slopes of the Sierra Nevada, Washington Cascades, and the Montana Rockies receive less precipitation and therefore retain less moisture than their western slope counterparts, they will tend to heat and cool much more quickly, causing differences in temperatures at similar elevations (Barry, 2008). Thus the impacts of the rain-shadow effect on temperature appear to be stronger at lower elevations.

### *3.3.10 Land Surface Temperature: Latitude and Continentality*

Temperature in the Western U.S. is largely influenced by broad-scale patterns of solar radiation, which are controlled primarily by latitude (Barry, 2008; Mock, 1996; Shinker, 2010). Thus land surface temperatures during early April in higher latitude focus regions cross the zero degree threshold ( $\overline{LST}_{0^{\circ}\text{C}}$ ) at lower elevations than the mid-latitude focus regions (Figure 3.3; Table 3.5). For the focus regions, latitude appears to exhibit a stronger influence on the land surface temperature-elevation relationship than the maritime effect (Table 3.5). This is not to imply that mesoscale continental effects do not impact land surface temperatures, but rather the influence of continentality may not be readily apparent in the spatial transects employed in this study. The influences may also be masked by the predominant influences of latitude, elevation, and potentially other factors not considered in this study. Serreze et al. (1999) found that as a whole the Pacific Northwest (which includes Oregon and Washington) maintained, on average, much warmer winter/spring temperatures than interior locations at similar latitudes; however the maps included in Serreze et al.'s (1999) study showed that the Washington Cascades are clearly much colder than the Oregon Cascades and the more coastal Olympic mountains. The winter/spring temperatures in the Washington Cascades are more similar to the interior

continental climates than the coastal maritime climates. During winter months, the Washington Cascades are subject to easterly atmospheric circulation, which brings cold continental air through mountain passes (Steenburgh et al., 1997). The Washington Cascades may be far enough away from the coast that latitude and/or easterly atmospheric circulation, rather than continentality, dominates in determining land surface temperatures. Furthermore, due to the snow-albedo feedback effect and the insulating nature of snow, snow cover on the ground may promote cooler springtime temperatures than would be observed in patchy snow covered areas that warm more quickly. Thus it is observed in the Washington Cascades that snow cover remains higher and land surface temperatures remain colder for longer duration than the other three study sites.

**Table 3.5** Early April elevation of  $\overline{LST}_{0^{\circ}\text{C}}$  for the western and eastern slopes of the Washington Cascades, Montana Rockies, Sierra Nevada, and Colorado Rockies.

<b>Elevation Where <math>\overline{LST} &lt; 0^{\circ}\text{C}</math></b>		
<b>Focus Study Site</b>	<b>W Slope (m)</b>	<b>E Slope (m)</b>
Washington Cascades	1700	1700
Montana Rockies	1700	1800
Sierra Nevada	3100	3600
Colorado Rockies	3300	3300

### 3.3.11 Land Surface Temperature: Standard Deviation

There are several cases in which peaks in land surface temperature standard deviation correlate well to peaks in the snow covered area standard deviation. This is observed for early April through early July on the western slope Sierra Nevada and western slope of the Washington Cascades, early April and early May on the eastern slope Sierra Nevada and the western and eastern slope of the Colorado Rockies, and early April on the eastern slope of the Montana Rockies (Figure 3.5). The largest  $\overline{\sigma}_{LST}$  is observed at the low elevations in early April

on the eastern slope of the Montana Rockies, possibly due to frequent spring storm events. We postulate that the observed coincidence between the peaks in snow covered area and land surface temperature standard deviation are a result of the influence of the presence/absence of snow and its influence on energy fluxes.

### 3.4 SUMMARY

Snow cover, land surface temperature and precipitation differences throughout the Western U.S. are not controlled by one single variable but rather are the result of multiple factors operating at several spatial scales. Mock (1996) and Shinker (2010) describe three spatial scales on which different phenomena operate that impact temperature and precipitation variations in the Western U.S.: broad-scale synoptic patterns, meso-scale impacts of continental location, and local-scale influences of physiography. These three operating scales of control on precipitation and temperature in the Western U.S. help explain the spatial differences in precipitation and the spatial and temporal differences in land surface temperature and snow cover observed in this study. Focus region elevation and eastern/western slope variations in snow cover highlight differences caused by local-scale phenomena, maritime versus continental divisions illustrate mesoscale differences, and the latitudinal divisions show the impacts of broad-scale synoptic patterns.

Considerable spatial variability in precipitation and spatiotemporal variability in land surface temperature and snow covered area is observed between the focus regions. Precipitation tends to increase in maritime environments, higher latitudes, and on western slopes. It is not surprising, therefore, that the greatest precipitation amounts are observed on the western slope of the higher-latitude maritime Washington Cascades. Snow cover tends to increase with increasing elevation and is lower in elevation in maritime environments, at higher latitudes, and

on (most) western slopes. Land surface temperatures tend to be lower at higher elevations, higher latitudes, and on (most) western slopes. The lowest elevation, longest lasting snow cover is observed on the western slope of the Washington Cascades, and the highest elevation snow cover is observed in the Colorado Rockies. Likewise, the coldest temperatures are observed in the Washington Cascades. For all focus regions, land surface temperature decreases with increasing elevation; however at the highest elevation this rate of decrease slows. The largest divergence in western and eastern slope snow cover is observed in the Sierra Nevada, likely due to the extreme rain shadow effect caused by the very high elevation mountain range. Little to no divergence is observed in the snow cover of the Colorado Rockies, likely due to springtime upslope storm events and storms from the Gulf of Mexico. Interestingly, snow cover on the eastern slope of the Sierra Nevada is more similar to the continental Colorado Rockies than the adjacent western slope Sierra Nevada. The greatest divergences in western and eastern slope land surface temperature are observed in the Sierra Nevada, while very little divergence between western and eastern slope land surface temperature is observed in the Colorado Rockies. The eastern/western slope differences in land surface temperature, or lack thereof, are probably due to the influences of snow and moisture content on land surface temperature. Latitude appears to play a stronger role than continentality in explaining regional differences in land surface temperature, although this may be the result of atmospheric circulation or other factors not considered in this study.

All focus regions demonstrate a decreasing rate of change of  $\overline{Q50}$  versus elevation, likely due to the fact that higher elevations tend to warm more slowly than lower elevations. The Colorado Rockies, and to a certain extent, the eastern slope Montana Rockies and the Sierra Nevada, exhibit an interesting change in behavior of snow cover loss timing at high elevation, in

which snow cover depletes to 50% within a very short time period (e.g. 10 days in the Colorado Rockies). The progressive increase in elevation of the peaks in snow cover standard deviation through the ablation period is indicative of the variability in the snow line from year to year. Land surface temperature standard deviation tends to peak at similar elevations as the snow cover standard deviations for several focus regions, likely because of snow's impact on the land surface energy budget. Transitional areas between lower elevations with no snow and higher elevations with persistent snow can exhibit high variability each year, suggesting they are most sensitive to weather patterns. This high sensitivity to temperature at transitional elevations means the snow line progresses rapidly upward in elevation as the ablation season progresses. In contrast, higher elevations with persistent snow cover in the Colorado Rockies, Sierra Nevada, and on the eastern slope of the Montana Rockies have nearly synchronous snow loss late in the season when temperatures are warm everywhere.

Besides the geographic characteristics included in this study, many other factors (i.e. vegetation, slope) may be of importance in explaining climatic spatiotemporal variability in the Western U.S. Further studies considering other variables not considered in this study could help improve understanding of Western U.S. climate and may be of use to identify regions with particular snow cover and land surface temperature characteristics.

## CHAPTER 4: WESTERN U.S. ANALYSES

### 4.1 INTRODUCTION

The focus region analyses helped identify some of the factors that control spatiotemporal variability in SCA and LST. This chapter extends the focus region findings to an analysis of the spatiotemporal patterns in SCA and LST for the Western U.S. as a whole. The objectives of this portion of the study are to (1) determine the spatiotemporal variations in average snow cover in the Western U.S., (2) define zones of persistent, transitional, intermittent and seasonal snow, (3) determine the elevation of these zones, (4) assess the spatiotemporal variations of temperature within the seasonal snow zone, and finally, by using average wintertime LST (5) determine which areas in the seasonal snow zone may be more sensitive to changes in climate.

### 4.2 RESULTS

#### 4.2.1 *January to June Snow Cover*

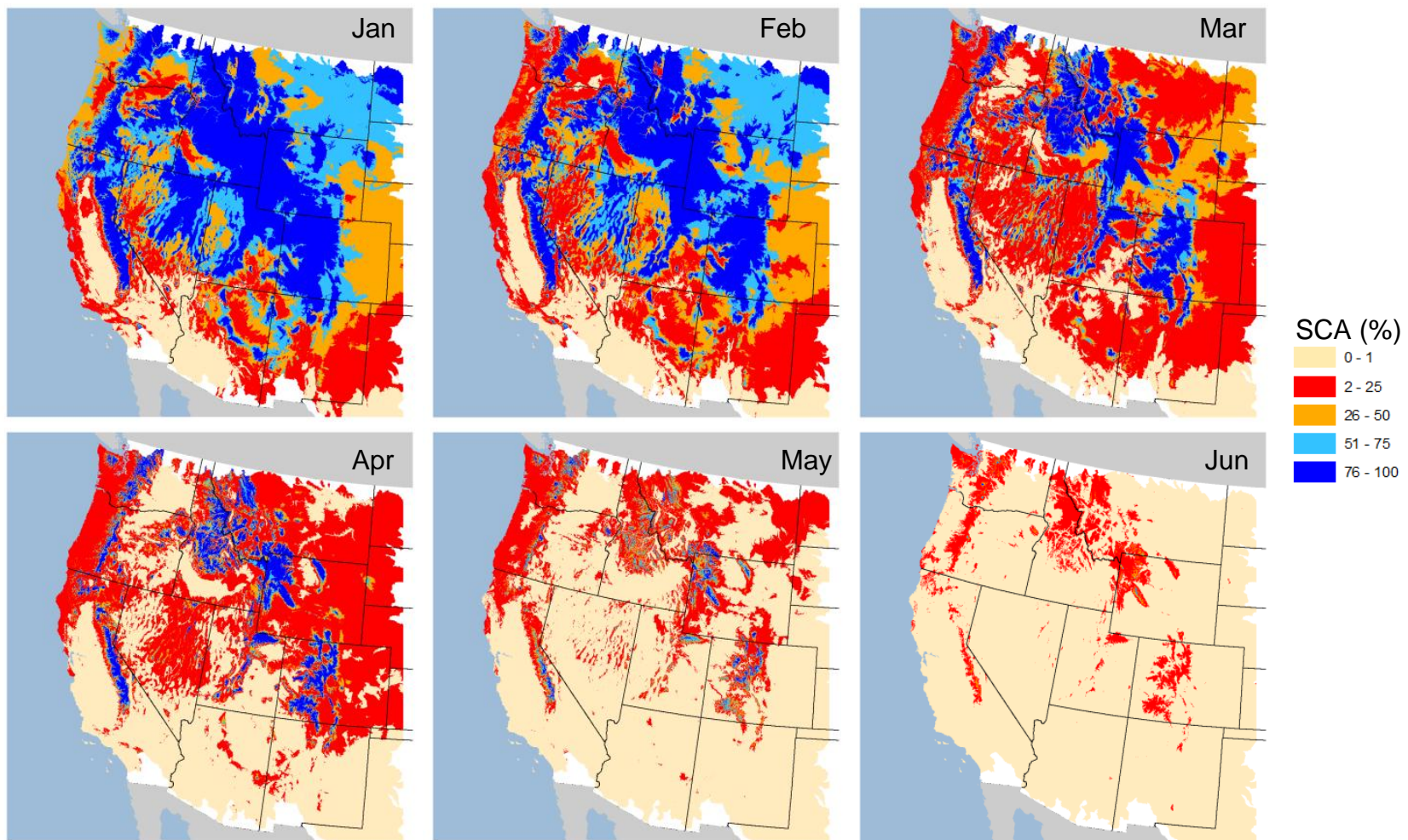
Average monthly January to June snow cover derived from MODIS for 2000 to 2010 is shown in Figure 4.1. In January snow covers much of the Western U.S., with greater than 50% SCA observed in low to high elevation Sierra Nevada and Cascades, the NW-SE axis of the Rocky Mountain cordillera, the Blue Mountains in Oregon, mid to high elevation northeastern and central eastern Nevada, the Upper Gila Mountains of Arizona and New Mexico, and most of the northern Great Plains. February snow covers a similar extent as observed in January; most snow cover loss in February is at lower elevation areas. By March more snow cover loss is apparent. Snow cover has decreased to less than 25% in lower elevations of coastal Oregon and

Washington and has retreated to middle elevations in the Cascades and Sierras and to higher elevations in Arizona/New Mexico. Mid-elevation Utah and Northwestern Nevada show substantially less snow cover, and the northern Great Plains shows loss of snow cover to less than 50%. By April the Western U.S. is divided strongly into areas with greater than 50% snow cover and areas with less than 50% snow cover. Areas retaining greater than 50% snow cover include the middle and higher elevation Cascades and Sierra Nevada, higher elevation ranges along the axis of the Rocky Mountain Cordillera, the Washington Olympics, the Utah Wasatch and Uinta, high elevation Nevada ranges, and sparse areas in the higher elevation Upper Gila Mountains in Arizona/New Mexico. By May snow cover has retreated to even higher elevation zones, but retains a strong presence in the higher elevation Sierra Nevada, Cascades, Washington Olympic Mountains, Utah Wasatch and Uinta, and the middle and higher elevation axis of the Rocky Mountain cordillera. By June most snow cover has depleted to less than 50%; however small areas of greater than 50% snow cover still remain in the highest elevations of the Cascades and the high elevation interior Rockies in northwest Wyoming.

#### 4.2.2 *Snow Zones*

Western United States snow persistence (SP) is shown in Figure 4.2, and the elevation of the seasonal snow zone (SSZ), persistent snow zone (PSZ), transitional snow zone (TSZ) and intermittent snow zone (ISZ) are shown in Figure 4.3. Detailed maps of the snow zones are included in Appendix B for California/Nevada (CA/NV), Oregon/Washington (OR/WA), Idaho/Montana/Wyoming (ID/MT/WY), and Utah/Colorado/Arizona/New Mexico (UT/CO/AZ/NM).



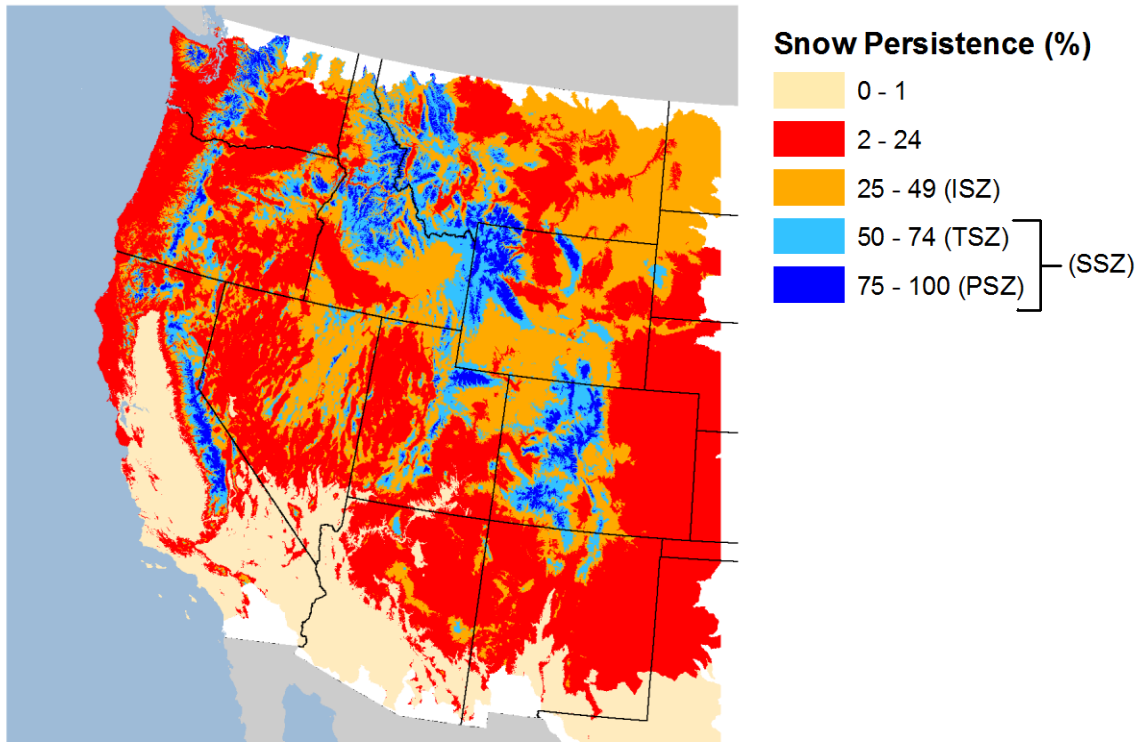


**Figure 4.1** MODIS 2000 – 2010 average monthly snow covered area from January to June in the Western United States.

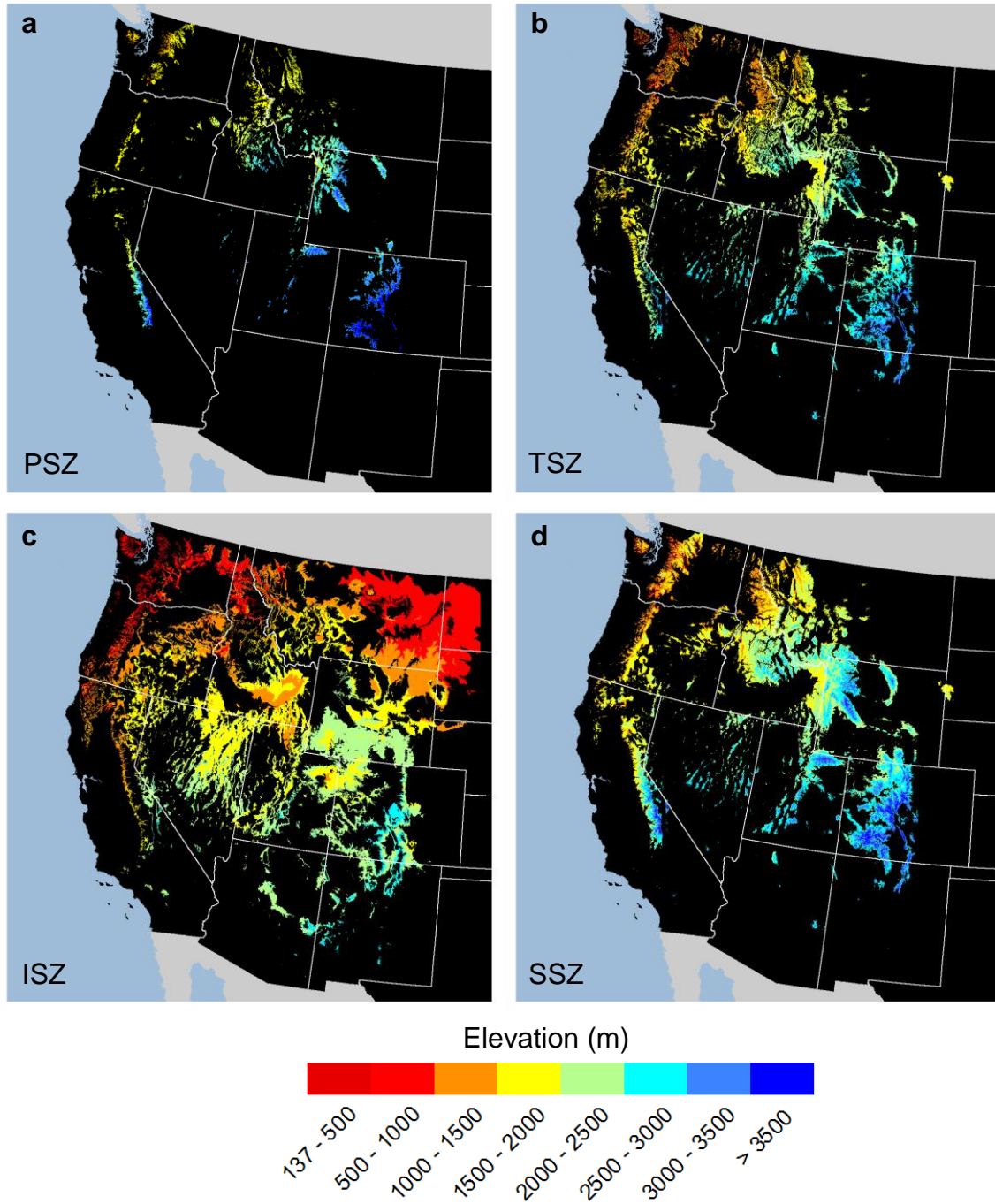
The seasonal snow zone (SSZ; Figure 4.3d) covers an area of 448,135 km<sup>2</sup> and makes up 13.2% of the total area (Table 4.1). The persistent snow zone (PSZ; Figure 4.3a) is 114,410 km<sup>2</sup> and constitutes 3.4% of the total area and 25.5% of the SSZ (Table 4.1). The lowest elevation PSZ (1000 to 1500 m) is observed in the Washington Cascades/Olympic Mountains, whereas the highest elevation PSZ (>3000 m) is observed in the Colorado Rockies, the southern Sierra Nevada, the Utah Uinta and Wasatch and the northwestern Wyoming/central Idaho/southwestern Montana Rockies. Between these two extremes there is a northwest/southeast elevation gradient. Lower elevation PSZ (1500 to 2000 m) is observed in the Washington Olympic, Washington/Oregon Cascades, northern California Mountains and the northern Idaho/northwestern Montana Rockies. Lower to mid elevation PSZ (2000 to 2500 m) is observed in northern Sierra Nevada and the central Idaho/western Montana Rockies. The middle to higher elevation PSZ (2500 to 3000 m) is observed in the central Sierra Nevada, the western slope southern Sierra Nevada, the Rockies of central Idaho, southwestern Montana, northwestern Wyoming, and northern Colorado, and small areas within the ranges of northeastern Nevada and northern Utah. Western and eastern slope differences in the snowline are observed in the southern Sierra Nevada (western slope at 2500 m versus eastern slope at 3000 m) and the Washington Cascades (western slope at 1000 m versus eastern slope at 1500 m). There is no PSZ observed in Arizona.

**Table 4.1** Area, percent of total area, and percent of seasonal snow zone (SSZ) for the persistent snow zone (PSZ), the transitional snow zone (TSZ), the intermittent snow zone (ISZ), the seasonal snow zone (SSZ), and the remaining area in the Western United States.

	<b>Area (km<sup>2</sup>)</b>	<b>Percent of Total Area (%)</b>	<b>Percent of SSZ (%)</b>
<b>PSZ – Persistent Snow Zone</b>	114,410	3.4	25.5
<b>TSZ- Transitional Snow Zone</b>	333,725	9.8	74.5
<b>ISZ – Intermittent Snow Zone</b>	845,246	24.9	NA
<b>SSZ – Seasonal Snow Zone</b>	448,135	13.2	NA
<b>Remaining Area</b>	2,106,899	62	NA



**Figure 4.2** Snow persistence (SP) and snow zones in the Western U.S. determined from 2000-2010 MODIS average snow cover averaged from January 1<sup>st</sup> to July 3<sup>rd</sup>. Four snow zones are defined: the intermittent snow zone (ISZ; orange; 25-49% average snow covered area), the transitional snow zone (TSZ; light blue; 50-74%), and the persistent snow zone (PSZ; dark blue; 75-100%). The seasonal snow zone (SSZ) is the combination of light and dark blue zones, with snow persistence from 50-100%.



**Figure 4.3** Elevation of the (a) persistent snow zone (PSZ), (b) transitional snow zone (TSZ), (c) intermittent snow zone (ISZ), and (d) seasonal snow zone (SSZ) for the Western United States.

The transitional snow zone (TSZ) makes up 9.8% of the total area of the western U.S. and 74.5% of the SSZ. It covers 333,725 km<sup>2</sup> (Table 4.1; Figure 4.3b). The elevation where the TSZ begins is the same as the elevation where the SSZ begins. The lowest elevation TSZ (750 to 1000 m) is observed in the Washington Olympic/Cascades ranges. The highest elevation TSZ (3000 to >3500 m) is observed predominantly in the southern Rockies of Colorado and New Mexico and is also observed in the southern Sierra Nevada, the southern Wasatch, the Upper Gila Mountains of Arizona and New Mexico, and the central Nevada Ranges. Between these two zones exists a northwest/southeast elevation gradient. TSZ between 1000 and 1500 m is observed predominantly in the Washington and Oregon Cascades, northern Idaho and northwestern Montana. The central to northern portions of the western slope Sierra Nevada, northern California, southern and eastern Oregon, central Idaho, and central-western Montana have TSZ between 1500 and 2000 m. TSZ from 2000 to 2500 m is observed predominantly in the Rocky Mountains of Idaho, Montana, and Wyoming, and in smaller areas of the southwestern slope Sierra Nevada and the ranges of northern Nevada, northern Utah, and northern Colorado. TSZ between 2500 and 3000 m is observed in the Colorado Rockies, the Utah Uinta and Wasatch, western Wyoming, and the central to southern portions of the eastern slope Sierra Nevada. Several areas show western and eastern slope differences in the lower elevation limit of the TSZ snowline, such as the southern Sierra Nevada (1500 m on the western slope versus 2500 m on the eastern slope) and the southern Oregon Cascades (1000 m on the western slope versus 1500 m on the eastern slope).

The intermittent snow zone (ISZ) is by far the largest snow zone, covering an area of 845,246 km<sup>2</sup> and making up 24.9% of the total area (Table 4.1; Figure 4.3c). The lowest elevation ISZ (<500 to 1000 m) is in the western slope Oregon Cascades, the Washington

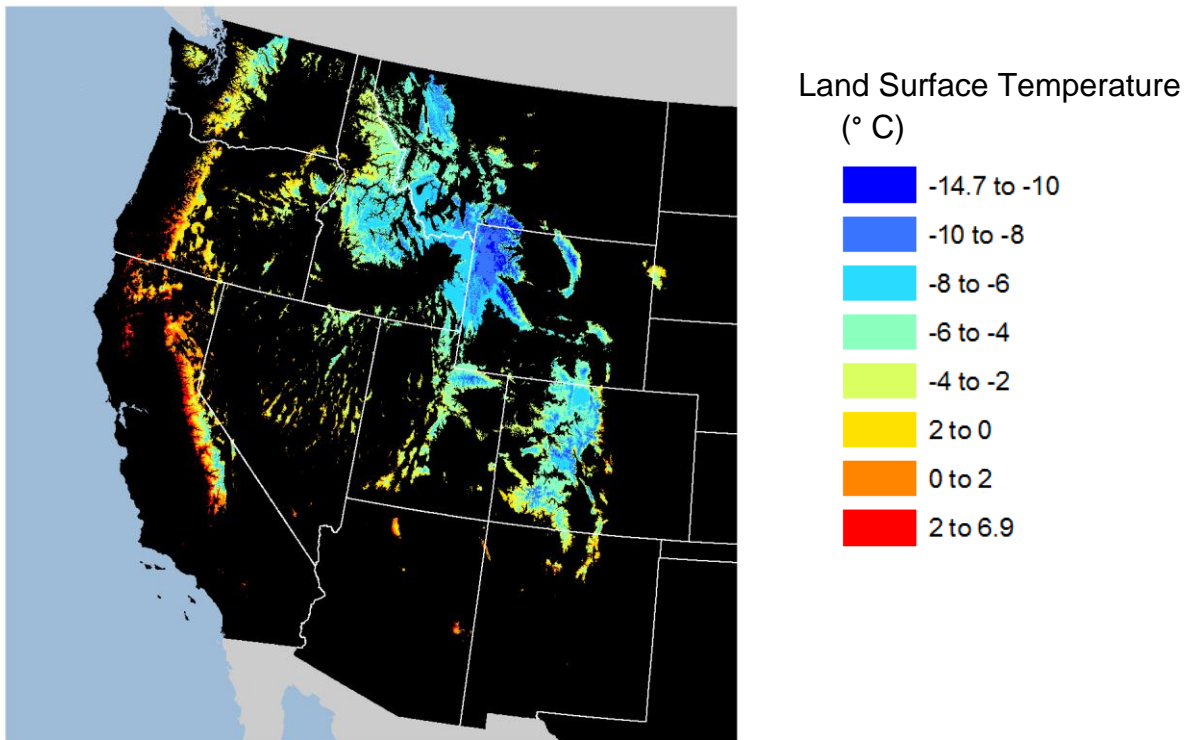
Cascades/Olympic/Rocky Mountains, the northern Idaho/ northwestern Montana Rockies, and the northern Great Plains. The highest elevation ISZ is observed between 2500 and 3500 m in the south-central Colorado/northern New Mexico Rockies, the central Utah Wasatch Mountains, the eastern slope of the Sierra Nevada, and the Upper Gila Mountains in New Mexico and Arizona. An elevational gradient is observed between these two zones. Lower-mid elevations ISZ (2000 to 2500 m) are observed on the eastern slope of the Sierra Nevada, the ranges within central Nevada, the southern Wyoming Great Plains, the Upper Gila Mountains in Arizona and New Mexico, the Utah Wasatch and Uinta, and the Colorado Rockies. Lower elevation ISZ (1000 to 2000 m) is observed on the western slope of the Sierra Nevada, northern California, eastern Oregon, the flatter basins within north-central Nevada, southwestern Montana, the northern Great Plains along southeastern Montana/northeastern Wyoming, and areas of western/northeastern Utah and northwestern Colorado. Western slope and eastern slope differences in the snowline of the ISZ are observed in the southern Sierra Nevada (west slope: 1500 m; east slope: 2000 m), the northern Sierra Nevada (west slope: 1000 m; east slope: 1500 m) and the Oregon Cascades (west slope: 500 m; east Slope: 1000 m).

#### 4.2.3 *Land Surface Temperature in the Seasonal Snow Zone*

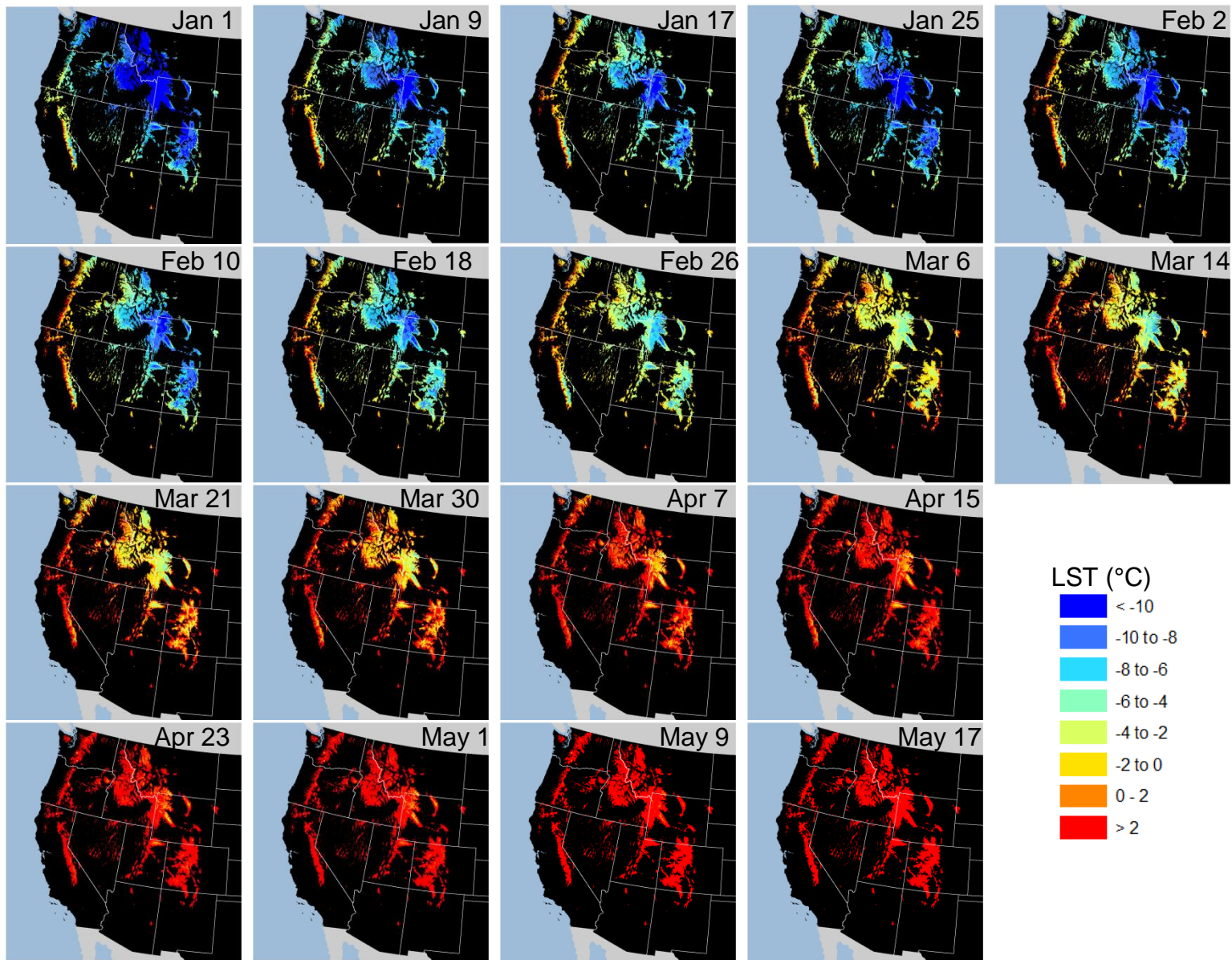
Average land surface temperature for January to March ( $LST_{JFM}$ ) within the seasonal snow zone (SSZ) of the Western United States is shown in Figure 4.4. The warmest areas ( $> 2^{\circ}\text{C}$ ) are at lower elevations of the Sierra Nevada in CA, CA/OR Cascades, CA/OR Klamath, and the higher elevations of the southern CA Mountains and Mt. Baldy/the Kaibab Plateau in AZ. The coldest areas ( $< -8^{\circ}\text{C}$ ) are observed predominantly in the interior northern Rocky Mountains clustered around the intersection of ID, MT, and WY; however, high elevation areas in the CO



Rockies, UT Wasatch and Uinta, CA Sierra Nevada, and a few areas in the OR Blue Mountains and northeastern NV have less than  $-8^{\circ}\text{C}$   $\text{LST}_{\text{JFM}}$ . There is a NE to SW temperature gradient that is observed between the colder interior regions and the warmer more coastal OR Cascades, CA Sierra Nevada, and the southwestern CA/AZ region. A similar gradient is observed in the temporal evolution of 8-day 2000 to 2010 average MODIS land surface temperature from January to May within the SSZ of the Western United States (Figure 4.5). In January the warmest areas are in the lower and mid elevation Sierra Nevada, followed by the lower elevations of the CA/OR/WA Cascades, southwest AZ/NM Upper Gila Mountains, the NV mountain ranges, the Blue Mountains of OR, and northern ID/western MT. The areas that remain the coolest longest are the highest elevations of the northern Rockies around ID/MT/WY, the CO Rockies, the Sierra Nevada, and a few peaks in the OR/WA Cascades.



**Figure 4.4** 2000 – 2010 average MODIS land surface temperature for January, February March ( $\text{LST}_{\text{JFM}}$ ) for the seasonal snow zone (SSZ) in the Western United States



**Figure 4.5** MODIS derived 8 day average LST for 2000-2010 in the seasonal snow zone (SSZ) of the Western U.S.



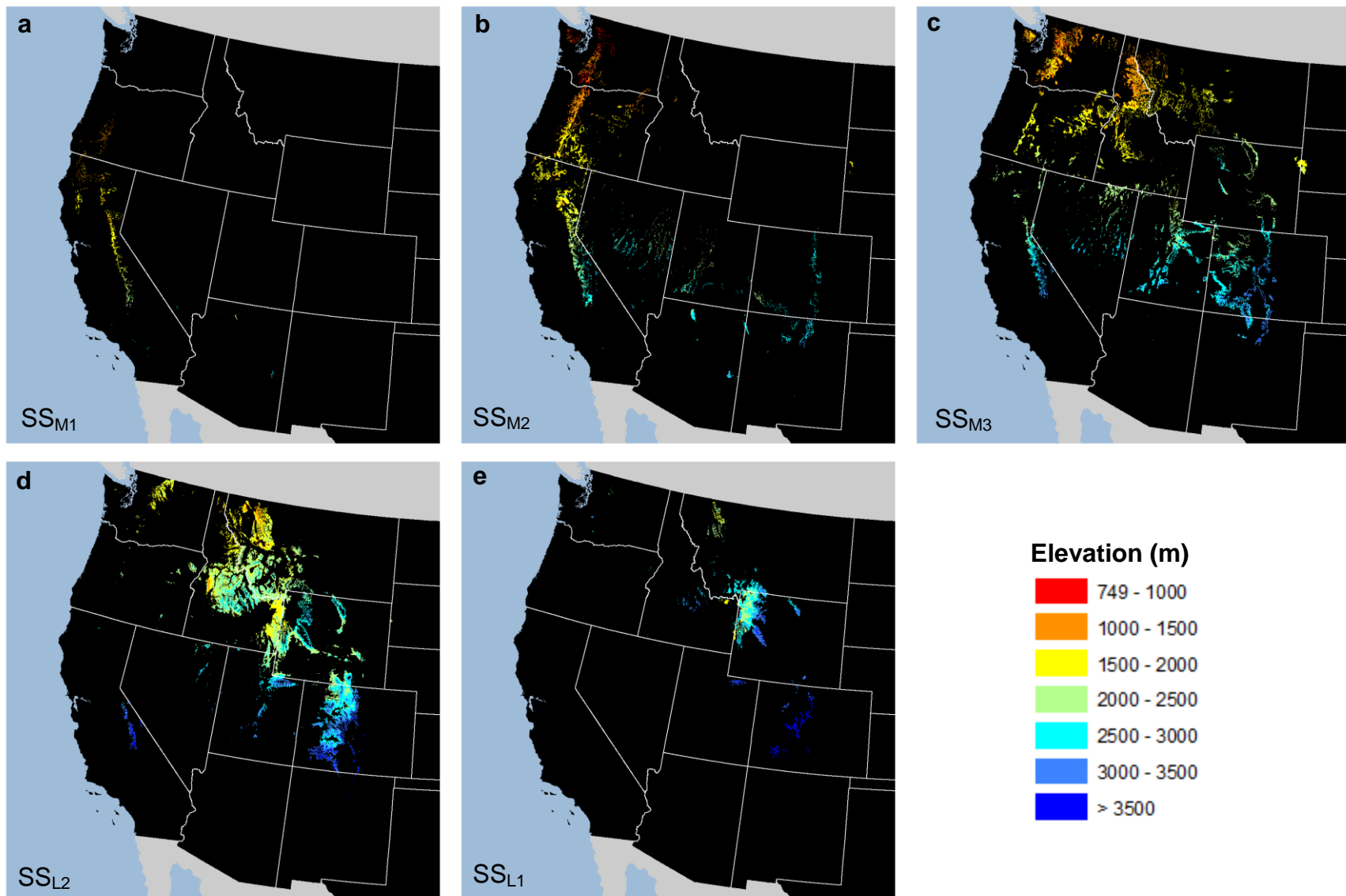
#### 4.2.4 Sensitive Snow Zones

The elevation of the sensitive snow zones in the Western United States, which are grouped by standard deviation (see Chapter 2), are depicted in Figure 4.6. Focus maps of the sensitive snow zones are included for California/Nevada, Oregon/Washington, Idaho/Montana/Wyoming and Colorado/Utah/Arizona/New Mexico in Appendix B.

The first most sensitive snow zone ( $SS_{M1}$ ) exists primarily in the lower elevations of the Sierra Nevada and the Oregon Cascades, covering 8,864 km<sup>2</sup> and making up 2.0% of the seasonal snow zone (SSZ) (Table 4.2; Figure 4.6a). The lowest elevation  $SS_{M1}$  (1000 to 1500 m) is observed in the Oregon Cascades and northern California. The highest elevation  $SS_{M1}$  (2500 to 3000 m) is observed in sparse areas of the mountains in southern California, southern Nevada, and central-eastern Arizona. There is an elevation gradient that runs from high elevation  $SS_{M1}$  in the south to lower elevation  $SS_{M1}$  in the north. In the southern Sierra Nevada and the Kaibab plateau in Arizona the  $SS_{M1}$  is observed between 2000 and 2500 m. In the central and northern Sierra Nevada and the northern California mountains  $SS_{M1}$  between 1500 and 2000 m is observed.

**Table 4.2** Area and percent of seasonal snow zone (SSZ) of the 1<sup>st</sup> ( $SS_{M1}$ ), 2<sup>nd</sup> ( $SS_{M2}$ ), and 3<sup>rd</sup> most sensitive snow zones ( $SS_{M3}$ ) and of the 2<sup>nd</sup> ( $SS_{L2}$ ), and 1<sup>st</sup> least sensitive snow zones ( $SS_{L1}$ ) for the Western United States.

	Area (km <sup>2</sup> )	Percent of SSZ (%)
$SS_{M1}$	8,864	2.0
$SS_{M2}$	56,874	12.7
$SS_{M3}$	143,907	32.1
$SS_{L2}$	197,871	44.2
$SS_{L1}$	40,619	9.1



**Figure 4.6** Elevation of the (a) 1<sup>st</sup> most (SS<sub>M1</sub>), the (b) 2<sup>nd</sup> most (SS<sub>M2</sub>), the (c) 3<sup>rd</sup> most (SS<sub>M3</sub>), the (d) 2<sup>nd</sup> least (SS<sub>L2</sub>), and the (e) 1<sup>st</sup> least (SS<sub>L1</sub>) sensitive snow zones in the Western United States.

The second most sensitive temperature zone ( $SS_{M2}$ ) covers an area of 56,874 km<sup>2</sup> and constitutes 12.7% of the SSZ (Table 4.2; Figure 4.6b). The  $SS_{M2}$  is observed predominantly in the Sierra Nevada and the Cascades; however areas in Nevada, Idaho, Utah, Colorado, Arizona and New Mexico are also included. The lowest elevation  $SS_{M2}$  (750 to 1000 m) is observed in the Washington Cascades and Olympic Mountains. The highest elevation  $SS_{M2}$  (> 2500 m) is observed along the eastern slope Colorado Front Range and the southern Colorado/northern New Mexico Rockies, the southern Sierra Nevada, the southern California Mountains, the southern Utah Wasatch, the Upper Gila Mountains of Arizona/New Mexico, and the central to southern Nevada Ranges. Between the lowest elevation  $SS_{M2}$  in the northwest and the highest elevation  $SS_{M2}$  along the southern and southwest areas there is an elevation gradient.  $SS_{M2}$  between 2000 and 2500 m is observed in the central Sierra Nevada, central Nevada ranges, central Utah Wasatch, and southeastern Colorado Rockies. In the northern Sierra Nevada, northern California Mountains, southern Oregon Cascades and eastern Oregon Mountains  $SS_{M2}$  between 1500 and 2000 m is observed.  $SS_{M2}$  between 1000 and 1500 m is observed in the northern Oregon Cascades, northeastern Oregon Blue Mountains, and the Washington Cascades/Olympic Mountains. The western slope Sierra Nevada  $SS_{M2}$  covers a much wider area than the eastern slope, where  $SS_{M2}$  is only a thin band. A similar western/eastern slope difference is observed in the Oregon Cascades. No  $SS_{M2}$  is observed in Montana.

The third most sensitive snow zone ( $SS_{M3}$ ) is observed in all states in the Western U.S. The  $SS_{M3}$  covers 143,907 km<sup>2</sup> and makes up 32.1% of the SSZ (Table 4.2; Figure 4.6c). The lowest elevation  $SS_{M3}$  (750 to 1500 m) is observed in the Washington Olympic/Cascades Mountains and the Rocky Mountains of northwestern Washington, northern Idaho, and northwestern Montana. The highest elevation  $SS_{M3}$  (>3000 m) is observed in the Colorado

Rockies, the southern Sierra Nevada, the southern Utah Wasatch, the central Nevada ranges, and Humphrey's peak in Arizona. An elevation gradient exists from the lowest elevation  $SS_{M3}$  in the northwest to the highest elevations in the south and southeast.  $SS_{M3}$  between 2500 and 3000 m is observed in the Colorado/New Mexico Rocky Mountains, the southern/central Utah Wasatch and the northern Utah Uinta, the central Sierra Nevada, the central Nevada Ranges and small areas in Wyoming.  $SS_{M3}$  from 2000 and 2500 m is observed in the northern Sierra Nevada and northern California Mountains, the northern Nevada Ranges, the northern Utah Wasatch, northeastern Colorado, Wyoming, southern Idaho, southeastern Oregon and southern Montana. In the Oregon and Washington Cascades, eastern Oregon, the Washington Olympic Mountains, central Idaho and southwestern Montana the  $SS_{M3}$  is observed between 1500 and 2000 m. In the southern Sierra Nevada the lower limit of  $SS_{M3}$  tends to begin above 2500 m on the western slope and above 3000 m on the eastern slope, and  $SS_{M3}$  tends to be thinner on the eastern slope.

The second least sensitive snow zone ( $SS_{L2}$ ) exists predominantly in the Rocky Mountains, with sizable areas of  $SS_{L2}$  present in the northern Washington Cascades and the high elevation Sierra Nevada (Figure 4.6d). The  $SS_{L2}$  covers an area of 197,871 km<sup>2</sup> and makes up 44.2% of the SSZ (Table 4.2). The lowest elevation  $SS_{L2}$  is in northern Montana, Idaho, and Washington, whereas the highest elevation  $SS_{L2}$  is in the Colorado Rockies, the Utah Wasatch and Uinta, the southern Sierra Nevada, and parts of western Wyoming. The  $SS_{L2}$  in the southern Sierra Nevada is at the highest elevations (>3000 m). In the Washington Cascades the  $SS_{L2}$  is mainly observed between 1500 to 2000 m. In northeastern and central Idaho and northwestern Montana the  $SS_{L2}$  is observed as low as 1000 m and as high as 3000 m. Central portions of Idaho show  $SS_{L2}$  ranging from 1000 to 3000 m. In Wyoming  $SS_{L2}$  is mainly observed between 2000 and 3000 m.  $SS_{L2}$  in the northern CO Rockies ranges from 2000 to greater than 3500 m,

and in the southern CO Rockies  $SS_{L2}$  ranges from 2500 to greater than 3500 m. Utah  $SS_{L2}$  is confined to the mid to higher elevations (2000 to 3500 m) of the Wasatch and the Uinta Mountain ranges, with the southern Wasatch  $SS_{L2}$  ranging between 2500 and 3500 m, while the more northern Wasatch/Uinta  $SS_{L2}$  is observed as low as 2000 m in the Wasatch and up to 3500 m in the Uinta. Small areas of  $SS_{L2}$  are observed in the northeastern corner of Nevada (2000 to 3500 m), the OR Cascades, OR Blue Mountains and southeastern OR Mt. Steens (1500 to 3000 m), and the northern NM Rockies (>3000 m). No  $SS_{L2}$  is observed in Arizona.

The least sensitive snow zone ( $SS_{L1}$ ) is confined to the mid to high elevation interior Rockies and small areas within Oregon, Washington, and northern California (Figure 4.6e). The  $SS_{L1}$  covers 40,619 km<sup>2</sup> and constitutes 9.1% of the SSZ (Table 4.2). The predominant area of  $SS_{L1}$  is at the intersection of Idaho, Montana, and Wyoming, and ranges as low as 1500 m to greater than 3500 m. The high elevation Colorado Rockies boast the second most prominent area of  $SS_{L1}$ , which is above 3000 m. The Uinta range in northeastern Utah also maintains a sizeable area of  $SS_{L1}$  above 3000 m. Other smaller areas of  $SS_{L1}$  exist in northern CA (> 3000 m), northern OR (> 2500 m) and Washington (> 2000 m). There is no  $SS_{L1}$  in Nevada, Arizona, and New Mexico.

#### 4.3 DISCUSSION

The first objective of this thesis was to determine the spatiotemporal variations in average snow cover for the Western U.S.; this is achieved with the average monthly snow covered area for January through July. These maps provide insights into snow cover loss patterns in the Western United States. In January and February snow covers a large portion of the Western U.S. including higher-elevation topographically diverse sites, lower elevation foothills and flatter regions (Figure 4.1). In March much of the lower elevation foothills and flatter regions have

begun to lose snow cover due to increasing temperatures at lower elevations and more direct insolation due to less shading in the flatter regions/foothills (Price, 1986). By April, snow cover exists primarily in areas included within the SSZ, which is confined predominantly to the topographically diverse mountain ranges that receive less direct solar radiation due to increased shading (Price, 1986). Snow cover is found predominantly along two main axes: north to south along the Sierra Nevada/Cascades and northwest to southeast along the Rocky Mountain Cordillera. These high elevation, topographically diverse areas tend to remain colder longer (Barry, 2008; Price, 1986; Shinker, 2010) and have greater wintertime SWE accumulation (Fassnacht et al., 2003) than other areas in the Western U.S and therefore melt later in the ablation period (Clow, 2010). As the ablation season progresses through April and May, snow cover increases to progressively higher elevations as lower elevations warm and snow melts. Almost all snow cover in the Western U.S. is ablated by July. Areas with the longest lasting snow cover include the high elevation Sierra Nevada, Colorado Rockies, Utah Uinta and Wasatch, the mid to high elevation Washington/Oregon Cascades and coastal mountains, and the interior Rockies of Idaho, Montana, and Wyoming.

We complete the second and third objectives by using snow persistence to define zones of persistent, transitional, intermittent, and seasonal snow and evaluating the elevation within these zones. The snow zone analyses illustrate the influences of elevation, latitude, and continentality on snowpack in the Western U.S (Figure 4.3). As shown in the focus region analyses, decreasing latitude and increasing continentality result in an increase in the elevation of the snowline. Expanding to the regional perspective shows that this latitude effect, combined with the continental-maritime effect, result in a NW/SE axis of snowline, which starts at the lowest elevations in the Northwestern Washington Cascades and increases to the highest

elevations in the southwestern Rockies. This NW/SE snowline axis is observable in the ISZ, the TSZ, the PSZ, and the SSZ (Figure 4.3). The orientation of this gradient in a NW/SE axis rather than strictly N/S or W/E demonstrates the combined influences of broad-scale latitudinal patterns and meso-scale continental patterns (Barry, 2008; Ives et al., 1974; Shinker, 2010). Across similar latitudes, the SSZ begins at higher elevations in areas with increasing continentality. Across similar meridians, which tend to have similar continentality, the SSZ snowline is observed at lower elevations in higher latitudes. Thus the combined effects of latitude and continentality result in the NW/SE elevation gradient observed in the snowline (Barry, 2008; Ives et al., 1974).

Focus region analyses demonstrated that snow cover tends to occur at lower elevations on western slopes of the Sierra Nevada, the Washington Cascades, and the Montana Rockies. Similarly, in the expanded Western U.S. analyses, several areas readily illustrate the influence of local-scale differences caused by physiography of predominant slope, such as is observed in the ISZ and TSZ in the Sierra Nevada and Oregon Cascades, and the PSZ in the Sierra Nevada and Washington Cascades, which tend to have lower elevation snow lines on the western slopes. Focus region analyses illustrated the importance of elevation in determining local snow cover, with increasing elevations corresponding with increasing snow cover, yet also demonstrated the regional differences on snowline caused by the impacts of latitude and continentality. The Western U.S. analyses show the influence of elevation as a local-scale physiographic control on the presence or absence of snow. Latitude and continentality largely determine the elevation at which snow cover can exist for the SSZ, and in the southern portions of the Western U.S. snow cover exists because the local-scale physiography reaches high enough elevations. If not for the high elevation Rocky Mountains, much of the Western U.S. would probably not receive high

amounts of snow accumulation. This phenomenon is reflected in the lack of seasonal snow cover in New Mexico and Arizona, even though it is possible for seasonal snow cover to exist at very high elevations, as is shown by the presence of a SSZ in the highest mountaintops of New Mexico and Arizona. However, the elevation of the Rockies quickly drops off in New Mexico & Arizona, and therefore much less snow cover is observed in these states. While areas of the SSZ are primarily confined to mountainous terrain, areas within the ISZ are largely observed in flatter areas and the more undulating foothills of the Western U.S. As mentioned previously, these areas receive more direct insolation and lower amounts of SWE, which is likely the reason snow in these areas tends to melt quickly.

We assess the spatiotemporal variations of temperature within the seasonal snow zone by looking at the progression of the average 8-day LST within the seasonal snow zone during winter and spring, which was the fourth objective of this thesis. Finally, we complete the fifth objective by using the wintertime (January-February-March) average LST to define the relative sensitivity of the seasonal snow zone to climate change. Areas that warm first (last) generally coincide with areas highlighted as most (least) sensitive to climate change (Figure 4.4; Figure 4.5; Figure 4.6). Although sensitivity of snow to climate change depends on both precipitation and temperature, for this study we have used temperature only. While observed temperatures and those modeled in global climate models have tended to increase in most areas of the Western U.S., observed and modeled precipitation trends have been variable, largely due to the interannual variability of precipitation (Brown and Mote, 2009; Hamlet et al., 2005; Stewart, 2009). Previous studies have found that decreases in SWE and decreases in SWE to precipitation ratios have largely been explained by increases in temperature rather than by changes in precipitation amount (Brown and Mote, 2009; Hamlet et al., 2005; Stewart, 2009). Additionally, warmer MODIS land surface



temperatures have been found to coincide with periods of melt and mass loss on the Greenland ice sheet (Hall et al., 2008). We therefore used land surface temperature as a proxy for climate change sensitivity under the assumption that the warmest areas will be most sensitive to shifting from seasonal snow cover to intermittent snow cover. By using the data-derived seasonal snow zone, which represents areas that, based on an 11-year average, have greater than 50% snow cover on April 1<sup>st</sup>, we base the sensitivity analysis explicitly on areas with documented seasonal snow cover.

The snow sensitivity classifications in this study are defined based on the spread of the land surface temperature data, rather than fixed temperature thresholds, as has been implemented in other modeling and climate change studies (Brown, 1997; Brown, 2000; Brown and Mote, 2009; Casola et al., 2009; Lynch-Stieglitz, 1994; Nolin and Daly, 2006; Stewart, 2009). Modeling and climate change studies have often used an air temperature of 0°C to represent the threshold that determines whether precipitation falls as rain or snow (Brown, 1997; Brown, 2000; Brown and Mote, 2009; Fassnacht and Soulis, 2002; Lynch-Stieglitz, 1994). However, snow can fall at air temperatures greater than 0°C, and rain can fall at temperatures lower than 0°C due to the fact that snow formation depends on cloud temperature, rather than air temperatures near the ground surface (Lackmann et al., 2010). Just as air temperature will not necessarily reflect snowfall conditions, neither will land surface temperature; in fact land surface temperatures will be more complex in their relation to cloud conditions, and therefore snowfall, due to land surface coverage conditions (e.g. snow cover percentage, the presence/absence of vegetation, vegetation type and land use). Therefore, to define appropriate sensitivity thresholds, the methods in this study do not rely on relatively arbitrary land surface temperature values such as 0°C, but rather consider the mean and standard deviation of the land surface temperature

within the seasonal snow zone to define the sensitive snow zones. This presents a data-driven method for determining relative sensitivity of snow based on observed temperature conditions where seasonal snow cover is documented. One weakness of this approach is the relatively large (1 km) spatial resolution of the MODIS LST product, and the subsequent averaging over each huc100 spatial unit, which can cause the LST to be influenced by a wide variety of land cover types. Similarly, the LST used for sensitive snow zone definitions is averaged over a wide time period (January-February-March average LST for the 2000 to 2010 average). However, in viewing the similarities between the relatively warmer and colder areas observed in both the LST<sub>JFM</sub> map and the 8-day 2000 -2010 average temporal evolution of land surface warming from January to May, we feel confident that the defined sensitive snow zones give an accurate representation of the temperature conditions that would affect relative sensitivity of snow to climate warming in the Western U.S.

For the sensitive snow zones, there is a northeast to southwest gradient of snow sensitivity to climate change, with areas along the outer rim of the southwest and the maritime regions being some of the most sensitive (Figure 4.6a,b). The most sensitive snow zones are lower to mid elevation Sierra Nevada (1500 to 2500 m), followed by the low to mid elevation Cascades (750 to 2000 m), the mid-elevations (2000 to 3000 m) in southern and central Utah Wasatch Mountains, mid to higher elevations (>2500 m) of the southern California Mountains, the eastern Colorado Front Range, and the southern San Juan Mountains in Colorado (Figure 4.6a,b). Sensitive snow zones also exist in the Nevada Basin and Range. The least sensitive snow zones are found predominantly in the interior regions of the Western U.S. (the northern Rockies in Idaho/Montana/Wyoming) (Figure 4.6d,e). Other areas included in the least sensitive snow zones are the mid to high elevation (> 2500 m) Colorado Rockies and Utah Uinta/Wasatch,

the high elevation (> 3000 m) Sierra Nevada, and the low to high elevation (> 1500 m) Washington Cascades and mid elevation (> 2000 m) Oregon Cascades/Blue Mountains.

The sensitive snow zones highlighted in this study are consistent with other studies of climate change sensitivity in the West. The Pacific Northwest (Brown and Mote, 2009; Hamlet et al., 2005; Mote et al., 2005; Nolin and Daly, 2006; Serreze et al., 1999; Stewart, 2009) and the lower and mid-elevation Sierra Nevada (Brown and Mote, 2009; Hamlet et al., 2005; Mote et al., 2005; Stewart, 2009) are considered the most sensitive areas to climate change in the Western U.S. The southwest is also considered highly sensitive to climate change (Mote et al., 2005; Stewart, 2009). In this study, some areas in Colorado are also found to be sensitive snow zones, especially along the eastern Colorado Front Range and in the southern Colorado San Juan Mountains; these areas may not have been highlighted as sensitive in previous studies because ground-based monitoring networks at mid elevations in the Colorado Rockies are sparse. Although the Colorado Rockies are relatively cold, Clow (2010) suggested that Colorado may not be resistant to climate change and that the colder Colorado climate may initially mask the impacts of climate change. Further, in areas where snowfall depends more on precipitation than on temperature, the impacts of climate change may not be as obvious. Western U.S. maritime snowfall tends to be temperature dependent, while interior Western U.S. snowfall is precipitation dependent (Hamlet et al., 2005; Mote et al., 2005; Serreze et al., 1999). Likewise, higher elevation sites, such as the high elevation southern Sierra Nevada, maintain very cold wintertime temperatures, and snowfall is thus precipitation dependent (Hamlet et al., 2005; Mote et al., 2005). Sensitivity of snow zones to climate warming may be harder to predict in these precipitation-dependent regions. However, if the assumption that temperature is the primary driver of snow sensitivity to climate is correct in most locations, our results show that the initial

impacts of climate change on the seasonal snow zone will likely become apparent in the warmer areas of the Western U.S.- the lower to mid elevations of the maritime regions and the southwestern U.S. If warming continues to take place, colder interior climates may also become impacted by shifts from seasonal to intermittent snow cover.

Snow cover and to a lesser extent, land surface temperature, can vary markedly from year to year due to interannual variability and decadal oscillations (e.g. ENSO, PDO). Furthermore, a warming climate will change the hydrology and biota of the Western U.S. As mentioned in Chapter 1, decreases in snow water equivalent (Hamlet et al., 2005; Mote et al., 2005; Stewart, 2009), decreases in snow to precipitation ratios (Knowles et al., 2006; Pederson et al., 2010; Stewart, 2009) and earlier onset of streamflow rise (Cayan et al., 2001; Clow, 2010; Stewart, 2009) have already been documented in many areas throughout the west. Furthermore, in large portions of the western U.S., especially in Colorado and Wyoming, there has been massive tree die-off (in some areas with nearly 100% tree mortality) due to mountain pine beetle infestations (Samman and Logan, 2000), which may cause increased snow accumulation and more rapid spring melt (Mikkelsen et al., 2011). Due to the short time period covered by the MODIS data used in this study, we have not attempted to determine the influences of these processes on the results. The averages of SCA used in this study instead provide a climatological perspective on snow conditions. These averages will be influenced by the range of variability in climate and land cover conditions within the time period of analysis, and we suggest that as additional years of MODIS data become available, future studies should use these data sets to evaluate the influence of interannual and decadal variability, climate change, and bark beetle kill on snow cover and land surface temperature in the Western U.S.

#### 4.4 SUMMARY

The main areas of snow cover in the Western United States exist along two primary axes: from north to south along the Washington Cascades and the Sierra Nevada, and from northwest to southeast along the axis of the Rocky Mountain Cordillera. The snowline in the Western U.S. shows a strong elevation gradient from the northwest (low elevation) to the southeast (high elevation) caused by the combined influences of latitude and continentality. The seasonal snow zone is generally confined to higher-elevation, topographically diverse areas, while the intermittent snow zone often is observed in lower elevation foothills and flat areas, which receive more direct insolation. Within the seasonal snow zone, the warmest areas are the lower to mid elevation maritime mountain regions and the mid to higher elevation southwest, whereas the coldest areas are observed in the northern Idaho/Montana/Wyoming Rockies, the high elevation Sierra Nevada, the mid to high elevation Utah Wasatch/Uinta, the Colorado Rockies, the lower to high elevation Washington Cascades, and small areas in mid to high elevations of Oregon and Nevada. Areas in the seasonal snow zone that may be most sensitive to climate change are predominantly along the lower to mid elevation maritime ranges and the southwestern United States, while the least sensitive areas are the interior northern Idaho/Montana/Wyoming Rockies, the mid to higher elevation Colorado Rockies, the high elevation Sierra Nevada and the mid to high elevation Washington Cascades. Resource managers in these sensitive snow zones should take into consideration the potential consequences of climate change on snowpacks in order to prepare for an altered hydrologic regime.

## CHAPTER 5: CONCLUSIONS

Snow covered area, land surface temperature, and precipitation exhibit considerable spatiotemporal variability throughout the Western United States. This study illustrates the climatic variability caused by physiography and spatial location at the local-scale, meso-scale, and broad scale. In Chapter 3, the focus region analyses offer in depth understanding of the influences of latitude, continentality, elevation, and slope orientation on snow cover, land surface temperature, and precipitation. Latitude influences these variables on the broad scale (>100 km) through insolation inputs, which cause lower temperatures at higher latitudes, and global atmospheric circulation patterns, which cause higher precipitation at higher latitudes. The combination of greater precipitation and lower temperatures in the higher latitudes of the Western U.S. causes lower elevation, longer lasting snow cover. Continentality, or distance from water sources, creates variability in snow cover, land surface temperature, and precipitation at the meso-scale (1-100 km); areas with increasing continentality tend to receive less precipitation, due to decreasing moisture content of air masses as they move further into the continent, and as a result have snow cover at higher elevations.

Elevation and slope orientation influence precipitation, land surface temperature, and snow cover on a local-scale (1-10 km), primarily due to the impacts of orography. Higher elevations tend to have more precipitation, lower land surface temperatures, and greater snow covered area. Slope orientation (western versus eastern slope) can have substantial influences on precipitation, land surface temperature, and snow cover, as is observed in the Sierra Nevada, the

Washington Cascades, and the Montana Rockies, but such is not always the case, as is observed in the Colorado Rockies. In areas where western and eastern slope differences are observed, western slopes tend to receive more precipitation, have colder land surface temperatures, and have snow cover at lower elevations. Many of the storms that the Sierra Nevada, the Washington Cascades, and the Montana Rockies receive are westerly, whereas the Colorado Rockies is subject to frequent upslope storm events as well as southerly Gulf of Mexico storm tracks; this may explain the relative similarity in western and eastern slope snow cover for the Colorado Rockies compared to the other focus regions.

Analyses in Chapter 4 provide insights into the spatial and temporal differences in climatology and sensitivity to climate change of the Western U.S. Most of the snow cover in the Western U.S. is observed along two main axes: north to south along the Cascades and Sierra Nevada, and northwest to southeast along the Rocky Mountain cordillera. At the beginning of winter (January and February) snow cover is also observed in many of the flatter areas and lower, undulating foothills. These areas often receive lower snowfall as well as more direct insolation than the more topographically diverse mountain ranges and therefore lose snow cover sooner in the melt season. Although some areas in the higher elevation mountain ranges can receive high amounts of solar radiation due to decreased atmosphere, overall the colder temperatures and shading in these areas lead to snow patches that can last well into the melt season. As snowmelt progresses through the season snow is forced to higher elevations and lasts longest in the high elevation Rockies, Washington Cascades, and southern Sierra Nevada.

The maps depicting elevation of the snow zones illustrate the combined influences of latitude and continentality on the elevation of the snow line. All snow zones demonstrate an elevation gradient of the snowline, which runs from the lowest elevations in the northwest to the

highest elevations in the southeast of the Western United States. Latitude and continentality combine to form the diagonal northwest to southeast gradient, rather than a strict north to south gradient if latitude was the only factor or a west to east gradient if continentality was the only factor. The influence of local-scale topography on snow cover is also apparent in the snow zone maps. Much of the higher latitude areas have lower elevations overall, yet they remain snow covered because the theoretical snow line occurs at low elevations. The theoretical snow line in Colorado is very high in elevation, but because of the high elevations of these mountains, they still receive a substantial snowpack.

Within the seasonal snow zone, land surface temperatures highlight seasonally snow covered environments that may be sensitive to climate warming in the Western United States. Both the temporal progression of land surface temperature and the land surface temperature average for January-February-March reveal similar spatial patterns in the temperatures of the seasonal snow zone. The coldest areas are located predominantly in the northern interior Rockies around the intersection of Idaho, Montana, and Wyoming, and the high elevation Colorado Rockies; however cold areas are also observed in the high elevations of the Sierra Nevada and Utah Wasatch/Uinta, as well as low to high elevation Washington Cascades. If changes in seasonal snow cover are primarily driven by temperature, these cold areas are likely the least sensitive to climate change. In contrast, the warmest snow zones are located in the mid-elevation Sierra Nevada, along the lower to mid-elevation Cascades, in mountains along the southwest and on the eastern Colorado Front Range. Because these areas have average winter land surface temperatures at or even above freezing, they may be most sensitive to climate warming. Between these two zones exists a temperature/sensitive snow zone gradient from warmest (most sensitive) in the southwest to coldest (least sensitive) in the northeast.



Our findings on the sensitivity of snow zones to warming are similar to other climate change sensitivity studies (Brown and Mote, 2009; Clow, 2010; Hamlet et al., 2005; Mote et al., 2005; Nolin and Daly, 2006; Serreze et al., 1999; Stewart, 2009), illustrating the robustness of this data-driven approach to defining sensitive snow zones. Furthermore, the use of spatially continuous remotely sensed data allowed us to identify the spatial extent of these sensitive snow zones, as well as to identify sensitive areas such as the mid-elevation Colorado Front Range which could be missed by studies that relied on ground based measurement networks.

The methods of snow zone mapping and climate sensitivity analysis presented in this thesis take full advantage of a rich spatiotemporal dataset from MODIS. Climate trend analyses and climate sensitivity studies often utilize point data (Armstrong and Armstrong, 1987; Cayan, 1996; Clow, 2010; Knowles et al., 2006; Mizukami and Perica, 2008; Mock, 1996; Mote et al., 2005; Serreze et al., 1999), yet the network of climate and snow monitoring stations in the Western U.S. is sparse, and often higher elevation areas, which receive substantial amounts of snow accumulation, have no station data and are therefore under-represented in station-based analyses. This study clearly shows the utility of satellite imagery in spatiotemporal studies of land surface characteristics and climate change sensitivity.

This study has characterized the 11-year climatology and climate sensitivity of MODIS derived snow cover and land surface temperature using both an in-depth focus region and a regional-scale approach. However, 11 years is too short of a time period to determine any trends and is a fairly short time period for deriving true climatological averages considering the influences of interannual variability, decadal oscillations (e.g. ENSO and PDO), warming temperatures trends, and changes in vegetation (e.g. bark beetle kill in the Rocky Mountains). However, the utility of remote sensing to monitor snow zones is clearly apparent from our

results. We therefore suggest that programs such as MODIS continue to monitor the earth surface, and, as technology progresses, that new techniques for measuring snow and temperature be implemented. In future years we suggest that trend analyses be conducted on MODIS derived snow cover and land surface temperature to evaluate how these variables change over time.

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## **APPENDIX A: FOCUS REGION BASIN STATISTICS**

**Table A.1:** Huc ID, easting, northing, elevation minimum/maximum/mean/standard deviation and area for focus regions.

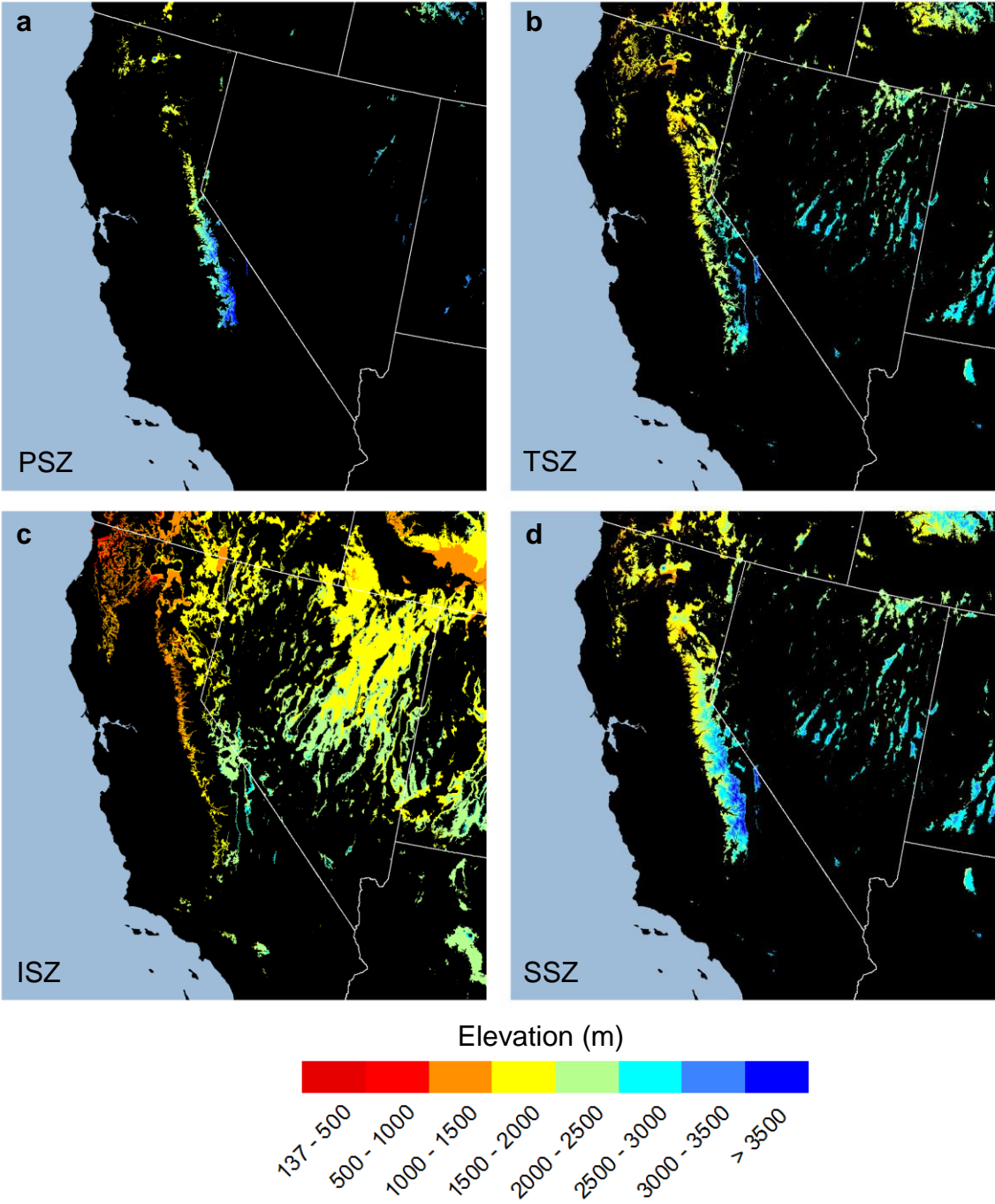
HUC ID	Centroid (m)		Elevation (m)				Area (km <sup>2</sup> )
	Easting	Northing	Min	Max	Mean	Std Dev	
<b>Western Slope Washington Cascades</b>							
17110006	-1876003	3052183	30	3162	1124	538	1921
17110009	-1899647	3014854	28	2388	840	487	2177
17110010	-1921911	2994445	29	2206	637	461	1797
17110012	-1948239	2998747	0	1593	264	324	1579
17110013	-1942259	2966163	0	1696	516	424	1231
17110014	-1952256	2940715	0	4366	865	621	2609
17110015	-1984450	2926031	0	4319	553	487	1826
17080004	-1954476	2881125	241	4365	1113	441	2679
<b>Eastern Slope Washington Cascades</b>							
17020008	-1789860	3065506	228	2659	1413	498	4737
17020009	-1822635	3036303	183	2785	1340	563	2412
17020010	-1818014	2951285	149	2742	826	473	3854
17020011	-1851923	2989125	183	2783	1194	479	3436
17030001	-1865951	2922370	363	2351	972	362	5542
17030002	-1903646	2897337	334	2437	1242	379	2891
17070106	-1924585	2816656	61	3575	906	397	3488
<b>Western Slope Montana Rockies</b>							
17010207	-1304561	2935468	960	3048	1716	351	2939
17010209	-1310621	2875705	914	2681	1771	338	4338
17010203	-1299526	2787775	1000	2800	1648	310	5990
<b>Eastern Slope Montana Rockies</b>							
10030202	-1238522	2962643	1006	2793	1311	185	3107
10030201	-1252817	2924985	1006	2864	1483	316	3335
10030205	-1193581	2864940	792	2743	1231	295	5308
10030104	-1233480	2837852	1026	2772	1506	397	4869
10030102	-1168788	2809896	792	2743	1219	262	6737
<b>Western Slope Sierra Nevada</b>							
18020128	-2085505	2054193	103	2971	1371	597	2620
18020129	-2088205	2022825	121	3041	1297	733	2201
18040013	-2106516	2005422	56	2330	818	552	1649
18040012	-2086091	1982939	91	3086	1318	773	2054
18040010	-2069543	1952186	62	3483	1585	826	2596
18040009	-2058474	1917466	65	3941	1719	935	4169
18040008	-2056933	1884295	121	3840	1588	905	2842
18040006	-2018017	1835036	133	4152	2077	965	4400



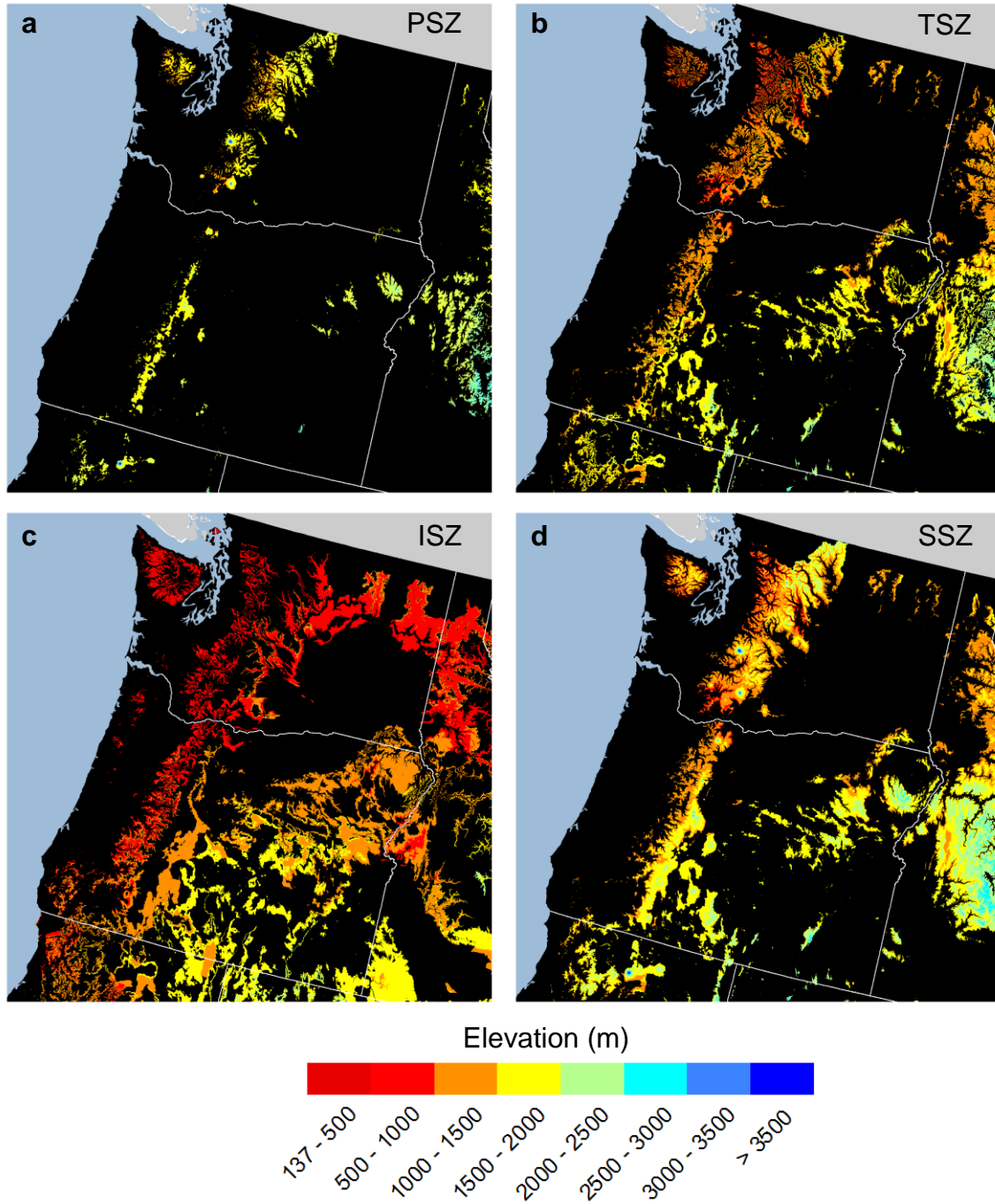
**Table A.1 (cont):** Huc ID, easting, northing, elevation minimum/maximum/mean/standard deviation and area for focus regions.

HUC ID	Centroid (m)		Elevation (m)				Area (km <sup>2</sup> )
	Easting	Northing	Min	Max	Mean	Std Dev	
<b>Eastern Slope Sierra Nevada</b>							
16050102	-2014540	2079453	1256	3230	1839	343	3122
16050101	-2038825	2038848	1889	3250	2123	274	1323
16050201	-2019219	2008221	1404	3413	2033	438	2435
16050302	-1997978	1979526	1386	3539	2072	495	2552
16050301	-1977489	1947870	1400	3719	2231	424	2839
18090101	-1973372	1899974	1945	3958	2340	369	2784
18090102	-1953480	1837436	1112	4302	2216	698	4986
18090103	-1937530	1728155	1055	4321	1760	698	3556
<b>Western Slope Colorado Rockies</b>							
14050001	-933834	1992671	1889	3745	2445	379	6796
14010001	-879269	1935755	1767	4092	2783	374	7533
14010002	-860224	1895170	2227	4312	3128	416	1771
14010003	-900936	1889922	1889	4143	2864	462	2518
14010004	-935818	1858599	1767	4280	2919	517	3754
14020001	-920962	1809852	2438	4197	3201	315	1998
14020003	-919586	1756468	2342	4279	2947	329	2859
13010004	-896149	1713520	2253	3991	2765	388	3498
<b>Eastern Slope Colorado Rockies</b>							
10180001	-863056	2004093	2385	3860	2706	262	3694
10190007	-776541	2007149	1401	4052	2140	561	4852
10190006	-782850	1970500	1425	4333	2224	624	2151
10190005	-785845	1934793	1448	4302	2193	659	2562
10190004	-804327	1900300	1554	4289	2703	658	1478
10190002	-787849	1857553	1568	4334	2448	529	4785
10190001	-833994	1828938	2255	4306	2944	300	4168
11020001	-851227	1775401	1647	4383	2853	499	7934

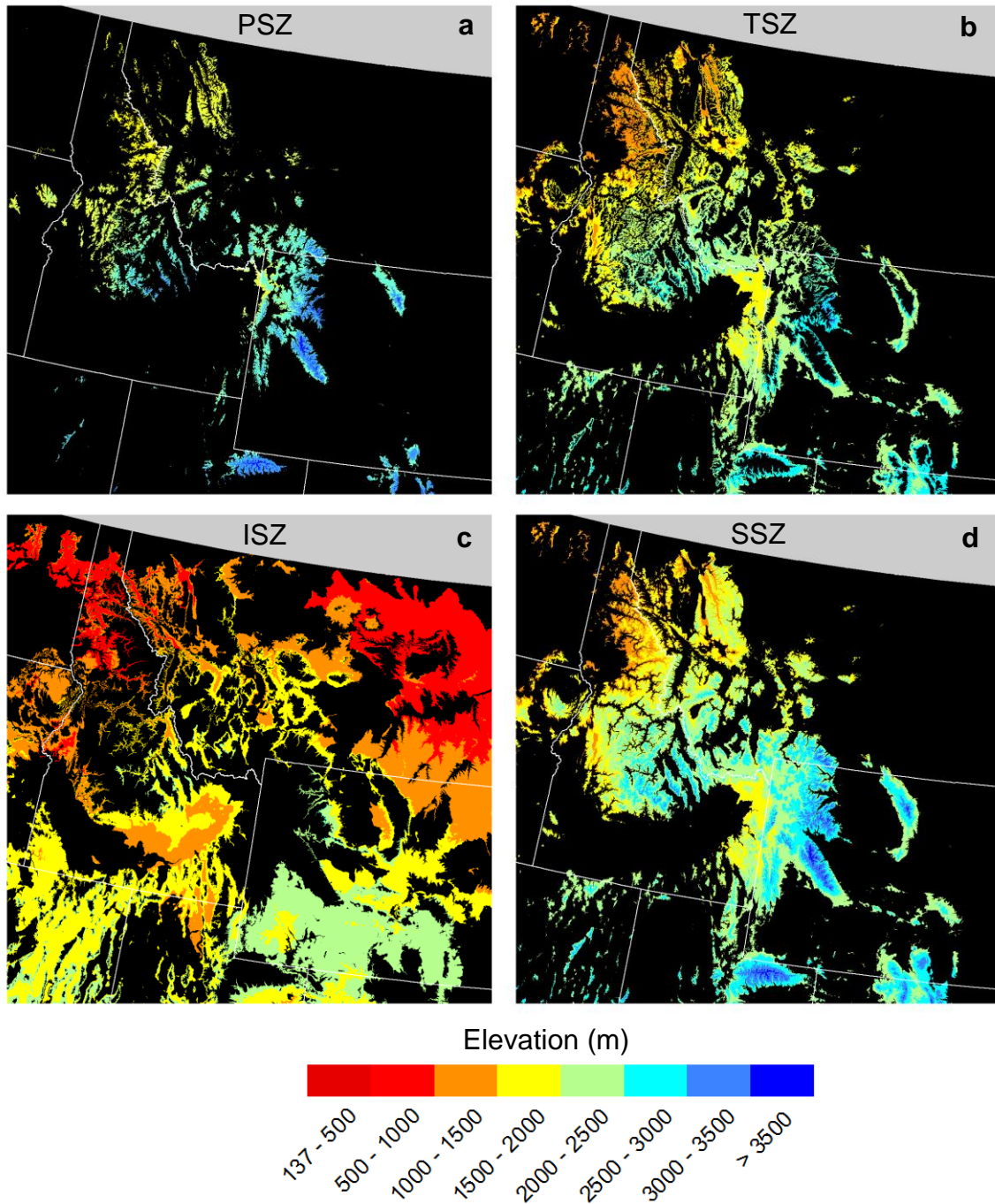
## **APPENDIX B: SNOW ZONES**



**Figure B.1** Elevation of the (a) persistent snow zone (PSZ), (b) transitional snow zone (TSZ), (c) intermittent snow zone (ISZ), and (d) seasonal snow zone (SSZ) for California and Nevada.

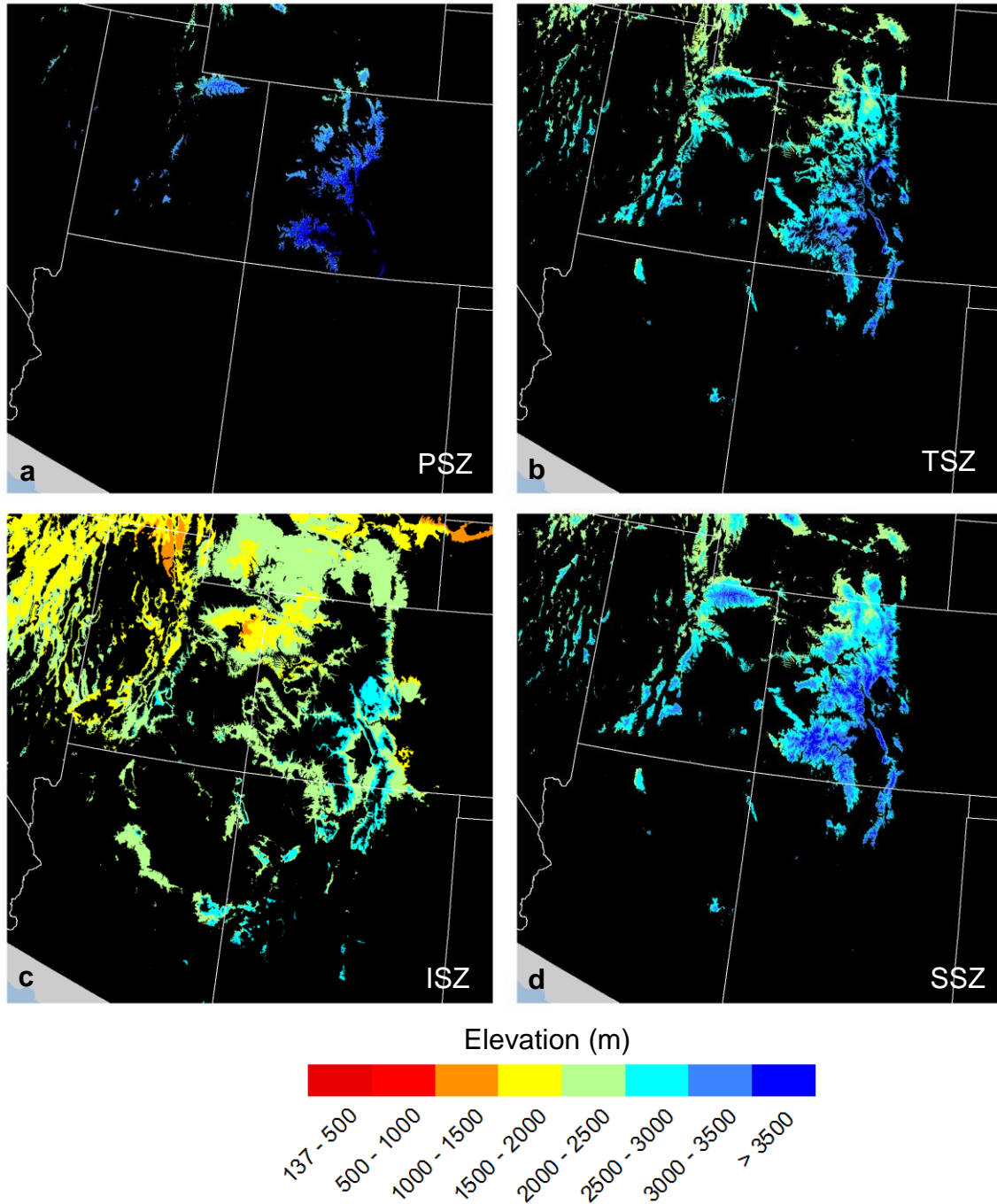


**Figure B.2** Elevation of the (a) persistent snow zone (PSZ), (b) transitional snow zone (TSZ), (c) intermittent snow zone (ISZ), and (d) seasonal snow zone (SSZ) for Oregon and Washington.

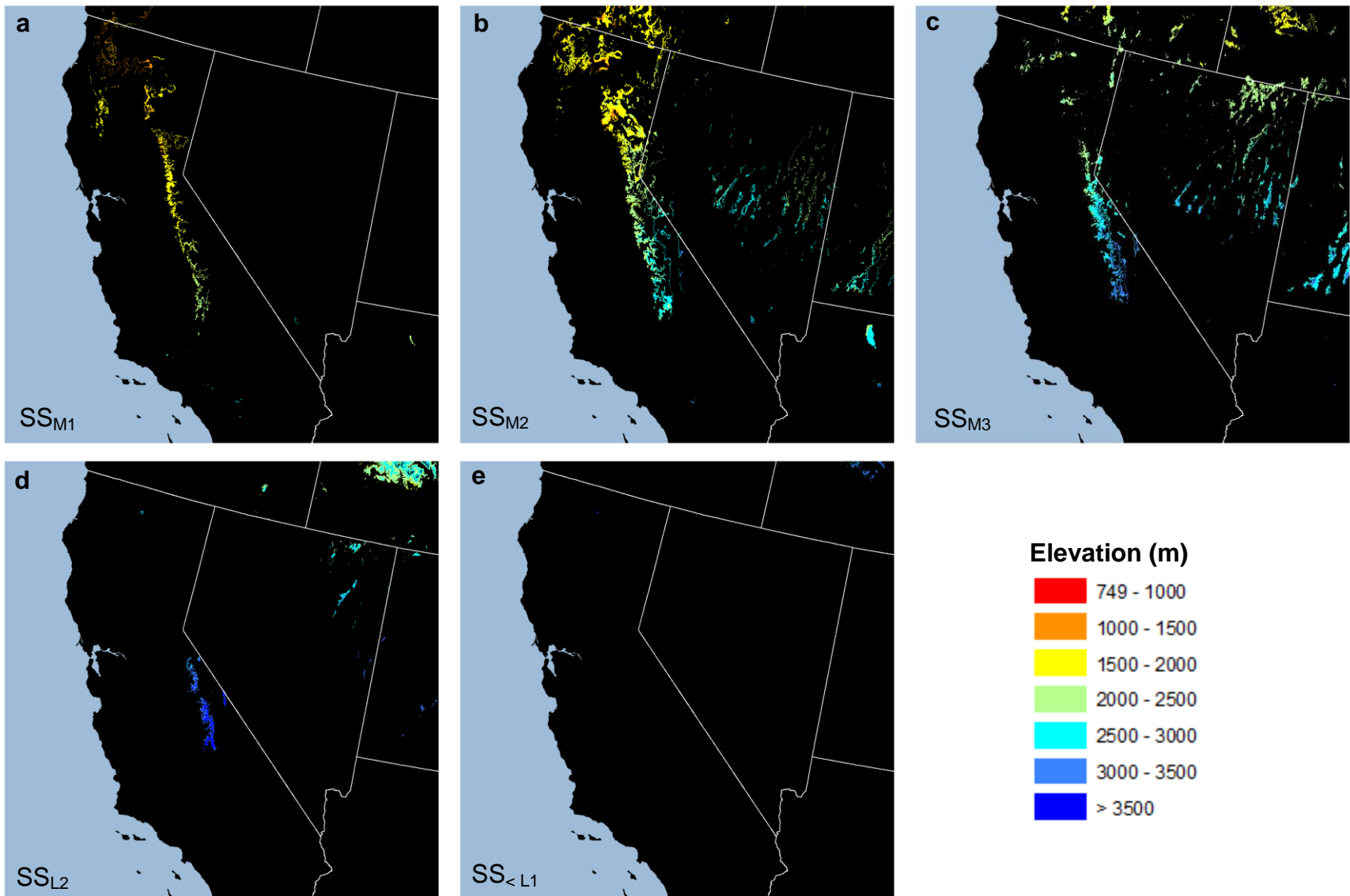


**Figure B.3** Elevation of the (a) persistent snow zone (PSZ), (b) transitional snow zone (TSZ), (c) intermittent snow zone (ISZ), and (d) seasonal snow zone (SSZ) for Idaho, Montana, and Wyoming.

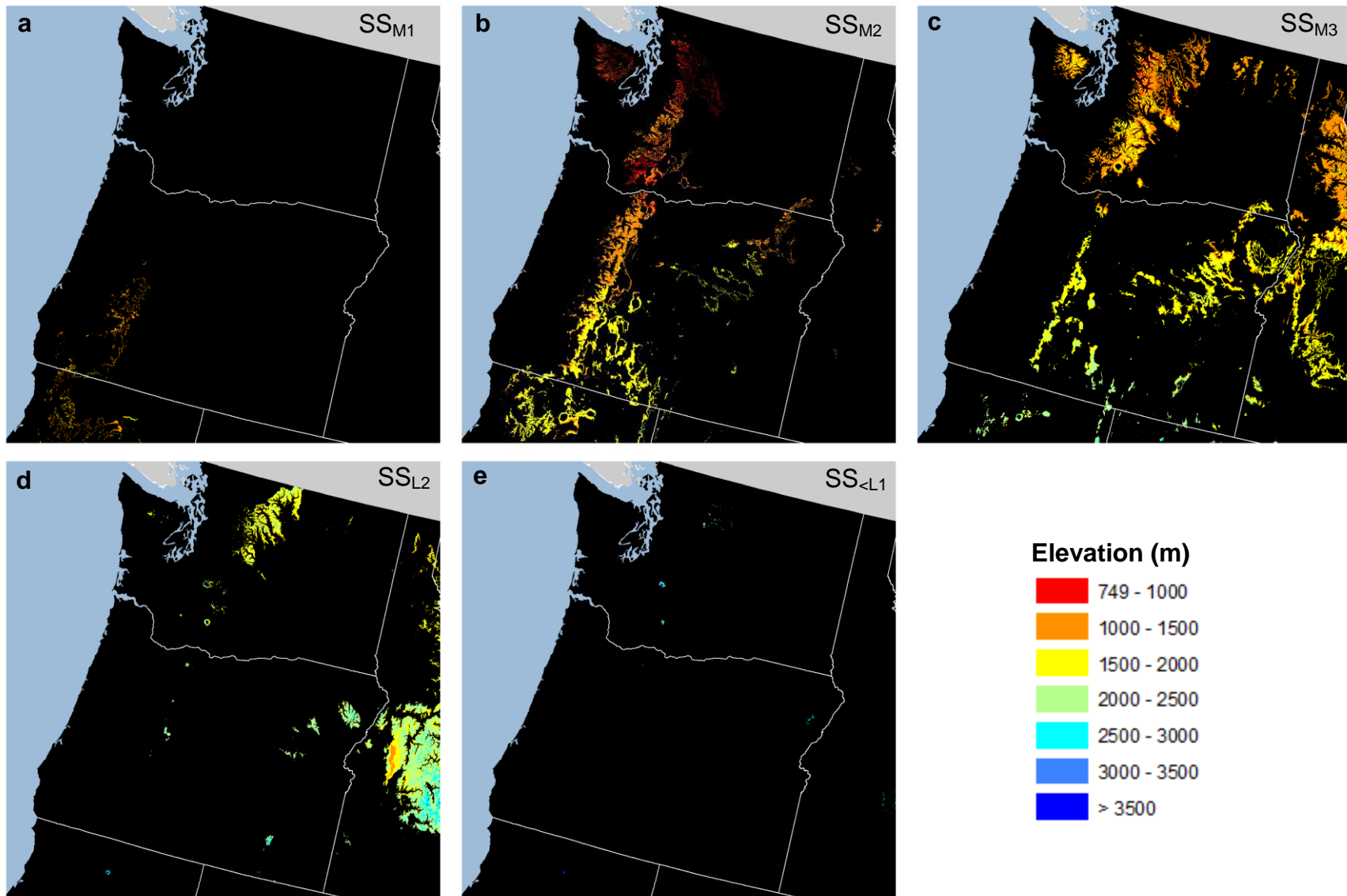




**Figure B.4** Elevation of the (a) persistent snow zone (PSZ), (b) transitional snow zone (TSZ), (c) intermittent snow zone (ISZ), and (d) seasonal snow zone (SSZ) for Utah, Colorado, Arizona, and New Mexico.

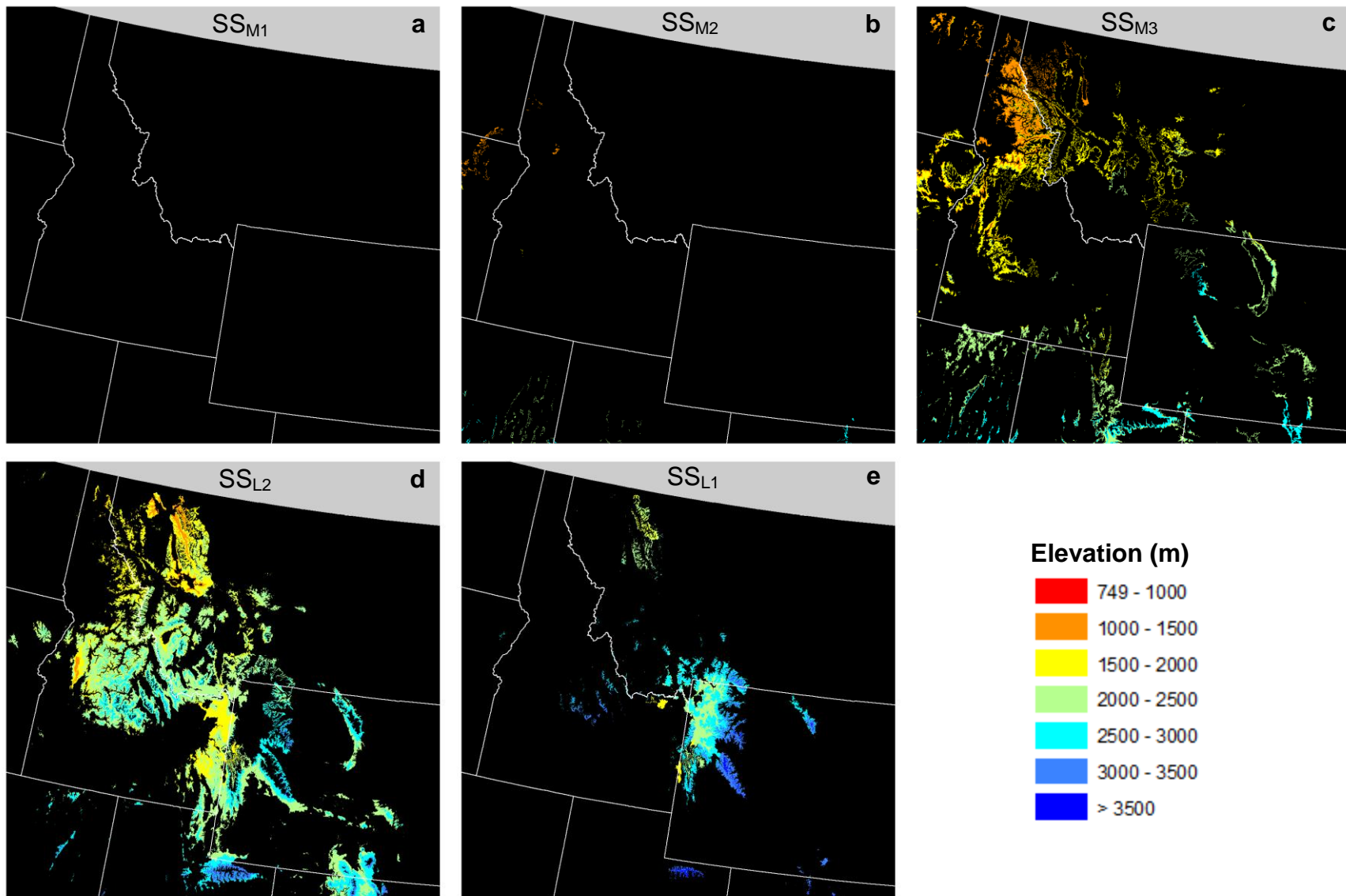


**Figure B.5** Elevation of the (a) 1<sup>st</sup> most ( $SS_{M1}$ ), the (b) 2<sup>nd</sup> most ( $SS_{M2}$ ), the (c) 3<sup>rd</sup> most ( $SS_{M3}$ ), the (d) 2<sup>nd</sup> least ( $SS_{L2}$ ), and the (e) 1<sup>st</sup> least ( $SS_{L1}$ ) sensitive snow zones in California and Nevada.

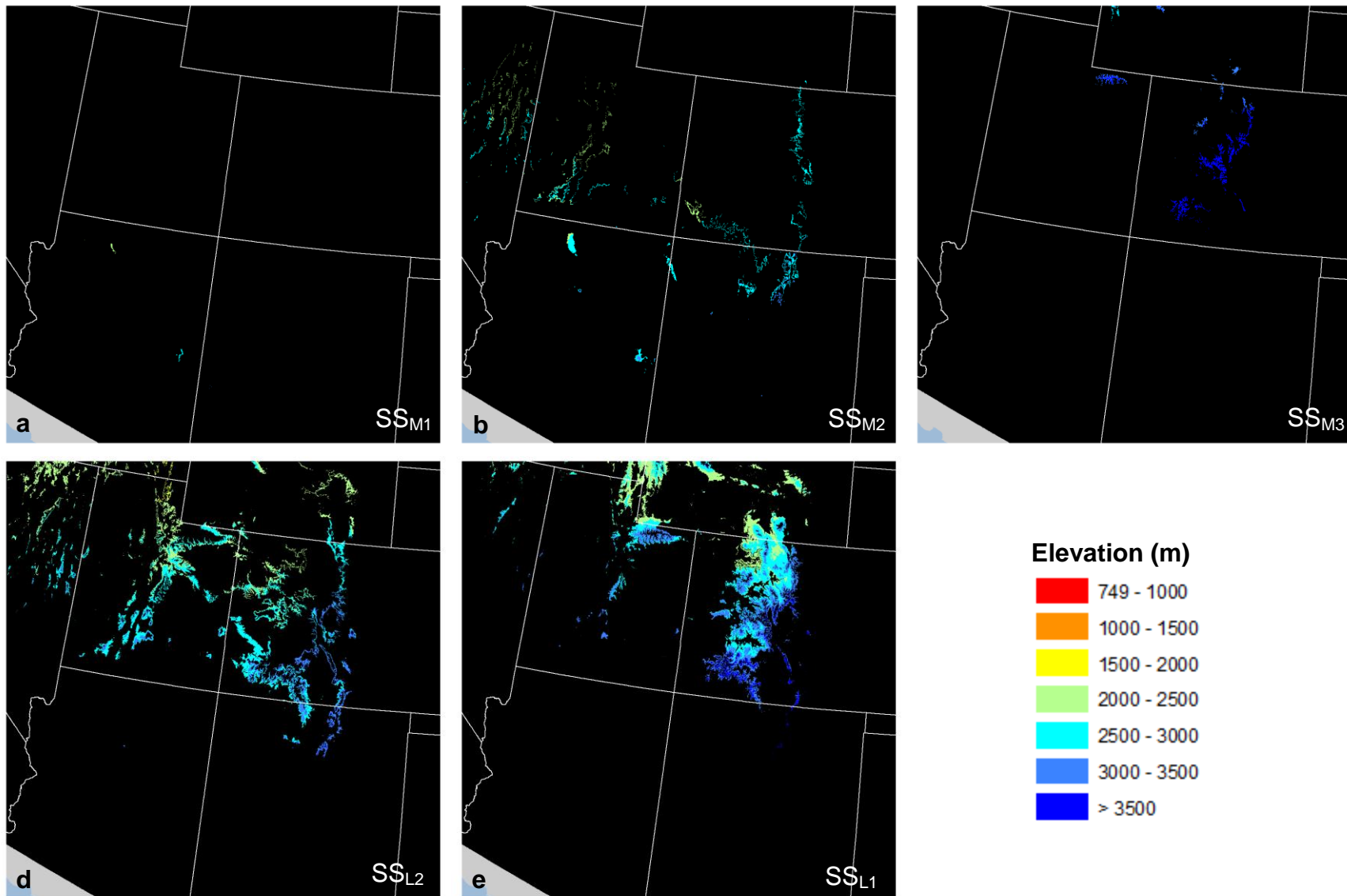


**Figure B.6** Elevation of the (a) 1<sup>st</sup> most ( $SS_{M1}$ ), the (b) 2<sup>nd</sup> most ( $SS_{M2}$ ), the (c) 3<sup>rd</sup> most ( $SS_{M3}$ ), the (d) 2<sup>nd</sup> least ( $SS_{L2}$ ), and the (e) 1<sup>st</sup> least ( $SS_{L1}$ ) sensitive snow zones in Oregon and Washington.





**Figure B.7** Elation of the (a) 1<sup>st</sup> most ( $SS_{M1}$ ), the (b) 2<sup>nd</sup> most ( $SS_{M2}$ ), the (c) 3<sup>rd</sup> most ( $SS_{M3}$ ), the (d) 2<sup>nd</sup> least ( $SS_{L2}$ ), and the (e) 1<sup>st</sup> least ( $SS_{L1}$ ) sensitive snow zones in Idaho, Montana, and Wyoming.



**Figure B.8** Elevation of the (a) 1<sup>st</sup> most ( $SS_{M1}$ ), the (b) 2<sup>nd</sup> most ( $SS_{M2}$ ), the (c) 3<sup>rd</sup> most ( $SS_{M3}$ ), the (d) 2<sup>nd</sup> least ( $SS_{L2}$ ), and the (e) 1<sup>st</sup> least ( $SS_{L1}$ ) sensitive snow zones in Utah, Colorado, Arizona, and New Mexico.