Climate Change Impacts on Snowpack Heterogeneity: Spatial and Temporal Variability at Multiple Scales

A Dissertation
Presented in Partial Fulfillment of the Requirements for the
Degree of Doctor of Philosophy
with a
Major in Water Resources
in the
College of Graduate Studies
University of Idaho
by
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May 2019

Authorization to Submit Dissertation

This dissertation of Adrienne M. Marshall, submitted for the degree of Doctor of Philosophy with a Major in Water Resources and titled "Climate Change Impacts on Snowpack Heterogeneity: Spatial and Temporal Variability at Multiple Scales" has been reviewed in final form. Permission, as indicated by the signatures and dates below, is now granted to submit final copies to the College of Graduate Studies for approval.

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Abstract

Throughout the western United States, seasonal snowpack is critical for water resources timing and availability and ecosystem function. Warming temperatures associated with climate change reduce snow accumulation and advance melt timing, with serious consequences for snow-dependent social and ecological systems. While many impacts of climate change on snowpack are well established, this dissertation investigates several elements of changing snowpack that have not been previously assessed. In particular, each chapter contributes to an improved understanding of the changing heterogeneity of snow under climate change. The first chapter tests the sensitivity of snow drifting to altered climate, using a physically-based hydrologic model and thirty years of hydroclimatological data at a site where aspen stands are subsidized by a wind-driven snow drift. We find a warming-induced reduction in snow drifting, increase in ecohydrologic homogeneity across the landscape, and altered interannual variability of hydrologic metrics. The second chapter assesses changes in interannual variability of snowpack magnitude and timing across the western United States, using downscaled global climate model data as forcing to the Variable Infiltration Capacity (VIC) model. We find that changes in interannual variability are spatially heterogeneous across the western U.S., but that interannual variability of annual maximum snow water equivalent (SWE_{max}) decreases in regions transitioning from snow- to rain-dominated precipitation regimes. Changes in the date of SWE_{max} are less spatially coherent, but agreement between general circulation models (GCMs) is most reliably found at relatively warm sites where the date of SWE_{max} variability increases. The third chapter assesses another element of snow heterogeneity by testing the effect of snowfall intensity on winter ablation. Using a statistical modeling approach with observational snow data, we find that higher snowfall intensity is associated with reduced winter ablation; projected changes in snowfall intensity will likely exacerbate warming-induced increases in winter ablation in the maritime mountains of the western U.S. and mitigate it in the cooler continental regions. Finally, a fourth interdisciplinary, collaborative chapter synthesizes research on climate change in the mountainous headwaters of the Columbia River Basin. Findings show that research in this basin is focused on climate change impacts, rather than adaptation or mitigation, that social and biophysical sciences are not well integrated, and that research priorities differ across an international boundary. Cumulatively, this set of studies advances knowledge of how the spatial and temporal heterogeneity of snowpack will respond to climate change in the western United States, with implications for snow-dependent social and ecological systems.

Acknowledgements

This dissertation would not have been possible without the support of many individuals and institutions. Most importantly, I appreciate the thoughtful mentorship and guidance of Dr. Timothy Link, who has worked tirelessly to improve the contents and presentation of this research. I am also greatly indebted to my committee members, Drs. John Abatzoglou, Jan Eitel, and Tara Hudiburg, for supporting and challenging me throughout the dissertation process. I am also grateful to Drs. Andrew Robinson, Danny Marks, and Gerald Flerchinger, who have provided advisory support on individual chapters.

My fellow students in the University of Idaho Water Resources Graduate Program have been a great source of support and camaraderie. I particularly appreciate the hard work and insights of my student co-authors on the collaborative chapter of this dissertation; these include Meghan Foard, Courtney Cooper, Paris Edwards, Shana Hirsch, and Micah Russell.

This dissertation relies heavily on publicly available data; individual data sources are acknowledged in each chapter but I would be remiss not to acknowledge the decades of hard work by many individuals that has made this research possible. Particularly important data sources include the SNOTEL network, hydroclimatological data at the Reynolds Creek Experimental Watershed, and the results of the Integrated Scenarios project.

I have also been fortunate to receive funding support from the National Science Foundation's Integrative Graduate Education and Research Traineeship (IGERT) program, the Department of the Interior's Northwest Climate Science Center, and the University of Idaho College of Natural Resources Dissertation Finishing Fellowship. These sources of support have provided me with the freedom to determine my own direction of research and pushed me to develop interdisciplinary products and maximize the relevance of this work for applied outcomes.

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Statement of Contribution

The Introduction (Chapter 1) and Conclusion (Chapter 6) are sole-authored. In Chapters 2-4, the co-authors listed primarily acted in advisory roles. Chapter 5 is a co-authored, collaborative project. With the exception of Dr. Timothy Link, the other authors contributing to Chapter 5 are fellow students in the National Science Foundation (NSF) Integrative Graduate Education and Research Traineeship (IGERT) program that funded much of this dissertation. For Chapter 5, the development of the conceptual framework, conducting research, and manuscript writing were all conducted as a team. I took a leadership role within the team, and am listed as first author for that reason.

Chapter 1: Introduction

Snowpack is critically important for water resources globally and in the western United States (U.S.). The most robust feature of global climate change - warming temperatures - is projected to reduce snowfall (Barnett et al., 2005) and advance runoff timing (Stewart et al., 2004). Indeed, anthropogenic warming has already yielded a reduction in snowpack in the western U.S. (Barnett et al., 2008; Hamlet et al., 2005; Knowles et al., 2006; Mote, 2006, 2018; Pierce et al., 2008) and advance in snowmelt and runoff timing (Stewart et al., 2005; Barnett et al., 2008). Despite uncertainty in the effect size and direction of changes in precipitation in the western U.S., these changes are temperature-sensitive and well-established. Changes to snowpack will affect ecological processes, including altered soil temperature and moisture that impact forest greening (e.g., Harpold et al., 2015; Maurer and Bowling 2014; Trujillo et al., 2012), and the timing and magnitude of downstream water resources provisioning (e.g., Dettinger et al., 2015). The potential economic losses from reduced snowpack under anthropogenic warming over the next century are valued in the trillions of dollars (Sturm et al., 2017).

Despite the rich body of literature assessing climate change impacts on snow, there are several elements of changes in snowpack heterogeneity in both space and time at multiple scales that have not been as widely examined. However, these changes in heterogeneity may also have critical consequences for snow-dependent social and ecological systems. Each of the first three chapters in this dissertation analyzes a different type of changing heterogeneity of snowpack, while the fourth chapter is an interdisciplinary synthesis of climate change research in a mountainous headwaters region.

At hillslope scales, wind redistribution of snow is one of the most important processes determining snowpack heterogeneity (Clark et al., 2011). In many regions, including in the sagebrush-steppe ecosystems common in the western United States, blowing snow can create hydrologic heterogeneity that acts as a critical moisture subsidy for species not otherwise found on the landscape, such as aspen (*Populus tremuloides*; Soderquist et al., 2018). Particularly in regions where precipitation is transitioning from snow to rain-dominated regimes, the warming temperatures associated with climate change may reduce wind redistribution of snow, effectively reducing hydrologic heterogeneity on the landscape and affecting the moisture subsidy to sensitive species. In the first chapter of this dissertation, I conduct climate change experiments, using a physically-based hydrologic model and thirty years of hydrometeorologic data in a small watershed where hydrologic heterogeneity is dominated by wind-driven redistribution of snow. Findings from these modeling experiments quantify the temperature sensitivity of wind redistribution of snow at this site and ecohydrologic effects of changes in wind redistribution, as well as the amount of precipitation change that would be needed to mitigate these changes. The long time series available also allows for projection of potential changes in interannual variability under climate change scenarios, and identification of how these changes in interannual variability may vary across the watershed.

The second chapter provides a deeper investigation into interannual variability of snowpack, mapping changes across the landscape at much larger scales. While previous studies have assessed changing variability of temperature and precipitation in portions of the western United States (e.g., Berg and Hall, 2005; Rupp et al., 2016; Swain et al., 2018), relatively few studies have assessed projected changes in snowpack variability (though Lute et al., 2015 is an exception). In this chapter, I compare the interannual variability of peak snow water equivalent (SWE) amount and timing in historical and future cases, using a daily, gridded snowpack product distributed across the western U.S. and forced with historical and future downscaled climate projections. I also assess changes in the frequency of consecutive years with very early or low peak SWE, expecting that these consecutive years of snow drought will have particularly important effects on water resources management or ecosystem function. Understanding changes in variability, and how they vary across the landscape, is a critical step for holistically assessing climate change impacts and potential adaptation needs.

Another important component of snowpack heterogeneity is potential changes in snowfall intensity. Increases in precipitation intensity are well established in both observations and projections (Giorgi et al., 2011; Min et al., 2011; Seneviratne et al., 2012), but changes in snowfall intensity have received much less study. Existing analyses of changing snowfall intensity suggest that large snowfall events will likely decrease less than average events (Danco et al., 2016; O'Gorman, 2014). One previous modeling-based study identified an important potential effect of changing precipitation intensity: increasing precipitation intensity may decrease winter snow ablation and increase peak SWE due to differences in snowpack energy balance when snow falls in a few relatively large, versus many small storms (Kumar et al., 2012). This chapter builds on this modeling-based work by empirically testing the association between snowfall intensity and winter ablation in a spatially distributed observational network of snowpack, using a statistical modeling framework. I also use future projections of snowfall and snowpack to assess the impacts of projected changes in snowfall intensity on winter ablation in future climates.

Finally, each of these changes in snowpack heterogeneity will have important consequences for snow-dependent social and ecological systems. Understanding the potential impacts of both changing snowpack and changing snow heterogeneity requires integrated synthesis of the science that has been conducted to assess climate change impacts. In a fourth, interdisciplinary and collaborative chapter, I work with a team of researchers to characterize the spatial and topical distributions of climate change research in the headwaters of the Columbia River Basin. We expect that the science that is conducted inevitably affects the set of climate change adaptation and mitigation actions that are possible and appropriate. Mapping the networks of existing climate change research will allow us to characterize what science has been done and what more may be needed in order to holistically understand climate change impacts in the region.

Each of the first three chapters will address a different element of changing snowpack heterogeneity, while the fourth places earlier chapters in a broader socio-ecological context. Together, these studies will contribute to a broader scientific understanding of processes in snow hydrology, climate change impacts on snow and snow-dependent systems. This improved understanding of climate change impacts is an important step in the ultimate identification of potential management needs and responses in a changing climate.

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Chapter 2: Warming alters hydrologic heterogeneity: simulated climate sensitivity of hydrology-based microrefugia in the snow-to-rain transition zone

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Published in *Water Resources Research*

Abstract

In complex terrain, drifting snow contributes to ecohydrologic landscape heterogeneity and ecological refugia. In this study, we assessed the climate sensitivity of hydrological dynamics in a semi-arid mountainous catchment in the snow-to-rain transition zone. This catchment includes a distinct snow drift-subsidized refugium that comprises a small portion (14.5%) of the watershed but accounts for a disproportionate amount (modeled average 56%) of hydrological flux generation. We conducted climate sensitivity experiments using a physically based hydrologic model to assess responses of a suite of hydrologic metrics across the watershed. Experiments with an imposed 3.5 °C warming showed reductions in average maximum snow water equivalent of 58-68% and deep percolation by 72%. While relative decreases were similar across the watershed, much greater absolute decreases in snowpack occurred in the drift-subsidized site than the surrounding landscape. In drift-subsidized locations, warming caused a shift from a regime that included both energy and water-limited evapotranspiration conditions to exclusively water-limited conditions. Warming also resulted in altered interannual variability of hydrologic metrics. The drift-subsidized unit was more sensitive to warming than the surrounding landscape, with reduced potential for the effects of warming to be offset by increased precipitation. Despite spatially homogeneous changes in climate, the effects of climate change on the hydrological dynamics in this watershed were spatially heterogeneous in this watershed due to the presence of lateral water transport in the form of drifting snow. These findings suggest an increase in hydrologic homogeneity across the landscape and relatively large changes in snow drift-subsidized refugia.

Introduction

In complex mountainous terrain, a myriad of processes contribute to spatially heterogeneous snowpack development and ablation; this subsequently affects spatiotemporal dynamics of evaporation, transpiration, and runoff generation. These processes include elevation-dependent precipitation and temperature, variable incoming shortwave radiation and other energy fluxes, and heterogeneous vegetation (Anderson et al., 2014; Clark et al., 2011; Tennant et al., 2017). On relatively small scales (i.e., less than 0.5 km^2), the interaction of wind with topography and vegetation can have a major influence on the spatiotemporal distribution of snow (Hiemstra et al., 2002; Winstral & Marks, 2002), producing drifts that persist late into the spring and control the magnitude and timing of streamflow (Chauvin et al., 2011; Hartman et al., 1999; Marks et al., 2002). In some catchments, these drifts affect vegetation distribution, providing a moisture subsidy that supports species not otherwise found in the catchment. In the semi-arid ecosystems of the western United States, quaking aspen (Populus tremuloides) are often found in these drift sites, and are important sources of biodiversity on the landscape (Chong et al., 2001; Johns, 1993; Soderquist et al., 2018).

The drifts that support the establishment and persistence of aspen can be considered a type of hydrology-based refugium, hereafter called drift-subsidized refugia. Refugia are habitats that are somewhat decoupled from larger-scale climate and may support biodiversity conservation in the context of anthropogenic climate change (e.g., Keppel et al., 2012; Taberlet & Cheddadi, 2002). McLaughlin et al. (2017) argued that hydrology can be an important determinant of refugia and identified three types of hydrologic microrefugia: stable refugia, which are insensitive to warming, relative refugia, which are about as sensitive to warming as the surrounding landscape, and transient refugia, which are more sensitive to warming than the surrounding landscape. Identifying the extent to which snow drifts act as stable, relative, or transient hydrology-based refugia will improve our understanding of climate change impacts in these semi-arid systems.

One potential threat to drift-subsidized refugia is a precipitation phase shift from rain to snow. Indeed, such a transition is a well-established effect of climate change in the western United States (Klos et al., 2014; Knowles et al., 2006; Nayak et al., 2010). When precipitation falls as rain rather than snow, it is not redistributed by wind, potentially leading to increased spatial homogeneity of moisture inputs to the landscape (Rasouli et al., 2015). This can reduce moisture subsidies to zones that have historically acted as topographic snow sinks, which in turn affects streamflow timing and magnitude

(Luce et al., 1998; Stephenson & Freeze, 1974). Changes in snowpack dynamics would likewise cascade into changes in seasonal soil water availability, with subsequent ecological impacts (Harpold & Molotch, 2015; Trujillo et al., 2012). The effects of a snow-to-rain transition may be expressed as increasing hydrologic homogeneity across the landscape, which we evaluate via a spatial analysis of the relative hydrologic sensitivity to warming and precipitation changes, changes in energy and water limitation, and changes in interannual variability.

First, the responses of hydrologic metrics and water balance components to warming and precipitation, individually and in concert, may vary spatially across the landscape. It is well established that warming results in decreased spring snow water equivalent (Mote et al., 2005; Pierce et al., 2008) and earlier streamflow timing (Stewart et al., 2004; 2005), and that the sensitivity of these outcomes to warming varies across the landscape. Projected changes in precipitation in the western U.S. are generally uncertain in both direction and magnitude (Dettinger et al., 2015; Rupp et al., 2016), and the potential for increased precipitation to mitigate the effects of warming on snow hydrology is greater in colder, high latitude environments than in warmer, lower latitude regions due to the common snow-to-rain transition in warmer regions (Rasouli et al., 2015).

The potential for increased landscape hydrological homogeneity with warming can also be expressed within the framework of energy and water limitations. Vegetation in water-limited environments is constrained by the need to minimize water stress (Caylor et al., 2009) with snow accumulation exerting a strong control on vegetation greenness in these environments (Trujillo et al., 2012). As a result, vegetation in semi-arid regions is thought to be highly sensitive to environmental change and display non-linear responses to environmental variability (Jenerette et al., 2012; Newman et al., 2006). In contrast, vegetation in energy-limited environments is more sensitive to variability in energy availability than water availability (Donohue et al., 2009), and may use water less efficiently than vegetation in water-limited environments (Troch et al., 2009). Controls on aspen productivity differ between energy- and water-limited environments; in energy-limited conditions, aspen photosynthesis is controlled by the duration of leaf-on conditions, but when water becomes limiting, both photosynthesis and respiration are reduced (Barr et al., 2007). Water and energy limitation are often assessed within a Budyko framework, in which the partitioning of available water into evapotranspiration and streamflow is controlled by the relative availability and seasonal synchronicity of water and energy (Budyko, 1974; Milly, 1994). Warming temperatures may shift a system towards water-limitation rather than energy-limitation, and an important question is whether, and to what degree, precipitation increases could mitigate this shift in regions of heterogeneous water availability.

Climatic constraints on vegetation productivity and impacts on hydrology are often examined using climatological averages; yet, interannual variability in energy- and water-limitations are likely important in hydrology and vegetation dynamics, including the establishment and persistence of refugia. Broadly, temporal environmental variability is thought to affect ecological processes through impacts on reproductive strategies, bet-hedging, and optimal phenotypes (Koons et al., 2008; Childs et al., 2010; Tuljapurkar et al., 2011). Observations of streamflow variability in the late 20th century suggest that in the Pacific Northwest, variance has increased because dry years have become drier (Luce & Holden, 2009), and in the western United States, the period since 1980 has had highly variable streamflow (Jain et al., 2005; Pagano & Garen, 2005). Likewise, cumulative winter snowfall water equivalent is projected to decrease and become relatively more variable from year-to-year across the western U.S. during the 21st century (Lute et al., 2015). The impacts of altered climate on hydrologic metrics, including characterizations of catchments as energy- or water-limited, may vary both inter-annually and between landscape units. Thus, understanding both the sensitivity of mean states as well as interannual variability in hydrologic metrics across the landscape is needed.

In order to assess the climate sensitivity of a snow drift-subsidized hydrology-based microrefugia along each of these dimensions, this study assesses the sensitivity of hydrologic metrics to changes in temperature and precipitation across a small catchment where the hydrology is strongly affected by wind-driven snow redistribution. Specifically, we address the following research questions: (1) How much do altered temperature and precipitation, alone and in concert, affect hydrological dynamics in three distinct landscape units, and therefore the hydrologic heterogeneity across the watershed? (2) Where and when do shifts from energy to water limitation occur with warming? and (3) How does interannual variability of hydrologic metrics change with warming in each of these locations? Understanding the climatic sensitivity of hydrologic metrics at meaningful spatial units across a watershed is critical for determining the persistence of hydrologic microrefugia, and coupled hydrological and ecological processes such as streamflow timing, evaporative dynamics, and vegetation growth and mortality. This work leverages a dataset of exceptionally long temporal extent and high resolution and a well-validated physically-based model to characterize these heterogeneities and advance a deeper understanding of climate impacts on semi-arid hydrological systems in complex terrain. We build on previous studies that have addressed the impacts of climate change on hydrology in watersheds with significant drifting snow to advance the understanding of changing variability in both space and time. Importantly, we also assess the potential for changing landscape hydrological heterogeneity despite spatially homogeneous changes in climate due to lateral snow transport processes.

Methods

Study site

These research questions required a study site in the snow-to-rain transition zone with wind-driven redistribution of snow and an observational record of adequate spatial and temporal resolution and extent to support a detailed climate sensitivity assessment. Upper Sheep Creek (USC) is a small (0.26 km2), semi-arid headwater catchment spanning a 200-m elevation range within the Reynolds Creek Experimental Watershed and Critical Zone Observatory in southwestern Idaho, USA, with an exceptionally long legacy of research (Chauvin et al., 2011; Flerchinger & Cooley, 2000; Flerchinger et al., 1996, 1998, 2016; Flerchinger & Seyfried, 2014; Luce et al., 1998, 1999; Stephenson & Freeze, 1974) (Figure 1). Between 1984-2013, thirty years of hourly meteorological data, including air temperature, precipitation, relative humidity, shortwave radiation, and wind speed, have been collected at multiple sites in USC, as described in previous work including Chauvin et al. (2011) and Flerchinger et al. (2016). In addition, annual snow surveys have been conducted in the catchment to characterize the distribution of snow and provide detailed data for model development and validation.

In previous studies (Chauvin et al., 2011; Flerchinger & Cooley, 2000; Flerchinger et al., 1998, 2016) and in the current study, USC was divided into three hydrologic response units (HRU) based on similarity in vegetation, soils, and snow accumulation. The "aspen" HRU is dominated by drift-subsidized aspen and willow (Salix spp.), occurs on the upper portion of northeast-facing slopes, and has deep, loamy soils (Flerchinger et al., 1998). The "big sage" (Artemisia tridentata) HRU occurs in the lower portions of northeast-facing slopes, with loess silt loam and low rock content. The "low sage" (Artemisia arbuscula) HRU comprises the southwest-facing slopes, with rocky, shallow soils having high clay content. The aspen and big sage HRU share a meteorological station just below 1900 m. At this location, mean annual temperature between 1985 and 2013 was 6.4 °C, and average annual precipitation was 570 mm, with approximately 60% occurring as snow. At the low sage HRU (meteorological station at 1870 m), mean annual temperature was 6.8 °C and mean annual precipitation over the same 30-year period was 426 mm. Streamflow leaving the catchment is intermittent, and averaged 48 mm annually over the same period.

For each HRU, data from snow surveys have been used to determine drift factors, which are multipliers that quantify the average amount of snow scour or drift subsidy for each HRU (Flerchinger & Cooley, 2000). Drift factor calculations accounted for ablation prior to snow surveys through the use of meltwater collectors and adjustments based on modeled ablation. These drift factors averaged 2.29 for aspen, 0.93 for big sage, and 0.98 for low sage, and ranged from 0.68 to 1.22 in the sagebrush HRUs and 1.80 to 3.20 in the aspen HRU between 1984 and 1994. Because no statistical relationship was found between the drift factors and perturbed climate variables (i.e. temperature and precipitation; see supplementary material Figure S4), average drift factors were used in this study. These were applied to each HRU on an hourly timestep as follows:

$$P_{eff} = d \times P_s + P_l$$

where d is a time-invariant drift factor and P_{eff} is drift-adjusted effective precipitation. P_s is solid precipitation and P_l is liquid precipitation. These are calculated using an empirical seasonally-varying precipitation-phase equation based on air temperature (Dai, 2008). This is a relatively simple approach to phase partitioning, but it is based on an extensive dataset of terrestrial measurements and is commonly used by the hydrologic modeling community (Harpold et al., 2017). For more detail on drift factor calculations, readers are referred to Flerchinger and Cooley (2000).

It is possible that differences in meteorological conditions and snowpack and/or vegetation characteristics could result in non-stationary values for the drift factors. The assumption of constant drift factors is one simplification in this study, the goal of which is to generally understand how this and similar systems are likely to be affected by a transition from a snow to rain-dominated regime.

SHAW Model

The Simultaneous Heat and Water (SHAW) model was used to test the sensitivity of hydrologic fluxes to altered climate conditions. SHAW is a one-dimensional model that simulates energy and water fluxes in multi-species canopies with multiple nodes for soil, vegetation, and snowpack, as described in Flerchinger et al., (1998) and Flerchinger et al., (2012). Using the drift factors described above, SHAW was applied to USC previously to determine spatiotemporal water balance variations

(Chauvin et al., 2011; Flerchinger & Cooley, 2000; Flerchinger et al., 1996; 1998), compare eddy covariance and model-based estimates of evapotranspiration (Flerchinger & Seyfried, 2014), and assess prescribed fire impacts on hydrology (Flerchinger et al., 2016).

Using the same model implementation as used in the present study, model validation was performed by Chauvin et al. (2011) with 24 years of data. When SHAW was validated against streamflow measurements, R^2 was equal to 0.85. Modeled evapotranspiration (ET) was within 10% of ET measured using a Bowen ratio system for most of the observed 24 to 73 day periods, but error was as high as 30% (74 mm measured versus 52 mm simulated) for one 27-day period. Modeled change in annual soil moisture storage was within 2-17 mm of measured values. These values suggested reasonable confidence in model performance. In addition to the previous validation, the physical basis of the SHAW model and the long period of record on which it was validated support its ability to model diverse climatic conditions. In this study, forcing data for SHAW were hourly meteorological data observed at the locations denoted in Figure 1. SHAW was run for water years 1984-2013.

Climate perturbations

Downscaled global climate models were used to guide the development of specific climate scenarios. We used data from 20 global climate models (GCMs) that were downscaled using a multivariate adaptive constructed analogue (MACA) method with the Livneh et al (2013) dataset to downscale models to 1/16th degree (~6 km) (Abatzoglou & Brown, 2012; Rupp et al., 2013). Using these data, we investigated scenarios that involved seasonally dependent changes in temperature and precipitation (Vano et al., 2015), and scenarios that altered warming rates contingent on precipitation occurrence (Rupp & Li, 2016). However, preliminary tests suggested that these additional scenarios resulted in relatively minor differences in this watershed (see supplementary material; Cleveland, 1979), and we ultimately built scenarios with constant temperature and precipitation changes, similar to the approach in Rasouli et al. (2015).

Climate sensitivity experiments for the watershed were designed by examining projected changes simulated across a range of GCMs for end of the 21st century (2070-2099) under forcing prescribed by representative concentration pathway (RCP) 8.5. A 3.5 °C warming for mean annual temperature above 1971-2000 conditions was projected for the watershed in the 20-model mean. We also considered experiments with incremental warming of 0.9 °C, 1.7 °C, 2.6 °C, and 4.3 °C to represent conditions for more proximal time periods and different emissions scenarios and climate sensitivities. Using the same scenario construction, multi-model mean change in annual precipitation by the end of the century was +7.7%, with a 20-model standard deviation of 6.3%. Despite these relatively small projected increases in precipitation, we adjusted precipitation in 20% increments to a minimum of -20% and maximum of +60% in order to ascertain the extent to which fairly extreme hypothetical changes in precipitation could mitigate the effects of warming in different HRUs. Scenarios were named for their temperature and precipitation change (e.g., T+3.5 P+20 would be a 3.5 °C temperature increase and 20% precipitation increase). Other meteorological variables were not perturbed.

In order to assess the sensitivity of our results to the drift factors used, we conducted additional experiments with altered drift factors. From the historical observational record, we selected the years with the 10^{th} and 90^{th} percentile empirical drift factor values for the aspen HRU, using drift factors from Flerchinger and Cooley (2000), updated with snow survey data from water years 1995-2012. While the mean aspen drift factor was 2.29, the 10^{th} percentile drift factor was 1.80, and the 90^{th} percentile drift factor was 3.20. In the low sage HRU, the corresponding values were 1.08 and 0.97; in the mountain big sage HRU, the corresponding values were 0.97 and 0.98. Using the empirical drift factors for these water years for all HRUs, we repeated model runs with the same set of altered temperature and precipitation scenarios as described above. We term these experiments "d10" and "d90" for the cases with the 10^{th} and 90^{th} percentile drift factor, respectively.

Analysis of results

For each climate scenario and HRU, daily values and annual summary statistics were extracted. Summary statistics were selected to assess the magnitude and timing of snowpack accumulation and ablation, and magnitude of water balance components. These included effective precipitation after drift redistribution (P_{eff}), snowfall liquid water equivalent as a fraction of precipitation (SFE/P), maximum snow water equivalent (SWE_{max}), date of SWE_{max} (DOMS), total ET, potential streamflow (Q_{pot}), and center of timing of potential streamflow (CT). Q_{pot} was defined as the sum of deep percolation and overland flow assuming that all water that percolates below the rooting zone will eventually become streamflow. CT was calculated as the center of timing of Q_{pot}. As SHAW is a one-dimensional model, Q_{pot} and CT should not be interpreted as a precise representation of streamflow, but Chauvin et al. (2011) found that SHAW explained 85% of variance in streamflow volume in USC, and suggested that the rest is lost to deep drainage. While this correlation is imperfect, it is suitable for the goals of this study, which are to advance an understanding of spatial and temporal sensitivity of hydrological fluxes in this system to changing hydroclimate, rather than to precisely predict streamflow.

For each scenario, changes in 30-year means of each variable were calculated. The nonparametric Kolmogorov-Smirnov (KS) test was used to assess equivalence between scenarios. Probability density functions of the annual distributions of hydrologic metrics were assessed visually to determine how interannual variability of each metric would change in the different HRUs. The interquartile range (IQR) of each metric over each 30-year simulation was calculated to determine how much interannual variability changed in response to climate change experiments.

Finally, the trajectory of each water year was plotted in Budyko space to determine the prevalence of transition from energy- to water-limited fluxes, and to identify the conditions under which this transition occurs. Reference crop potential evapotranspiration (PET) was calculated using a Penman-Monteith implementation, assuming no changes in vegetation or water use efficiency with increased atmospheric carbon dioxide concentration (Allen et al., 1998, 2005; Guo et al., 2016). The Penman-Monteith method of estimating PET, like other offline methods of PET estimation, is known to overestimate climate-driven changes in PET relative to global climate models due to the assumptions of constant surface roughness and stomatal conductance in common implementations (e.g., Milly and Dunne, 2016; 2017). Because the Penman-Monteith method is widely used and other PET estimation methods also overestimate changes in PET, we used the Penman-Monteith method but discuss the potential effects of this bias. Change in storage (S) was subtracted from precipitation (P) before calculating aridity and evaporative indices in order to accurately account for the water available to meet evaporative demands, as in Chen et al., (2013). We defined energy-limitation as water years where PET/(P-S) was less than one, and water-limitation as cases where PET/(P-S) was greater than one.

Results

Sensitivity to warming only

Warming in all HRUs reduced SWE_{max} and Q_{pot}, while ET increased and CT occurred earlier in the year (Figure 2a; S4 and S5; Table 1). Each of the three HRUs responded differently to warming perturbations. P_{eff} in the aspen HRU decreased by 21% in the T+3.5 scenario, but did not change significantly in the other HRUs (Table 1). Changes to P_{eff} in scenarios with warming are due to the drift factors applied to each HRU during snowfall events. SFE/P decreased by a similar percent across all three HRUs, with a slightly smaller percentage decrease in the aspen than the sage HRUs, due to the slightly colder temperatures at the aspen site. SWE_{max} decreased by a similar percentage across the HRUs (58-68%), but this was associated with a much greater absolute difference in the aspen HRU than in the two sage HRUs.

One important function of drift subsidies in the aspen HRU is the development of a snowpack that persists longer into the spring and summer than snow in the surrounding area, subsidizing soil moisture and producing streamflow later into the dry summer season and hence producing the hydrologic refugium. DOMS advanced by a similar amount in the aspen and big sage HRU. In the low sage HRU, DOMS advanced by a smaller amount and was not statistically significant relative to the T+0 scenario. This is likely because the low sage historically had an intermittent snowpack where DOMS was highly affected by individual storms; DOMS therefore had low interannual consistency, and was not as predictably affected by warming as in the other two HRUs.

Warming resulted in a slight ET increase in all three HRUs, with a similar magnitude of increase across all three units, which was associated with a smaller percentage increase in the aspen HRU. This increase occurred due to an increase in spring ET, paired with a decrease in summer ET; the springtime increase was large enough to compensate for the summer decline. The warming-induced increase in ET in the aspen HRU despite a decrease in P_{eff} suggests some evidence of energy limitation in this drift-subsidized unit. Q_{pot} decreased by a similar percentage across all three HRUs, which was associated with a much greater absolute decrease in the aspen HRU than in the two sage HRUs.

Combined effects of warming and altered precipitation

Altered precipitation may mitigate or exacerbate the spatially variable effects of warming on hydrology and the transience of refugia. Figure 3 depicts the hydrologic effects of combined changes in precipitation and temperature in the three HRUs. Relatively vertical contour lines indicate that a variable is primarily temperature-sensitive, while relatively horizontal lines indicate that a variable is primarily precipitation-sensitive. P_{eff} in the two sage HRUs was only minimally affected by warming; in the aspen HRU, the 21% decrease in P_{eff} associated with the T+3.5 scenario was negated by a 20-40% increase in precipitation. With altered precipitation alone, P_{eff} scaled with changes in precipitation.

Snowpack dynamics were relatively insensitive to precipitation changes in comparison with the effects of warming, as shown by the relatively steep isolines for SFE/P, SWE_{max}, and DOMS. None of the precipitation scenarios tested were large enough to completely offset the effects of a 2.6 °C or greater warming on SWE_{max} in all three HRUs. Even with relatively minor warming of 1.7 °C, a fairly extreme 40-60% increase in precipitation was needed to offset the effects of warming on SWE_{max}. The effects of warming on DOMS were particularly insensitive to concurrent increases in precipitation.

Increases in precipitation were better able to mitigate the effects of warming on Q_{pot} than on snowpack dynamics. However, the effects varied markedly across the three HRUs, with a +20% increase in P compensating for losses in Q_{pot} with +3.5 °C warming in the low sage HRU, whereas a +40% increase in P was needed to compensate Q_{pot} declines in the aspen HRU. The difference between sites is likely because the aspen HRU lost much of the drift subsidy when precipitation fell as rain rather than snow, while the sage HRUs slightly gained moisture due to the lack of drifting snow. The precipitation changes needed to compensate for the warming effects on snowpack dynamics and Q_{pot} are considerably larger than average precipitation changes projected by GCMs for this region.

Watershed-integrated changes

The hydrologic effects of warming and potential for precipitation to mediate these effects were integrated over the watershed to account for the disproportionate contribution of the different HRUs to catchment-scale fluxes. In the T+3.5 scenario, the suite of hydrologic changes resulted in an overall 70% decrease in watershed-integrated Q_{pot} and a 15-day advance in CT (Figure 4). By comparison, in the T+3.5 P+20 scenario there was a 15% decrease in Q_{pot} and a 14-day advance in CT. CT was more sensitive to decreases in precipitation than increases; for example, in the most extreme warming scenario (T+4.3), a 20% increase in precipitation delayed CT by two days, while a 20% decrease yielded an additional six-day advance.

Both warming and increased precipitation altered the relative contributions of each HRU to watershed Q_{pot} (Figure 5; change in absolute contributions in Figure S7). Despite the relatively small size of the aspen HRU (14.5% of the watershed), it historically contributed 56% of the watershed Q_{pot}, averaged across 30 water years. In the T+3.5 scenario, the aspen contribution to the watershed-integrated Q_{pot} decreased to 43%. However, the share of Q_{pot} contributed by each HRU varied enormously between water years, as did the effects of warming on relative Q_{pot} contribution between HRUs. The aspen HRU was a relatively large contributor to Q_{pot} in the coolest, wettest terciles of years, though interannual variability in relative contributions to Q_{pot} was very large (Figure 5a, 5b). In most of the preceding analyses, warming and increased precipitation had opposing effects. However, increased precipitation, like warming, resulted in a decrease in the average relative contribution of the aspen HRU to watershed Q_{pot} (Figure 5d). While Q_{pot} from the aspen HRU increased with increasing precipitation, area-weighted Q_{pot} from the other two HRUs increased more.

In the most extreme warming scenarios for the warmest and driest terciles of years, the decreasing trend in aspen contribution to Q_{pot} was reversed. In these years, the aspen HRU contributed a slightly larger percentage of Q_{pot} in the T+4.3 scenario than in the T+3.5 scenario. This is likely due to the fact that in the aspen HRU, the average absolute Q_{pot} in warm/dry years did not change much between the T+3.5 and T+4.3 scenarios, while in the low sage HRU, Q_{pot} continued to decrease.

Sensitivity to altered drift factors

Altered drift factors had the greatest effect in the aspen HRU, which had the greatest empirical variability in scaling factors. The sensitivity of each hydrologic metric to altered drift factors in the aspen HRU is depicted in Figure 6; the same analysis is provided for the other HRUs in the supplementary material. In the aspen HRU, P_{eff} was strongly affected by the drift factors. Indeed, the difference between P_{eff} modeled with the d10 and d90 scenarios in the historic case (482 mm) was greater than the difference in P_{eff} between the historic and warmest cases (266 mm). Change in P_{eff} with warming was also greatest in the d90 experiment. SWE/P ratio was also greater in the d90 experiment, but the rate of change in SWE/P with warming was roughly the same regardless of drift factor used. The same was true of DOMS. SWE_{max} was similar to P_{eff}; it was greatest and had the greatest change in the d90 experiment, although in the case of SWE_{max}, the difference between drift factor experiments was not greater than the maximum effect of warming.

Change in ET between drift factor experiments varied with both precipitation and temperature changes. With no precipitation change or warming, ET was lowest in the d90 scenario, but increased most rapidly with warming. The same was true in scenarios with increased precipitation. In the P-20 scenarios, ET was largest in the d90 case for all temperature scenarios, and had a non-linear response to warming. Q_{pot} was similar to SWE_{max}, with largest values and largest decreases in the d90 scenario. CT advanced with warming for all drift scenarios, and was latest in the d90 experiment. Drift scenario affected CT most strongly in the lowest precipitation scenarios, but was relatively unimportant with increased precipitation. This is likely due to the high frequency of water years with near-zero Q_{pot} in the P-20 cases. Very low values of Q_{pot} result in very early CT values, but small increases in Q_{pot}, such as those gained from a larger drift factor, generate much later CT values. This change is essentially the result of numerical inconsistencies of calculating CT with very low Q values, rather than a physical process.

A potential hypothesis for the impacts of climatic changes on snow drift redistribution is that warmer conditions create denser, more cohesive snow that experiences less redistribution. While we did not explicitly evaluate this effect, the gray dashed lines in Figure 6 indicate the trajectory of change from a historic case with the base drift factor to a T+4.3 case with the d10 drift factor. For most variables, the trajectories of change with altered drift factors are very similar to the trajectories of change with constant drift factors, though hydrologic changes are generally exacerbated by the reduced drift factor.

Energy and water limitation

Figure 7 depicts trajectories in Budyko space from historic to projected future cases, where displacement from left to right suggests a shift towards water-limitation, while upward displacement indicates that a greater fraction of available water was evapotranspired in the simulation, with less available for streamflow. Many of the effects of warming and precipitation change on hydrology discussed above suggest that the drift-subsidized aspen refugium may have been historically energy-limited but will likely become water-limited with warming. In Budyko space, this is indicated by the wetter years, drawn as arrows that begin in the energy-limited region but extend into the water-limited region with warming and altered precipitation scenarios (Figure 7a). In contrast, the two sage HRUs were historically water-limited, and continued to be water-limited with warming (Figure 7b, green and yellow arrows).

In this historical case, 21 of 30 water years in the drift-subsidized aspen HRU were water-limited, whereas all 30 years under the T+3.5 scenario were water-limited. This shift was somewhat mitigated by increases in precipitation; in the T+3.5 P+20 scenario, only 27 out of 30 water years were water-limited. In the big and low sage HRU, all water years were historically water-limited and remained water-limited in warmer scenarios with moderate precipitation increases or decreases (Figure 7b; data for individual years not shown). Our findings regarding changing water balances suggested that precipitation increases were better able to mitigate the effects of warming in the two sage HRUs than in the aspen HRU; this also appears to be true in the case of energy and water limitation. In the warming scenarios, the trajectory of change in the aspen HRU was in a similar direction whether precipitation increased or decreased by 20%, though the precipitation changes moderated the effect size. In contrast, in the two sage HRUs, the direction of change in precipitation governed the direction of movement through Budyko space.

The Penman-Monteith method of estimating changes in PET likely overestimates changes in PET in climate change scenarios, and the bias may be approximately as large as the estimated change (Milly and Dunne, 2017). For that reason, these findings may be an overestimate of shifts from energy to water limitation. If we assume that the Penman-Monteith calculation overestimates change in PET by a factor of two to account for the bias calculated by Milly and Dunne (2017), all water years still become water-limited in the T+3.5 scenario. In the more moderate T+1.7 scenario, there is a slight difference; the smaller change in PET results in only 26, rather than 27 water years that are water-limited. These minimal changes in estimates of water-limitation are likely due in part to the decrease in effective precipitation with warming in the aspen HRU. However, even with relatively large increases in precipitation, our findings are only minimally sensitive to errors in PET: in the T+3.5 P+60 scenario, 17 years are water-limited with the original Penman-Monteith calculation, while only 15 are water-limited if we assume the Penman-Monteith calculation doubles estimates of changing PET. The general finding that warming results in a shift from interannual variations in energy- and water-limitation to consistent water limitation in the aspen HRU is robust regardless of potential errors in PET estimation.

Changes in interannual variability

Even though temperatures were increased by the same amount in each water year, different hydrologic variables in each HRU showed distinctly different changes in interannual variability (Figure 8; S7 for changing precipitation scenarios). Moreover, for many variables, the change in interannual variability was more dramatic than the mean change due to warming temperatures. The interannual IQR of P_{eff} in aspen decreased from 457 mm to 347 mm in the T+3.5 scenario (Figure 8). This relatively large decrease is likely due to the loss of years with large drift redistributions. In contrast, changes in IQR in P_{eff} in the big and low sage HRU were relatively small. Each of the snowpack metrics assessed varied differently between HRUs as well. SFE/P did not show major changes in interannual variability in any of the HRUs. In contrast, SWE_{max} in the aspen HRU historically had the largest amount of variability, and the IQR decreased with warming in all HRUs. DOMS variability increased as temperatures warmed in the aspen and big sage HRU. In aspen, historic IQR of DOMS was 23 days, but it increased to 40 days in the T+3.5 scenario, suggesting a shift towards a much more intermittent and sporadic snowpack with reduced interannual consistency in the drift-subsidized refugium. The low sage HRU initially had large DOMS variability and maintained this large range in warmer scenarios.

The shifts in distributions of Q_{pot} did not reflect the shifts in distributions in P_{eff} but followed similar patterns as changes in SWE_{max} variability; with T+3.5, the IQR of Q_{pot} values decreased dramatically for all three HRUs, from 391 to 137 mm in aspen, 71 to 22 mm in big sage, and 67 to 10 mm in low sage. When these changes were integrated over the entire catchment, the IQR of Q_{pot} decreased from 76 to 30 mm (-61%), suggesting increasing interannual homogeneity of streamflow volumes under warmer temperatures due to the lack of years with higher flows. Changes in the interannual variability of CT varied spatially; CT in the aspen HRU became more variable, while CT in the low sage HRU became less variable with warming.

The changes in ET IQR were different from the other variables heretofore discussed. In aspen, the initial distribution of ET was tightly clumped and was greater than in the other HRUs. While mean ET values changed slightly in warming scenarios, the IQR of ET in aspen increased from 46 mm to 128 mm in the T+3.5 case. The big and low sage HRUs had similar increases in ET IQR to the aspen, with an increase from 67 to 102 mm in the big sage and 65 to 132 mm in the low sage.

Discussion

Climate change sensitivity of drift-subsidized hydrologic refugia

The sensitivity of drift-subsidized and scoured landscape units to warming and altered precipitation suggests that the drift-subsidized hydrology-based refugium is sensitive to warming, and that increased hydrologic homogeneity across the landscape can be expected with warming. Evaluating the sensitivity of the drift-subsidized aspen HRU in the context of the stable, relative, and transient refugia described by McLaughlin et al. (2017) suggests that it predominantly acts as a transient refugium. The larger decrease in SWE_{max} and Q_{pot} and the exaggerated change in P_{eff} in the aspen HRU support its status as a transient refugium, with a greater sensitivity to warming than the surrounding landscape.

Our results also suggested that the drift-subsidized refugium generally became a less important contributor to Q_{POT} as temperatures warmed, even as overall watershed-integrated Q_{pot} decreased. The potential for drift-subsidized sites to act as important contributors to streamflow that persists later into the summer season has long been recognized (Hartman et al., 1999; Luce et al., 1998; Marks et al., 2002), but the extent to which this role will be maintained in warmer scenarios is less well established. Berghuijs et al. (2014) found that a precipitation shift from snow towards rain reduces streamflow, but did not directly establish a mechanistic explanation. This study supports the mechanistic explanation that, at least in watersheds with large amounts of drifting snow, watershed homogenization due to reduced drifting may be a contributor to decreases in Q_{pot}. To the extent that the aspen HRU's status as a refugium is associated with its ability to generate streamflow, this change adds to the evidence that it acts as a transient refugium, with greater sensitivity to warming than the surrounding landscape.

The potential for increased precipitation to offset the effects of warming on snowpack varied between HRUs. One previous study found that in a cool, mid-latitude catchment, the effects of 1 °C of warming on SWE_{max} could be compensated for by a 20% increase in precipitation, while the same precipitation increase could compensate for 3 °C of warming in a colder, higher elevation catchment (Rasouli et al., 2014, 2015). Our findings were quantitatively similar to those of Rasouli et al. (2015), with the additional finding that the potential for precipitation increases to offset the effects of warming varied on smaller scales in the absence of substantial differences in climatic conditions. A greater increase in precipitation was needed in the drift-subsidized aspen refugium to offset the effects of warming on Q_{POT} than in the sage sites; this is because warming yielded a significant decrease in P_{eff} in the drift-subsidized site, but not in the scoured sage sites. While Rasouli et al. (2015) found differences in the amount of precipitation needed to offset the effects of warming across large climatic gradients, we found that in the presence of drifting snow, this also varied across small climatic gradients and spatial scales. This finding suggests further evidence for increasing hydrologic homogeneity across the landscape and that the aspen HRU acts as a transient refugium, exhibiting more sensitivity to warming with lower potential for that impact to be mitigated by increased precipitation, than the sage HRUs.

The shift from energy to water limitation in the drift zone added to evidence that the aspen HRU acts as a transient refugium. While the two sage HRUs were water-limited in both historic and warmer scenarios, the shift to consistent water limitation in the aspen HRU suggested a fundamental change in water availability in the drift-subsidized refugium. Moderate to severe warming scenarios with no precipitation change or small precipitation increases may result in conditions where aspen is much more consistently water-limited than it was historically. In light of the differences in plant adaptation strategies (Caylor et al., 2009) and sensitivity to rapid environmental change (Newman et al., 2006) in water-limited systems, the aspen HRU may hence be more susceptible to decreased productivity and mortality. Stephenson (1990) found that deciduous forests rely on relatively high supplies of both energy and water that are seasonally synchronized with each other; in this case, the historically present late-melting snow drift delayed the effective water supply to aspen, generating conditions appropriate for deciduous forests that may not otherwise exist. The shift towards earlier, reduced snowmelt decreases water supply and synchrony between energy and water, making this site less suitable for deciduous vegetation. Moreover, the shift from energy- to water-limitation in the aspen HRU suggests another type of increasing hydrologic homogeneity and loss of refugia over the landscape: with projected warming, the entire watershed becomes water-limited, in contrast to the historical case, where a portion of the landscape was energy-limited, if only intermittently.

While our results generally suggest increasing spatial homogeneity with warming, one surprising finding was that changes in interannual variability depended strongly on the hydrologic metric and landscape unit of interest. These changes occurred despite the fact that the temporal variability of climate inputs was not explicitly altered. Globally, interannual temperature variability is not projected to change consistently with warming (Huntingford et al., 2013), but precipitation variability over daily to decadal timescales is projected to increase by 4-5% per degree Celsius of warming over land (Pendergrass et al., 2017). We found that interannual variability of hydrologic metrics was altered even without taking altered climate variability into account. For example, the decrease in SWE_{max} IQR suggested landscape homogenization with respect to SWE_{max} over both space and time. In contrast, we found an increase in ET variability in the drift-subsidized aspen site, which could be accounted for by two mechanisms: first, in water years in which the aspen HRU was historically energy-limited, increasing temperatures could increase ET. Second, in water years that were historically water-limited, ET might decrease due to lower P_{eff} in the warmer scenarios. In this context, in which snow drift subsidies provide water availability for a species and isolated drift-based ecosystem not otherwise found on the landscape, increasing interannual variability of water used for ET suggests more uncertainty for species that depend on these isolated, yet ecologically important systems.

The variability of DOMS also increased in the aspen and big sage HRUs until it approximately matched the variability of DOMS in the low sage HRU, with a more intermittent snowpack. This, as well as the decrease in annual difference in DOMS between aspen and low sage, suggest the homogenization of deep drainage timing across the landscape, though effects on streamflow will depend on the landscape distribution and timing of infiltration and subsurface flow. These findings highlight the importance of considering the changing variability of water year conditions that are likely in future climate scenarios, rather than relying on changes in means alone.

Assumptions and Limitations

There are several assumptions and limitations that should be considered in the interpretation of these results. Likely the most important is the use of constant, empirically-derived drift factors. In reality, drift factors may vary intra- and inter-annually, depending on climate conditions, snow age, precipitation amounts, vegetation conditions, and wind speed and direction. For example, smaller drift factors and therefore smaller effective precipitation differences between HRUs would generally be expected for warmer snowfall events; therefore, the results presented here may overestimate the spatial heterogeneity of SWE across the watershed under warmer conditions. However, despite attempts to establish a relationship between the drift factors and temperature and precipitation, no statistically significant relationship was found (Figure S4). Drift factors may also be sensitive to variations in wind speed and direction, which we did not directly test.

Sensitivity tests of different scaling factors suggested that for most response variables, the drift factors used did not affect the direction of change due to warming and general findings of the study since the major changes in the spatial heterogeneity of effective are driven by a shift towards higher rain:snow ratios. There were however, some variables (P_{eff}, Q_{pot}, and SWE_{max}) for which the effects of warming were greater with larger drift factors. The magnitude of the drift factors had a large effect on ET, and the nature of this effect changed between precipitation scenarios. These changes are generally in agreement with our findings that ET in the drift-subsidized refugium is on a threshold between energy- and water-limited conditions. For example, in high precipitation scenarios, the d90 experiment had the lowest ET, which may be due to energy limitation exacerbated by a longer snow-covered duration. In the lowest precipitation scenario, the d90 experiment had the largest ET and smallest reduction in ET with warming, which is in accordance with increased water limitation in warm, dry scenarios. These findings suggest that caution is warranted in the interpretation of changes in ET given uncertainties in drift factors.

These experiments with altered drift factors do not address the potential for changing drift factors in warmer conditions. As a greater fraction of snowfall occurs at warmer temperatures and deposited snow ages, denser, more cohesive snow is less likely to be redistributed (Li and Pomeroy, 1997). Given that the drift factors in this study were not reduced under warmer conditions or for older snow, the results may be skewed towards greater landscape heterogeneity under the warmer scenarios. In a nearby watershed, Rasouli et al. (2015) used a physically-based snow drift model to assess snowpack sensitivity to warming. As discussed above, their results were both quantitatively and qualitatively similar to ours, which lends confidence to the empirical approach used here. A few empirical studies have evaluated interannual variability of snow drifting. Flerchinger and Cooley (2000) found increased landscape variability of snowpack in wetter years in Upper Sheep Creek. In contrast, in the much colder Green Lakes Valley, Colorado, Erickson et al. (2005) found that terrain (and potentially snow drifts) introduced snow redistribution most strongly during dry winters.

Although the trapping of snow by shrubs was noted to be an important process in Upper Sheep Creek, topography was found to exert the primary control on snow distribution patterns (Prasad et al., 2001). It is important to note that this could change in the event of climate-induced changes in vegetation structure; for example, a shift from sagebrush to juniper could result in reduced redistribution (Kormos et al., 2017). Similarly, some of the ecohydrological changes identified in this study may create unsuitable conditions for existing vegetation, particularly aspen; the potential feedbacks associated with these changes are not assessed here but could be important areas of further study. Prasad et al. (2001) also found that empirical drift factors resulted in acceptable simulations with relatively small errors, even with relatively large changes in precipitation. In this study, snow patterns are similarly controlled primarily by topography, and changes in precipitation redistribution are controlled by phase changes with warming. Further study of additional physical mechanisms for snow drift sensitivity to climate change is needed to more fully understand hydrologic sensitivity to climate changes in complex terrain.

Conclusions

We assessed the hydrologic sensitivity of three different HRUs to perturbed temperature and precipitation in a small, mid-latitude catchment with hydrology largely influenced by wind-driven redistribution of snow. Despite minimal spatial variability of climatic conditions and no changes in the spatial variability of climate, three landscape units displayed very different ecohydrologic responses to climate drivers due to the presence of drifting snow. Simulations suggested that the drift-subsidized hydrologic refugium was generally more sensitive to climatic changes than the other landscape units, which suggests that it likely acts as a transient refugium. The specific changes identified in this study include increased spatial homogeneity in hydrological dynamics, shifts from energy- to water-limitation in the drift-subsidized refugium, and altered interannual variability of hydrologic metrics. These changes will vary across diverse climate and landcover types, though similar processes are likely present in other regions with drifting snow and a snow-to-rain transition. Synthesizing landscape-level changes in spatial and temporal variability and the effects of reduced snow redistribution on flow regimes are important directions for future research. The importance and distinctive response of the drift-subsidized refugium revealed by this work also emphasizes the importance of accurate simulation of processes that produce spatiotemporal variations in snowcover over larger regions to more comprehensively understand how climatic changes will be manifested in complex montane regions.

Acknowledgements, Samples, and Data

The authors have no conflicts of interest. We wish to acknowledge the decades of data collection at Reynolds Creek Experimental Watershed that made the present study possible. The authors also wish to express their gratitude to Dr. Adam Winstral who helped to inspire this research as a result of various discussions over the years. The comments of three anonymous reviewers and two editors, including Dr. Charles Luce, greatly improved this manuscript. This work was supported by the Department of the Interior Northwest Climate Science Center (NWCSC) through a Cooperative Agreement G14AP00153 from the United States Geological Survey (USGS) and by the National Science Foundation's IGERT Program (Award #1249400). Data products are available at https://doi.org/10.21429/C9PD2Z. The R software (3.3.2) was used for data analysis (R Core Team, 2017).

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Figures

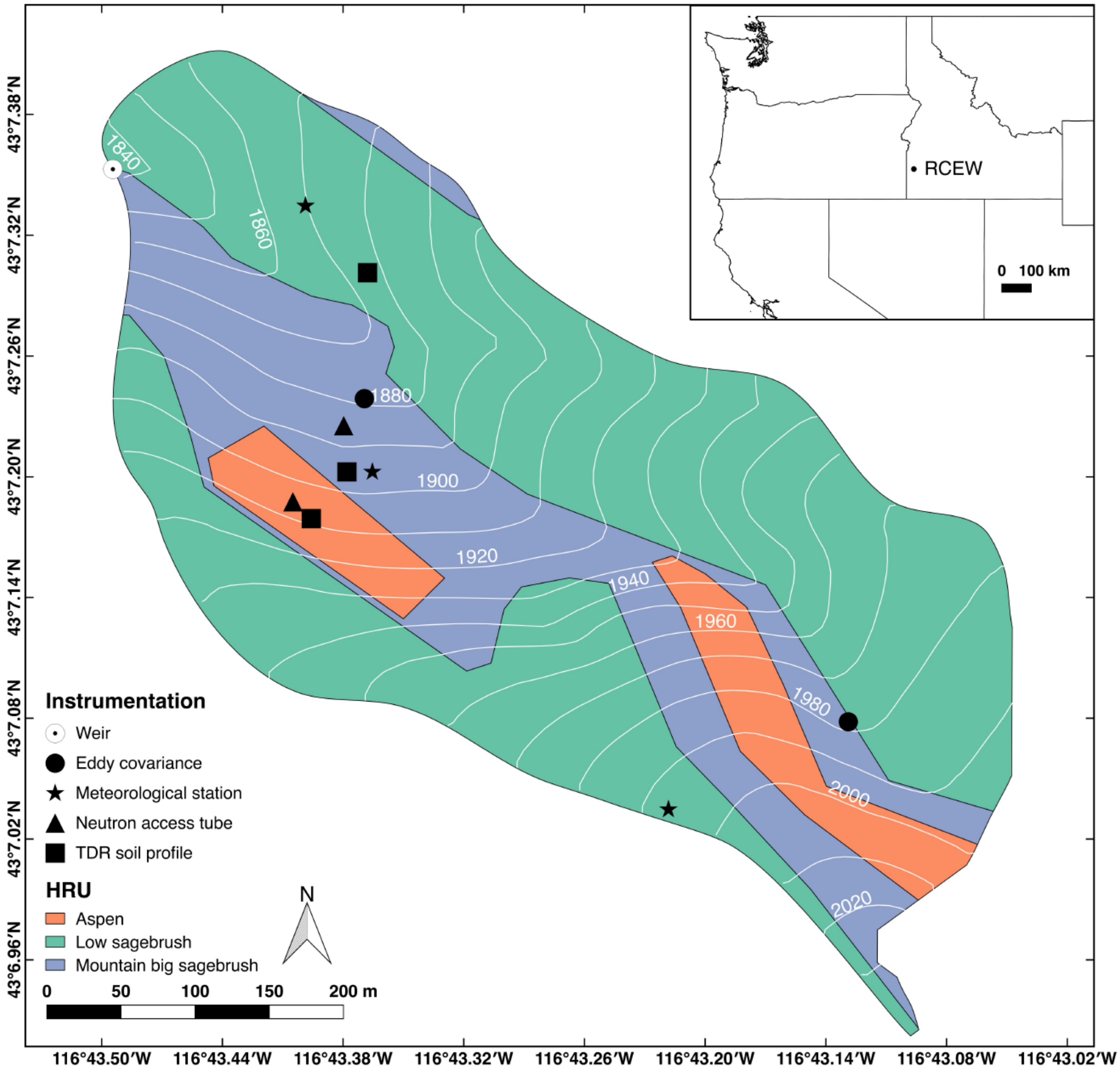

Figure 2.1 Upper Sheep Creek vegetation, topography, and instrumentation. Inset shows context, with Reynolds Creek Experimental Watershed (RCEW) identified.

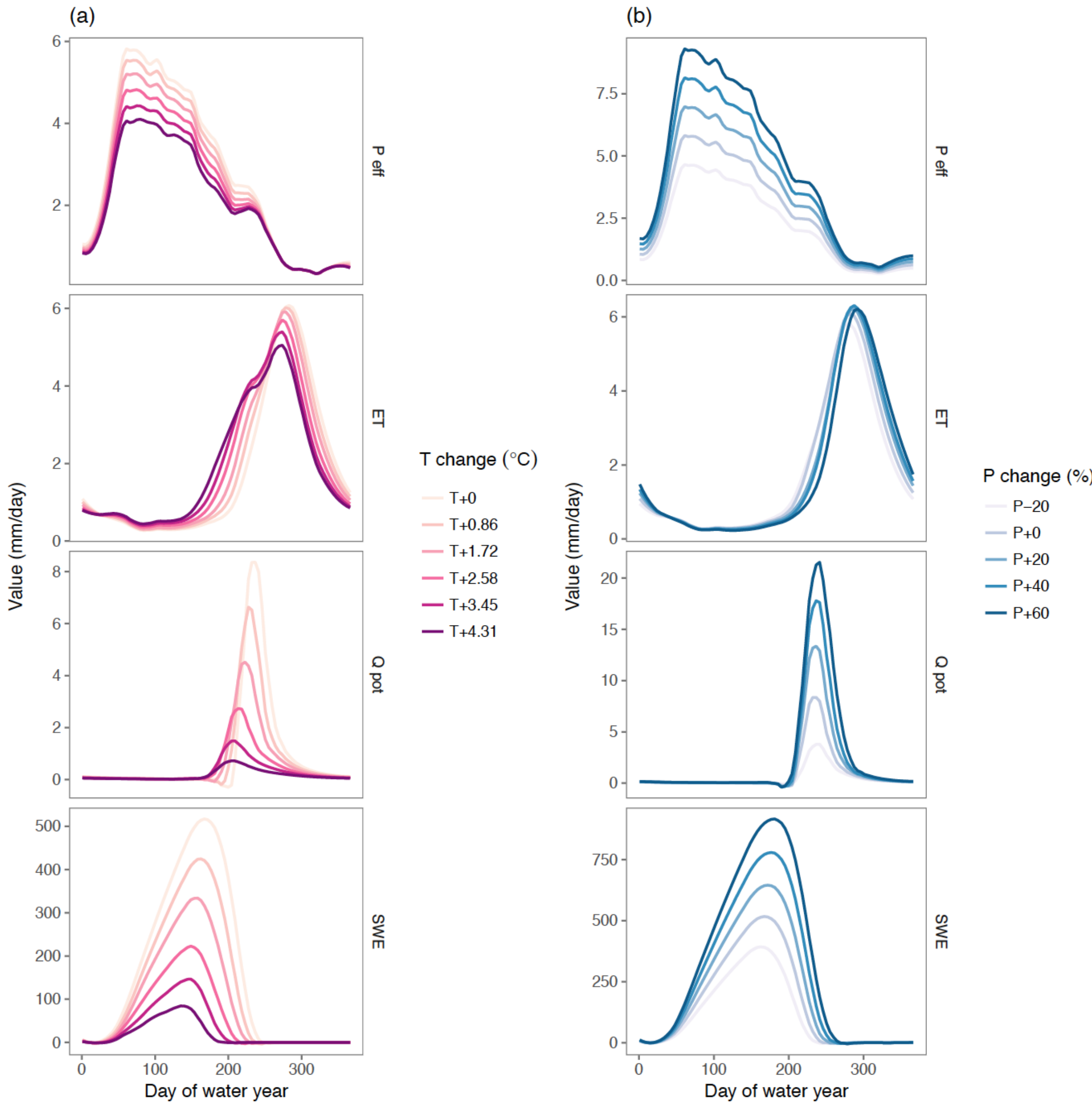

Figure 2.2 Changes in P_{eff}, ET, Q_{pot}, and SWE in the drift-subsidized aspen site under (a) warming and (b) precipitation change. Curves are loess-smoothed 30-year mean values. Note the varying axis extents.

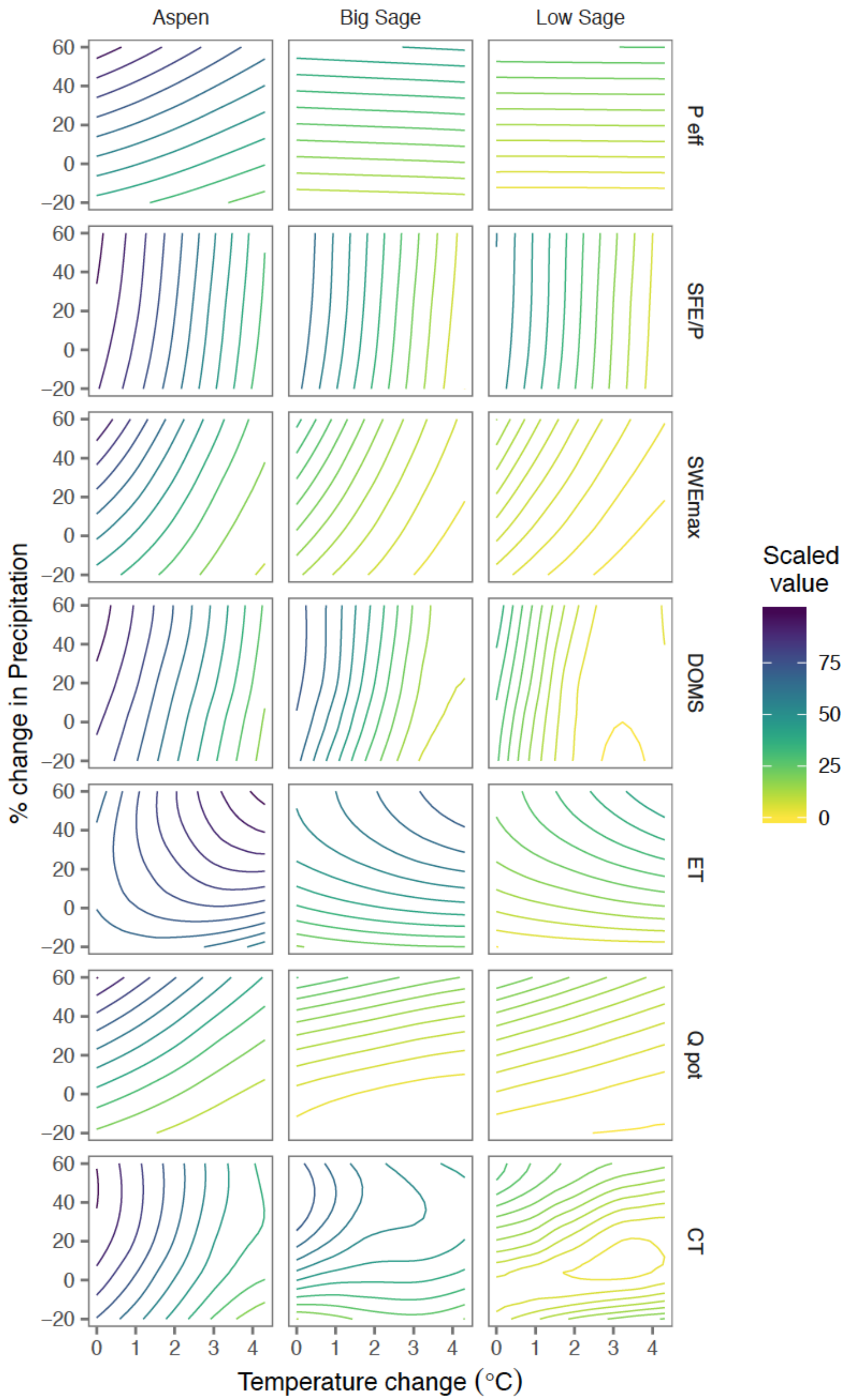

Figure 2.3 Sensitivity of hydrologic variables to temporally constant and seasonally variable changes in temperature and precipitation. The values of each variable have been normalized to range from zero to 100.

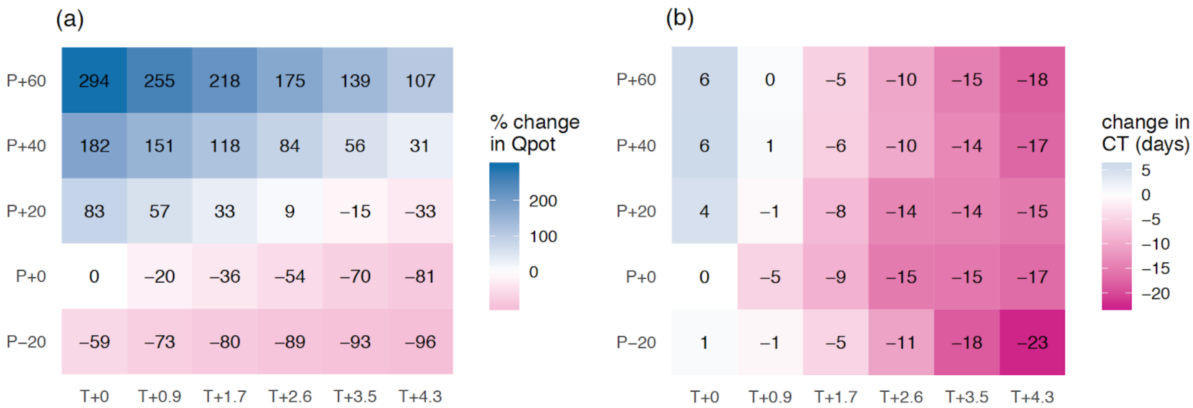

Figure 2.4 Watershed-integrated changes in (a) Q_{pot} and (b) CT with temperature and precipitation change scenarios.

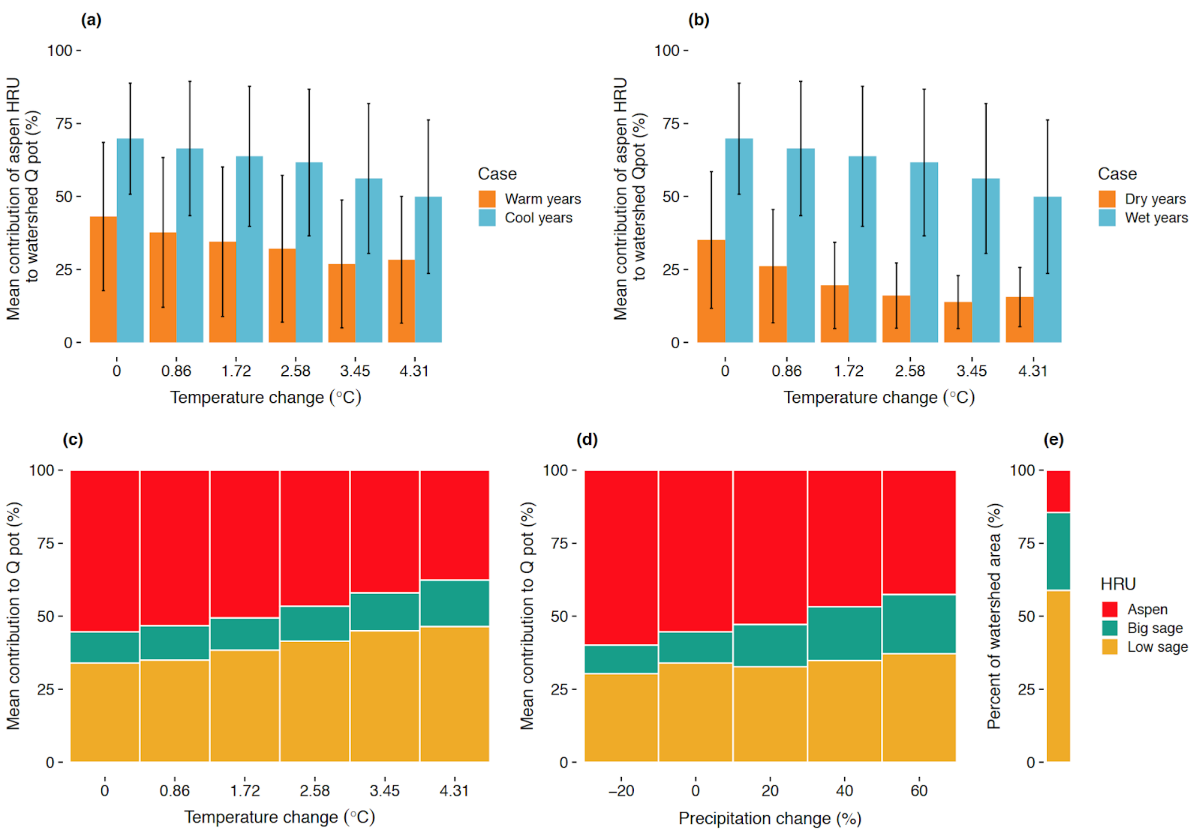

Figure 2.5 (a, b) Contribution of aspen HRU to watershed Q_{POT} in (a) warmest and coolest and (b) wettest and driest tercile of years. Error bars represent 10-year standard deviation. (c) 30-year average contribution of each HRU to warming with increasing temperature, and (d) same as (c) with altered precipitation instead of temperature. (e) displays the percent of watershed area occupied by each HRU.

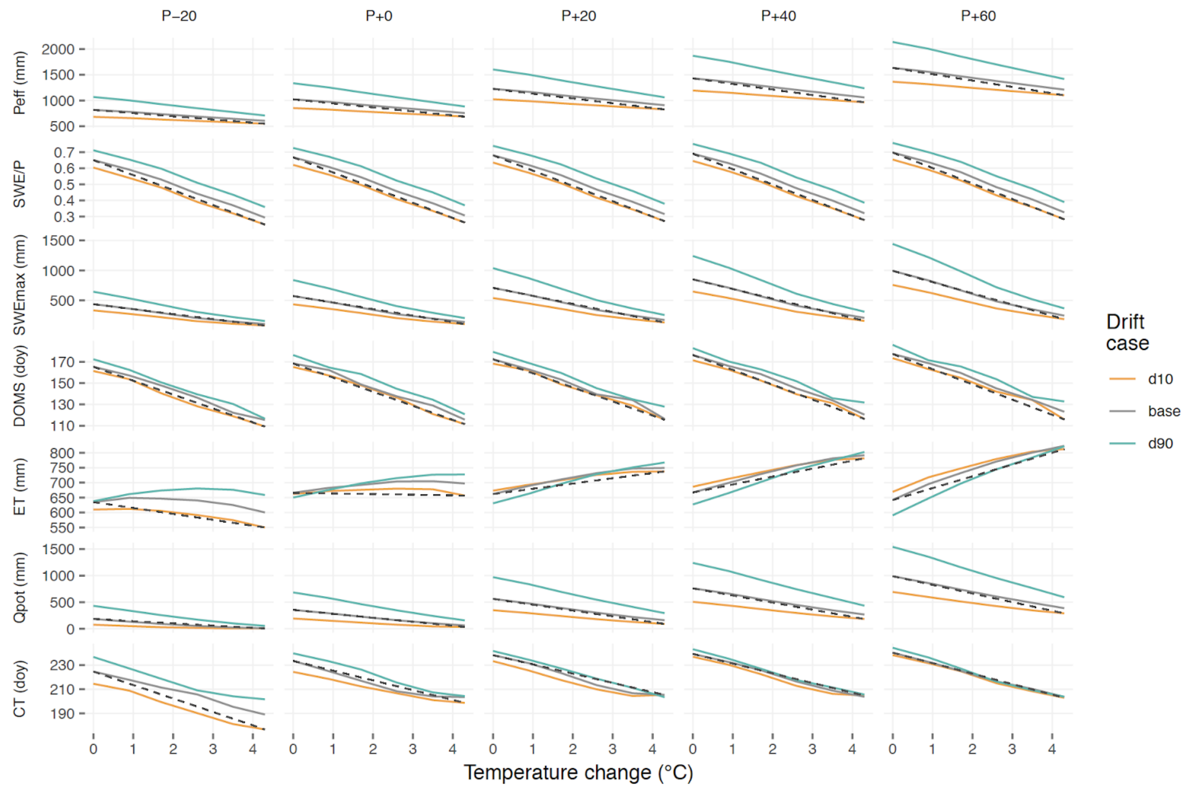

Figure 2.6 Change in each hydrologic variable with altered temperature (x-axis) and precipitation (columns) in three drift factor scenarios. "Base" is the mean drift factor. "d10" and "d90" represent scenarios with the 10th and 90th percentile values of aspen drift factors. Dashed lines indicate the change that would occur if drift factors shifted from the base case to the d10 case in warmer scenarios.

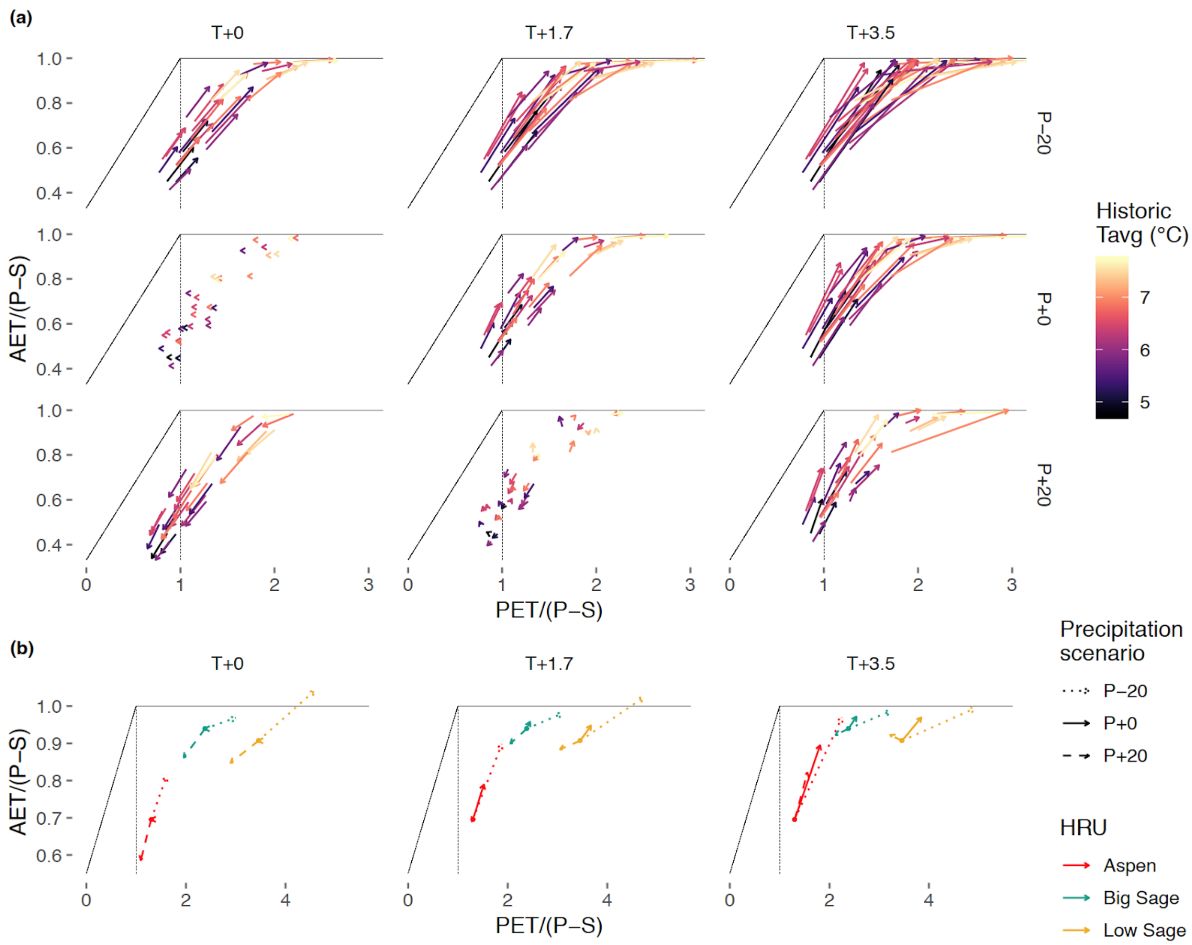

Figure 2.7 (a) Individual water year trajectories, where the base of each arrow represents the historic value in Budyko space and the head represents the value in climate-perturbed scenarios in the drift-subsidized aspen HRU. Individual water years are colored by their historic annual average temperature. (b) 30-year mean trajectory through Budyko space for each of the three HRUs with three precipitation scenarios.

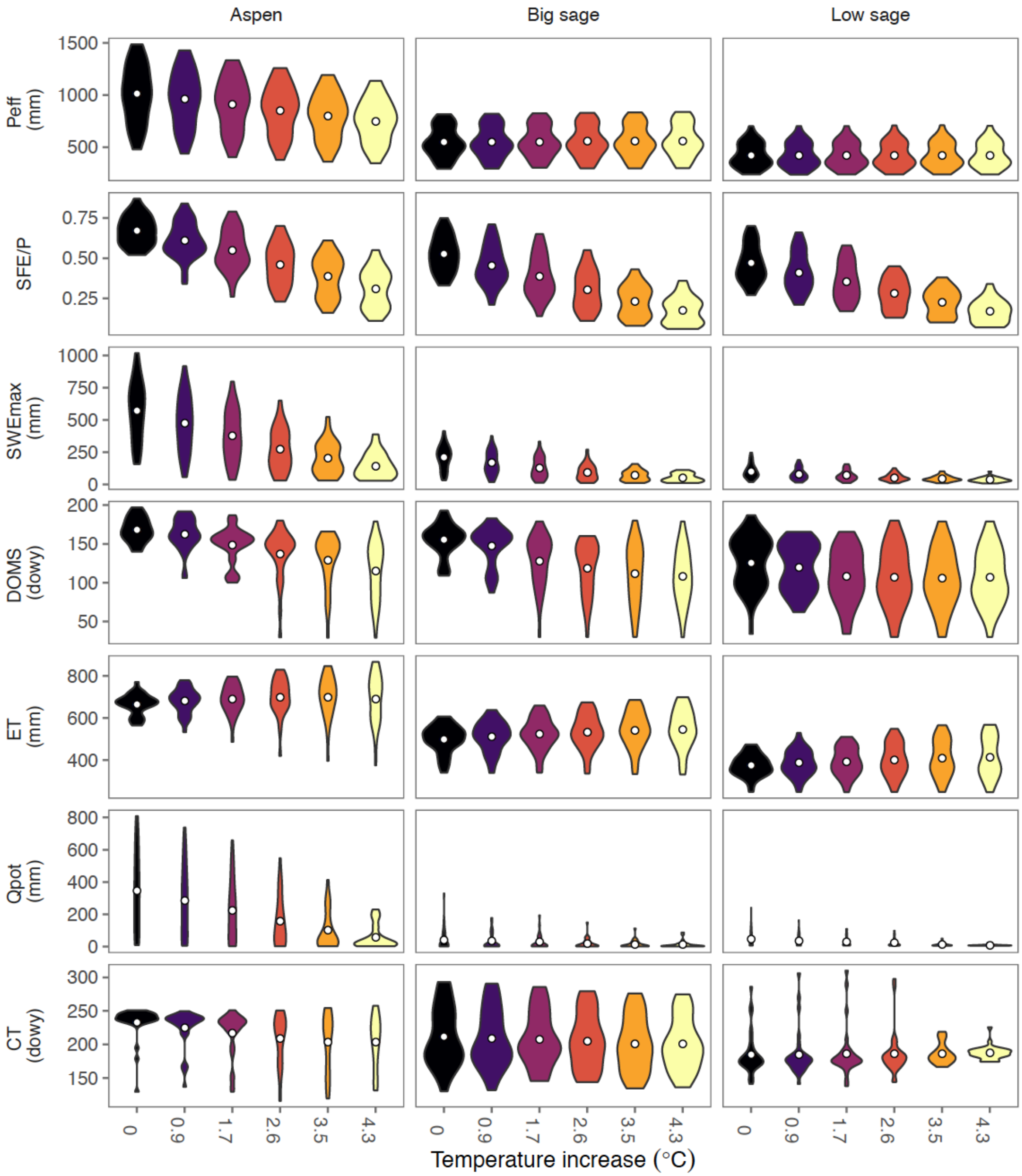

Figure 2.8 Interannual variability of annual summary variables for each HRU under temperature increases. Each shape represents a density function of values; points within shapes represent mean values. Color indicates increase in temperature as denoted on the x-axis.

Tables

Table 2.1 Changes in 30-year mean values of annual hydrologic metrics with T+3.5 scenario in different HRUs. Differences are the warmer scenario minus the historic scenario (e.g., a negative number means decrease with warming). Percent changes are not reported for variables that are measured on an interval scale. Significance in changes based on two-sided K-S test: * = p < 0.05, ** = p < 0.01, *** = p < 0.001.

Variable	HRU	Historic value	Value with T+3.5	Difference	Percent change (%)
P_{eff} (mm)	Aspen	1022	806	-216**	-21
	Big sage	554	566	12	2.1
	Low sage	428	429	1	0.3
SFE/P	Aspen	0.67	0.38	-0.29***	-43
	Big sage	0.52	0.23	-0.29***	-56
	Low sage	0.47	0.22	-0.24***	-52
SWE_{max} (mm)	Aspen	573	197	-376***	-66
	Big sage	212	67	-145***	-68
	Low sage	102	43	-59***	-58
DOMS (day)	Aspen	168	129	-40***	N/A
	Big sage	156	111	-44***	N/A
	Low sage	125	105	-19	N/A
ET (mm/year)	Aspen	666	705	39*	5.8
	Big sage	499	542	43	8.6
	Low sage	376	413	37	9.8
Q_{pot} (mm/year)	Aspen	355	104	-252***	-71
	Big sage	48	15	-32	-68
	Low sage	47	13	-33***	-71
CT (day of water year)	Aspen	233	204	-30***	N/A
	Big sage	211	204	-6	N/A
	Low sage	186	187	1*	N/A

Chapter 3: Projected changes in interannual variability of peak snowpack amount and timing in the Western United States

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In review at *Geophysical Research Letters*

Abstract

Interannual variability of mountain snowpack has important consequences for ecological and socioeconomic systems, yet changes in variability have not been widely examined under future climates. Physically-based snowpack simulations for historical (1970-1999) and high-emission scenario (RCP 8.5) mid-21st century (2050-2079) periods were used to assess changes in the variability of annual maximum snow water equivalent (SWE_{max}) and SWE_{max} timing across the western United States. Models show robust declines in the interannual variability of SWE_{max} in regions projected to transition from snow- to rain-dominated precipitation regimes. The average frequency of recurrent snow drought years (SWE_{max} <historical 25th percentile) is projected to increase from 6.6% to 42.2% of years. Models also project increases in the variability of SWE_{max} timing, suggesting reduced reliability of when SWE_{max} occurs. Differences in physiography and regional climate create distinct spatial patterns of changes in snowpack variability that will require adaptive strategies for water and environmental resource management.

Introduction

Anthropogenic climate change is altering water resources in the western United States, with decreased mountain snowpack storage across the region (Pierce et al., 2008; Mote et al., 2018) and earlier runoff timing in basins that supply water to humans and ecosystems (Barnett et al., 2008). The effects of warming on changes in average snowpack conditions are well characterized for historical (Hamlet et al., 2007; Knowles et al., 2006; Mote 2006; Pierce et al., 2008; Siler et al., 2019) and future conditions (Hamlet et al., 2005; Klos et al., 2014; Gergel et al., 2017; Kapnick and Delworth, 2014; Fyfe et al., 2017; Rhoades et al., 2018a), but there is a relative paucity of information on how the interannual variability of snowpack amount and timing could shift as the climate changes.

Changes in interannual variability of snowpack amount and timing would impact ecological, socioeconomic, and coupled social-ecological systems that rely on snow cover and melt, although these impacts are not as well established as the impact of changes in mean conditions. For example, the magnitude of interannual variability affects the reliability and predictability of reservoir inflows (Rhoades et al., 2018b), hydroelectric power generation (Fleming et al., 2012), and tourism (Scott et al., 2008). For each of these cases, low interannual variability may be associated with greater reliability, whereas high variability increases the potential for high snowfall and runoff years to offset the negative consequences of drought years. High interannual precipitation variability has been associated with reduced groundwater depletion because occasional high precipitation years can break a positive feedback between groundwater pumping and reservoir depletion (Apurv et al., 2017), though the contribution of variability in snow-related patterns, processes, and fluxes to surface and groundwater withdrawals is not as well established.

One dimension of variability that may be particularly important is the degree of change between consecutive years. Precipitation whiplash — the occurrence of an extremely dry winter immediately followed by or preceding an extremely wet winter — is projected to increase with climate change in California (Swain et al., 2018), though snowfall patterns will not reflect those of overall precipitation due to increasing temperatures. Recent multi-year snow droughts in the western United States, whether caused by unusually warm winters and/or low winter precipitation, have drawn attention to the causes and impacts of chronic snow droughts (Hatchett and McEvoy, 2018; Cooper et al., 2016; Ullrich et al., 2018). While studies of snow droughts have predominantly addressed snowpack amounts, consecutive years with early snow accumulation and melt timing may also be an important control on water resources (Jefferson et al., 2008). Current flood operations are guided by "static" rule curves that require reservoir drawdowns during fall months and neglect antecedent moisture conditions beyond the current season (Willis et al., 2011). The combination of required drawdowns and the potential for multi-year drought is a widespread threat to water availability from managed reservoirs.

Snowpack variability also affects ecological processes on seasonal and interannual timescales. For example, snow is important for wildlife, such as wolverine (Copeland et al., 2010), and vegetation dynamics, such as timing of forest greenness (Trujillo et al., 2012). Earlier snow melt timing advances peak soil moisture timing (Harpold and Molotch, 2015) and flowering plant phenology (Dunne et al., 2003), increases vegetation water stress (Harpold, 2016), and is associated with increased wildfire activity (Westerling, 2016). While the importance of interannual variability and consecutive years with early snowmelt timing has not been formally established in this context, we suggest that it may exacerbate stress on vegetation or affect plant community composition and productivity.

Previous studies have examined projected changes in interannual variability of temperature, precipitation, and snowpack across portions of the western U.S. For example, in the Columbia River Basin, interannual temperature variability is projected to increase during summer and decrease in winter (Rupp et al., 2016). Interannual precipitation variability in the western U.S. is projected to increase, especially toward the end of the 21st century (Berg and Hall, 2015; Swain et al., 2018). Snowfall accumulation variability is projected to decrease in warmer-maritime regions and increase in colder-continental regions (Lute et al., 2015). However, spatially explicit assessments of changes in the interannual variability of snowpack magnitude and timing are limited. These changes in snowpack variability may vary spatially in both direction and magnitude over relatively fine scales.

In this study, we assess projected changes in interannual variability of snowpack magnitude and timing, measured as annual maximum snow water equivalent (SWE_{max}) and date of SWE_{max} (DMS), across the western United States. These variables are selected to characterize the total amount of snow available to contribute to runoff (SWE_{max}) and the timing of the snow accumulation season (DMS). We also assess the frequency of two consecutive years with SWE_{max} below the historical 25th percentile and discuss its importance for water resources management. Finally, we conduct a spatially explicit assessment of the frequency with which DMS occurred in specific months. This study is the first to assess how the magnitude and direction of change in variability are expected to vary spatially and differ between SWE_{max} and DMS across the western United States. These findings provide important information for improving assessments of climate change impacts on water resources for socioeconomic, ecological, and coupled social-ecological systems.

Methods

Daily SWE data for the western United States were obtained from the Variable Infiltration Capacity (VIC) model (Liang et al., 1994) forced with downscaled climate model outputs. Ten global climate models (GCMs) from the Fifth Climate Model Intercomparison Project (CMIP5; Taylor et al., 2011) were selected based on their ability to credibly simulate temperature and precipitation patterns and variability of the northwestern US (Rupp et al., 2013; Abatzoglou and Rupp, 2017). The credibility of models based on these same criteria were likewise comparable for the southwestern U.S. While historical performance may not predict future accuracy, culling of models is often used to guide impact-based modeling (e.g., Vano et al., 2015). Projected changes in seasonal temperature and precipitation patterns for the subset of 10 models were largely consistent with a broader sample of CMIP5 outputs for the region.

Each GCM was run for historical (1950-2005) and high-emissions representative concentration pathway (RCP) 8.5 (2006-2099) climate forcings. Daily GCM outputs were statistically downscaled to 1/16th degree resolution using multivariate adaptive constructed analogues (MACA; Abatzoglou and Brown, 2012) with the training dataset of Livneh et al. (2013) and used to drive VIC. Comparisons of VIC simulations with observed SWE from the SNOTEL network found good agreement, particularly with respect to interannual variability (Gergel et al., 2017), and have been widely used in hydroclimate research (e.g., Li et al., 2017). Study area maps are provided in the supplementary material (Figures S1-S2).

For each GCM, water year, and grid cell, annual SWE_{max} and DMS were calculated from daily SWE for the water year beginning on October 1 for historical (1970-99) and future (2050-79) periods. In cases where multiple days had the same value of SWE_{max}, the last occurrence was recorded as DMS. The data were subset to grid cells where historical mean SWE_{max} was greater than 100 mm; all subsequent calculations were conducted using this spatial domain. The interquartile range (IQR; 75th minus 25th percentile) of SWE_{max} and DMS was calculated over the 30-year periods for the historical and future time periods for each GCM and grid cell. Mean values of SWE_{max} and DMS are presented in the supplementary material (Figures S3-S5).

We also calculated changes in the frequency of snow droughts, which we define as two consecutive years with SWE_{max} less than or DMS earlier than the historical 25th percentile value. We then calculated the frequency of cases where two consecutive years experienced snow drought for each time period. We term these “consecutive low SWE years” and “consecutive early SWE years” for SWE_{max} and DMS, respectively. In cases where three consecutive years had SWE_{max} below the first quartile, two years would be tallied. Sensitivity analyses with four-year durations were conducted, though we urge caution in interpreting GCM results pertaining to lower frequency events (Abatzoglou and Rupp, 2017).

We use a bootstrap approach to test statistical significance of changes in variability. For each GCM and grid cell, we resampled with replacement 100 samples from the historical distribution and calculated variability metrics for each sample. Differences were deemed significant where the 2050-2079 variability fell outside of the historical 5^{th}-95^{th} percentiles. As GCM variability has been cited as a key source of uncertainty for early- to mid-century regional climate projections (e.g., Chen et al., 2011; Hawkins and Sutton, 2011), we consider changes to be robust when significant changes are observed in at least five of ten GCMs. Results are also presented as supplemental material in an interactive data visualization tool at: https://snowvariability.nkn.uidaho.edu/.

Results and Discussion

SWE_{max} variability

The historical SWE_{max} IQR is largest in high elevation, cold regions of the Sierra Nevada and Cascades, and lower across the colder interior mountains. Lower elevations throughout the study area exhibit lower IQR due to lower upper quartile values of SWE_{max} (Figure 1a). Changes in SWE_{max} IQR from historical to future periods show distinct spatial patterns (Figure 1b). In lower elevations of maritime mountains, SWE_{max} IQR decreases due to greater declines in the upper versus bottom quartile of SWE_{max} distributions, suggesting that under the scenarios considered here there will be fewer years with what would historically be considered an above average snowpack. In higher elevations of the Sierra Nevada, Cascades, Northern Rockies, and Idaho Batholith SWE_{max} IQR increases. In the southern Rockies, both quartiles decreased (Figure S6), such that SWE_{max} IQR was relatively unchanged. Across the entire domain, the largest decreases in SWE_{max} IQR occur at sites where historical average winter (November-March) temperatures are greater than 0 °C (Figure S7) and SWE_{max} is highly sensitive to warming. At colder sites (historical winter temperature ~< -3 °C), changes in SWE_{max} IQR are not well explained by temperature. Historical winter precipitation and changes in both winter precipitation and temperature IQR were generally not strongly linked to changes in SWE_{max} IQR (Figures S8-S11).

At least five out of ten GCMs simulate significant decreases in SWE_{max} IQR for 24.9% of grid cells, while 13.0 % of grid cells meet this criteria for increases (Figure S13-14). Sites with significant decreases in SWE_{max} IQR are predominantly in warmer regions that are likely to experience a transition towards more rain-dominated precipitation regimes (Klos et al., 2014).

Historically, the frequency of consecutive low SWE years showed minimal spatial variability with an average of 6.6% of water years identified as part of a two-year or longer snow drought (Figure 1c). In 2050-2079, an average of 42.2% of water years classify as snow droughts. These changes are greatest in maritime regions and across the large area that comprises the lower elevations of the northern Rockies (Figure 1d). Spatial patterns of change in consecutive low SWE years are broadly similar to percentage changes in mean SWE_{max} (Figure S5). The average frequency of four-year snow droughts increased from 0.26% of water years to 25.0% of water years.

To illustrate the spatially complex nature of changes in variability, SWE_{max} IQR is depicted for three grid cells along a transect in the Sierra Nevada for a single GCM (Figure 2; other GCMs in Figures S16-17). Distributions of annual SWE_{max} for the historical and mid-21st century cases reveal distinctly different patterns of change across the transect. At the low- and mid-elevation grid cells, which were historically near the winter 0°C isotherm, zero or near-zero SWE_{max} values become increasingly common in the future, and both the upper and lower quartile values decrease (Figure 2b). At the lowest grid cell, the upper quartile decreases much more than the lower, decreasing the SWE_{max} IQR by over 60%. At the mid-elevation grid cell, the lowest quartile decreases more than the upper, doubling the SWE_{max} IQR. Finally, at the highest grid cell, the upper quartile increases, likely due to increasing winter precipitation (Rupp et al., 2017), and the lower quartile decreases, likely due to warmer years with low SWE_{max}, producing a 130% increase in SWE_{max} IQR.

These findings suggest that SWE_{max} IQR decreases and consecutive low SWE years increase in areas that are near the historical 0°C isotherm, where warming causes a shift from snow to rain (Klos et al., 2014), primarily due to the loss of years with exceptionally deep snowpack. This is in agreement with Lute et al. (2015), who found a maritime-continental gradient of changes in SWE_{max} standard deviation, with increases in colder continental inland ranges and decreases in warmer maritime regions. The continental-scale patterns of historical SWE_{max} IQR and large decreases in SWE_{max} IQR in maritime regions are likely due to the contribution of snowfall intensity and extreme events to interannual variability (Lute and Abatzoglou, 2014). In maritime regions, larger SWE_{max} years depend on a few large events (Guan et al., 2010), which are susceptible to warming and precipitation phase shifts from snow to rain (Lute et al., 2015) and increased winter ablation (Kapnick and Hall, 2012). The warming-induced loss of a few large snowfall events in years that would otherwise have large SWE_{max} values would reduce SWE_{max}, producing a large decline in IQR.

DMS variability

Historically, DMS IQR was largest at lower elevations in the Sierra Nevada and Cascades and lower in the colder Rocky Mountains (Figure 3a). This pattern illustrates that peak snowpack timing was historically most variable in warmer regions with high interannual precipitation variability and relatively intermittent snowpack. Conversely, changes in DMS IQR display complex spatial patterns (Figure 3b) that are not well explained by historical climate or changes in climatic variability (Figures S8-11). DMS IQR generally decreases in the highest elevation, coldest regions and increases in warmer areas, though there are some exceptions to this pattern, such as the foothills of the Oregon Cascades. More areas exhibit significant increases in DMS IQR (29.2% of pixels) than decreases (6.2%; Figure S13).

The complex spatial patterns of change in DMS variability can be better understood through inspection of grid cells along a transect (Figure 2d). In the lowest elevation Sierra Nevada pixel, DMS was historically quite variable, but the earliest quartile advances more than the latest, so DMS IQR increases from 34 to 57 days in the future case. In the mid-elevation pixel, the first and third quartiles change by the same amount, and variability is unchanged. At the highest elevation, the latest quartile of years changes much more than the earliest quartile, and DMS IQR decreases from 31 to 19 days.

The increase in DMS IQR in warmer regions is likely indicative of sites in the snow-to-rain transition zone that historically had relatively low DMS IQR. As temperatures warm and snowpack declines, DMS may become increasingly dependent on the synchrony of precipitation events and sub-freezing temperatures, and thus more variable. For example, in the Sierra Nevada and Cascades, a few large storms deliver a large fraction of the annual snowfall (Lute and Abatzoglou, 2014), which may be of heightened importance in a warmer climate with fewer days conducive to snowfall (e.g., Lute et al., 2015). Late DMS years may be heavily affected if even one large storm produces rain, rather than snow. To the extent that DMS is related to runoff timing, increasing variability suggests increased variability of runoff timing, though these impacts will be mediated by post-DMS ablation rates and runoff generation processes, which may in turn be affected by climate change (e.g., Barnhart et al., 2016; Musselman et al., 2017).

Historically, an average of 6.4% of water years were consecutive early SWE years; this number increases to 56.7% in the mid-21st century case. As with SWE_{max}, the historical frequency of consecutive early DMS years does not show obvious spatial patterns (Figure 3c). Consecutive early SWE years increase across the domain, with a pattern of change that is similar to change in mean DMS (Figure S4), with greatest changes in the northern Rockies and Cascades, but without the maritime-to-continental climate gradient seen in changes in SWE_{max}. For four-year durations, the average frequency of consecutive early DMS years increases from 0.27% to 38.0%.

A frequency analysis of the historical and potential future timing of DMS summarizes DMS variability (Figure 4). We define "reliable DMS" as cases where a grid cell has at least 50% of DMS values in a given month. April was the predominant month in which SWE_{max} occurred historically (29.0% of grid cells had reliable DMS in April). March DMS was more common at lower elevations, and May was relatively common at higher elevations, particularly in the continental interior. In 2050-2079, April is no longer the most common month in which peak SWE occurs, with only 15.5% of grid cells having reliable April DMS. Pixels with reliable DMS in May decrease from 9.9 to 1.2% of grid cells. Instead, peak SWE values in March and February become increasingly common. These findings are broadly consistent with existing literature that shows that DMS has shifted earlier and is projected to continue to do so in future climates (Kapnick and Hall, 2010; Montoya et al., 2014), though here we add more spatially explicit and detailed projections. Moreover, these changes reflect increasing variability of DMS, as the total percentage of pixels that had no months with reliable DMS increased from 37.1 to 51.1%.

Model agreement on significant changes in SWE_{max} and DMS IQR varies regionally. On average, model agreement on changes in SWE_{max} and DMS IQR is lower than model agreement on changes in mean values (Figure S13). Warming is a robust feature of modeled future climates, while changes in precipitation and temperature variability exhibit much greater uncertainty and model disagreement (Rupp et al., 2016; 2017). Snowpack variability may be affected by warming, changes in precipitation magnitude, and spatial and temporal variability of both temperature and precipitation, as well as other contributors to the snow energy balance, such as shortwave radiation (Musselman et al., 2017; Painter et al., 2017) and atmospheric humidity (Harpold and Brooks, 2018). Different snow models may affect results but have previously been identified as a relatively small source of uncertainty (Chen et al., 2011). To the extent that GCMs agree on the direction of change in snowpack variability, we propose that these changes are likely incurred due to warming, but future work should quantitatively assess physical mechanisms for changes in snowpack variability.

Implications and Conclusions

Interannual variability of SWE_{max} in the western United States is projected to change, with large decreases in IQR for regions transitioning from snow- to rain-dominated climates, particularly in maritime regions, and smaller changes in cooler continental climates. In contrast, DMS may become more variable across much of the western U.S. Spatial patterns of the sign and magnitude of these trends are critical for understanding their likely impacts. Further work may be needed to assess the robustness of these results given the multiple sources of uncertainty, including climate forcing due to inter-model, inter-scenario, and internal variability (e.g., Hawkins and Sutton, 2011), downscaling approach and reference observational data (Alder and Hostetler, 2018), and choice of hydrologic model (e.g., Chen et al., 2011) which were beyond the scope of this investigation. Despite these uncertainties, there are several potential implications of these findings.

For water resources operations, regions with increases in interannual variability of runoff volume and timing that have large engineered or natural storage may be more resilient to changes than those with less storage, particularly when storage exceeds the average annual discharge (Langbein, 1959). The impacts of snowpack magnitude and timing variability on water resources also depend on the combined effects of snow and rain on hydrological regimes, particularly as previously snowmelt-dominated systems experience increasing contributions from rainfall (e.g., Knowles et al., 2006; Kormos et al., 2016). Changing snowpack variability will likely be very different from changes in variability of precipitation, which is generally projected to increase (e.g., Konapala et al., 2017; Pendergrass et al., 2017; Swain et al., 2018), and the combinations of these changes will determine changes in water resources dynamics. The increased frequency of consecutive low SWE years will also likely affect water resources operations and require improved early drought detection methods (AghaKouchak et al., 2015) and optimization of reservoir operation rule curves to account for antecedent storage (Anderson et al., 2008; Ralph et al., 2011; Willis et al., 2011).

Recreational activities, such as ski resort operations, that depend on a minimum amount and relatively early snowpack accumulation as well as reliability of snow conditions coinciding with peak visitation periods (e.g., Scott et al., 2008), will likely be affected by altered interannual variability. Our results suggest low snowpack years will be more common, with reduced interannual variability, and that the number of consecutive years of low SWE_{max} will increase for ski resorts at lower elevations where more precipitation will likely fall as rain rather than snow. Similar findings apply for ecosystem functions that are influenced by interannual snowpack variability. For example, relatively high snowpack facilitates subalpine seedling establishment (Andrus et al., 2018). Reduced variability of snowpack magnitude and the loss of high SWE_{max} years could limit seedling establishment and alter successional dynamics. High snowpack years also limit early season fire activity in many mountainous regions of the West (Westerling 2016; Abatzoglou and Kolden, 2013); loss of these years could enable more consistent early onset of fire activity in flammability-limited regions, barring increased spring and early-summer rainfall.

While studies of the importance of average snowpack conditions for water resources and ecosystems abound, the impacts of changing variability on these systems are less well established. Our results suggest that snowpack variability will be substantially altered in the future climates considered here, with robust increases in the frequency of recurrent snow drought and reduced interannual variability of annual SWE_{max}. To the extent that changes in snowpack variability affect water resources and ecosystem function, climate change impact studies and adaptation planning efforts should account for future changes in snowpack variability.

Acknowledgments, Samples, and Data

This work was supported by the National Science Foundation's IGERT Program grant #1249400 (AM, TL) and NOAA grant NA150AR4310145 (JA) Data is publicly available at http://climate.nkn.uidaho.edu/IntegratedScenarios/. Comments from Valeriy Ivanov and two reviewers greatly improved this manuscript.

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Figures

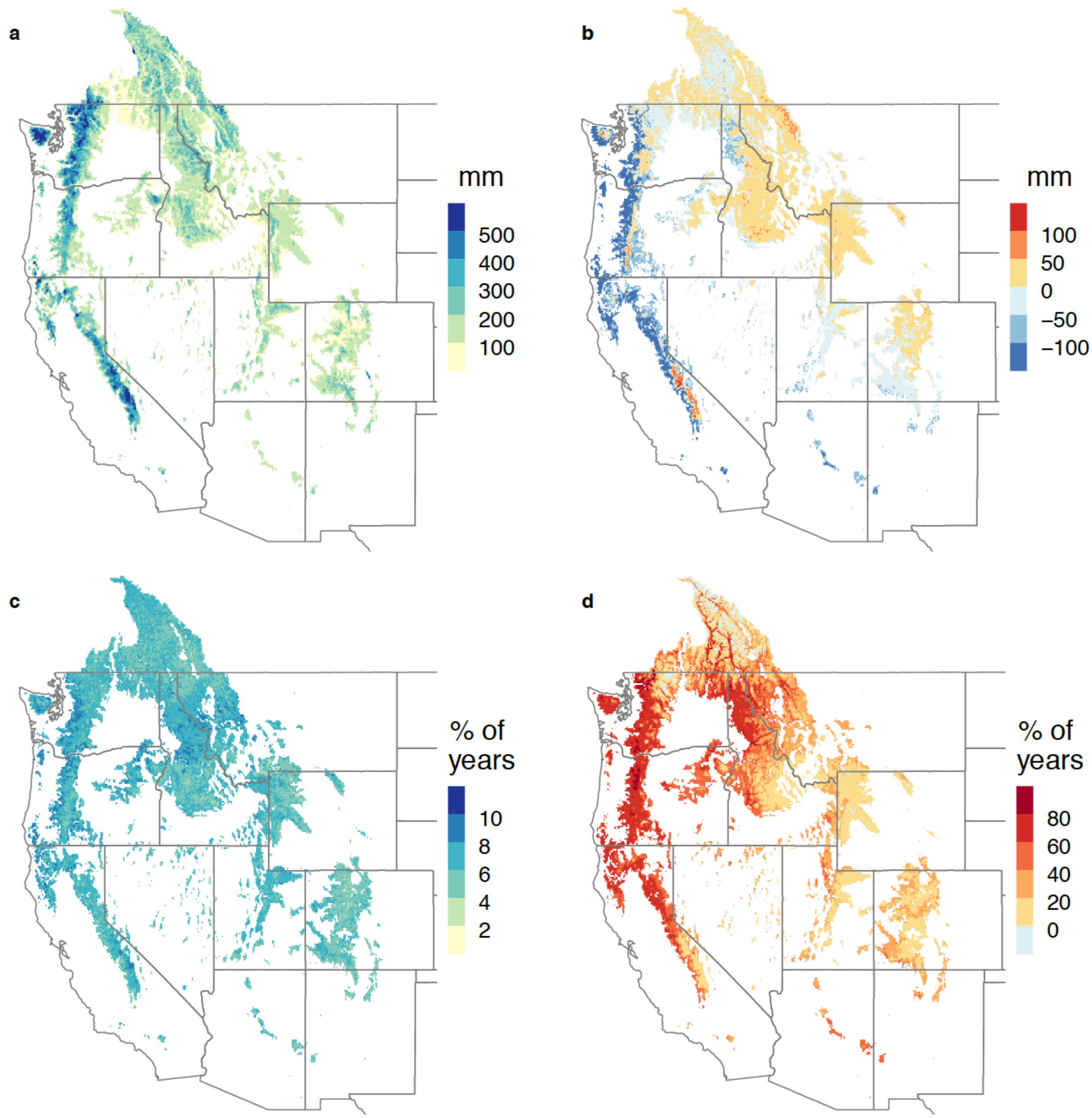

Figure 3.1 (a) Historical and (b) change in SWE_{max} IQR. (c) Historical and (d) change in percent of water years classified as two-year consecutive snow droughts.

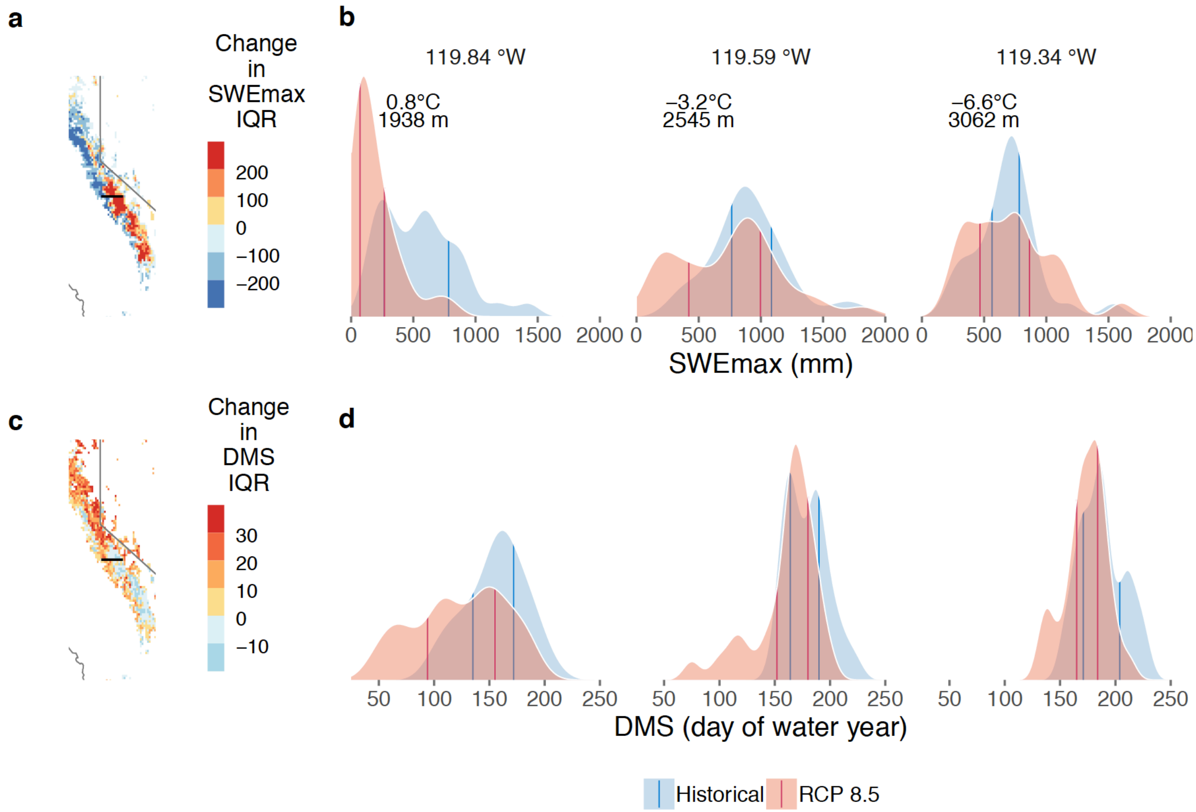

Figure 3.2 (a, c) Maps of changes in SWE_{max} and DMS IQR for CNRM-CM5 with transects marked at 38.0°N (Figure S15 for area map). (b, d) Distributions of (b) SWE_{max} and (d) DMS in CNRM-CM5 for three points on the transect marked in (a) and (c). Vertical lines indicate first and third quantiles in the historical and mid-21st century cases. Historical average winter temperature and elevation are noted.

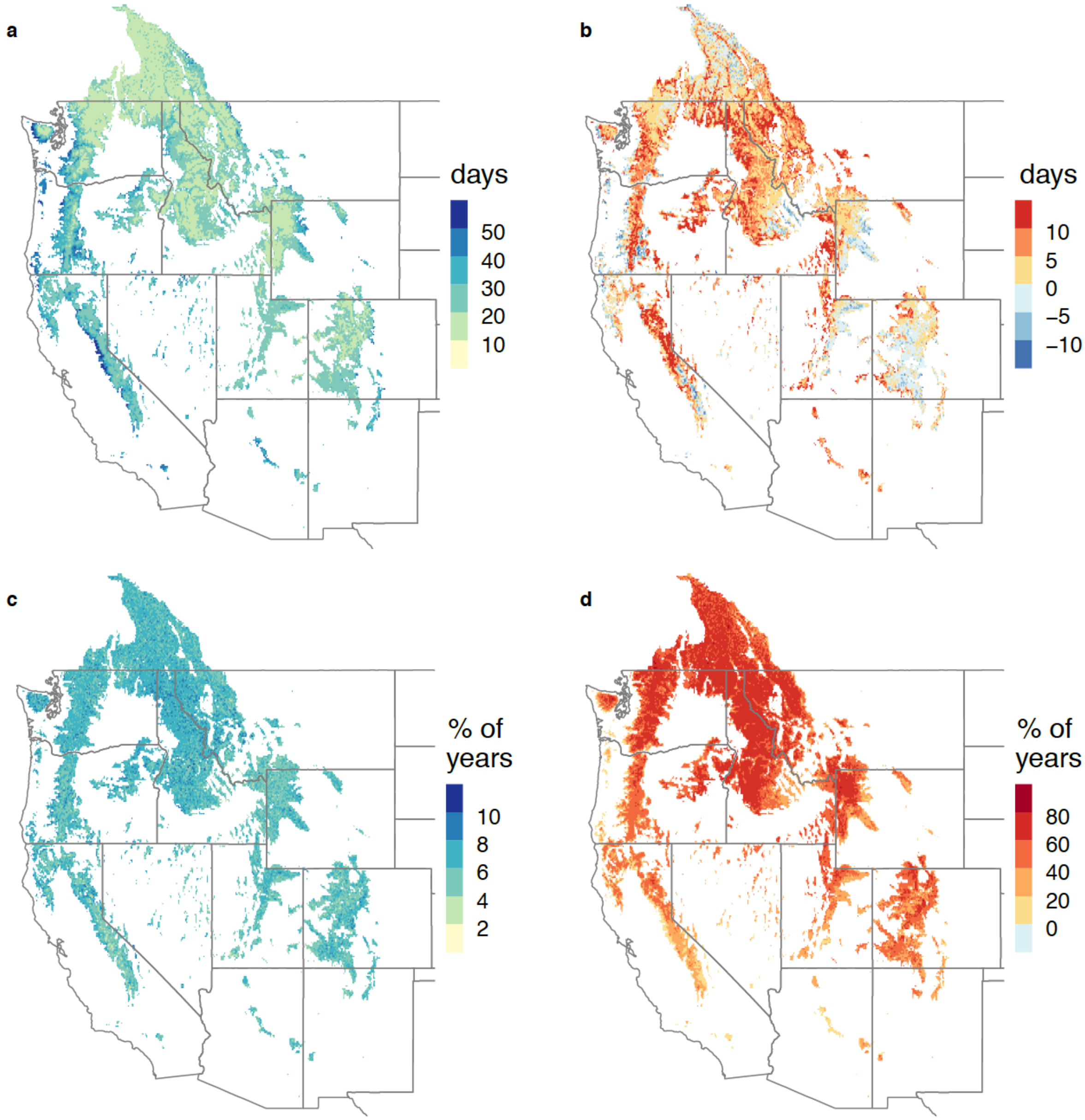

Figure 3.3 (a) Historical and (b) change in DMS IQR. (c) Historical and (d) change in frequency of consecutive early SWE years.

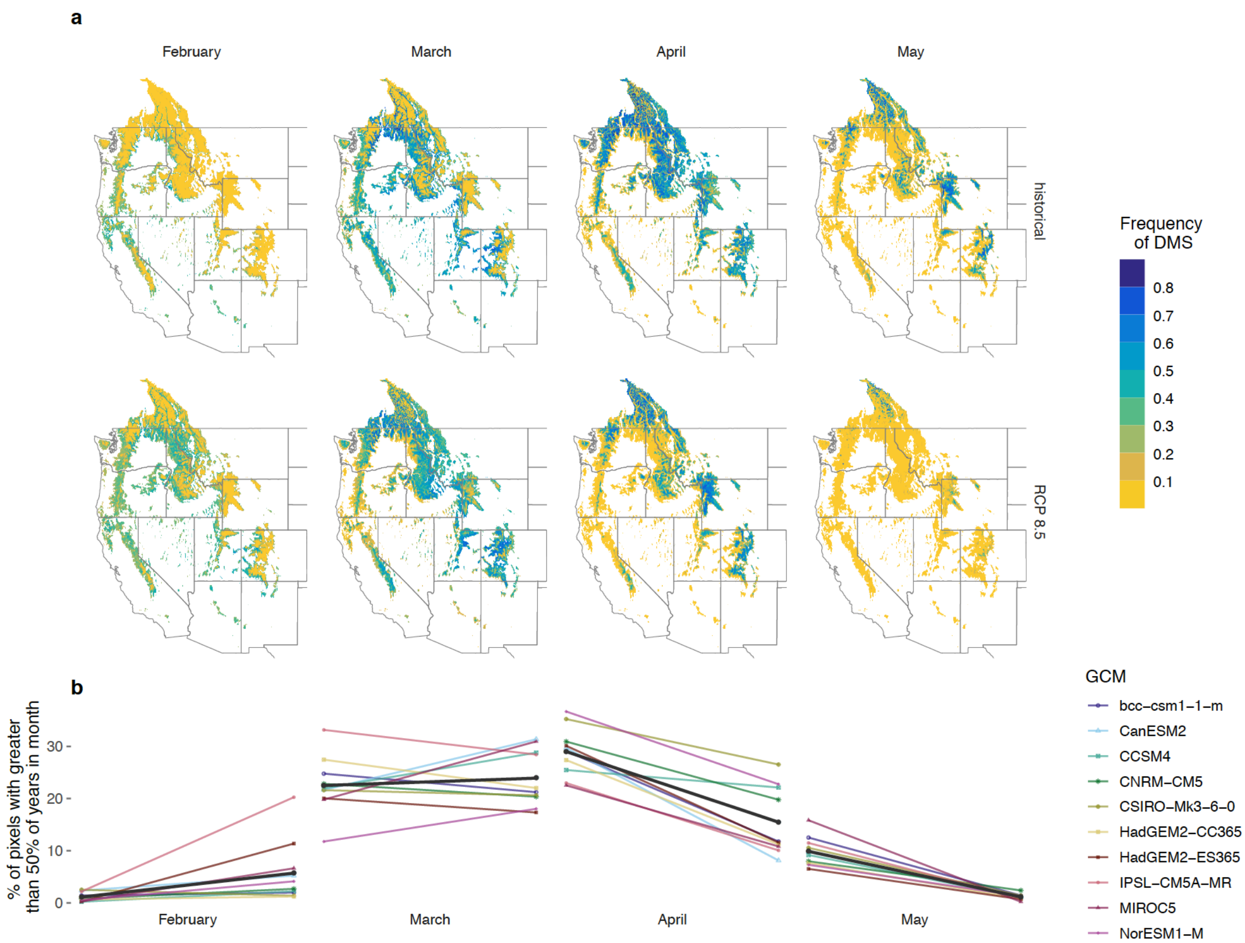

Figure 3.4 (a) Historical and mid-21st century frequency of date of peak SWE occurring in each month. (b) Change in percent of pixels from historical (left) to mid-21st century (right) for which DMS occurs in a given month the majority of the time, with 10-GCM mean in black.

Chapter 4: Higher Snowfall Intensity Reduces Warming Impacts on Winter Snow Ablation

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Abstract

Warming temperatures are altering winter snowpack accumulation and ablation. Physically-based snowpack simulations have indicated that increasing precipitation intensity may buffer the impacts of warming on annual maximum snow water equivalent. Here, we assess this relationship using an observational dataset from the western U.S. and show that greater snowfall intensity reduces winter ablation, particularly in warmer conditions. We also use output from a hydrologic model to evaluate the effect of snowfall intensity on ablation in future climate scenarios. Snowfall intensity is projected to increase in the continental interior, slightly mitigating the effects of warming on winter ablation, and decrease in maritime climates, exacerbating the effects of warming on ablation. The average effect of the trend in snowfall intensity on winter ablation varies spatially, ranging from -7.7 to +7.8%. These results indicate the importance of considering snowfall intensity in climate change impacts on snow-dependent ecosystems and water resources.

Introduction

Two major features associated with anthropogenic climate change in the western United States are decreasing snowpack(Barnett et al., 2008) and increasing precipitation intensity (Giorgi et al., 2011; Min et al., 2011; Seneviratne et al., 2012). In the western United States, warming temperatures have already resulted in large-scale declines in spring snowpack (Barnett et al., 2008; Hamlet et al, 2005; Knowles et al, 2006; Mote et al, 2018; Pierce et al., 2008) due to both a transition from snow to rain(Klos, Link, & Abatzoglou, 2014) and increased winter ablation of snow. Decreases in spring snowpack are projected to continue into the 21st century (Fyfe et al., 2017; Gergel et al, 2017; Rhoades et al, 2018), and runoff timing in snowmelt dominated watersheds will continue to advance (Stewart, Cayan, & Dettinger, 2004). The impacts of decreases in spring snowpack and advances in melt timing include reduced water supply and increased conflict between users (Berghuijs et al, 2014; Dettinger et al, 2015), and altered soil moisture dynamics (Harpold & Molotch, 2015; Maurer & Bowling, 2014), forest phenology (Trujillo et al, 2012), and carbon sequestration (Arnold et al, 2014).

Another robust feature of anthropogenic climate change is increasing precipitation intensity (Seneviratne et al., 2012), characterized by both more frequent dry days (Polade et al, 2014) and more precipitation occurring on wet days (Giorgi et al, 2011; 2014; Orlowsky & Seneviratne, 2012). Both observations (Alexander et al., 2006; Frich et al., 2002; Karl & Knight, 1998; Kiktev et al, 2003) and projections based on climate models (Meehl et al, 2005; Min et al., 2011; Tebaldi et al, 2006) indicate increases in precipitation intensity, including in the western U.S. (Kim, 2005), though observed changes in precipitation intensity and extremes are generally not very spatially coherent, with variable statistical significance. Although most research conducted on precipitation intensity has focused on rainfall, snowfall intensity will respond differently due to the contrasting impacts of decreasing number of snowfall events due to warming and greater atmospheric humidity on snow days. Both theory and climate models suggest that the most extreme snowfall events will decrease much less than mean snowfall, and in some locations may even increase (Danco et al, 2016; Lute et al, 2015a; O'Gorman, 2014, 2015).

Although changes in snow accumulation are predominantly caused by a warming-induced shift from snow to rain, changes in snowpack ablation are the result of changes in the snowpack energy balance components and are subject to the alteration of processes that affect net short- and longwave radiation and turbulent fluxes, including snow albedo (Painter et al, 2017; Skiles et al, 2018), temperature, humidity (Harpold & Brooks, 2018), and wind (Mott et al, 2018; Pohl et al, 2006). Changes in snowfall intensity may alter these energy balance dynamics. In particular, changes in the simulated snow energy balance under more extreme precipitation regimes have been found to reduce winter ablation, thereby partially mitigating the effects of warming on annual maximum snow water equivalent (SWE; Kumar et al, 2012). This is for three general reasons: first, deeper snowpacks require larger energy inputs to reach a cold content of zero and initiate melt. Second, given otherwise identical environmental conditions, deeper snowpacks have higher thermal capacity and warmer average temperatures than shallower snowpacks; this results in reduced net longwave, sensible, and ground heat fluxes into the snowpack. Third, in relatively intense, intermittent snowfall regimes, available energy is not optimally used for snowmelt due to the higher probability of complete melt-out early in the accumulation season. While the sensitivity of SWE ablation to snowfall intensity has been established in simulation experiments, it has not been tested with observational data, nor has the spatial distribution of these effects been assessed.

In this study, we test the empirical evidence for the effect of snowfall intensity on winter ablation, using a network of snow telemetry (SNOTEL) sites distributed across the western United States. In order to assess the importance of this effect in future climate conditions, we also evaluate whether this effect is evident in spatially distributed snowpack simulations. Specifically, we use the outputs of the physically-based Variable Infiltration Capacity (VIC) model, forced with historical and Representative Concentration Pathway (RCP) 8.5 climate data from 10 global climate models (GCMs) from the Fifth Coupled Model Intercomparison Project (CMIP5) and downscaled using the Multivariate Adaptive Constructed Analogues (MACA) v2 algorithm trained using a historical gridded dataset (Abatzoglou & Brown, 2012; Liang et al, 1994; Livneh et al., 2013; Taylor et al, 2011). In conjunction with previous physically-based modeling simulations of snowfall intensity impacts (Kumar et al., 2012), we use evidence from both observed and simulated data to enhance scientific understanding and confidence in our findings.

Historical snowfall intensity impacts on winter ablation

Previous work on precipitation intensity has identified many ways to represent both intensity and extremes (Alexander et al., 2006; Frich et al., 2002); we use the simple daily intensity index, in which total precipitation is divided by the number of days on which precipitation occurred. We apply the simple daily intensity index only to snow liquid water equivalent accumulation, rather than total precipitation, as an index of snowfall accumulation intensity (SAI). Daily temperatures were extracted for each SNOTEL site from the TopoWx dataset to avoid known inhomogeneities in the SNOTEL temperature record (Oyler et al, 2015a; 2015b; 2016). Average winter temperatures (T_{avg}) were calculated as the mean daily average temperature from November through March.

Figure 1 shows the spatial distribution of the snowpack variables used in our statistical model. Winter ablation, the response variable, was calculated as the fraction of SWE measured by the snow pillow that ablated between the onset of snow accumulation and the date of maximum SWE. Mean winter ablation is generally lowest in the continental western United States and higher in the lower elevations of the Cascades as well as mountains of Arizona and New Mexico. SAI is generally greater in maritime regions and lower in continental regions, likely due to the significant contributions made by atmospheric rivers to overall winter precipitation in maritime regions (Lute & Abatzoglou, 2014).

We use a generalized additive mixed modeling approach to test the relationship between SAI and winter ablation, and its sensitivity to winter T_{avg} (Wood, 2017). Our statistical model estimates winter ablation as a function of a smoothed interaction between SAI and T_{avg}. To account for the spatial and temporal structure of the data, we include site-specific random effects and interaction terms between elevation and region, as defined by Serreze et al. (1999), as well as water year and geographic coordinates. Each observation was weighted by annual maximum SWE in order to reduce the influence of very low snow years. To test the importance of SAI, this model was compared to a null model where SAI was excluded. Model diagnostics are presented in Figures S1-S3.

Results of statistical modeling with SNOTEL data show that greater SAI is associated with reduced winter ablation, particularly in regions where winter temperatures exceed 0° C (Figure 2). For example, on average across sites, when winter T_{avg} is 3 °C and SAI is 20 mm/day, the model predicts that 15.0% (±6.1% s.d.) of accumulated SWE will ablate before the date of peak SWE. If SAI is reduced to 5 mm/day, the model estimates winter ablation of 24.0% (±11.1%); this suggests an average difference of 9.0% (±2.5%). In contrast, when winter T_{avg} is -3 °C, the same calculation suggests only a 2.7% (±1.1%) difference in ablation. Comparing AIC, R^2, and deviance explained with a null model supports the hypothesis that including SAI in a model of winter ablation improves the model, though these differences are relatively small (Table S1). This is because snowfall intensity has the largest effects at temperature ranges that comprise a relatively small portion of the data.

To assess the effect of snowfall intensity in only warmer conditions, we limited the model to observations where winter T_{avg} was greater than 0 °C (Figure S4). In this case, we found stronger evidence for the full model than in the case with all data (Table S1). In contrast with model results from the full dataset, there is some evidence for non-monotonic effects of SAI in the model with warmer conditions, though these effects occurred only in regions with very warm winter temperatures, low SAI values, and very low peak SWE. These findings indicate that snowfall intensity is particularly important in warm conditions, and may therefore be increasingly important in warmer climates.

Snowfall intensity impacts based on simulated snowpacks

To assess the importance of snowfall intensity in projected future climates, we applied an identical statistical model framework to simulated snow dynamics from VIC forced with 10 GCMs in historical and RCP 8.5 conditions from 1951-2099. We conducted this analysis using data from grid cells on which SNOTEL sites are located.

Similar to the analysis of observed data, results using modeled VIC output indicate that SAI affects mid-winter melt (Figure 3). All models suggest that SAI affects winter ablation at warmer temperatures, with minimal effect at cooler winter temperatures. Some GCM results have the same general shape as results using SNOTEL data; others suggest a non-monotonic effect of SAI and winter T_{avg} on winter ablation at very low values of SAI and warm winter T_{avg}. As with SNOTEL data, the data in the regions contributing to non-monotonic tendencies have relatively warm winter T_{avg} (>2-3 °C) with low SAI (<5-7 mm/day); peak SWE in these observations is generally 4-9% of the average peak SWE of the full dataset.

Comparisons with null models also support the hypothesis that SAI affects winter ablation. For all 10 GCMs, the differences in AIC range from 6870 to 11179, and support the full model in all cases (details in Table S1). The added R^2 from the full models ranges from 3.8% to 6.3% for the 10 GCMs. Differences in deviance explained range from 3.3% to 5.9%. These results suggest a larger difference between the full and null models than did the findings with SNOTEL data. This may be due to the fact that the VIC data include more warm observations in future climates, conditions for which SAI is more likely to affect winter ablation.

Snowfall intensity importance in future climates

Projected changes in SAI exhibit distinct geographic patterns that are fairly consistent between GCMs (Figure 4; Figure S6). SAI is generally projected to decrease in the maritime western and southwestern mountains, where it was historically largest, and increase in the cooler interior mountain west. The spatial patterns observed here are in agreement with previous findings regarding the magnitude of extreme snowfall events (Lute et al., 2015b). Although the spatial pattern of changes in SAI is consistent between GCMs, the magnitude of increases and decreases varies. The projected decreases in SAI in maritime regions are likely due to warming temperatures and a transition from snow to rain. In contrast, at cooler continental sites, the change in SAI may more closely reflect changes in overall precipitation intensity, as the sites are less subject to transitions from snow to rain. Winter T_{avg} is projected to increase consistently, with the greatest increases in the continental interior. Winter ablation is predominantly projected to increase, with a few isolated sites showing decreases; these may be sites that receive very minimal snow accumulation in future climates.

In order to determine whether projected changes in SAI exacerbate or reduce the effects of warming on winter ablation, for each GCM and SNOTEL site, we built a difference-adjusted time series of SAI values with the difference estimates depicted in Figure 4 removed. We then used our statistical models to estimate fitted mid-winter melt with the original time series and altered time series. Figure 5 shows the average difference in estimated mid-winter melt from 2070-2099 between the cases with original versus detrended values of SAI, averaged over GCMs. Average differences in winter ablation with and without a trend in SAI range from -7.7 to +7.8%. Positive values indicate that winter ablation was greater when the trend predicted by the GCM was included; negative values indicate that winter ablation was greater in the detrended time series. That is, the positive values in the Sierra Nevada and Cascades indicate that projected increases in winter ablation will be exacerbated by the generally decreasing trends in SAI, while the negative values in the Rocky Mountains indicate that projected increases in mid-winter ablation will be reduced by the generally increasing trends in SAI.

Discussion

Using both empirical and simulated snowpack data in historical and future climates, we found that snowfall intensity significantly affects winter snow ablation at relatively warm winter temperatures. Higher snowfall intensity is associated with reduced winter ablation, which aligns with theoretical expectations and a previous sensitivity assessment of physically-based snow simulations at the point scale (Kumar et al., 2012). Accounting for projected future changes in snowfall intensity suggests that increased snowfall intensity will reduce the effects of warming on winter ablation in the continental interior of the western United States, while decreased snowfall intensity will exacerbate the effects of warming on relatively warm, maritime western snowpacks.

The findings of this study suggest that the physical mechanisms described in previous studies (Kumar et al., 2012) by which snowfall intensity alters winter ablation dynamics may be responsible for the observed relationships between winter ablation and snowfall intensity. We also propose two additional mechanisms. First, snowfall intensity may alter canopy interception. If snowfall occurs in relatively few, intense storms, then the forest canopy may more often reach its maximum interception capacity. Higher snowfall intensity would therefore be associated with reduced interception; this mechanism is in agreement with the direction of the effect suggested by Kumar et al. (2012) and observed in this study. This effect should be present in the VIC model, which includes forest canopy interception, but to a much lesser degree in SNOTEL data because SNOTEL sites are usually placed in small forest gaps. Second, snow albedo dynamics may change with altered snowfall intensity. On average, there should be longer intervals between storm events with higher snowfall intensity; this would be associated with greater albedo decay and thus greater potential for winter ablation. This mechanism would act in the opposite direction of the effects observed in this study, and should be present in both SNOTEL data and the VIC model (Andreadis et al, 2009). It may be complicated by the fact that small snowfall events may not completely mask the albedo of the underlying surface (Baker et al, 1991), which would be reflected in observational but not modeled data.

Different characteristics of changing precipitation intensity, as well as different climatic contexts, may affect the relevance of our findings for different locations. Changes to projected snowfall intensity vary globally, with projected increases in much of the central United States (Notaro et al, 2014), Japan (Kawase et al., 2016), China (Zhou et al, 2018), and the Pyrenees (López-Moreno et al, 2011), though these increases in intensity are often paired with decreasing total snowfall or number of daily snowfall events. In much of the western U.S., the change in number of dry days and changing precipitation intensity on wet days have competing effects on average annual precipitation change in CMIP5 projections (Polade et al., 2014). Some of the mechanisms by which we suggest precipitation intensity affects winter ablation, such as non-optimal use of energy and albedo decay, are primarily affected by the number of dry days, while others, such as canopy interception and differential energy balance dynamics between a deep and shallow snowpack, are more likely affected by event size. The combination of these two factors in different locations would likely affect the applicability of our results in locations outside of our study domain.

Conclusions

We found that increased snowfall intensity is associated with reduced mid-winter melt, particularly in relatively warm conditions. Analysis of observational data and historical and future modeled data confirmed this effect in both cases, and suggested that snowfall intensity will become increasingly important in warmer climates. The increasing importance of snowfall intensity in warmer climates for determining snowpack energy balance dynamics and ablation suggests a need for high confidence estimates of changes in snowfall intensity, and represents one example of a mechanism creating non-stationarity between seasonally-averaged climatological variables and snowpack accumulation and subsequent runoff. Moreover, snow accumulation and melt models that account for a full snow energy balance will reflect the effects of snowfall intensity, while simpler temperature index models may not.

Given spatially varying projections of changes in snowfall intensity in the western U.S., altered snowfall intensity may reduce the effects of warming on winter snow ablation in the colder continental ranges, and exacerbate warming's effects in relatively warm maritime ranges, with average changes due to trends in snowfall intensity varying spatially from -7.7 to +7.8%. Particularly in regions where large sectors of the economy depend on a limited supply of water, or where aquatic species are on the margins of viability, these differences in winter ablation may be critical. These findings demonstrate the power of integrating multiple lines of evidence to enhance confidence in scientific results in a sector with important implications for snow-dependent social and ecological systems. In conjunction with other climate-induced alterations to the snow energy balance, changing snowfall intensity is an important factor to consider in projections of climate-induced changes in snowpack and water resources, as well as associated adaptation planning.

Acknowledgements

This work was supported by the National Science Foundation's IGERT Program (Award #1249400).

Methods

Data

Daily SWE data were obtained from the National Resources and Conservation Service Snow Telemetry Network (SNOTEL). Quality assurance and control is conducted by the NRCS. We evaluated two levels of additional quality control. In a minimal quality control approach, we only removed water years with greater than 10% of data missing. We also considered more stringent quality control measures as described by Serreze et al. (1999) and Lute and Abatzoglou

(2014). These procedures removed water years with greater than 10% of data missing, removed data with missing values or exceptionally large precipitation values at the beginning of a water year, large snowfall events (>63.5 mm SWE/day) immediately followed by large losses, snowfall events with with air temperatures greater than 5.7°C, and snowfall events with greater than 254 mm of SWE per day. These more stringent measures presented some concern that large snowfall events, which were of particular interest in this study, were erroneously removed from the data set. Comparisons of model results with the more and less stringent quality control approaches indicated that the level of quality control applied did not affect the overall study conclusions; we report results from the more minimally filtered data, where water years with greater than 10% of data missing were removed. These procedures resulted in 705 stations and 20249 station-years of data, with an average of 29 (±10) years of data per station. For each site and water year, data were extracted from the first date of snow accumulation until the date of SWE_{max}, and snowmelt and intensity metrics were calculated on this subset.

Snowfall intensity was calculated as:

$$SAI_i = \frac{\sum_{w=1}^{W} SWE_{wi}}{W}$$

where SAI_i is the snowfall accumulation intensity for the ith water year at a site, SWE_{wi} is the daily snowfall on days where snowfall is greater than 1 mm, and W is the number of days on which snowfall occurs, from the first snowfall greater than 1 mm in a water year until the date of peak SWE. SNOTEL has a resolution of only 2.54 mm SWE per hour; for this reason the effective threshold is 2.54 mm for SNOTEL data.

SNOTEL temperature data was not used because of known inhomogeneities in the temperature record (Oyler, Dobrowski, et al., 2015). As an alternative, TopoWx is a 30 arcsecond (~800 m) gridded dataset of daily minimum and maximum temperatures from 1948 - 2015, built with weather station data, elevation-based adjustments, and spatial interpolations of remotely sensed land skin temperature (Oyler, Ballantyne, et al., 2015; Oyler et al., 2016). Daily T_{min} and T_{max} TopoWx values were obtained for each SNOTEL site over the period of record, and T_{min} and T_{max} were averaged over November-March.

Model results with the full dataset indicated that SAI was most important at warmer temperatures. To more specifically examine these effects, we also conducted our analysis with only the parts of the dataset with winter Tavg greater than 0 °C. This resulted in 3505 observations, 292 unique sites, and 12 ± 12 years of data at each site.

To determine the effects of snowfall intensity on winter ablation under future climates, we used publicly available modeled SWE data under historical and future climates. Ten global climate models (GCMs) from the CMIP5 project (Taylor et al., 2011) were run for historical conditions and representative concentration pathway (RCP) 8.5. These model outputs were downscaled to 1/16th degree grid cells using MACAv2-LIVNEH downscaling (Abatzoglou & Brown, 2012; Livneh et al., 2013); we obtained daily average temperature from these scenarios from 1951-2099. The Variable Infiltration Capacity (VIC) model was run with these downscaled results (Liang et al., 1994). We obtained daily SWE data for each SNOTEL site from the VIC outputs. As with the SNOTEL data, T_{avg} was averaged over November-March, and the snowfall intensity metrics were calculated for each water year. One important difference is that due to SNOTEL data resolution, the effective threshold for SAI calculations was 2.54 mm for SNOTEL, while a 1 mm threshold was used for VIC. When SAI was calculated with a 2.54 mm threshold with the VIC data, SAI values were highly correlated (correlation coefficient = 0.91) with those calculated with the 1 mm threshold, suggesting this decision has a minimal impact on our results.

Statistical models

We used a generalized additive mixed modeling approach to test the relationship between SAI and winter ablation, as well as the temperature dependence of this relationship (Wood, 2017). Specifically, our model for fraction of SWE that melted before date of maximum accumulation for the jth site and ith water year was constructed as follows:

$$g\left(f_{melted\ i,j}\right) = \alpha + \beta_j + f_1\left(wy_j, N_i, E_i\right) + f_2\left(elev_j, region_j\right) + f_3\left(SAI_{i,j}, T_{avg\ i,j}\right) + \varepsilon_{i,j},$$

$$\beta_j \sim N(0, \psi_\theta), \varepsilon \sim N(0, \Delta\sigma^2)$$

where g indicates a beta family of model with a logit link, α is an intercept, β_j is a site-specific random effect, and f_1 , f_2 , and f_3 are tensor product smooths with cubic regression bases. In addition to average winter T_{avg} and SAI, the model includes northing (N), easting (E), elevation (elev), and region (as defined in Serreze, 1999). Each observation was weighted by annual maximum SWE in order to reduce the influence of very low snow years. A test of the model without the weights included suggested that weighting the model did not alter our overall conclusions. To test the importance of SAI, this model was compared to a null model where SAI was excluded.

Using the mgcv package in R (R Core Team, 2018; Wood, 2017), this model was implemented as:

```
mod <- gam(f_melted ~ s(site_id, bs = "re", k = 10) +
      serreze_region +
      s(elevation, by = serreze_region, bs = "cr", k = 10) +
      te(mean_tavg, sai, bs = "cr", k = 10) +
      te(N, E, wy, k = c(10, 10), bs = "cr"),
      data = dat,
      family = betar(),
      method = "REML",
      gamma = 1.4,
      weights = swe_weights)
```

Tensor product smooths were selected for their anisotropic nature; similar approaches have been applied previously in references (Augustin et al., 2009; Augustin et al., 2013; Montoyaet al., 2014). The beta distribution is conceptually appropriate for this dataset but requires values to be inside the interval (0, 1). We reset zero values to be equal to half the smallest non-zero value; tests of this approach with zero values set to one-tenth the smallest non-zero value indicated that this threshold had no impact on the study conclusions. The site random effect was included to account for site-specific effects, such as topographic or vegetation effects, and f_1 and f_2 were included in order to account for spatiotemporally varying factors that were not otherwise included in the model. Model diagnostics indicated that the necessary assumptions were met (Figures S1-S3). The effect of SAI on winter ablation was determined via comparison with a null model, in which the SAI term was removed. We included R^2, deviance explained, and AIC in our comparisons.

Estimated effect size of changes in SAI

To assess the impacts of snowfall intensity on SWE_{max} under future climates, the same statistical models were applied to the VIC data as were used for SNOTEL data. Models were applied separately for each of the 10 GCMs, and fitted on a dataset that included water years from 1951-2099.

We also estimated the impact of projected trends in SAI on changes in mid-winter melt. For each site and GCM, we calculated the 30-year average value of SAI in historical (1970-1999) and late-21st century (2070-1999) climates. We subtracted the mean difference between the historical and late 21st century cases from the late 21st century data in order to build a future time series with the trend in SAI removed. Then, winter ablation was modeled using the generalized additive mixed model fitted in equation 1 with both the observed and detrended SAI time series. Winter ablation calculated with the detrended SAI was subtracted from that calculated with SAI from the original VIC time series for the years 2070-2099 in order to estimate the effects of projected changes in SAI on winter ablation in the late 21st century. These values were averaged over GCMs and water years.

Data availability

All data used in this study is publicly available through the Integrated Scenarios Project (https://climate.northwestknowledge.net/IntegratedScenarios/), TopoWx repository (http://www.scrimhub.org/resources/topowx/), and the Natural Resources Conservation Service (https://www.wcc.nrcs.usda.gov/snow/).

Code availability

Computer code needed to reproduce these analyses is available from the corresponding author upon request.

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Figures

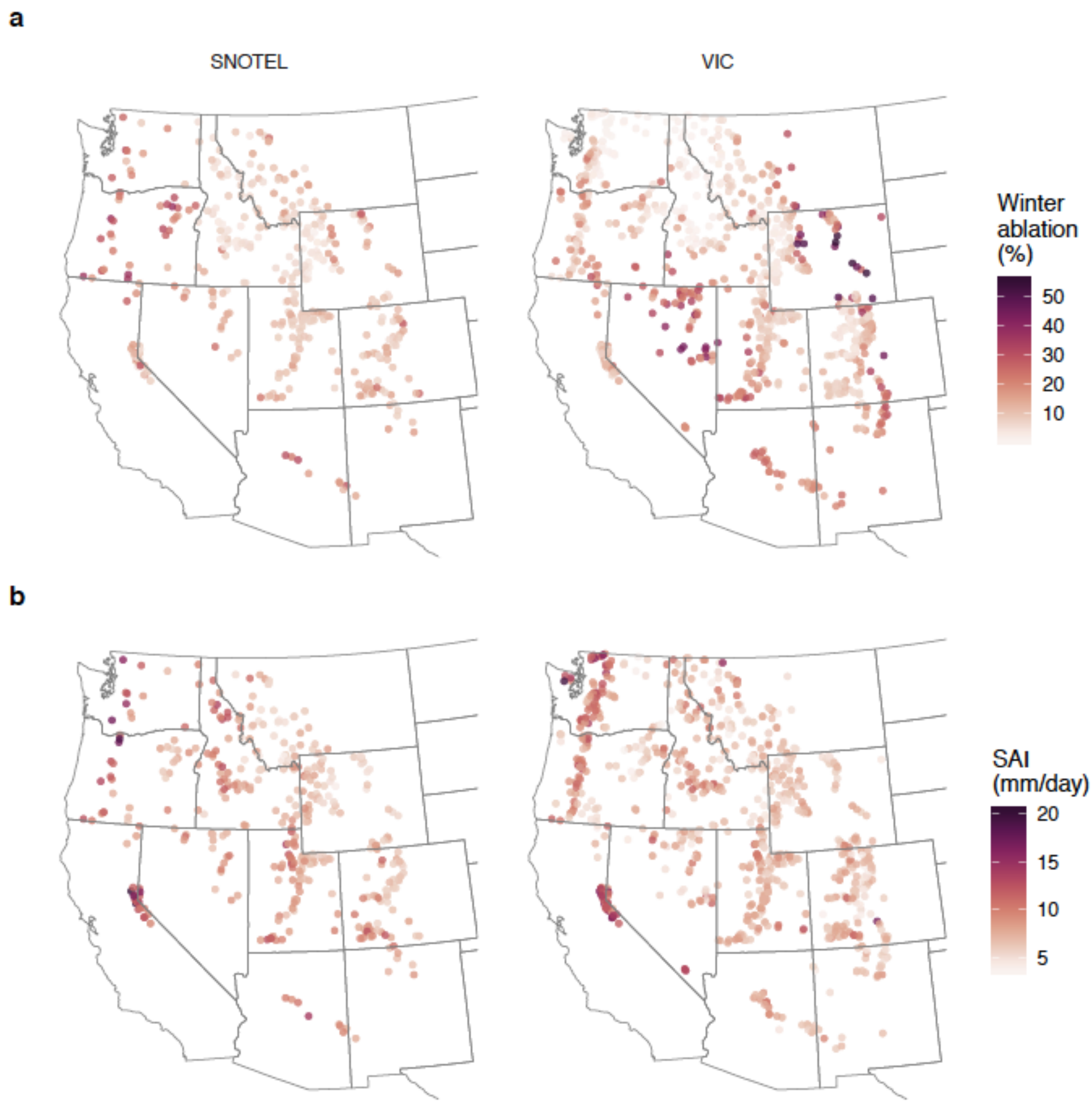

Figure 4.1 Distribution of (a) winter ablation and (b) SAI over the historical period from 1996-2015 shown for both SNOTEL and VIC data. Only SNOTEL sites with 20 years of data over that period are included in the figure.

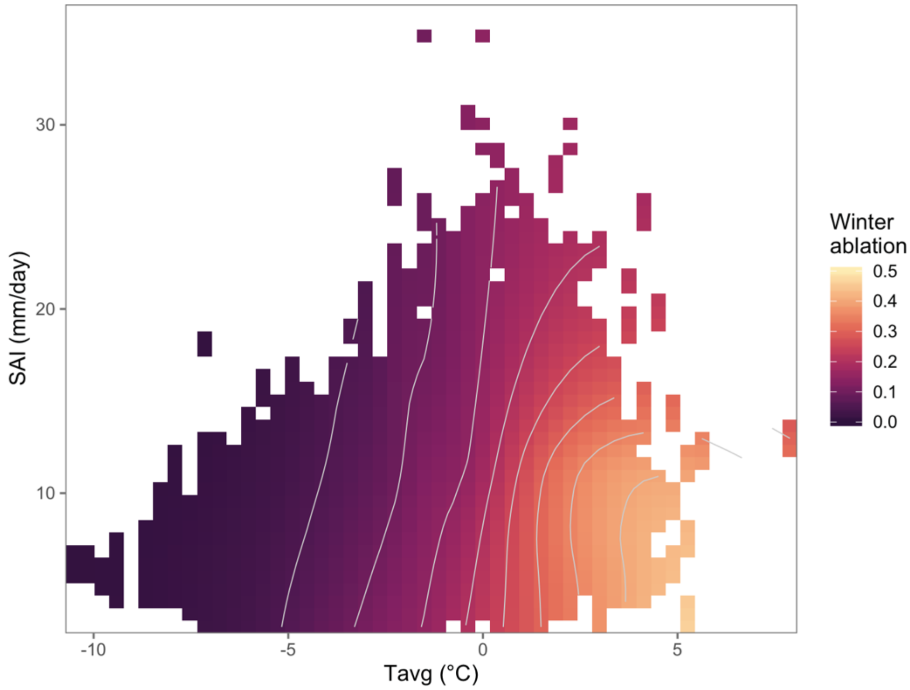

Figure 4.2 Contours represent the fraction of total snowfall that ablates during before the date of peak SWE, fitted with the statistical model for all SNOTEL data, filled only for regions with data in SNOTEL record. For variables not plotted, fitted values are estimated at at the median values, and the site with the median random effect was used. Changing the factor variables alters absolute values but does not change the contour shapes (Figure S5).

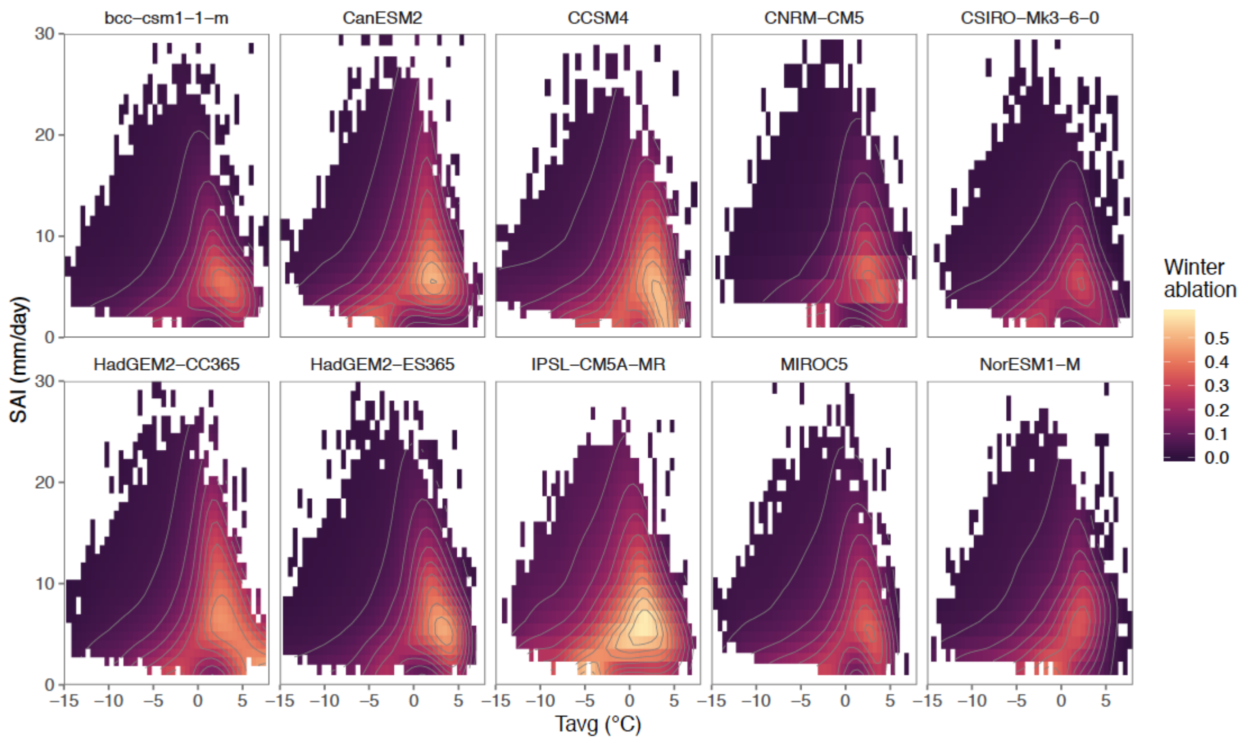

Figure 4.3 Contour plots of fitted winter ablation for model built with VIC data forced with each of 10 GCMs. Contour lines represent intervals of 0.05 (5%) change in winter ablation.

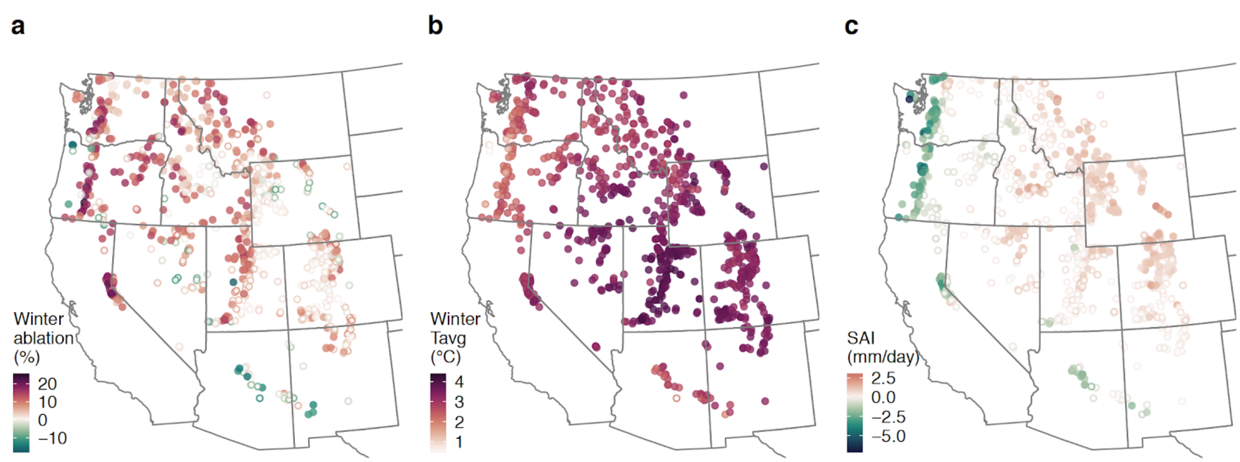

Figure 4.4 Change in 30-year average (a) winter ablation, (b) winter T_{avg}, and (c) SAI from 1970-1999 to 2070-2099, averaged across GCMs (individual GCMs for each variable in supplemental material). Points are filled if at least 50% of GCMs agreed on a statistically significant change (two-sided t-test $p < 0.05$).

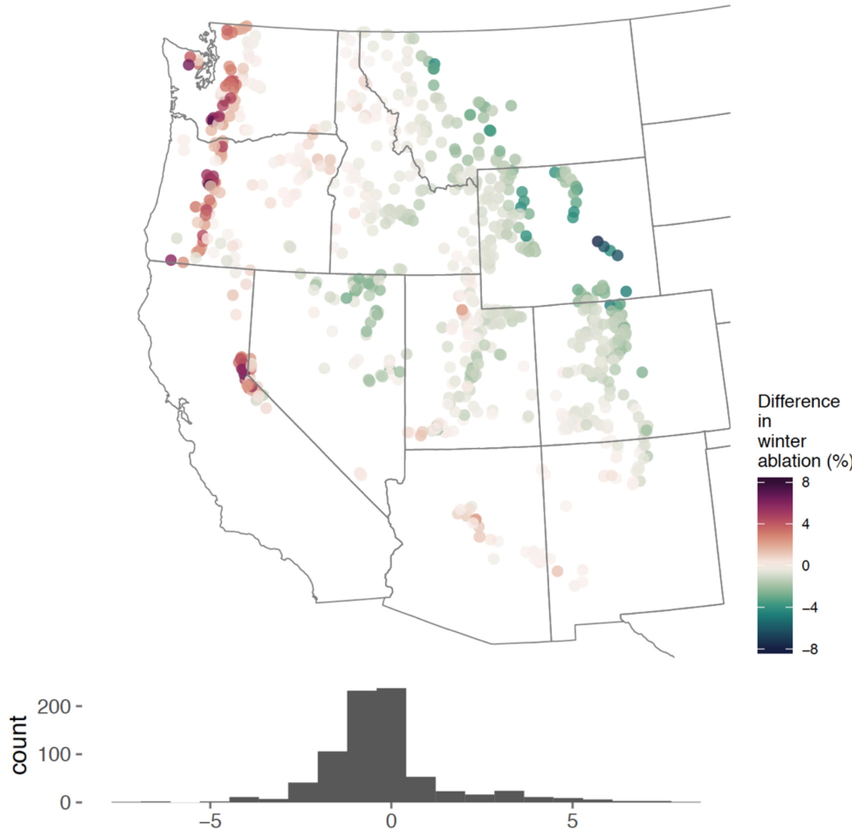

Figure 4.5 Fitted absolute changes in winter ablation percent with difference-adjusted SAI in contrast to GCM-projected SAI, averaged over 10 GCMs and water years 2070-2099 (results for individual GCMs are in Figures S9-S10). Histogram shows distribution of mapped values.

Chapter 5: Spatial and topical distributions of climate change research in the mountainous headwaters of the Columbia River Basin

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Abstract

This study reviews multidisciplinary thematic content and spatial distributions of climate change research in mountain headwaters of the Columbia River Basin (CRB). Like other transboundary basins, the CRB encompasses diverse ecosystems and cultures. Climate change presents interacting biophysical and social threats to existing ecosystem services such as the provisioning of seasonal snowmelt to maintain freshwater supplies, suggesting a need for complex adaptation and mitigation strategies. Results from this systematic review suggest that climate change research in the CRB focuses on impacts more frequently than adaptation, while mitigation is rarely a focus. Most studies assess trends at large spatial extents, rely on secondary data, and make projections of climate change impacts rather than observations or predictions. The spatial distribution and thematic content of research varies across the international border, with greater concentrations of research in the United States than Canada and few examples of transboundary collaboration. A general scarcity of social science research, and limited interaction between social and biophysical content, reinforces the need for increased collaboration between disparate disciplines. This content analysis illuminates knowledge gaps and calls for more research related to climate change adaptation and mitigation, increased integration of social and biophysical sciences, and collaborations that bridge the international border for a more integrated basin-wide focus. Results help inform new research both in mountainous regions and in the CRB specifically, while increasing the potential for science and management communities to co-produce actionable science and effective responses to climate change.

Introduction

Climate change research in mountains

Climate change in mountainous regions is projected to have serious consequences for social and ecological systems due to impacts on spatiotemporal snowpack dynamics, fire regimes, and biodiversity and ecosystem function, many of which are already occurring (La Sorte & Jetz, 2010; Nogués-Bravo, Araújo, Errea, & Martinez-Rica, 2007; Viviroli et al., 2011). These remote environments are critically important for many societies; for example, one-sixth of the global population resides in areas that depend on mountain meltwaters (Parry, 2007). Despite their importance, research in mountainous landscapes is relatively limited due to sparse monitoring networks and challenges associated with modeling complex terrain (Dobrowski, 2011; Strachan et al., 2016; Viviroli et al., 2011; Young et al., 1999). Systematic reviews are an important way to more holistically understand changes in these environments. In this study, we synthesize climate change research in a mountainous study area and identify key research gaps to benefit regions impacted by climate change.

Impacts of science on management

One motivation for research syntheses is that scientific findings frequently have implications beyond their own disciplines. Research methods, spatial and temporal resolution and extent, amount of research conducted, and disciplinary diversity used to understand specific environmental issues influence how knowledge is applied. This includes not only how the natural environment is perceived, but also how it is valued and managed (Bocking, 2004; Yearley, 2008). In an ideal world, the relationship between environmental knowledge production and application may be one of supply and demand, where resource managers (science consumers) express knowledge needs that are then fulfilled by scientific research (science producers); however, many barriers exist that create a more complex relationship between science and management (Bisbal, 2018; deCrappeo et al., 2018). These barriers include difficulties identifying relevant actors and management priorities and challenges associated with aligning scientific and management priorities (Bisbal, 2018). With regard to complex environmental resource problems, characterizing research from multiple disciplines and understanding how these different fields of knowledge intersect is a critical step in synthesizing knowledge to ultimately make it useful for management needs. Doing so enables society to respond and adapt to the myriad challenges introduced by non-stationary climate regimes (Hulme 2010; Milly et al., 2008).

Calls for approaches to systematically assess climate change research are ubiquitous (Hulme 2010; Petticrew et al., 2011), yet conducting comprehensive reviews is challenging because the scope of climate change involves synthesis across multiple disciplines (Lenhard et al., 2006). Previous global reviews of climate change and water resources in mountainous areas have identified topical research priorities, highlighted the importance of environmental monitoring, and concluded that more detailed regional studies and linkages between disciplines are needed (Viviroli et al., 2011). Several other recent reviews synthesize knowledge in specific regions and river basins. In the Po River basin (Italy) and the Red River basin (Vietnam), Pham et al. (2019) apply a comparative freshwater ecosystem services framework to review basin-scale climate impacts on freshwater. Lima and Frederick (2019) review the evolution of primary environmental threats and stressors in the Athabasca River Basin in Alberta, Canada. They identify a gap in studies that explicitly link climate change to other stressors, such as mining, dams, or land use change. A synthesis in Bangladesh identified a bias towards economic, rather than environmental or social issues, and call for more transdisciplinary studies to support evidence-based public policy (Tuihedar Rahman et al., 2016). A review of adaptive capacity and climate change across Himalayan River basins concludes that adaptation projects take place mostly at local scales, emphasize disaster risk management, and are led by government agencies (Sud et al., 2015). Finally, systematic efforts to review climate change in the Canadian Arctic report that scholarship in this region would benefit from increased involvement of the social sciences and humanities and more research related to adaptation (Ford et al., 2012; Ford and Pearce, 2010). With these works as inspiration, this study aimed to quantify thematic and spatial gaps in climate change related research for the mountainous headwaters of a large and complex watershed, offering recommendations for future research and a transferable model for performing a research synthesis.

Study Area

The CRB as a model for climate change

The Columbia River Basin (CRB) provides a useful case for understanding the state of knowledge about climate change in a mountain region that is profoundly affected by a non-stationary climatology. It is among many large, complex, and transboundary river basins with diverse ecosystems, complex socio-political histories, and a dependency on seasonal snowmelt to maintain water supplies and ecosystem function (Mankin et al., 2015). The region's water resources generate over half of the United States' hydroelectric power production, position the CRB as the leading producer of 22 key agricultural commodities, and sustain a population growing at more than twice the rate of the national average (EIA, 2018; USDA, 2018; US Census Bureau, 2017). The region depends on mountain water for environmental and economic well-being, has the scientific research and policy-making infrastructure to support extensive research, and engages in climate change adaptation and mitigation efforts through, for example, government-led vulnerability assessments (Muccione et al., 2016; Olson et al., 2017).

Climate change and water resources in the CRB

Warming temperatures in the CRB cause a suite of hydrologic changes, including decreasing snowpack, warmer stream temperatures, increasing precipitation in the northern parts of the basin, and uncertain changes in total flow volume (Elsner et al., 2010; Ficklin et al., 2014; Hamlet & Lettenmaier, 1999; Isaak et al., 2017; Rupp, Abatzoglou, & Mote, 2016; Schnorbus, Werner, & Bennett, 2014). In many areas, climate change contributes to earlier streamflow timing with lower summer flows (Stewart et al., 2005; Luce & Holden, 2009). Continued population growth and concomitant increasing water demands are expected across the CRB (Bilby et al., 2007; Huddleston et al., 2014). As is the case throughout much of the western United States, limited water availability invokes conflict among numerous actors who require this water for domestic, agricultural, navigation, hydropower, and municipal uses (Cotter and Sihota, 2015; Dettinger et al., 2015), as well as for supporting important cultural and spiritual values for tribal communities (Cosens et al., 2018). In addition to human water demand, watercourses and water bodies supply critical habitat to many aquatic species listed under the Endangered Species Act (ESA); the value of fisheries in the CRB is estimated at $150-$600 million (Cotter and Sihota, 2015).

Climate change and forest resources in the CRB

Forests represent another source of critical ecosystem goods and services in the CRB that are strongly affected by climate change. In mountainous regions, federally-managed forests cover the greatest percentage of land, and they provide ecosystem services valued around $149 billion annually (Flores et al., 2017). Forests also supply habitat to many terrestrial species, such as the threatened Northern Spotted Owl (Thomas et al., 2006) and wolverine (Copeland et al., 2010), which was listed due to its sensitivity to climate change. Impacts of climate change on forests include more frequent high-severity wildfires, range expansion of invasive bark beetles, and fluctuations in water availability (Westerling, 2006; Kemp et al., 2015). Despite extensive study of climate change impacts on forests, adaptation remains difficult due to uncertainty in projecting specific local impacts and managers' limited time for integrating current climate change science in management plans (Kemp et al., 2015).

Multi-disciplinarity in the CRB

Natural resource management and governance in the CRB is complex and requires interdisciplinary approaches that address numerous interacting physical, social, economic, ecological, and technical factors (Cosens et al., 2016; Hand et al., 2018). For example, mountain forests influence the water cycle, water quality, streamflow, sediment transport, diverse habitats, and specific silvicultural practices that in turn affect biophysical dynamics (Price et al., 2013), and all should be considered in the context of climate change. Governance in the CRB is also complex, involving the United States (US) and Canadian governments, distinct indigenous sovereign nations with differing objectives, as well as many state, provincial, local, and other management agencies (Hamlet, 2011). These complex interacting social and ecological systems underscore the necessity to address climate change from an interdisciplinary perspective using spatially explicit approaches (Alessa et al., 2015). Synthesizing information related to climate change in the CRB serves to identify key disciplinary and geographic knowledge gaps while improving access to existing information for policymakers, scientists, and citizens (Pullin and Stewart, 2006). While the results of this approach are specific to the CRB, they may also generate hypotheses for other regions and provide a model for future research.

Research Questions

The general objective of this research is to expand the breadth and depth of existing knowledge by identifying the thematic content and spatial attributes of peer-reviewed research related to climate change in the mountainous regions of the CRB. The specific questions that we address are: (1) What are the common thematic foci and relative deficiencies in this body of research? (2) What are the spatial scales and distribution of climate change research in the headwater regions of the CRB? (3) Is the thematic content of research clustered spatially or conducted at specific scales in a way that suggests a need for further study of particular topics in specific places? The primary outcome of this work is the elucidation of knowledge gaps in areas of scientific inquiry that are strategically beneficial to improving our understanding of changing mountain landscapes. These outcomes are accomplished with a systematic review of peer-refereed literature to improve the potential for identifying research needs and untapped opportunities of greatest potential benefit. By extension, this improves the potential for the co-production of actionable science and management-relevant science, and facilitates a more tailored "call and response" relationship among science producers and science consumers or decision makers (DeCrappeo et al., 2018).

Methods

Literature acquisition

Our literature acquisition methods identified studies that (1) are in the Columbia River Basin, (2) specifically address anthropogenic climate change impacts, adaptation, or mitigation, and (3) address mountainous environments (Figure 1). We used a multi-database search, incorporating literature from the Web of Science, Cabdirect, Proquest, and Crossref databases (see Supplemental Text 1 for specific search terms). We assessed each of the articles for inclusion in the corpus of literature based on their titles and abstracts, and used full texts when necessary. Articles were included if they were peer-reviewed and included a substantial focus on climate change impacts, mitigation, or adaptation in mountain regions of the CRB. Articles were excluded if they did not address climate change, studied paleoclimate, or were conducted at a spatial extent greater than the western United States (Figure 1).

Literature content analysis

Each article was analyzed to determine its spatial extent, location, and thematic content. We used a Google-form software questionnaire and a detailed codebook to ensure consistency among reviewers (see Supplemental Text 2 for codebook). To record location, we selected the US Geological Survey six-digit hydrologic unit codes (HUC-6) to identify the watershed(s) where each study took place (Figure S1). If a study included data from fewer than six individual locations, the latitude(s) and longitude(s) were recorded. Spatial extent, which we defined as the largest area to which findings were extrapolated within the western United States and British Columbia, was selected from seven classifications. We also selected the biome(s) where each study took place from a list of global biomes from Woodward et al. (2004). Freshwater biomes were added to distinguish studies between aquatic and terrestrial biomes.

We developed several categories to analyze the topical and disciplinary content of the research. Using definitions from the Intergovernmental Panel on Climate Change, studies were categorized based on whether the primary knowledge contribution of each article was related to climate change impacts, adaptations, or mitigation (IPCC, 2007; Table 1). If the article addressed impacts, we determined whether evidence was presented regarding observed historic impacts and/or modeled projected future impacts. Finally, we specified the primary discipline(s) and topics addressed in each article (Table 2). Discipline was determined based on the article and journal titles, primary author's discipline, and the the primary knowledge contribution of the article, while topics were selected more inclusively and included any important knowledge contribution. Topics that occurred extremely infrequently were binned into more inclusive categories.

Data analysis

Summary statistics were calculated to summarize frequencies for each of the content categories. To assess interdisciplinarity, we calculated the frequency of disciplinary co-occurrence to derive a network map. To explore the relationships among topics we conducted a hierarchical cluster analysis (HCA), using topics that occurred in at least five articles. We used Ward's least square error method of clustering because it is less susceptible to noise and outliers, and it yielded the highest agglomerative coefficient (Tan, 2007). This method groups topics into similar nested clusters and minimizes the similarity between clusters. Topics that co-occur more frequently are joined early in the clustering process. Inclusive clusters are joined together by branches in a dendrogram.

The relationships between different coding categories were assessed using correspondence analysis. This technique calculates factor scores for two categorical variables and converts them to Euclidean distances, which can be mapped together to visualize relationships in two-dimensional space. The spatial proximity between variables indicates the frequency with which they are researched together (Abdi & Williams 2010).

In order to test the strength of our findings regarding the frequency of co-occurrence, we used a text mining analysis on the article abstracts. Abstracts were available for 515 out of our total corpus of 558 studies. Common stop words and words that occurred less than 20 times were removed, and Pearson correlation coefficients for each remaining pair of words were calculated based on the frequency of co-occurrence in each abstract. Correlations are only reported for cases where Pearson's $p < 0.05$. For cases where other analyses suggested that topics were particularly likely or not to co-occur, we used these correlation coefficients as an additional line of evidence to test our results.

To compare studies that occurred only in Canada, the U.S., or spanning the international boundary, we used a Fisher's exact test. This method identifies whether there were significant differences in the topical distributions of national and transboundary studies. The Fisher's exact test was selected because we had small sample sizes. Results from a Chi-squared test were then used to determine which topics contributed to the differences.

Results and Discussion

The remainder of this manuscript is organized as follows: sections 5.1 to 5.5 describe the thematic content of the research, and frequency with which research themes co-occur. Section 5.6 describes the spatial distribution of research, and sections 5.7-5.8 present results on how specific thematic content is distributed spatially, including international comparisons. Section 5.9 describes important assumptions and limitations. Finally, the conclusions highlight our most important results and opportunities for further research.

Research in the CRB includes an abundance of studies on physical and ecological disciplines and topics.

Articles in the corpus are generally focused on physical and ecological disciplines. The most commonly identified disciplines are ecology (204 articles), hydrology (160), climatology (120), and forestry (108), as shown in Figure 2. We found 156 (28%) articles with two or more disciplines and 402 single-discipline articles (72%). The most common combinations of disciplines are hydrology and climatology (39), and ecology and forestry (24) (Figure 2a); however, these disciplines are closely related and hence do not represent integration across truly disparate disciplines.

We identified an average of 6.12 (±2.5 s.d.) topics per article. The six most common topics are temperature (86% of articles), precipitation (76%), forest ecology (47%), snow (40%), management (40%), and streamflow (37%). The frequency of these topics suggests a dominance of forest ecology and water issues, with fairly frequent discussion of management. The prevalence of management as a topic is important to note, due to the paucity of policy or management as a discipline (8%). This discrepancy arises because our methods were relatively exclusive when coding for discipline and inclusive when coding for topic, and suggests that many studies tend not to have management or policy as a primary focus, but still address management to some extent.

The hierarchical cluster analysis (HCA) illustrates the tendency for topics to be researched together. Physical science topics related to physical hydrology, precipitation, water quantity, streamflow, and snow are clustered together (cluster 1, Figure 3). The appearance of these topics in the first cluster demonstrates that hydrological topics are common in the corpus and confirms that they are consequential in relation to climate change in mountainous regions. The word correlation analysis of article abstracts provides supporting evidence for the HCA findings. Indeed, words associated with topics within cluster 1 (precipitation, streamflow, and snow) are positively correlated.

Some disciplines and topics appear to be infrequently researched together, suggesting an opportunity for further disciplinary integration.

Several lines of evidence indicate that some disciplines and topics are relatively infrequently researched in conjunction with each other. These include the frequency of disciplinary co-occurrence (Figure 2), the hierarchical cluster analysis (Figure 3), and correlational analysis of abstract texts.

One area of research where deeper disciplinary integration may be needed is the relationship between terrestrial and aquatic processes. For example, the disciplines of hydrology and forestry show a fairly strong negative correlation. In the HCA, topics related to forest ecology and water resources form two distinct clusters in branches two and three, also suggesting separation between these topics. The abstract text analysis also supports the idea that forest and aquatic issues are not well integrated; for example, word pairs with negative correlations include forest/fish and fire/fish. Of the minority of articles that do integrate topics related to forests, fires, and fish, five out of seven model the additive effects of climate change, altered forest vegetation, wildfire, and/or other disturbances on aquatic habitat (Davis et al., 2013), stream temperatures (Holsinger et al., 2013; Isaak et al., 2010) or sediment delivery (Neupane and Yager, 2013; Rugenski et al., 2014). All five articles conclude that that combined effects of climate change and forest disturbances are detrimental to aquatic habitat. The other two articles about fish, fire and forests do not directly investigate these topics, but instead consider their confounding influence on stream diversions (Walters et al., 2013) or as determining indicators of climate change (Klos et al., 2015). These studies reinforce the interconnection of forests, fires, and stream habitat and highlight both the necessity and further opportunities to integrate forest disturbances into climate change research on aquatic habitat.

Similarly, studies of fire and snow do not tend to be well integrated, as demonstrated by their distinct clusters in the HCA. The terms fire and snow are also negatively correlated in the abstract text analysis. The topic of snow appears in 42% (236) of the corpus studies, while the topic of fire appears in 20% (113) of articles. However, articles including both snow and fire make up only 5% (27) of the total. Given that snowpack and summer moisture deficit are thought to be leading causes of increases in large wildfire occurrence (Westerling, 2006; 2016), this may indicate an area where further thematic integration is needed. Several combined snow-fire studies address climate change impacts on fire severity or frequency, while integrating snow as an explanatory variable or discussing the importance of snowpack (e.g., Littell et al., 2010; Morgan et al., 2008; O'Leary et al., 2016). Several broader analyses appear in the corpus that address the impacts of changes in a broad suite of environmental variables, including both snow and fire (e.g., Brown et al., 2006; Holsinger et al., 2014), but only one study analyzes the impacts of fire on snowpack dynamics (Gleason et al., 2013). This suggests that there is an opportunity for more detailed analyses regarding potential fire-snow feedbacks in the context of rapidly changing climate.

Our findings also suggest that biophysical disciplines are generally not studied in conjunction with social science disciplines. Community resilience and attitudes and beliefs are separated from all other clusters in the HCA, indicating that they are more frequently discussed within the same publications than they are with other topics (Figure 3). This also appears to be true in the analysis of disciplinary co-occurrence. Of the five most commonly studied disciplines, none show positive correlations with social science disciplines, such as sociology, policy, or economics. However, the number of studies linking these pairs of disciplines suggests that there is at least some research linking these subjects. Many studies link biophysical subjects and policy issues; these include several studies of water resources engineering and supply management issues (e.g., Lee et al., 2009; Hatcher and Jones, 2013). Deeper integration across disciplines is very rare; only five studies represent sociology or policy in conjunction with biophysical disciplines. For example, these include agent-based modeling for planning around future watershed conditions (Nolin, 2012), a synthesis of biophysical climate change indicators and feedback from resource managers (Klos et al., 2015), and an analysis of forest managers' responses to climate change (Blades et al., 2016). These findings are generally in agreement with Bjurström and Polk (2011), who analyzed interdisciplinarity within climate change research through a co-citation analysis of the IPCC Third Assessment report and found that closely related disciplines commonly co-occurred, while more disparate disciplines were clearly separated.

The interaction of scientific and legal institutions influences research conducted in the CRB. Some biophysical topical areas indicate strong connections with policy and management, while these connections are weaker among other topics. Forest ecology and policy and management are closely affiliated, as evidenced by the abstract correlation and cluster analyses (Figure 3). The second branch of the HCA includes many forest ecology topics, as well as policy and management. Topics within the cluster are positively correlated in abstract texts, as well. For example, management is positively correlated with: policy, fire, timber, forest, and ecological, confirming that management studies often focus on forest systems. Many of the aforementioned forestry terms also correlate highly with policy, suggesting that policy is frequently related to forest systems. There are 21 articles in our corpus from the United States that refer to forest ecology and/or silviculture as well as policy; of those, 81% (17) state that they are motivated by various policies related to forests, fires, and wildlife management, such as the Wilderness Act, National Fire Plan, National Environmental Protection Act, or the Northwest Forest Plan.

Fish species, habitat, and restoration commonly co-occur, and an analysis of the articles in our corpus suggests that the Endangered Species Act (ESA) may motivate the coupling of fish and critical habitat restoration. The ESA emphasizes restoration of “critical habitat” for endangered species throughout many river systems in the CRB. In our corpus, 56% (23) of the articles about fish and habitat suggest that the ESA motivates this research; for example, Leibowitz et al. (2014) write, “the threatened and endangered status of many of these stocks under the Endangered Species Act (ESA) often drive water and basin management in the region.” Despite the fact that fish research in the region is often motivated by the ESA, policy is not the main focus of the research. Only 11% (12) of policy articles with the corpus pertain to aquatic habitat, while 25% (133) of the articles relate to aquatic habitat or fish. This may indicate that researchers are focused on habitat restoration rather than new policy changes to reestablish or protect the listed fish species.

The relationships between forestry and policy, and aquatic habitat and policy indicate “the co-production of science and law”—whereby science is needed to support legal action, and the resulting policies, in turn, mandate science to be conducted (Jasanoff, 2004). This can be done in straightforward ways, such as allocating funding to scientific work, which determines the goals and priorities for science. For example, the ESA requires fish and wildlife agencies to develop Biological Opinions (BiOps) that determine the ecological impacts from operation of hydroelectric dams. This relationship between the ESA and scientific research is commonly observed in our corpus. Furthermore, much of the early riparian habitat monitoring in the Pacific Northwest has been carried out within the field of forestry, driven by concerns about the impacts of forest operations and the regulatory framework of the Northwest Forest Plan (Thomas et al., 2006). In these ways and many others, the regulations, institutions, and organizations in place in the CRB are determining factors in the kind of scientific work that is conducted in the basin.

Studies on climate change impacts are much more common than those on adaptation or mitigation. Articles analyzing climate change impacts are much more common than those addressing adaptation or mitigation: 88% (489) primarily focus on climate impacts, while 10% (56) focus on adaptation and 2% (13) are on climate change mitigation. Ford and Pearce (2010) observe an increasing "adaptation gap," where the number of studies addressing climate change impacts is much larger than those addressing mitigation, and the gap between the two has grown over time, particularly as the number of studies on impacts has increased. The studies in our corpus similarly reflect an adaptation gap; comparing the 10-year periods from 1996-2005 and 2005-2015 shows that the gap between the number of adaptation and impacts papers has increased from 63 to 302, though adaptation papers represent a larger portion of the corpus in the later period than earlier, increasing from 3% to 11% of papers. A similar gap exists for mitigation studies.

Studies primarily assessing climate change impacts, adaptation, and mitigation have distinctly different patterns of disciplinary and topical distributions (Figure 4). Articles on climate change impacts tend to be associated with the disciplines of hydrology, climatology, and ecology, and are relatively evenly distributed among the top 20 most common topics. In contrast, studies of climate change adaptation are most commonly associated with the disciplinary categories of policy, forestry, biology, ecology, and sociology. The topics represented by adaptation articles are heavily skewed towards water quantity, silviculture, species range shifts, attitudes and beliefs, and pests and disease. A relatively small percentage of adaptation articles address groundwater (9%), climate oscillations (2%), or carbon cycling (4%); no adaptation articles studied glaciers. The relative lack of adaptation studies on these topics may suggest important knowledge gaps and hence opportunities for adaptation research.

Mitigation studies are disciplinarily concentrated in biology, ecology and forestry, and topically focused on carbon cycling, forest ecology, wildfire, silviculture, and management. These findings reflect established understanding that forest management and wildfire are large components of carbon budgets in mountainous regions (Schimel et al., 2002). However, this also suggests potential research needs. For example, freshwater and soil respiration impacts on carbon budgets appear to be poorly represented in our corpus, despite their demonstrated importance (Cole et al., 2007; Falk et al., 2005). Only two studies in the corpus address climate change mitigation and soils (Wilson et al., 2013; Jauss et al., 2015); these are both focused on forested environments. Aside from forest management, human activities that affect carbon emissions appear to be under studied. Examples include recreational activities, carbon footprint analyses of mountain communities, and carbon emissions impacts of montane hydropower operations (Deemer et al., 2016). While policy research is needed to identify effective means for reducing carbon emissions (Klein et al., 2005), few of the mitigation articles in our corpus explicitly address policy. Instead, we identify several topic areas related to mitigation that could benefit from integrating policy analysis. Specific examples from the corpus include a study estimating the potential effects of prescribed burning on carbon emissions (Wiedenmeyer and Hurteau, 2010), and a quantification of carbon stored in wood products (Stockmann et al., 2012). Both prescribed burning and carbon stored in wood products are identified as complex policy issues related to reducing carbon emissions in mountainous regions of the CRB (Law et al., 2018).

Climate change impacts, adaptation, and mitigation are also studied at different spatial extents (Figure 5). Correspondence analysis indicates that the first dimension is driven by impacts and mitigation, and explains 64% of the variability, while the second dimension is mostly driven by adaptation, and explains 25% of the variability in the dataset. This analysis also demonstrates the relationship between the type of impact studied and spatial extent. Climate change implications are most closely clustered with the smallest scale in our study. Mitigation is associated with relatively small scales, while observed and projected impacts are associated with relatively large scales. Most mitigation studies provide analyses of forest carbon cycles; the small spatial extent suggests that this information is often process-oriented at specific sites and is typically not upscaled to the landscape level. In contrast, adaptation studies tend to fall within the medium to large extents.

Research has predominantly focused at relatively large scales, made projections of future rather than observed conditions, and used existing rather than new data.

Articles included in the corpus ranged in spatial extent from point or plot scale to the western U.S. The Pacific Northwest (660,000km^2) and the Western U.S. extents are the most common and include 37% of articles (205). Another 22% of articles (121) span between 40,000km^2 and the Pacific Northwest (660,000km^2). The remaining 42% of articles (232) report on studies at spatial extents less than 40,000km^2. Different disciplines are generally associated with different spatial extents (Figure 6). For example, articles with climatology as a discipline tend to occur more frequently at larger extents. This is to be expected, given the nature of the discipline, though it may raise questions about whether microclimates and refugia are adequately studied from a climatological perspective (e.g. Curtis et al., 2014). Furthermore, the lack of small scale climate studies suggests that there may be a lack of knowledge about regional climate processes (Salathé et al., 2008), changes in microclimates (Daly et al., 2010), and rapid changes (Wiens, 1989). For example, rapid changes in vegetation, especially in ecotones, result from regional climatic changes which are often undetectable at larger scales (Allen & Breshears, 1998; Kelly & Goulden, 2008).

Projections of climate change impacts are more common than observations. Of the 507 articles that study climate change impacts, 28% (139) observe an environmental trend and discuss its attribution to climate change, while 35% (171) make formal projections of climate change impacts, and 42% (205) assess a climate change impact but do not explicitly observe or project a trend. Reporting on new field data is also relatively uncommon; only 34% (188) of studies include new data. The frequency with which studies include observed or projected impacts vary by discipline (Figure 7). Articles with disciplines categorized as ecology, forestry, biology, policy, or geology include more cases where implications are studied than observations or projections. In contrast, hydrology and climatology have more studies of projected and observed climate change impacts. For most disciplines, excluding forestry and sociology, studies of projected impacts are more common than studies of observed impacts.

To a certain extent, the predominance of studies about projected impacts relative to observational impacts is expected: observed impacts require decades of data to establish, and these long-term in *situ* observations are often unavailable in many locations (Strachan et al., 2016). In some cases, climate change impacts remain difficult to detect, given the range of internal variability (Hegerl et al., 2006).

While climate change implications studies may sufficiently provide the information needed to make projections, they may exclude important or unexpected changes that are only identifiable with long term observations (Hegerl et al., 2006). For example, long-term monitoring in other mountainous regions has identified paradoxical relationships between warming and frost damage of flowering plants (e.g., Inouye, 2008); these findings would not have been possible without long-term observations and may inform predictive modeling. This finding indicates the importance of long-term data collection and continued monitoring of environmental changes to assess the observable impacts of climate change across a range of disciplines and scales, and develop adaptation strategies to enhance the resilience of natural systems within the context of a non-stationary climate.

The frequency of projected rather than observed studies, large spatial extents, and relative infrequency with which new data is collected for research also reflects the increasing use of computer modeling. Computer modeling has become critical in scientific work aimed at understanding large-scale, climatic change (Edwards, 2010). The relative preponderance of studies at fairly large scales raises questions about whether these large-scale findings are well-supported by field data, which is usually collected at much smaller scales (e.g. McKelvey et al., 2011). The frequency of studies without field data reflects larger trends in scientific work, as understanding global environmental change increasingly relies on distributed and modeled datasets (Edwards, 2010). Further, modeling is increasingly employed over field-based studies in order to meet the challenge of predicting global change and managing uncertainty (Mauz & Ganjou, 2013). These trends also indicate a movement towards the use of “big data,” which can create challenges as it disrupts old knowledge structures and methods, but also creates opportunities for novel forms of interdisciplinarity and collaboration (Plantin et al., 2017).

The quantity of research conducted varies spatially, and is influenced by institutions, geographical features, and disturbance history.

Research is unevenly spatially distributed across the CRB (Figure 8). The quantity of research we identified is much smaller in Canada (84) than in the U.S. (405). For studies conducted at smaller extents, research activities are concentrated at several locations that appear to be fairly well explained by geographical features, such as the location of long-term research sites. For example, notable concentrations of research appear to occur at the H.J. Andrews Experimental Forest, Mount Rainier National Park, and in the Reynolds Creek Experimental Watershed. Another relatively high concentration of studies occurs in the Okanagan Basin, Canada, though these are not clustered at a particular research site.

This spatial distribution of research points to the “distributional consequences” stemming from the knowledge infrastructure in the CRB, as well as the importance of considering both physical and sociotechnical aspects of research (Edwards et al., 2013). Large-scale investments in data-intensive knowledge infrastructures can have lasting effects on the type of science conducted, as data is made re-usable by other scientists (Bowker, 2000), and long-term research sites become a focus for intensive study. Research infrastructure includes more than the material aspects of technology that enable science to be conducted. The organizational and relational aspects of scientific work such as protocols, standards, and systems of field-gathered and remotely-sensed data are also important (Star & Ruhleder, 1994). Moreover, the particular histories of land use and management policy can affect the distribution of research; for example, one content analysis focused on treeline found that land use designations, such as National Parks, affected the type of treeline research conducted (Whitesides and Butler, 2011). Multiple aspects of research and legal infrastructure have legacy effects on the production of science in a particular location such as the CRB.

Biophysical context influences the spatial distribution of the thematic content of research.

The thematic content of research is unevenly distributed across HUC-6 watersheds. Correspondence analysis reveals groupings of watersheds and disciplines (Figure 9). The first dimension accounts for 36.2% of the variability and the second dimension accounts for 18.5% of the variability. Research in the Upper Snake and Snake Headwaters tends to encompass the same disciplines and is closely associated with policy and ecology. Sociology is frequently coupled with the Okanagan (Canada), Columbia (Canada), and Spokane watersheds, with sociology studies in Canada commonly focused on social issues shaping forest management (Goemans & Ballamingie., 2013; Furness & Nelson, 2015; Carolan & Stuart, 2016). Hydrology is also associated with Okanagan (Canada), Columbia (Canada), Spokane, Yakima, and John Day watersheds. Forestry is closely coupled with the Willamette, Kootenai, and Upper Columbia River watersheds, though the topic’s central location within the correspondence analysis graph indicates that it is researched frequently within most watersheds. Maps of the spatial distribution of selected topics support the correspondence analysis and demonstrate that the topical distribution of research varies in space (Figure 10). For many topics, the variability between the U.S. and Canada is much larger than within-country differences; however, we focus our discussion here on within-country differences followed by discussion of transboundary differences.

Disturbance history influences the the topical distribution of research. For example, the preponderance of forest ecology and wildfire studies in the Greater Yellowstone Ecosystem may be due to the 1988 Yellowstone Fires, as evident in the many studies that reference these fires (e.g., Romme et al., 2011; Donato et al., 2016; Seidl et al., 2016; Zhao et al., 2016). Studies of pests and disease are also relatively common in the Greater Yellowstone Ecosystem, as well as the Salmon River watershed (Figure 10). Many of these studies are focused on pine bark beetle outbreaks (e.g. Buotte et al., 2016; Logan, MacFarlane, & Willcox, 2010; Seidl, Donato, Raffa, & Turner, 2016; Simard, Powell, Raffa, & Turner, 2012). A qualitative comparison with a remote sensing analysis of bark-beetle induced tree mortality suggests that there are relative hotspots of bark beetle outbreaks within this region, particularly in the Salmon River watershed (Hicke et al., 2016). However, Hicke et al. (2016) also identify relatively high beetle mortality in parts of the North Cascades. In our data, the North Cascades do not appear as a hotspot for beetle studies, suggesting that the distribution of research is only partially explained by disturbance history.

The Lower Snake and Yakima watersheds have the highest fractions of research pertaining to streamflow. Two long-term research sites are located in the Lower Snake (Reynolds Creek Experimental Watershed and Dry Creek Experimental Watershed), while the Yakima watershed is an important region for irrigated agriculture (e.g., Vano et al., 2010), which may contribute to the prevalence of streamflow research. While streamflow in much of the region is snowmelt-driven, the spatial distribution of research related to snow is slightly different from that on streamflow. Similar to streamflow, snow-related research is also common in the Yakima watershed; however, snow studies in the Snake River headwaters are relatively lacking.

There is a relatively high frequency of research that addresses management implications within the Upper Snake and Snake Headwaters. Articles addressing management in this area predominantly focus on interactions between water resources management and biophysical conditions under climate change (Loinaz et al., 2014; Qualls et al., 2013; Ryu et al., 2017; Sridhar and Anderson, 2017); forest and terrestrial ecosystem management, often specific to unique species such as whitebark pine (Logan et al., 2010; Macfarlane et al., 2013); or sagebrush steppe communities (West and Yorks, 2016). Interestingly, despite the relative prevalence of management topics in these two watersheds, adaptation studies are about as common as in the entire corpus. This finding suggests that many impacts-focused studies in this region also address management implications, which is perhaps a result of the long history of conservation planning efforts in the Greater Yellowstone Ecosystem (Clark et al., 1991).

Within Canada, management is frequently researched in the Upper Columbia watershed, where Lake Okanagan is located (85% of Canadian policy articles, n=18). Of these management articles, 67% (12) focused on forests (e.g., Nitschki & Inns, 2008; Goemans et al., 2013; Seely et al 2015), 17% (3) on wildlife (Bunnell, Kremsater, & Wells, 2011; Festa-Bianchet, Ray, Boutin, Côté & Gunn, 2011; McNay, Sutherland & Morgan, 2011), 11% (2) are about human dimensions (Turner & Clifton, 2009; Furness & Nelson, 2016), and less than 1% (1) avalanches (Sinickas & Jamieson, 2016). Water management topics are not addressed in this subset of articles even though the topic is critically important due to the high demand for irrigation water in the Okanagan watershed (Neilsen et al., 2006).

There are significant differences in the thematic content of research among studies in Canada, the United States, and transboundary studies.

We compared thematic content of articles exclusively in the U.S., in Canada, and those that are transboundary. The comparison suggests that the topical distributions of articles in these three categories are significantly different from each other (Fisher's exact test $p < 0.001$). The prevalence of articles addressing pests and disease and glaciers in Canada are the largest contributors to this difference, though topics related to human dimensions (policy, management, attitudes and beliefs, community resilience) are also more common in Canada than in the U.S. The extensive forested areas and recent pest outbreaks in the Canadian headwaters of the CRB may explain the greater research focus on forest pests and disease impacts. Climate change contributes to the rapid expansion of new bark beetle species at these latitudes, raising concerns for forest health in Canada (Anderegg et al., 2015; Bentz et al., 2010). The topical focus on glaciers in Canada within the corpus is likely due to the relatively high prevalence and hydrologic importance of glaciers (Moore et al., 2009).

Publications focused on Canadian regions also include more articles with topics relevant to human dimensions of climate change such as policy and management (Figures 9 and 10, e.g., McDaniels et al., 2012; Murdock et al., 2013; Parkins and MacKendrick, 2007). For example, Murdock et al. (2012) report on a bio-economic model intended to inform forest management decisions in a changing climate. Concerns about forest health issues due to the close proximity of communities and forests in Canada may influence the more frequent occurrence of topics related to the human dimensions of climate change (e.g., Furness and Nelson, 2016; Parkins, 2008; Parkins and MacKendrick, 2007). The relatively high frequency of studies addressing social issues in Canada may provide research models that would be beneficial to apply in the U.S. parts of the CRB; studies analyzing transboundary social issues under climate change may also be needed in the region.

Many of the thematic differences between research in the U.S. and Canada relate to biophysical topics and likely emerge from differences in landscape characteristics such as latitude and land cover. However, differences in laws and policy between the two countries may also play a role. The greater U.S focus on topics related to hydrology, aquatic habitat, and wildfire is likely linked to differences in policy and management within the two countries. As discussed in section 3.3, in the U.S., the ESA often motivates research on issues related to restoring federally listed fish habitat (Beechie et al., 2013). Snow-related research is more prevalent in the U.S. than Canada, which may be due to the fact that an important long-term snowpack dataset, SNOTEL, operates within the U.S. only, or could be due to expectations that snowpack in the colder Canadian portions of the CRB is more resilient to warming than in the warmer ranges found in the United States, and that precipitation is likely to increase in this part of the CRB (Hamlet et al., 2013).

Transboundary studies are distinguished by a relatively high frequency of studies addressing climate oscillations, streamflow, anadromous fish, and restoration, and a relatively low frequency of studies on policy, forest disturbances, silviculture, and carbon cycling. These include studies about climate change models across the entire CRB (e.g., Rupp et al., 2016); downscaled impacts of climate change on hydrology such as hydro-climatological models (e.g., Hamlet et al., 2013); comparative streamflow and water temperature modeling (e.g., Ficklin et al., 2014); and models of declining snowpack (e.g., Abatzoglou, 2011). Reconstruction of historical flows or trends are also common across transboundary studies (e.g., Wapples et al., 2008). Only five transboundary studies explicitly address policy and management issues (Sopinka and Pitt, 2014; Beechie et al., 2013; Schwandt et al,.

2010; Lee et al., 2009; Bisson et al., 2009). These studies focus on flood control, streamflow, and anadromous fish issues. A potential issue in interpreting the thematic content of these transboundary studies is that many transboundary studies tend to occur at relatively large scales (70% were larger than the Pacific Northwest, in contrast to only 37% in the full corpus). Therefore, there may be a confounding effect between topics that tend to be researched at large scales and those that are of particular interest across international borders.

Assumptions and limitations

There are several assumptions and limitations that should be considered in the interpretation of our findings. Our methods required that each article was categorized as either adaptation, mitigation, or impacts. Therefore, while there may be studies that address both mitigation and adaptation, these would have been coded in only one of these categories. We have also identified several areas of thematic content for which we argue that two important topics or disciplines are not well integrated. To support these conclusions, we use multiple lines of evidence where possible, but it is important to note that these analytical methods identify research integration that is *relatively* infrequent. We support these with discussion of the few studies that do address these potential gaps, but determining which areas are true and important knowledge gaps, and which are not studied because they are not particularly relevant, is ultimately subjective. Moreover, while we used multiple databases to identify research, there are likely some relevant articles that were omitted. We used multiple coding and multiple lines of evidence to support our conclusions, but as in any such investigation, errors in coding may occur. It is also important to note that our literature search was conducted in December 2016; while there are undoubtedly many new studies available, we expect that the general patterns and trends characterizing the science conducted in this region are not likely to have changed substantially in the intervening time.

Conclusions

Science produced in mountainous headwaters of the CRB affects our understanding of climate change impacts on social and ecological systems, as well as our understanding of potential adaptation and mitigation strategies. Results from this study suggest that climate change research in the CRB focuses on impacts much more frequently than adaptation, while mitigation is rarely a focus. Studies focused on adaptation and mitigation most commonly address forest ecology and carbon cycling. Most of the articles we reviewed assess trends at large extents, rely on secondary data, and make projections of climate change impacts rather than observations or predictions. The spatial distribution of research varies across the U.S./Canada boundary and is influenced by the placement of long-term research centers and to a lesser extent, national parks, as well as legacy disturbance events. Thematic content analysis indicates that studies most commonly focus on physical and life sciences, such as ecology, hydrology, and climatology. Studies that integrate biophysical and social disciplines or terrestrial and aquatic ecosystems are relatively rare. The limited interaction between social and biophysical content reinforces the opportunities for increased interdisciplinary collaborations and suggests that this is needed in areas where social and ecological systems are both tightly coupled and particularly sensitive to climate changes.

Assessing how knowledge about climate change is created and produced in the CRB headwaters highlights areas of under-represented knowledge as well as the management responses that inform the production of science. By assessing existing research, we also highlight gaps and areas of opportunity for future research (Boxes 1, 2). Implementing similar analyses elsewhere would expand understanding of gaps in climate change research and knowledge structures in mountainous regions. Comparing existing research may allow science and management communities to leverage resources more effectively, increasing the potential for the co-production of actionable science and effective responses to climate change.

Sidebars

Box 1. Climate change research and integration

- Closely related disciplines commonly co-occur (e.g., ecology and biology).
- Hydroclimatology is a top focus of climate change research in the CRB.
- Forest ecology and policy and management are frequently studied together.
- Fish species, habitat, and restoration are commonly co-occurring topics.
- Forestry and aquatic ecology research demonstrate the co-production of science and law.
- Adaptation studies frequently focus on forestry issues.
- Articles with climatology as a discipline tend to occur at relatively large scales.
- Climate change impacts are more frequently researched than adaptation or mitigation.
- Studies of projected impacts are more common than studies of observed impacts.
- The spatial distribution of research is partially explained by geographical features.
- The U.S. has more studies of hydrology, aquatic habitat, and wildfire than Canada.
- Canada has a greater focus on pests and disease, glaciers, and social issues.
- Transboundary studies tend to address climate oscillations, streamflow, anadromous fish, and restoration.

Box 2. Knowledge gaps and opportunities for research integration

- Studies of adaptation were relatively infrequent, suggesting a need for more adaptation research.

- Studies of mitigation were extremely infrequent. Given the large forest carbon stocks and propensity for disturbance in actively managed lands, this suggests an important missed opportunity.
- There is a clear need for deeper integration of social science with biophysical research.
- Basin-wide and/or transboundary research collaborations are needed.

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Tables

Table 5.1 Definitions used to assess area of primary knowledge contribution

*2007 Intergovernmental Panel on Climate Change definition

Term	Definition used in study
Adaptation	Adjustment in human systems in response to actual or expected climatic stimuli or their effects, which moderates harm or exploits beneficial opportunities.*
Mitigation	An anthropogenic intervention aimed at reducing the anthropogenic forcing of the climate system.*
Impacts	The effects of climate change on natural and human systems.* We categorized impacts as *observed*, in which trends were noted in empirical data and attribution to climate change was discussed, *projected,* in which the impacts of climate change were quantitatively modeled for future scenarios, and *implications*, in which the climate sensitivity of a system was assessed.

Table 5.2 Definitions of disciplines used in study, listed in alphabetical order

Discipline	Definition used in study
Biology	The study of life, including anatomy, physiology, animal behavior, genetics, morphology, growth, and more.
Climatology	Studies of weather and/or climate, including atmospheric and oceanic patterns and processes.
Ecology	The study of the interaction of biotic and abiotic factors in an ecosystem.
Economics	The study of the production, distribution, and consumption of monetary goods and services.
Engineering	The study of physical design and construction of functional structures.
Forestry	Studies that broadly include forest ecology and forest management.
Geology	The study of earth processes, plus rock & soil science.
Hydrology	The study of water processes, both above and below ground.
Sociology	Any study focused on human populations, human behavior, relationships, culture, and society.
Policy	Any studies related to rulemaking and decision making at an administrative level, including management.
Toxicology	Any branch of chemistry and toxicology that focuses on interactions among biological and chemical processes in the environment.

Figures

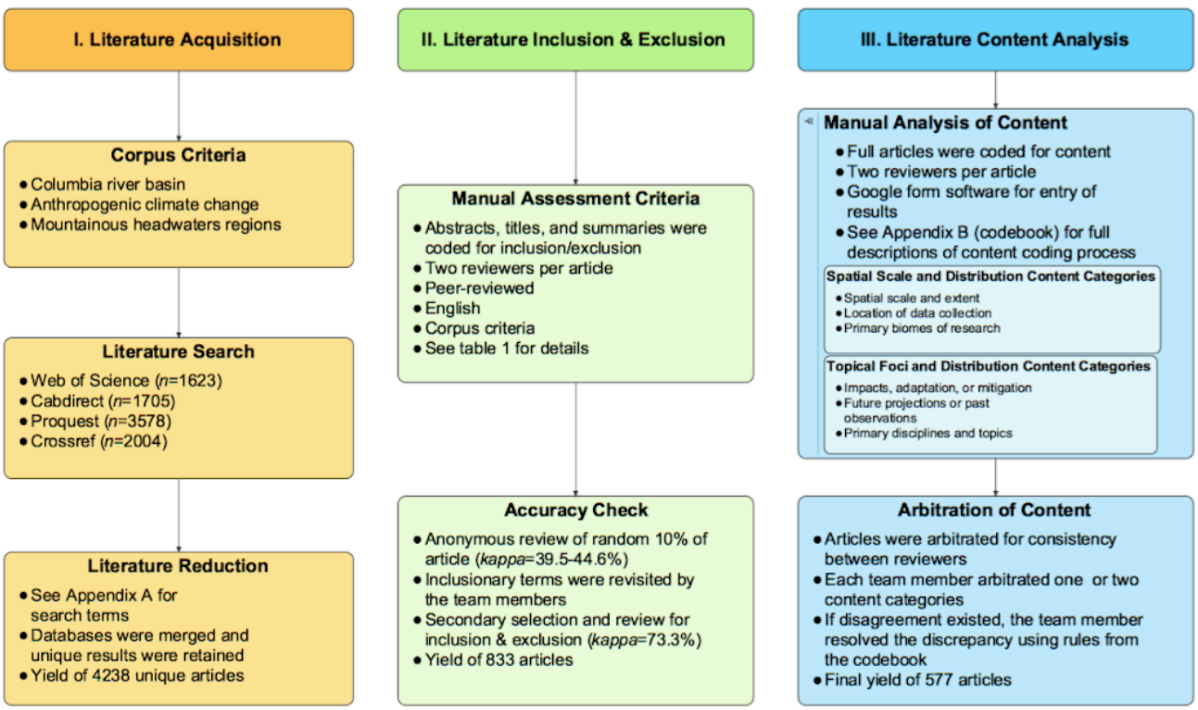

Figure 5.1 Flowchart for methods of literature acquisition, inclusion, exclusion, and content analysis

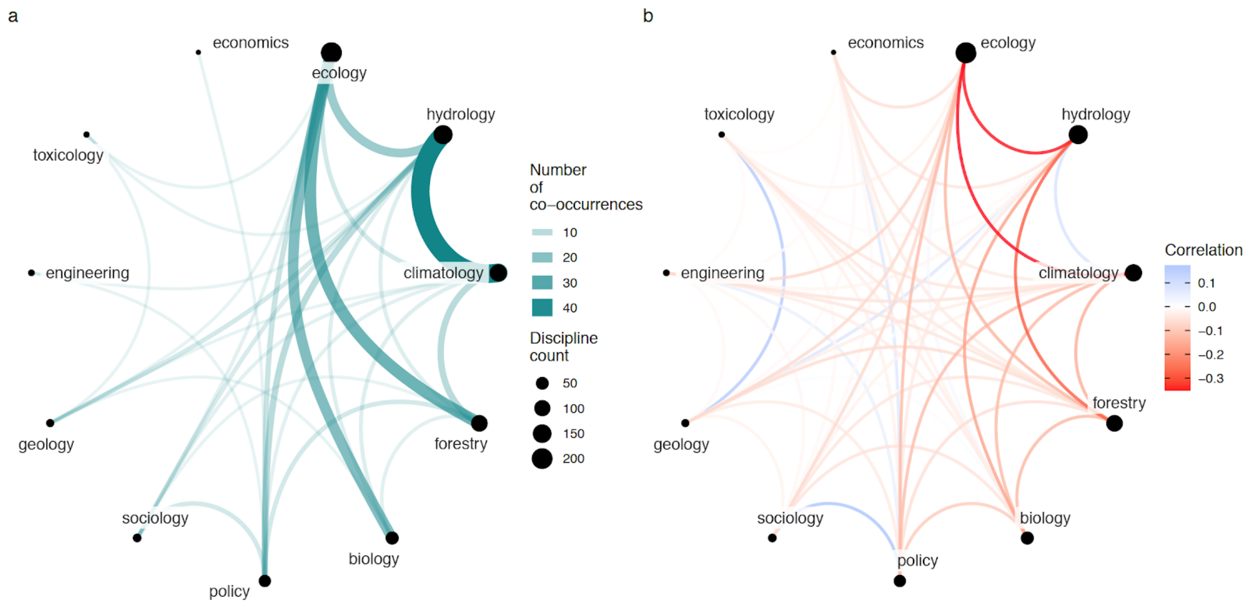

Figure 5.2 Network map of co-occurring disciplines, showing (a) number of co-occurrences, indicated by edge width and color, and (b) correlation coefficients between disciplines. Size of points indicates number of times each discipline occurred.

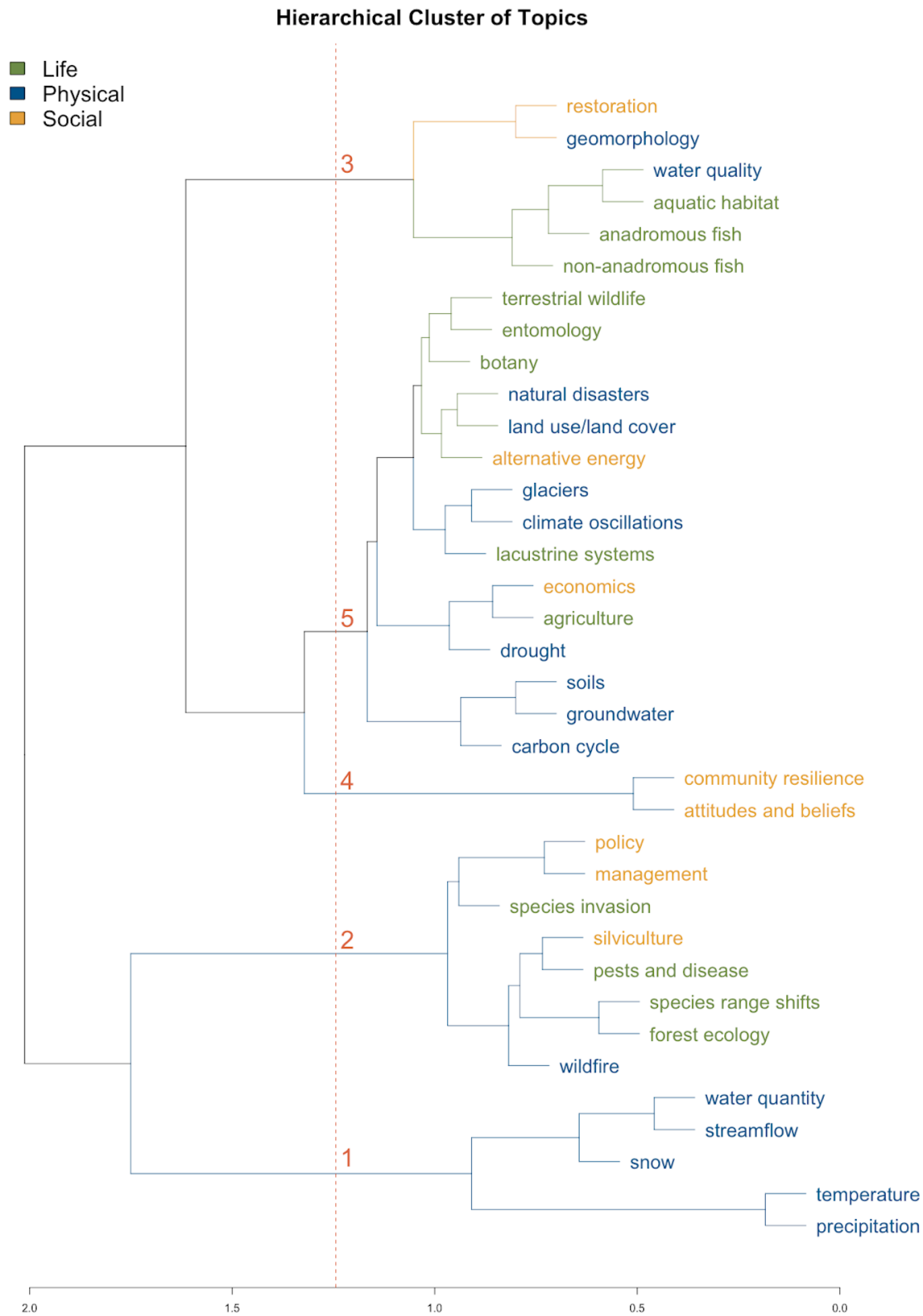

Figure 5.3 Dendrogram of hierarchical cluster analysis (HCA) of topical co-occurrences. The HCA measures the dissimilarity between variables and represents them in nested clusters. The x-axis shows the dissimilarity between topics. Topics that are grouped together near the right (distance = 0) are frequently coupled in the literature. Cluster numbers in red are referenced in the text. Colors of topics indicate whether each topic was classified as primarily related to the social (yellow), life (green), or physical (blue) sciences.

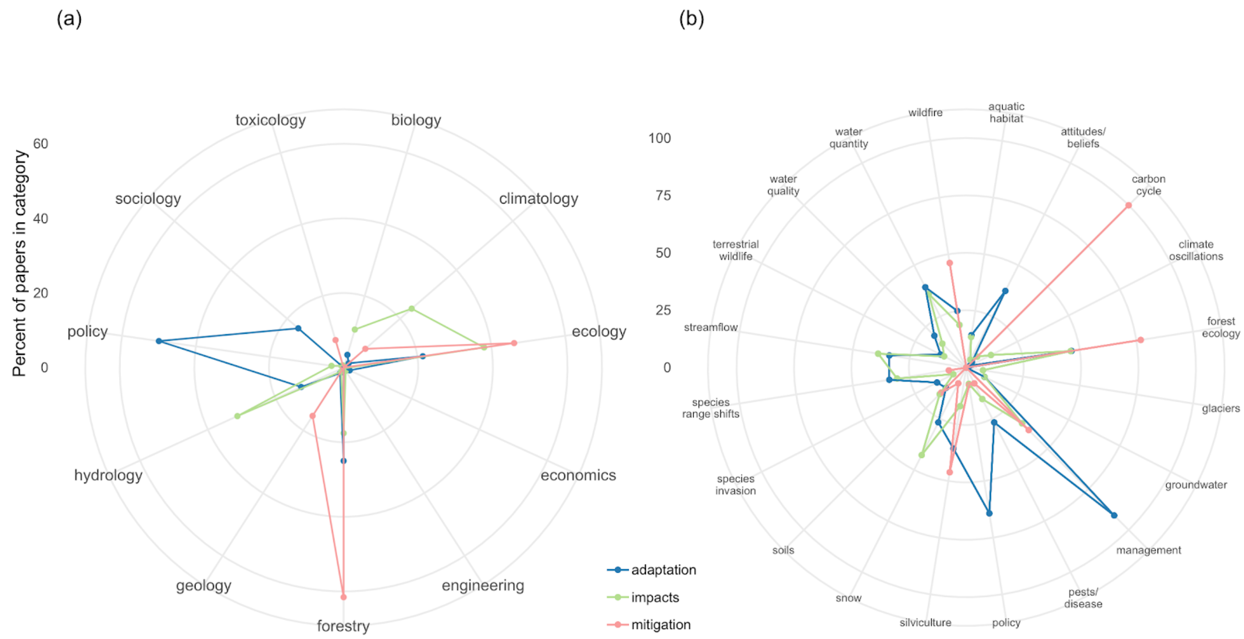

Figure 5.4 Radar plots showing the distribution of adaptation, impacts, and mitigation paper by (a) discipline and (b) topic. Axis displays the percent of papers in the adaptation, mitigation, and impacts categories that address a particular topic or discipline. Figure S2 shows numbers of papers, rather than percentages.

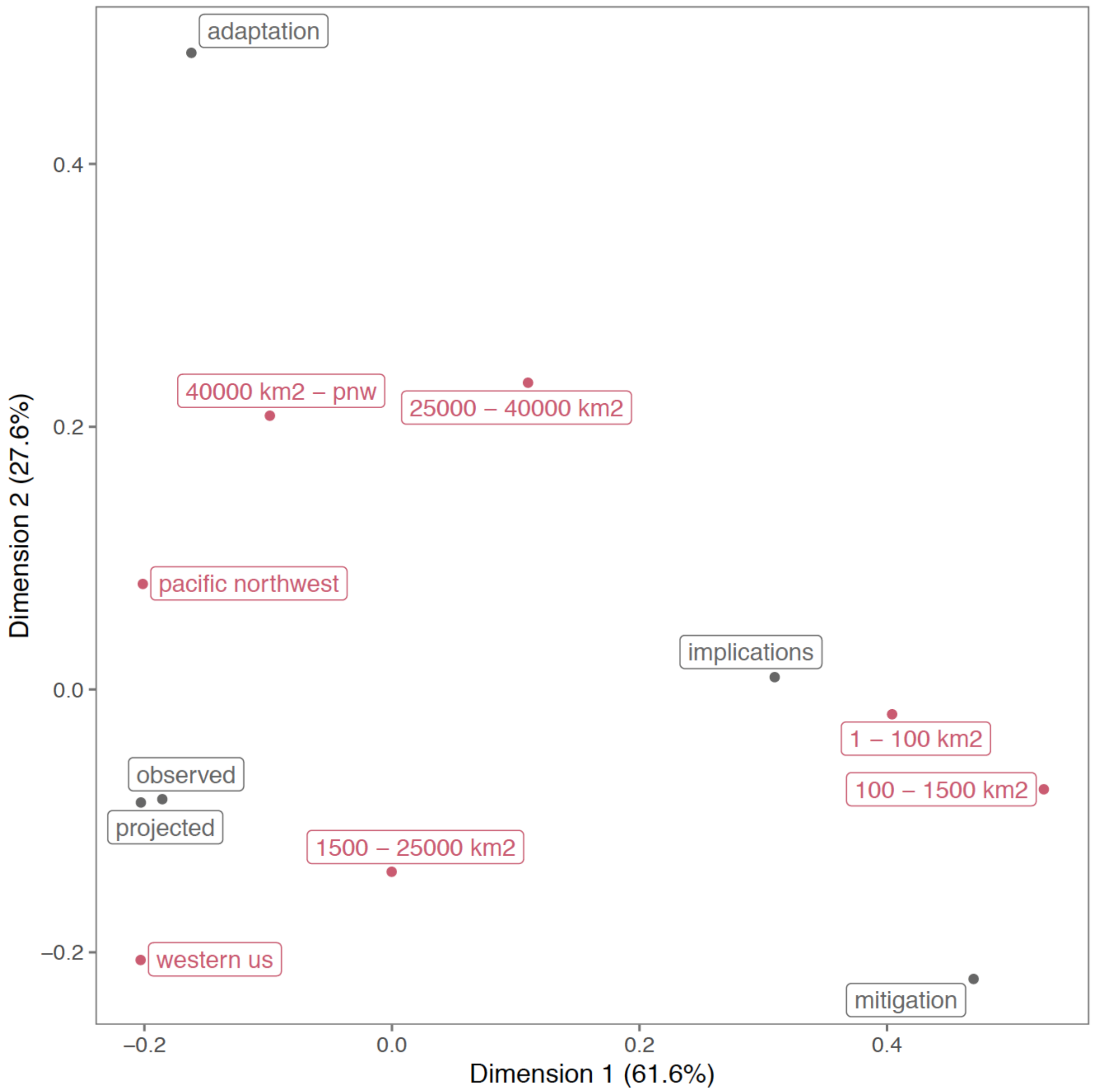

Figure 5.5 Biplot of correspondence analysis of impacts (observed, projected, or implications), adaptation, and mitigation (black labels) vs. spatial extents (red labels). variables that are close in Euclidean space are frequently coupled in the literature.

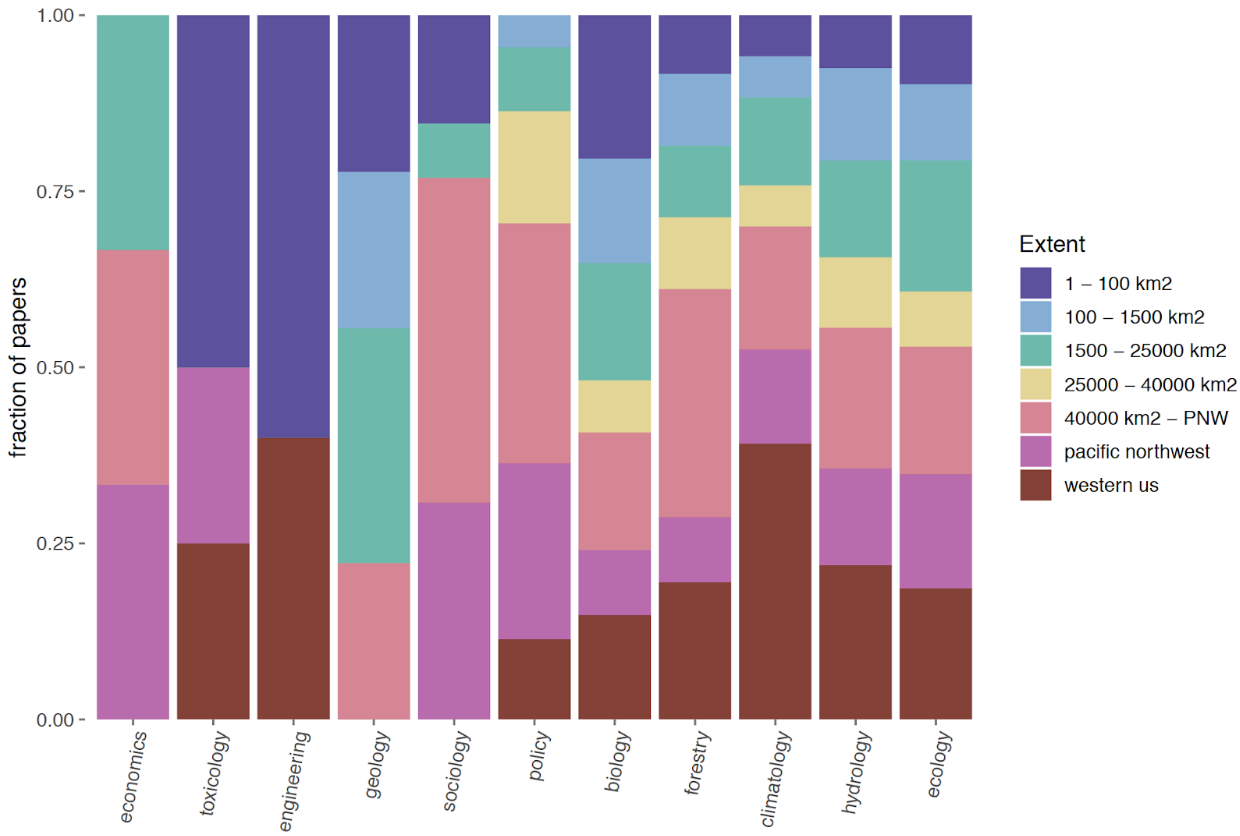

Figure 5.6 Spatial extent of disciplines. Disciplines are arranged in ascending order of frequency within the dataset.

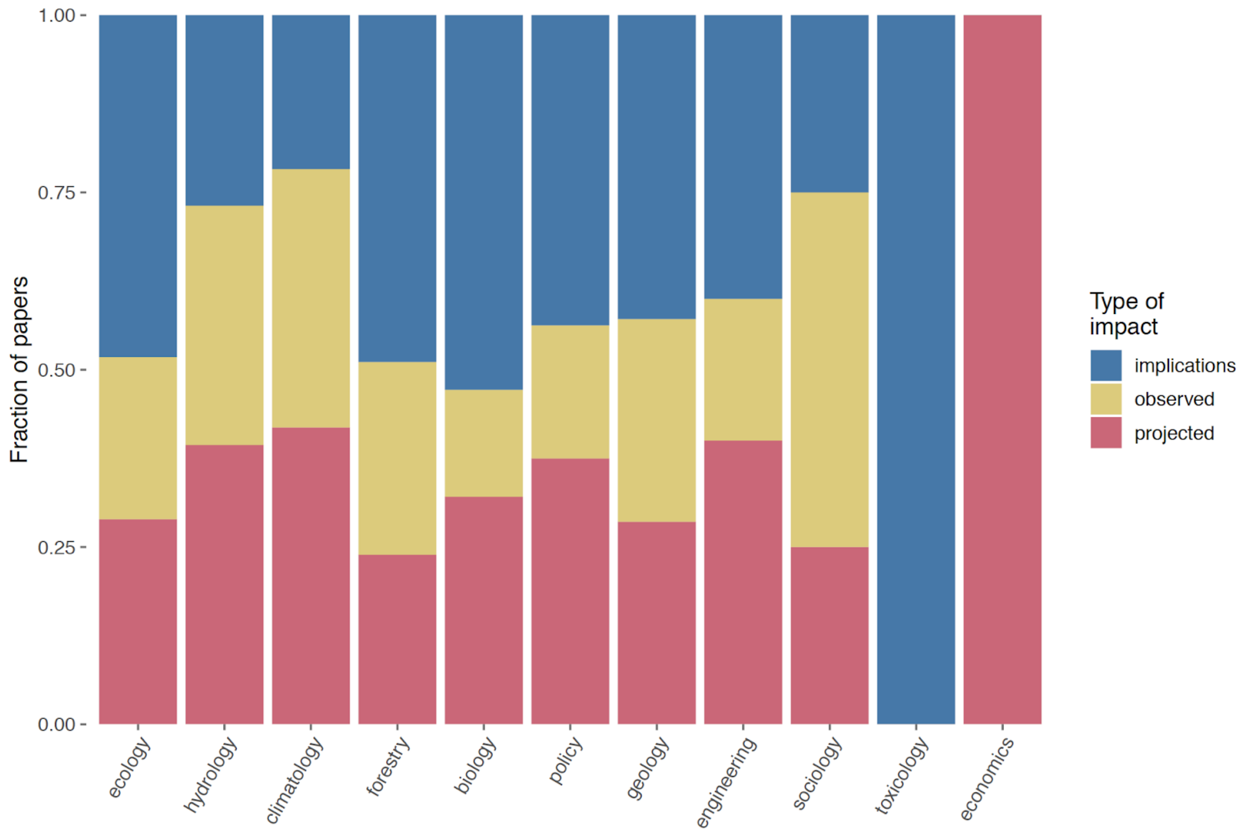

Figure 5.7 Studies of climate change impacts that identify climate change implications or observed or projected impacts, by discipline.

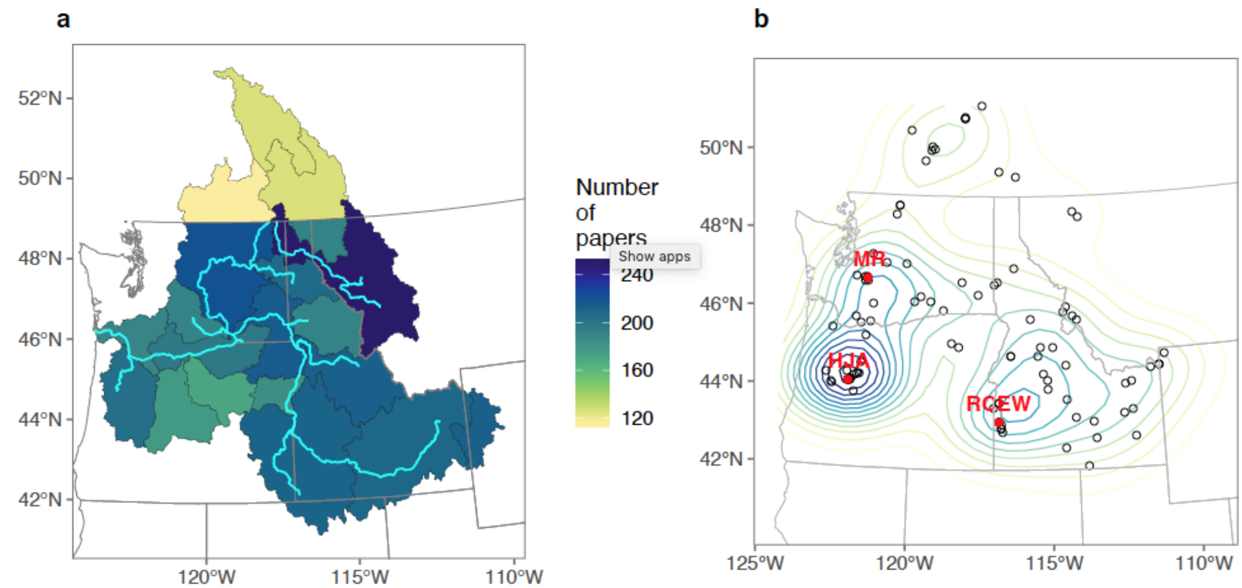

Figure 5.8 Spatial distribution of literature, displayed as (a) total number of papers per HUC-6 watershed and (b) point locations for studies with spatial extents less than 1500 km^2, with contours showing estimated density of studies. Rivers are displayed in cyan; points of interest with high concentrations of research are in red. MR = Mount Rainier; HJA = H.J. Andrews Experimental Forest; RCEW = Reynolds Creek Experimental Watershed.

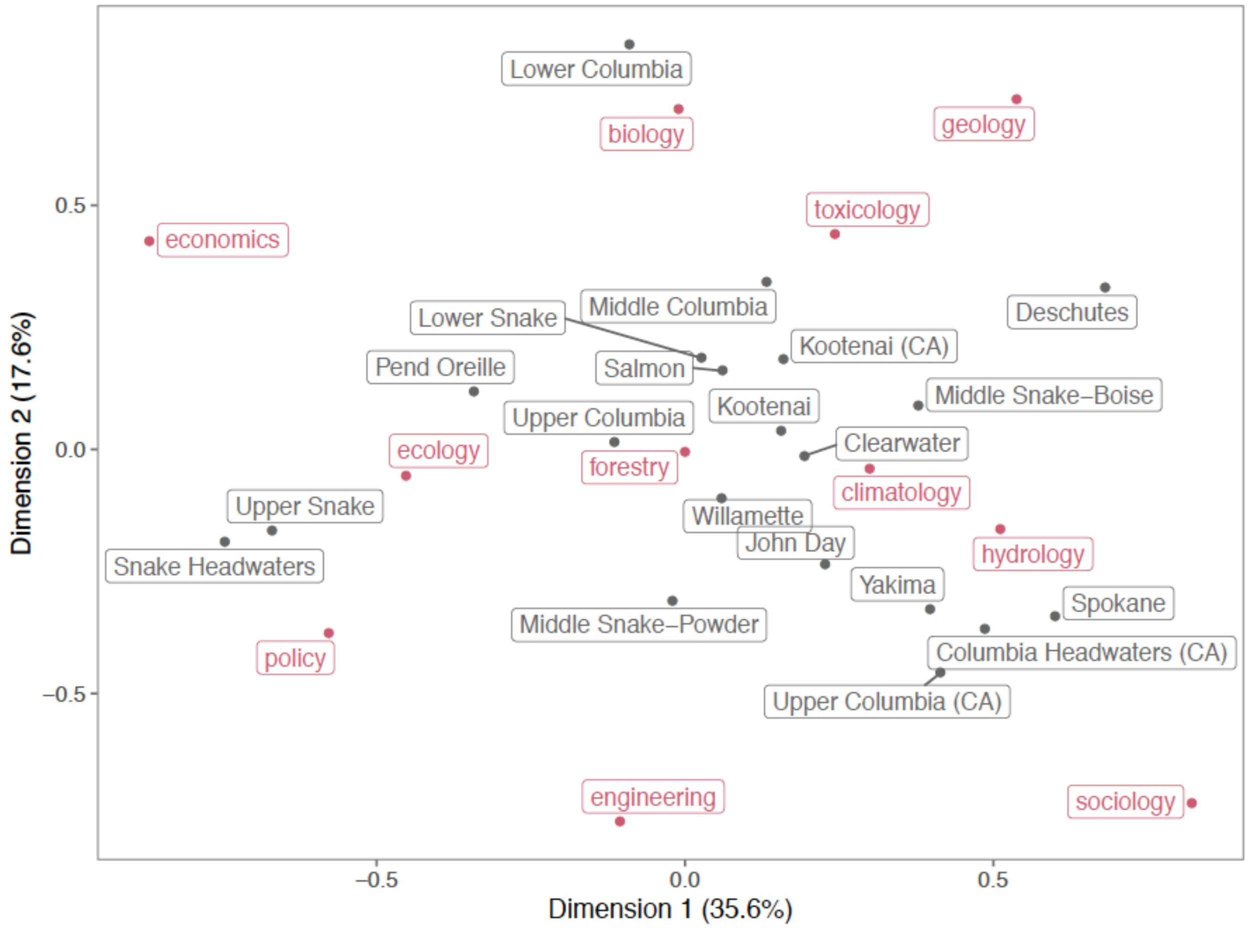

Figure 5.9 Biplot of correspondence analysis of watersheds (black labels) and disciplines (red labels). When variables appear close in Euclidean space, they are frequently coupled in the literature.

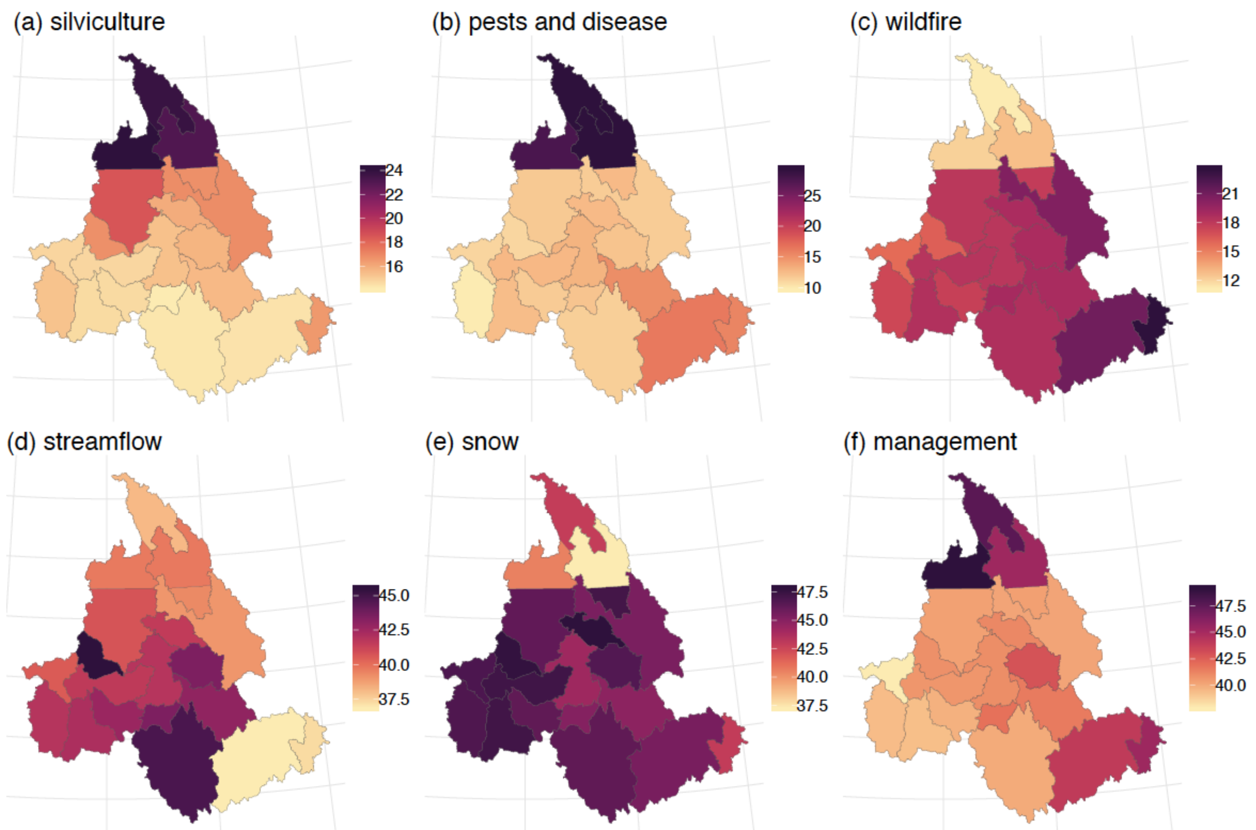

Figure 5.10 Spatial distribution of selected topics by HUC. Each legend shows the percent of papers in a given HUC that addresses the topic.

Chapter 6: Conclusion

The three disciplinary chapters included in this dissertation each identified and analyzed different aspects of changing snowpack heterogeneity due to anthropogenic climate change, while the fourth chapter synthesized the body of research addressing climate change in the Columbia River Basin. The findings of each of these studies could be extended through further analyses in order to enhance our understanding of climate change impacts on water resources and ecosystems and enable novel approaches to adaptation.

The first chapter assessed the impacts of altered temperature and precipitation on drifting snow in a sagebrush-steppe environment, where snow drifts subsidize aspen stands not otherwise found on the landscape. While the study provided detailed analysis at a specific location, there is a clear need for information that would enable scaling of these results. Further analyses could identify the prevalence of wind-driven redistribution at larger scales, as well as cases where aspen specifically are subsidized by wind-driven redistribution. These could usefully be developed through remote sensing analyses, building on methods developed by Wayand et al. (2018) and paired with remotely sensed vegetation distribution information to identify aspen stands. Such remote sensing analyses of wind redistribution of snow could be paired with climate data to develop statistical models to better inform our understanding of how snow redistribution may be altered by climate change. One potential concern is that it may be difficult to distinguish wind-driven redistribution from other mechanisms that increase snow persistence on the landscape, such as avalanching or shortwave radiation inputs, though it may be possible to control for these factors with terrain analyses. These analyses could also usefully be paired with spatially distributed physically-based modeling using models such as the Cold Regions Hydrological Model (Pomeroy et al., 2007) or SnowModel (Liston and Elder, 2006). This would allow for improved understanding of how generalizable the conclusions from Chapter 1 are across larger scales.

In the second chapter, I assessed changes in interannual variability of snowpack across the Western U.S. The results from this chapter suggested three major directions for potential future work: issues at smaller scales, changes in spatial variability, and formalization of the importance of snowpack heterogeneity for water resources and ecosystems. At smaller scales, a different suite of processes may be relevant than those responsible for the patterns observed at relatively large scales in Chapter

Two. For example, the wind redistribution studied in Chapter One, as well as topographic shading, avalanching, and influences of vegetation on snow accumulation and ablation may be more important at meter- to-hillslope scales (Clark et al., 2011), may operate in ways that affect changes in interannual variability at smaller scales. The larger scales studied in Chapter Two are likely more important for water resources outcomes, but smaller scale processes may be important for ecosystem function. These relatively small scale processes, as well as the climatic and topographic gradients responsible for the patterns identified in Chapter Two, may also have important effects on changes in spatial variability, which may in turn be important for basin-average changes in snowmelt, snowpack feedbacks to the atmosphere via radiative and turbulent fluxes (Essery, 1997; Liston, 1995), and landscape ecology via, for example, spatiotemporally heterogeneous availability of plant species for pollinators. Recent modeling efforts suggest increased spatial synchrony in spring phenology across the northern hemisphere (Liu et al., 2019), but the role of spatial variability of snowpack in these effects was not explicitly evaluated, despite strong evidence that snowmelt timing is a major contributor to spring phenology (Dunne et al., 2003). The reduced snow accumulation projected in much of the western U.S. may result in decreased spatial variability of accumulation and melt timing at small scales; at the basin scales that primarily affect water resources provisioning, changes in spatial variability of snow accumulation and melt may depend on the position of the watershed relative to the historic and new snow-to-rain transition elevation. Finally, there are several elements of the impacts of change in spatial and temporal variability of snowpack that are not well established. For example, reservoir operations and reliability of downstream streamflow magnitude and timing, as well as potential tradeoffs between water users (e.g., hydropower, instream flows, and municipal uses) are likely affected by interannual variability of snowpack and streamflow, but these impacts have not been studied formally. An integrated watershed model could be used to assess the importance of interannual variability for these outcomes, and identify potential climate change impacts and reservoir management strategies that would support climate change adaptation.

The third chapter tested the importance of snowfall intensity for changes in winter ablation of snow. This chapter used two disparate lines of evidence in order to strengthen its conclusions, using both empirical and modeled data. While the study was based on ideas from a one-dimensional physically-based modeling study (Kumar et al., 2012), both lines of evidence in this chapter ultimately used a statistical modeling approach to test the importance of snowfall intensity. One further line of evidence that would allow for better understanding of the impacts of snowfall intensity on winter ablation in current and future climates would be a spatially distributed study using a physically-based model.

Model experiments with differing levels of snowfall intensity could be used to determine how the importance of snowfall intensity for winter ablation and SWE_{max} varies over space and time.

The fourth chapter analyzed the spatial and topical distribution of climate change research in the Columbia River Basin, including assessment of how topics are distributed in space. There are a few potential natural next steps in this research as well. While this study was a relatively high-level analysis of a large body of literature, further insight could be gained through more detailed qualitative analysis of components of this literature. For example, our synthesis identified particular patterns of research that could be explored in more depth: these include a potential disconnect between terrestrial and aquatic studies, and disparities in research foci between the United States and Canada. More detailed qualitative investigation could help to identify more specifically the information needs created by these general gaps. Another potential area for future investigation would be to better characterize interdisciplinary research in this corpus by analyzing a subset of papers that were determined to address multiple disciplines. This analysis could identify the specific problems that interdisciplinary research has been used to address as well as the findings of these studies, in order to better understand how interdisciplinary efforts currently operate and could operate in the future.

While each individual chapter suggests possible avenues for future work, there are also potential syntheses that span the themes of individual chapters. While early studies of climate change impacts on snow hydrology tended to primarily evaluate the effects of changing temperature and precipitation on annual snow metrics, such as April 1 SWE (e.g., Hamlet et al., 2005; Mote 2006), more recent work has evaluated a more complete suite of energy balance contributors, including humidity (Harpold and Brooks, 2018), seasonality of shortwave radiation (Musselman et al., 2017), and changing snow albedo (Skiles et al., 2018). These studies have generally evaluated changing energy balance components in a spatially explicit manner, but further synthesis is needed to identify where, and under which interannually-varying conditions, different contributions to energy balance changes may be most important. Moreover, such a synthesis could also include analysis of how energy balance changes affect different components of annual SWE metrics, using, for example, the SWE triangle introduced by Rhoades et al. (2018).

It may also be useful for a synthesis to target the impacts of changing snowpack. For example, recent studies of the effects of changing snowpack on runoff magnitude and timing have debated the most important mechanisms by which altered snow hydrology affects streamflow (e.g., Barnhart et al., 2016; Berghuijs et al., 2014). A thorough literature review, combined with physically-based modeling and empirical observations, could better establish relationships between changing snowpack and streamflow, with particular attention to which processes dominate in different spatio-temporal contexts.

While each of the chapters in this dissertation has contributed to improved knowledge of how climate change impacts changing snowpack, much remains to be learned in order to better understand changing heterogeneity at multiple spatial and temporal scales, as well as the impacts of these changes on water resources and ecosystems. Holistically improving this set of knowledge will provide enhanced understanding of the possible and appropriate set of climate change adaptation tools, and ultimately assist decision making for sustainable resource management.

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Appendix A - Supplemental Material for Chapter 2

Supporting Information for: Warming alters hydrologic heterogeneity: climate sensitivity of snow drift and scour dynamics in the snow-to-rain transition zone

Introduction

This supplemental material includes two major components: (1) a description of climate scenarios that were constructed but not ultimately included in the primary manuscript because they were very similar to scenarios in which simple delta functions were applied, and (2) figures that supplement the results presented in the main manuscript. These materials are provided because they help to provide a more in depth understanding of the sensitivity of specific components of this system to climate change and variability but are not critical to include to effectively convey the key findings of the study.

Text A1. Seasonally variable scenarios

To develop seasonally variable scenarios, average daily mean temperature and precipitation were calculated for each day of the year in future and historic scenarios, using a historic time series from 1950-2005, and an RCP 8.5 time series from 2006-2099. For each of 20 GCMs and for future and historic scenarios, the daily mean time series were smoothed using a locally weighted regression with a Gaussian kernel and span of 0.08 (8% of the data) for temperature, and 0.1 for precipitation. These values were selected to choose the smallest span that smoothed the data, and to minimize the pattern of the absolute value of residuals when plotted against the fitted values (Cleveland, 1979) (Figures S2, S3). For each of the 20 GCMs, the difference between historic and future conditions for each day of the year was calculated by subtracting the delta function in the case of temperature change, or calculating a percent change in the case of precipitation change. The mean of these functions was then calculated across all 20 GCM. The delta function and 20-model standard deviation is represented in Figure S1.

Text A2. Precipitation-dependent warming scenarios

To determine differential temperature change during precipitation events, historical mean daily temperatures were subtracted from future precipitating/dry daily temperatures. A day was defined as dry if it had less than 0.254 mm of precipitation, after Rupp and Li (2016). Temperature differences were then regressed against weather conditions (precipitation), and month to determine the dependence of temperature change on precipitation condition (Table S1, S2). The results suggested that for every 100 mm of precipitation, warming was reduced by 1.7°C ($p < 0.0001$). This was of comparable magnitudes to the findings of Rupp and Li (2016), who found that in the Washington Cascades, when P = 100 mm of precipitation per day, change in Tmin was 1.2 °C less than in cases with trace amounts of precipitation.

Text A3. Influence of climate variables on drift factors

We investigated the possibility of using a statistical model to account for the influence of perturbed air temperature and precipitation on the empirical drift factor. The relationship between winter air temperature and precipitation and empirical drift factors in each HRU are presented in Figure S4 for years where data were available to compute drift factors. In all cases, drift factors are not statistically significantly related to climate variables ($p > 0.05$). We also investigated simple multiple linear regressions using an interaction between air temperature and precipitation; these were also not significant ($p > 0.05$). Thus, we were unable to construct a satisfactory statistical model to scale the drift factor based on the climate change scenario.

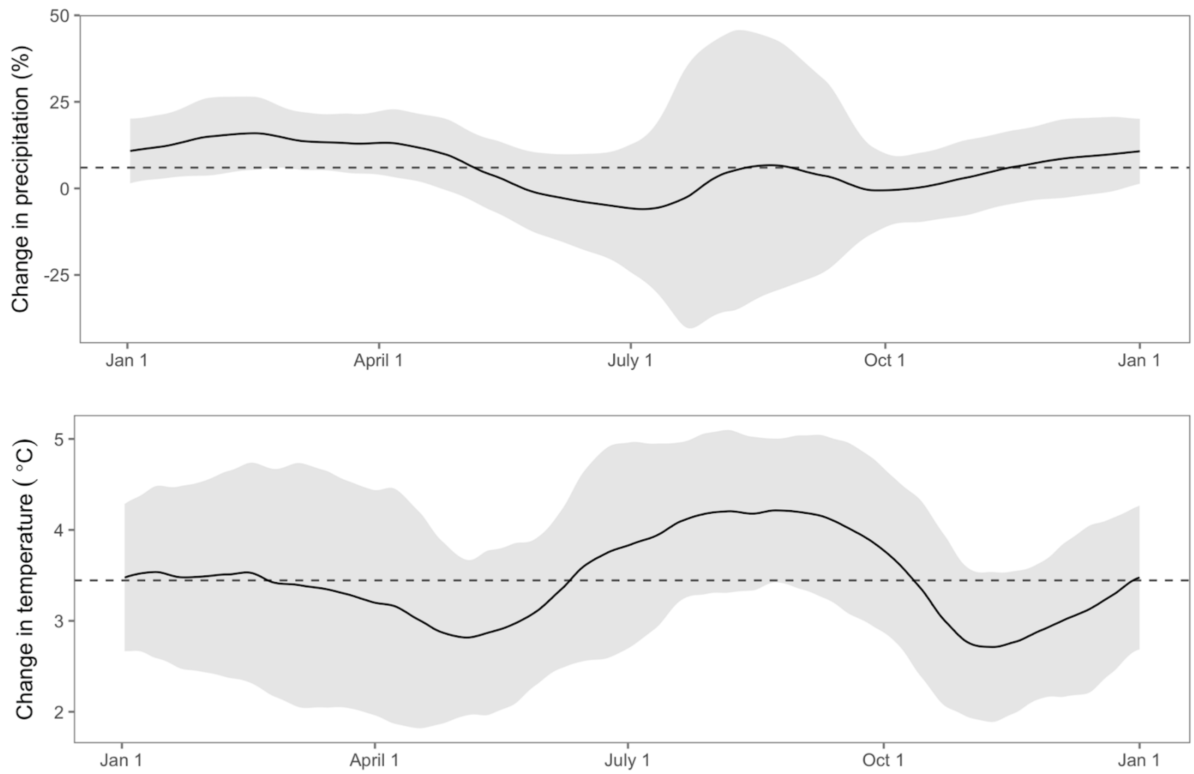


Figure A.1. Seasonally variable temperature and precipitation changes. Dashed lines represent means of the seasonally variable delta function. Solid lines represent the seasonal change function, and gray area represents 20-model standard deviation.

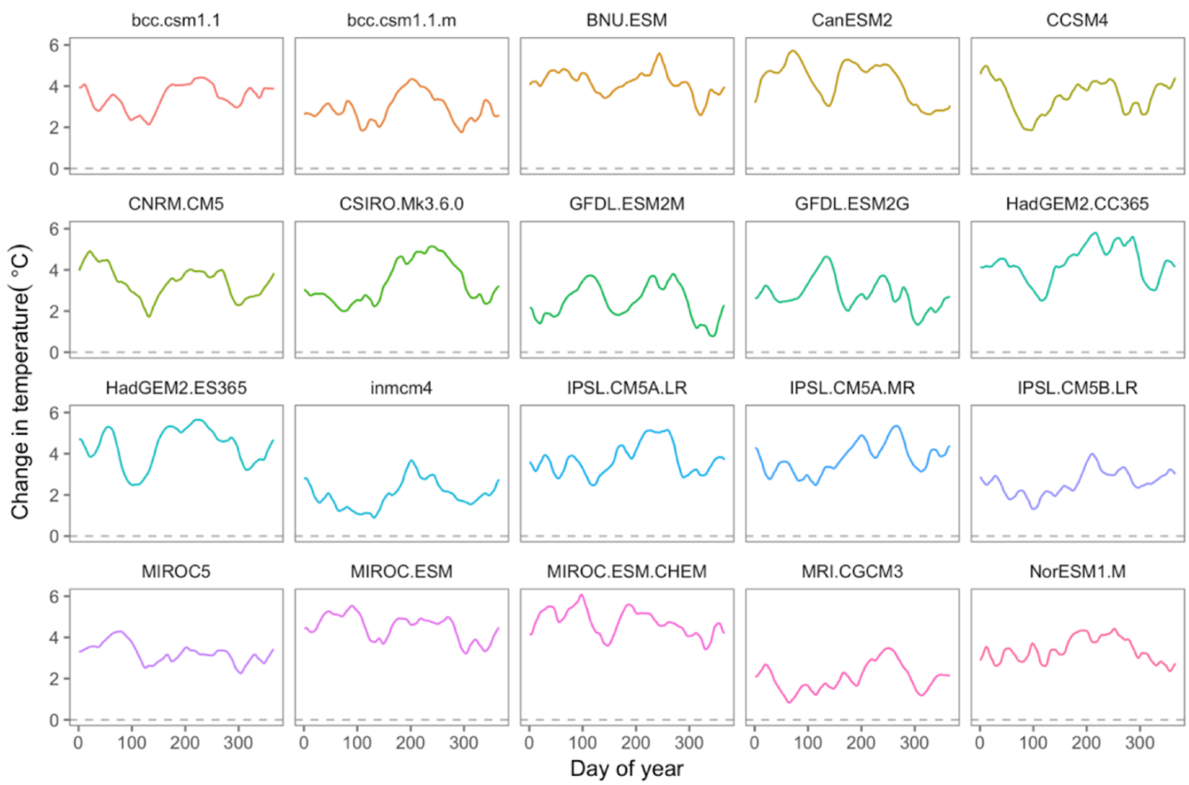


Figure A.2. Loess-smoothed daily change in temperature for the 20 models used in this study.

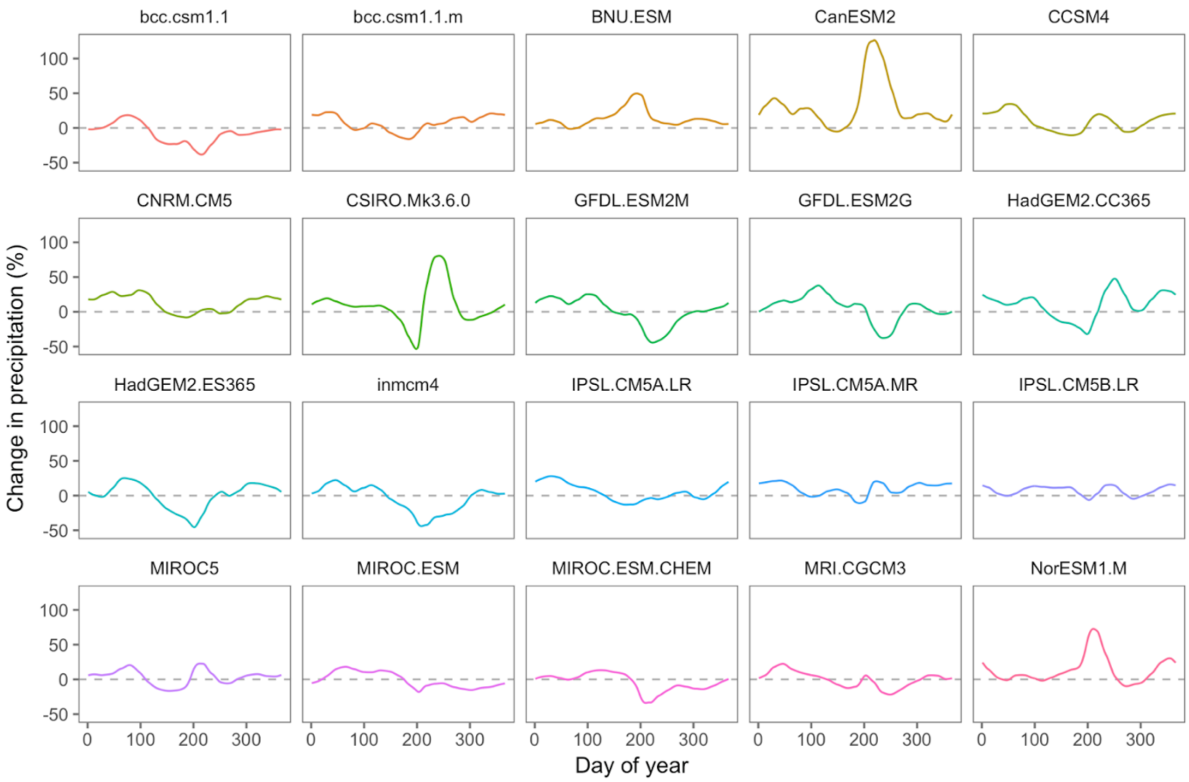


Figure A.3. Loess-smoothed daily change in precipitation for the 20 models used in this study. Dashed line represents zero value for reference.

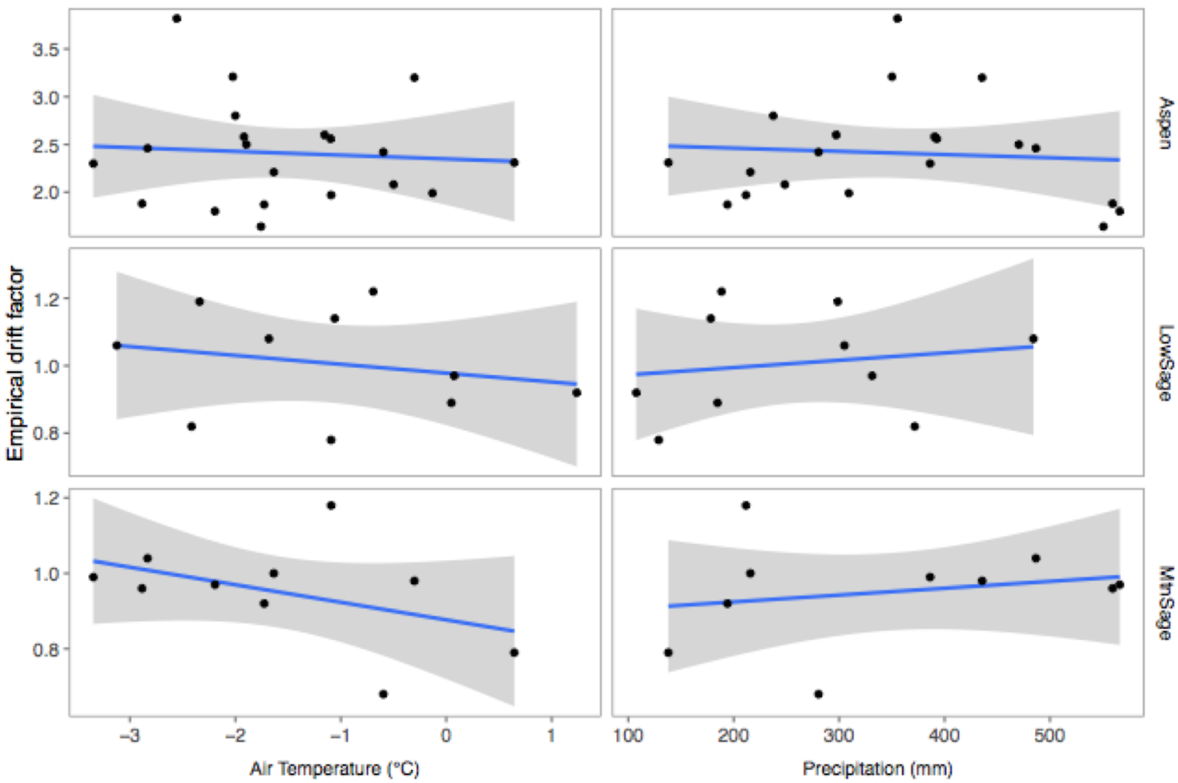


Figure A.4. Empirical drift factors as a function of November-March average air temperature and cumulative precipitation. The relationship between drift factor and climate is not statistically significant in any case. For the aspen HRU, data is from 1984-1993 and 2004-2013. For the mountain big sage and low sage HRUs, data is from 1984-1993.

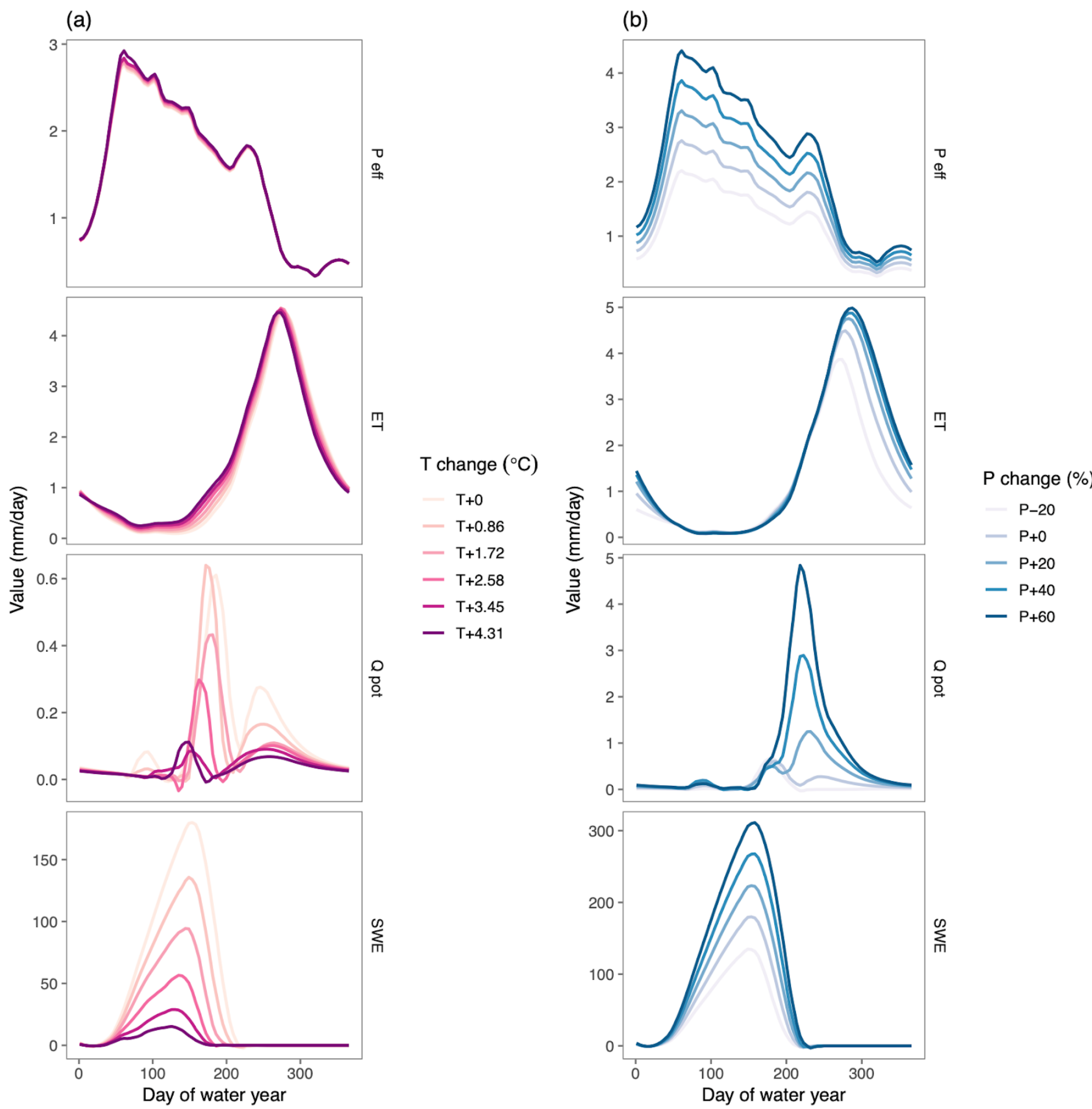


Figure A.5. Changes in Peff, ET, Qpot, and SWE with warming and altered precipitation in the mountain big sage HRU.

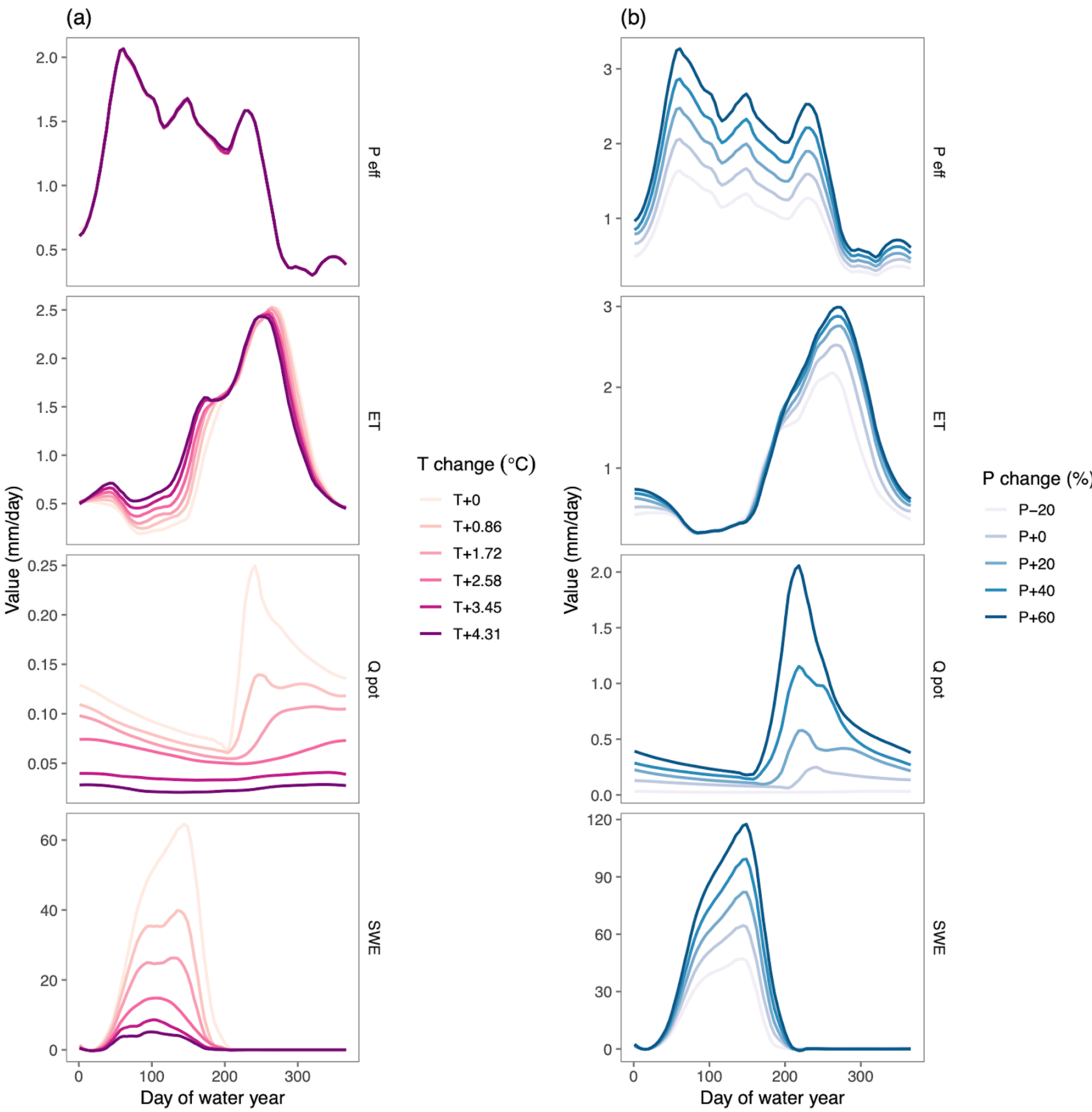


Figure A.6. Changes in Peff, ET, Qpot, and SWE with warming and altered precipitation in the low sage HRU.

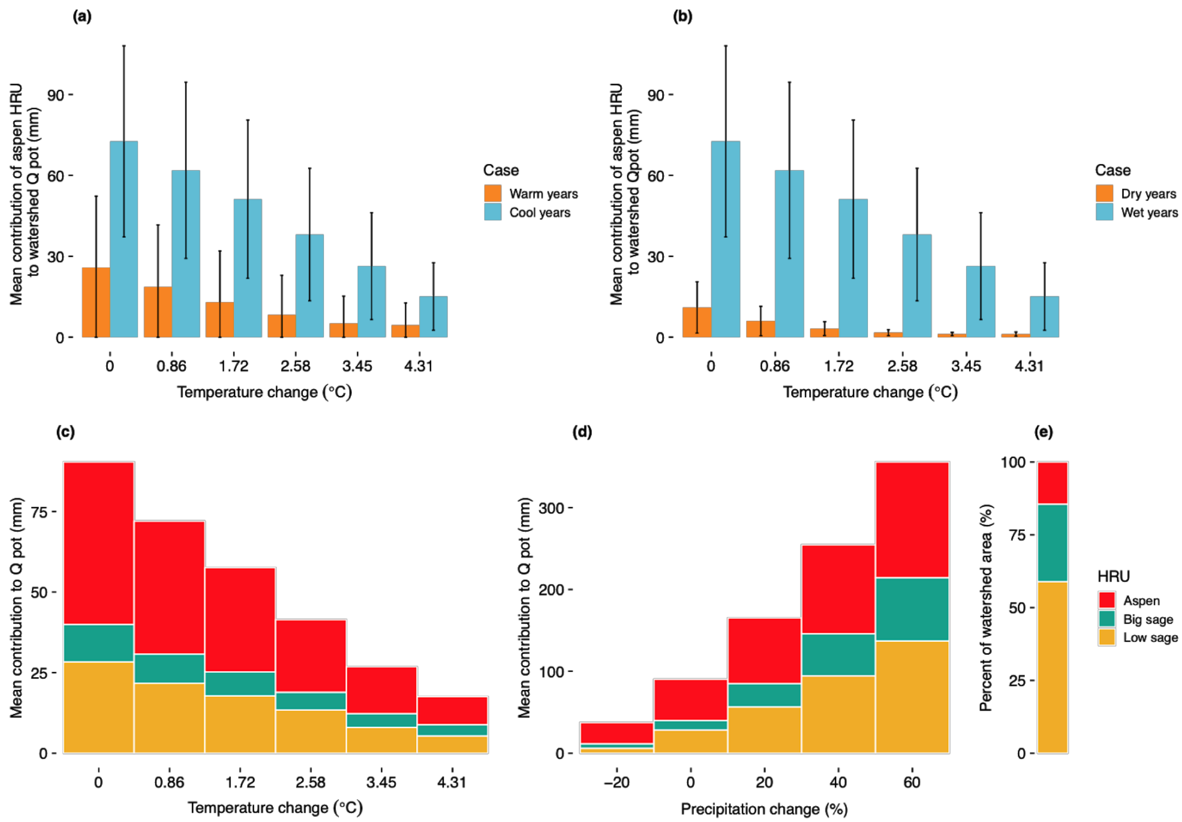


Figure A.7. (a, b) Contribution of aspen HRU to watershed QPOT in (a) warmest and coolest tercile of years and (b) wettest and driest tercile of years. (c) 30-year average contribution of each HRU to warming with increasing temperature, and (d) same as (c) with altered precipitation instead of temperature. (e) displays the percent of watershed area occupied by each HRU.

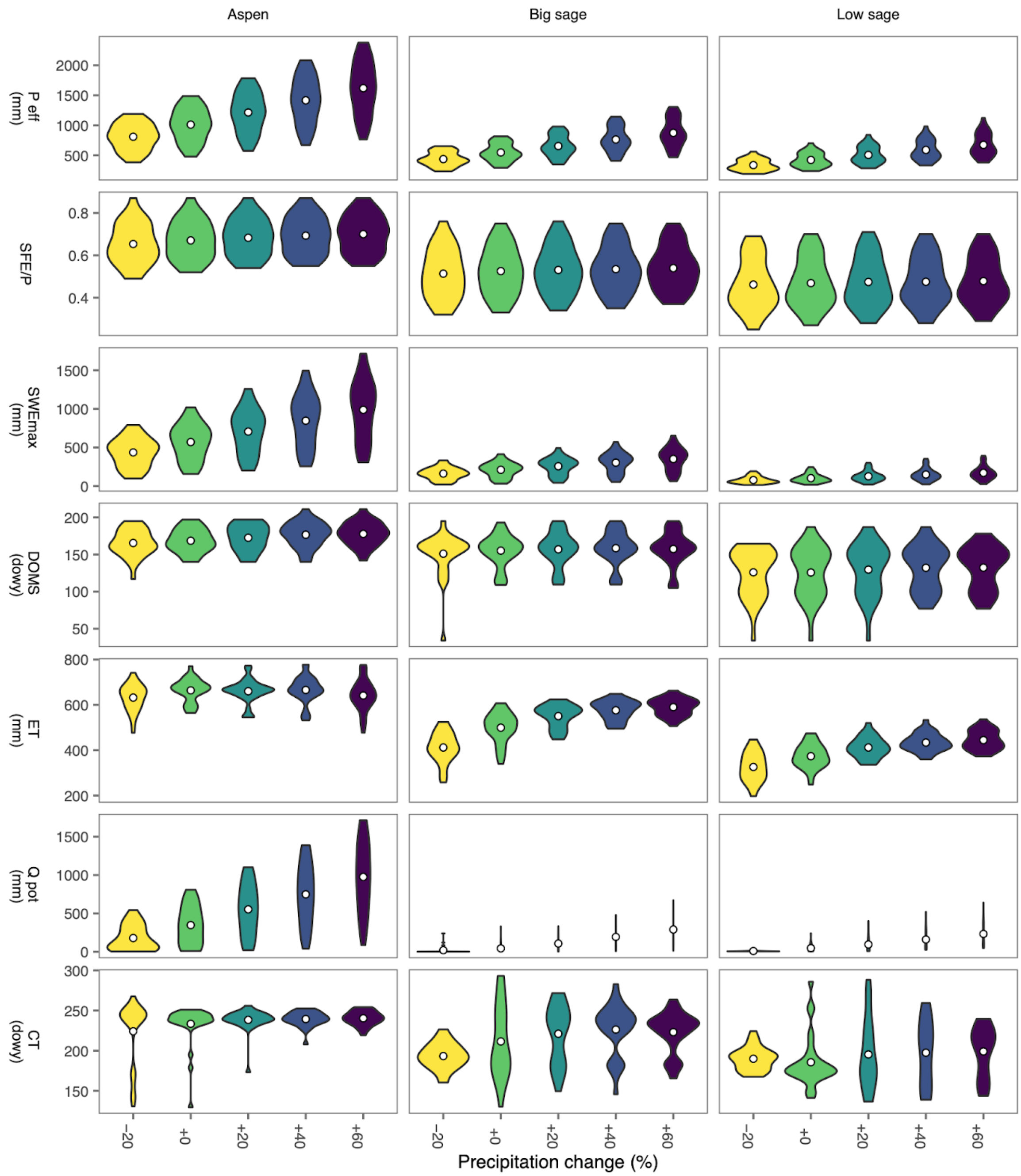


Figure A.8. Interannual variability of annual summary variables for each HRU under changing precipitation. Each shape represents a density function of values; points within shapes represent mean values. Color indicates increase in temperature as denoted on the x-axis.

Appendix B – Supplemental Information for Chapter 3

The following supplementary figures illustrate geographic context (Figures B1-B2; B15), historical, future values, and changes in mean values and first and third quartile values of SWE_{max} and DOMS (Figures B3-B7, B16-B17), relationships between changing variability of snowpack and climate (Figures B8-B11), additional measures of variability (Figure B12), statistical significance of findings (Figures B13-B14), and transect plots for additional GCMs (Figures B18-B10). Finally, supplemental material is also presented as an interactive data visualization tool at: https://snowvariability.nkn.uidaho.edu/.

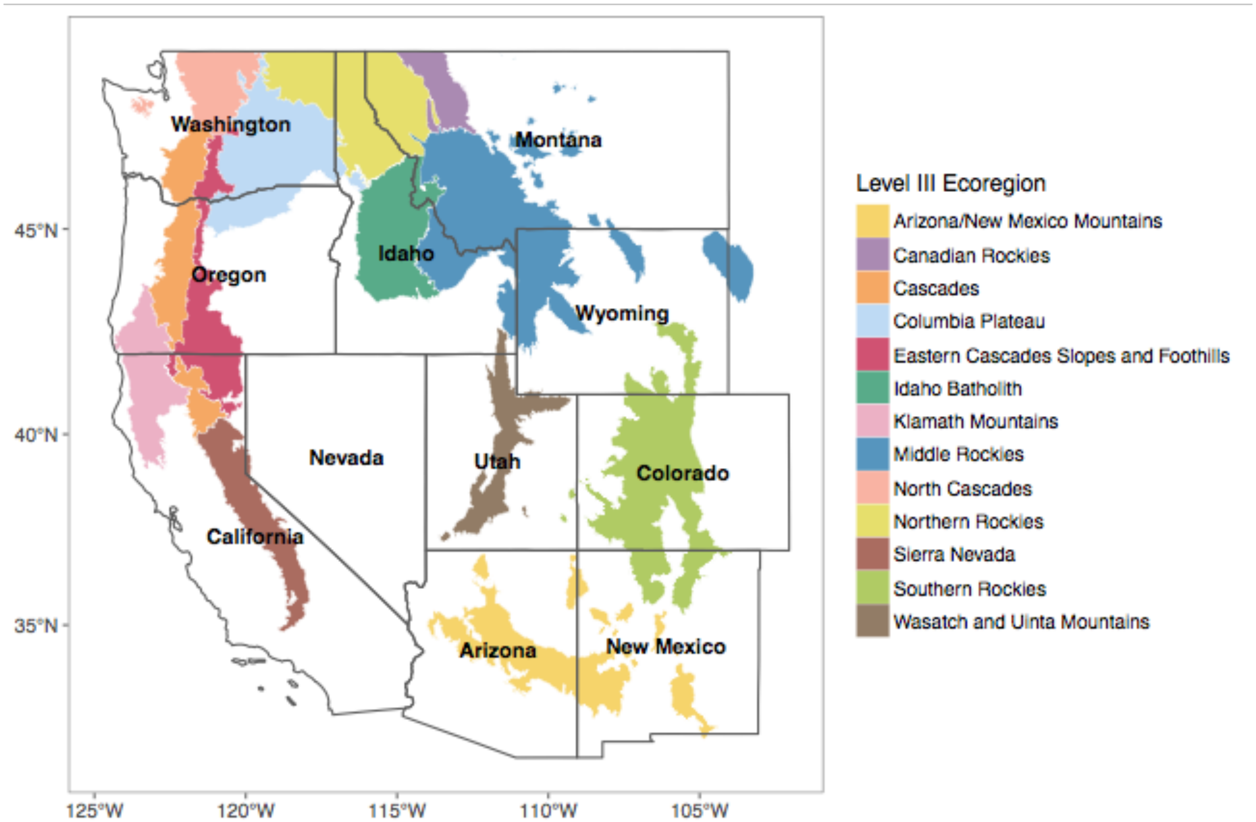


Figure B.1 States and regions referred to in the text. U.S. Level III Ecoregions primarily correspond to mountain ranges in the western United States, so these are used as a reference point for discussion.

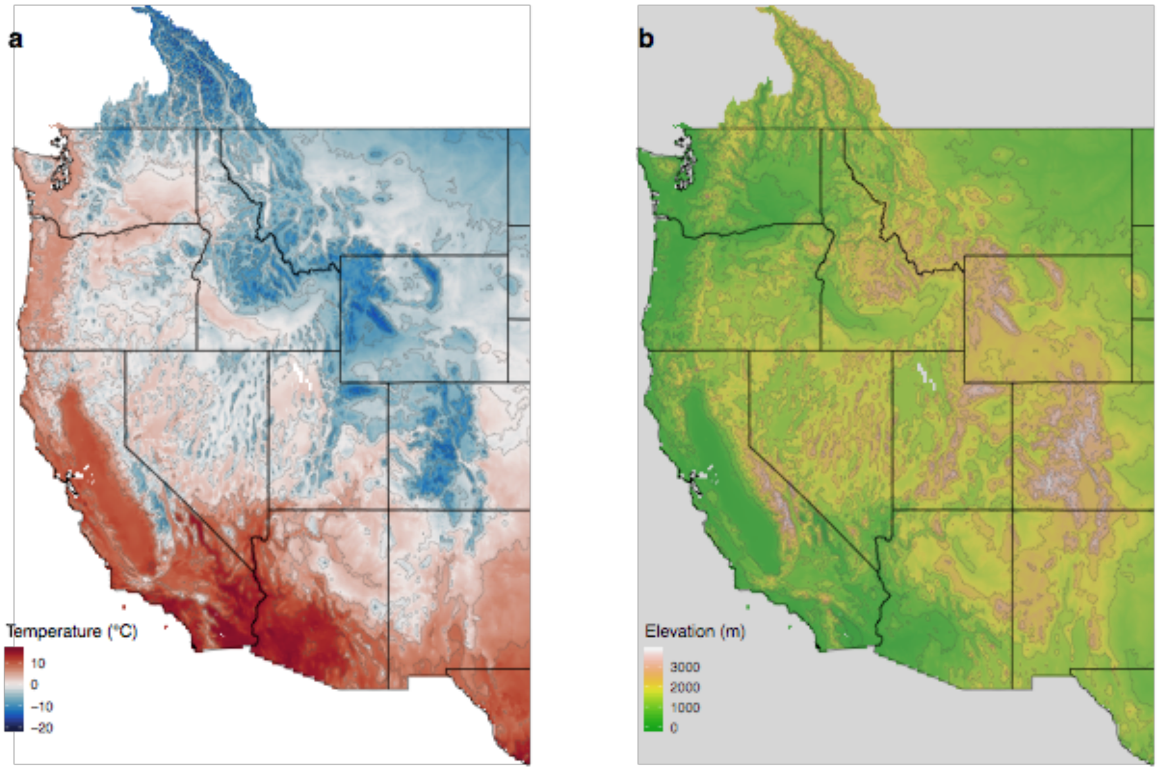


Figure B.2 (a) Historic average November-March temperature and (b) elevation. These figures provide context for comments in the text about regional temperature and elevation.

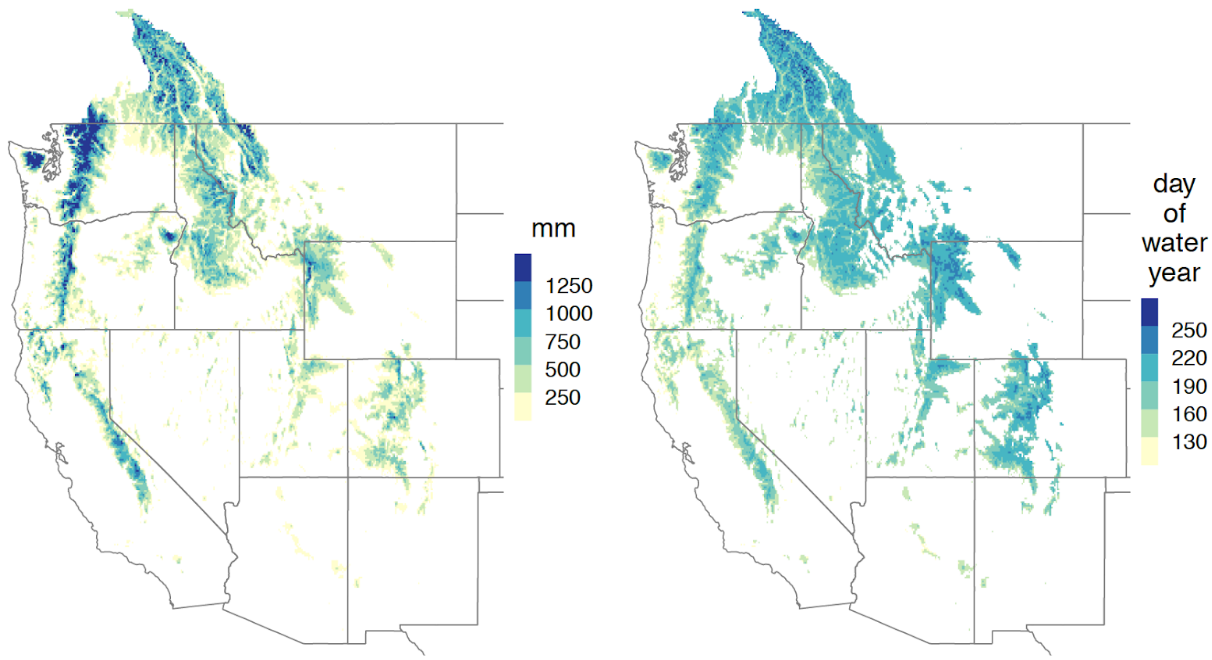


Figure B.3 Historical (1971-2000) (a) mean SWE_{max} and (b) DOMS.

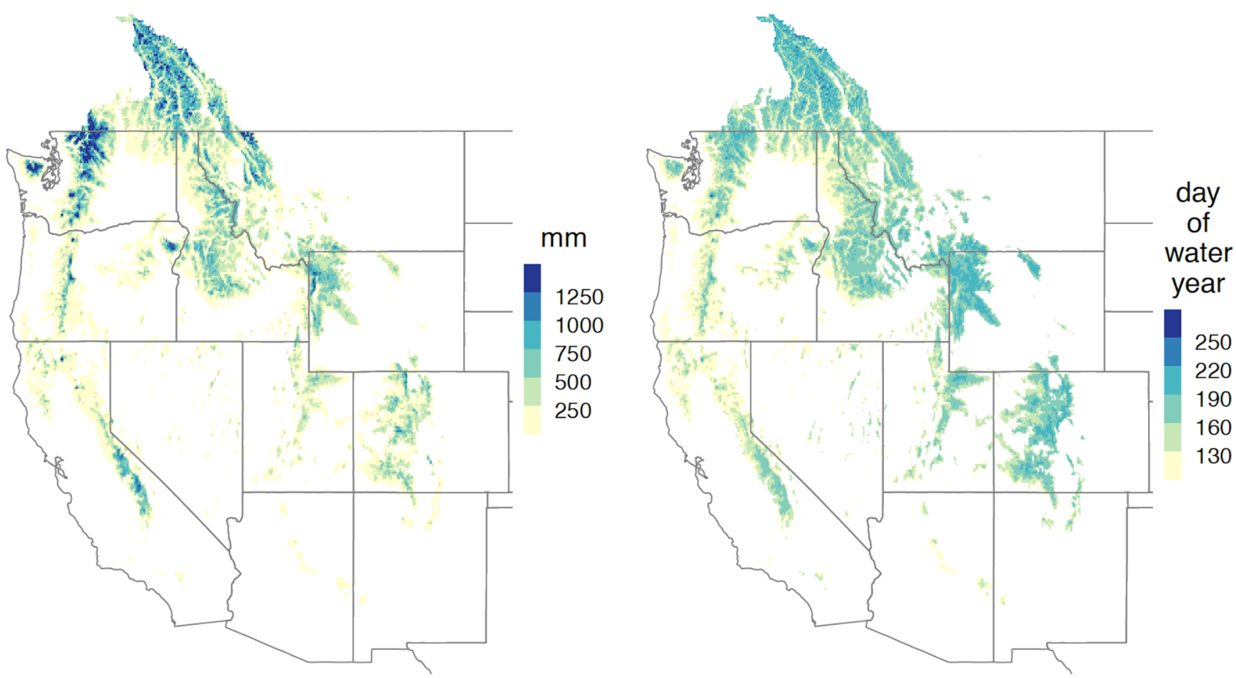


Figure B.4 RCP 8.5 (2050-2079) (a) mean SWE_{max} and (b) DOMS.

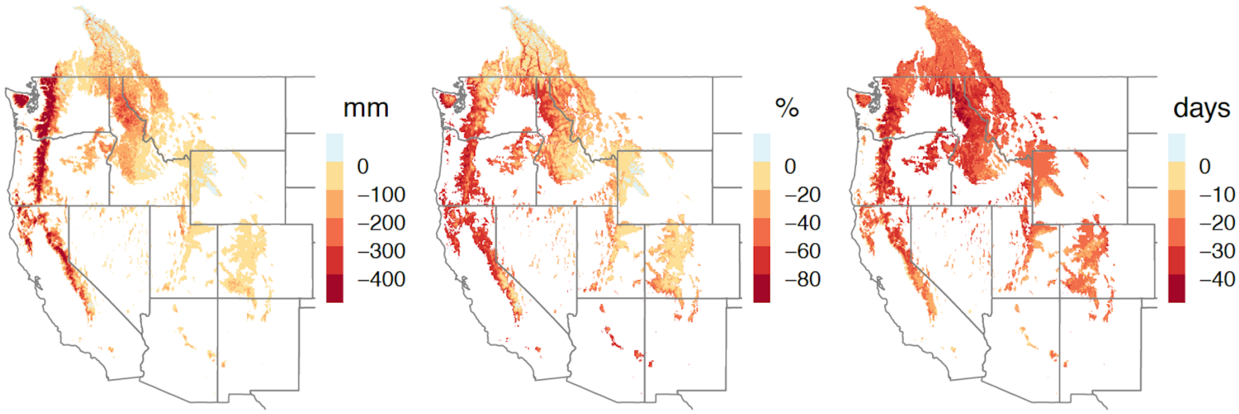


Figure B.5 (a) Absolute change in mean SWE_{max}, (b) percentage change in mean SWE_{max}, and (c) change in mean DOMS between 2050-2079 (RCP8.5) and 1971-2000 (historical forcing).

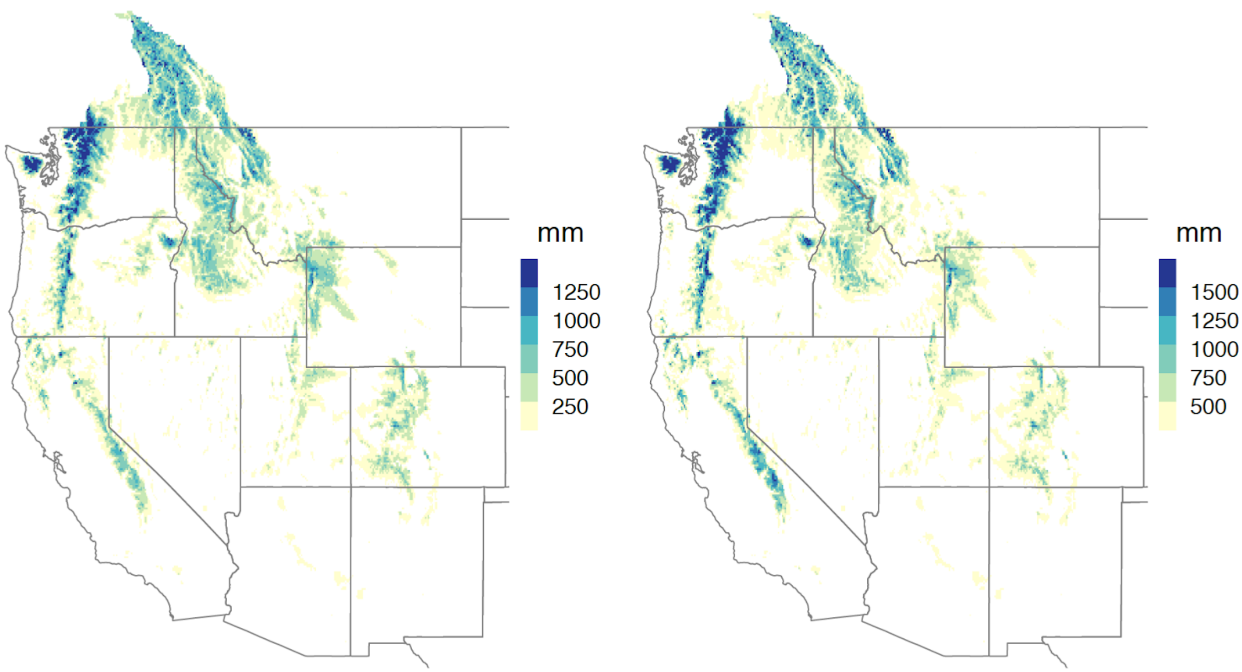


Figure B.6 Historical values of (left) 25th and (right) 75th percentiles of SWE_{max}.

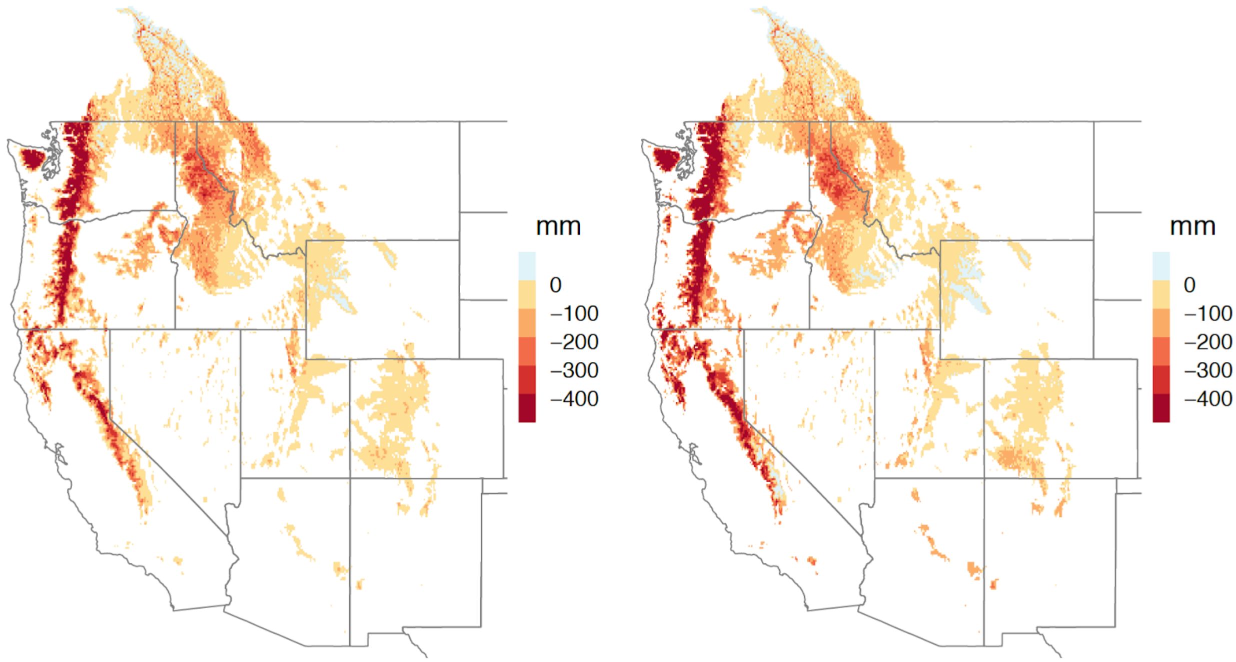


Figure B.7 Changes in values of (left) 25th and (right) 75th percentiles of SWE_{max}.

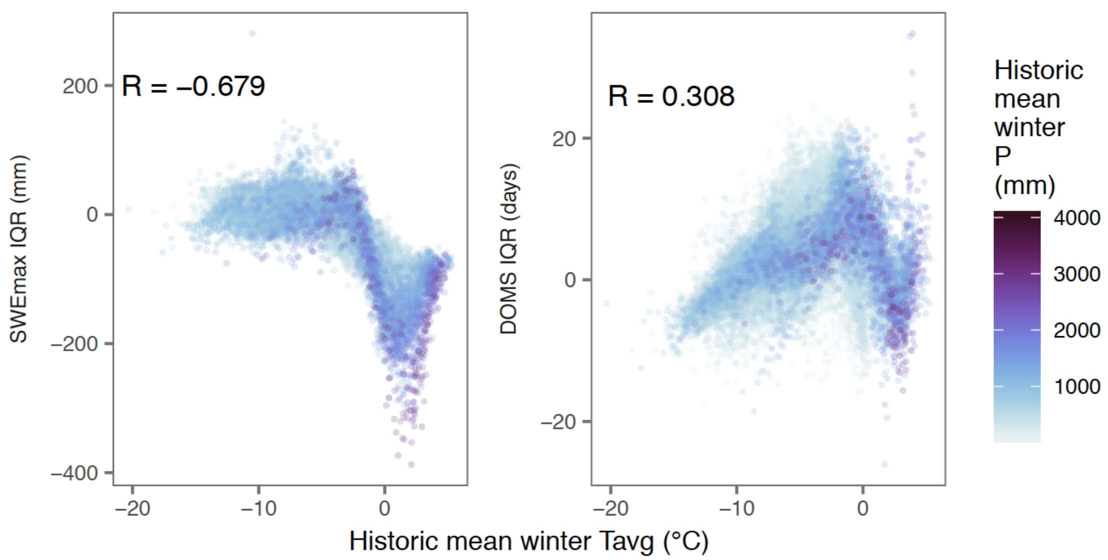


Figure B.8 Change in SWE_{max} and DOMS IQR as a function of historical average mean November-March temperature. Change in SWE_{max} IQR has a non-linear temperature-dependence, with largest and relatively consistent decreases in warm sites. Change in DOMS IQR is not clearly related to historical average winter temperature. Correlation coefficients indicate a fairly strong relationship for change in SWE_{max} IQR and a weaker, positive relationship for change in DOMS IQR, though these should be interpreted with caution given the spatial autocorrelation of observations and correlations with other climate variables.

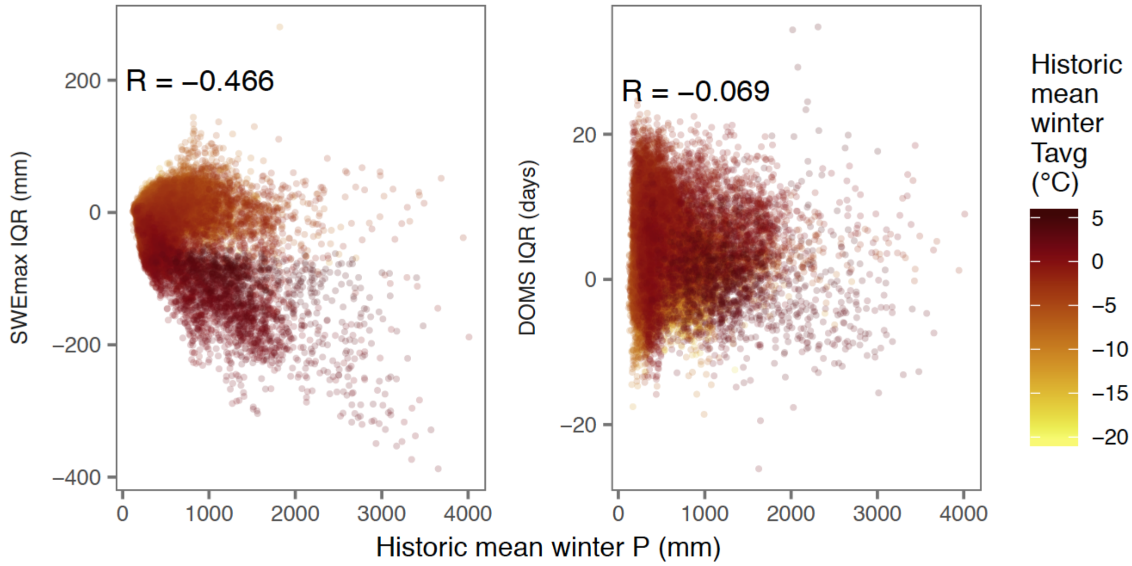


Figure B.9 Change in SWE_{max} and DOMS IQR as a function of historical average mean November-March precipitation. Changes in variability metrics are not strongly associated with historical mean winter precipitation, though the correlation coefficient for change in SWE_{max} IQR suggests that SWE_{max} IQR tended to decrease in areas with high winter precipitation.

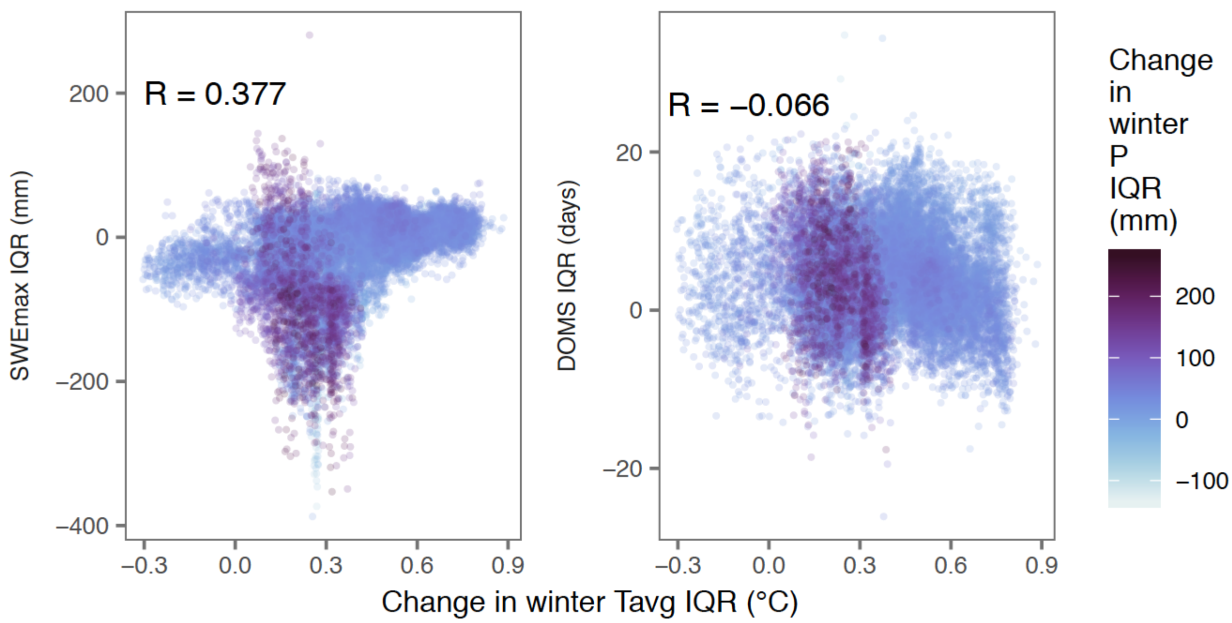


Figure B.10 Change in SWE_{max} and DOMS IQR as a function of change in mean winter T_{avg} IQR. There is some evidence of a positive association between change in winter Tavg IQR and SWE_{max} IQR, particularly in areas that do not have large changes in precipitation variability.

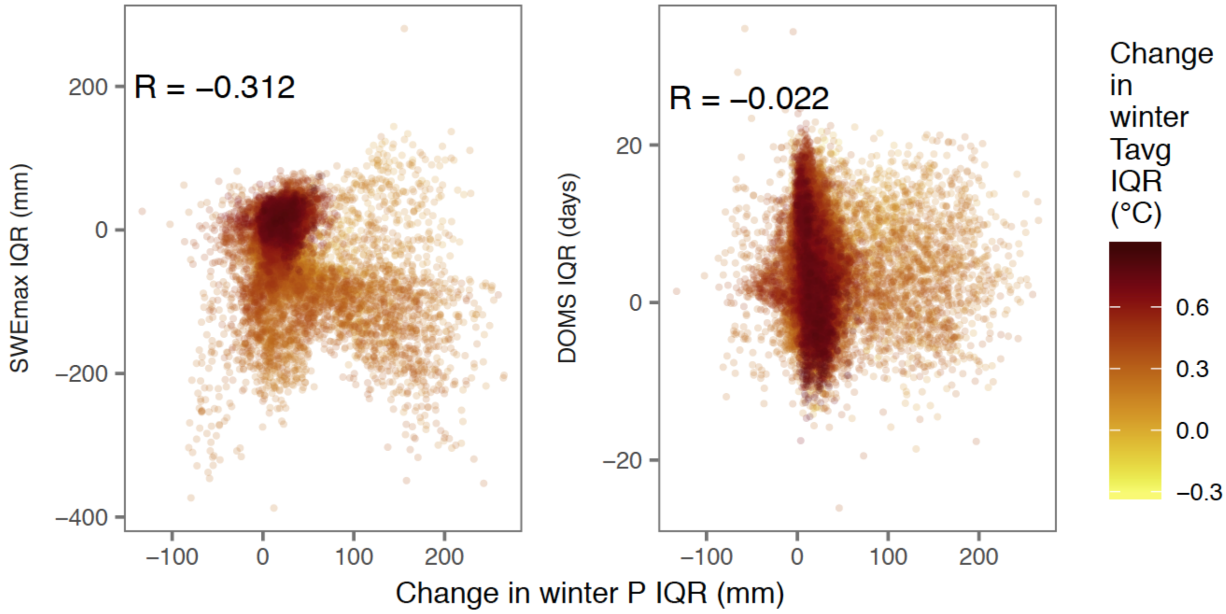
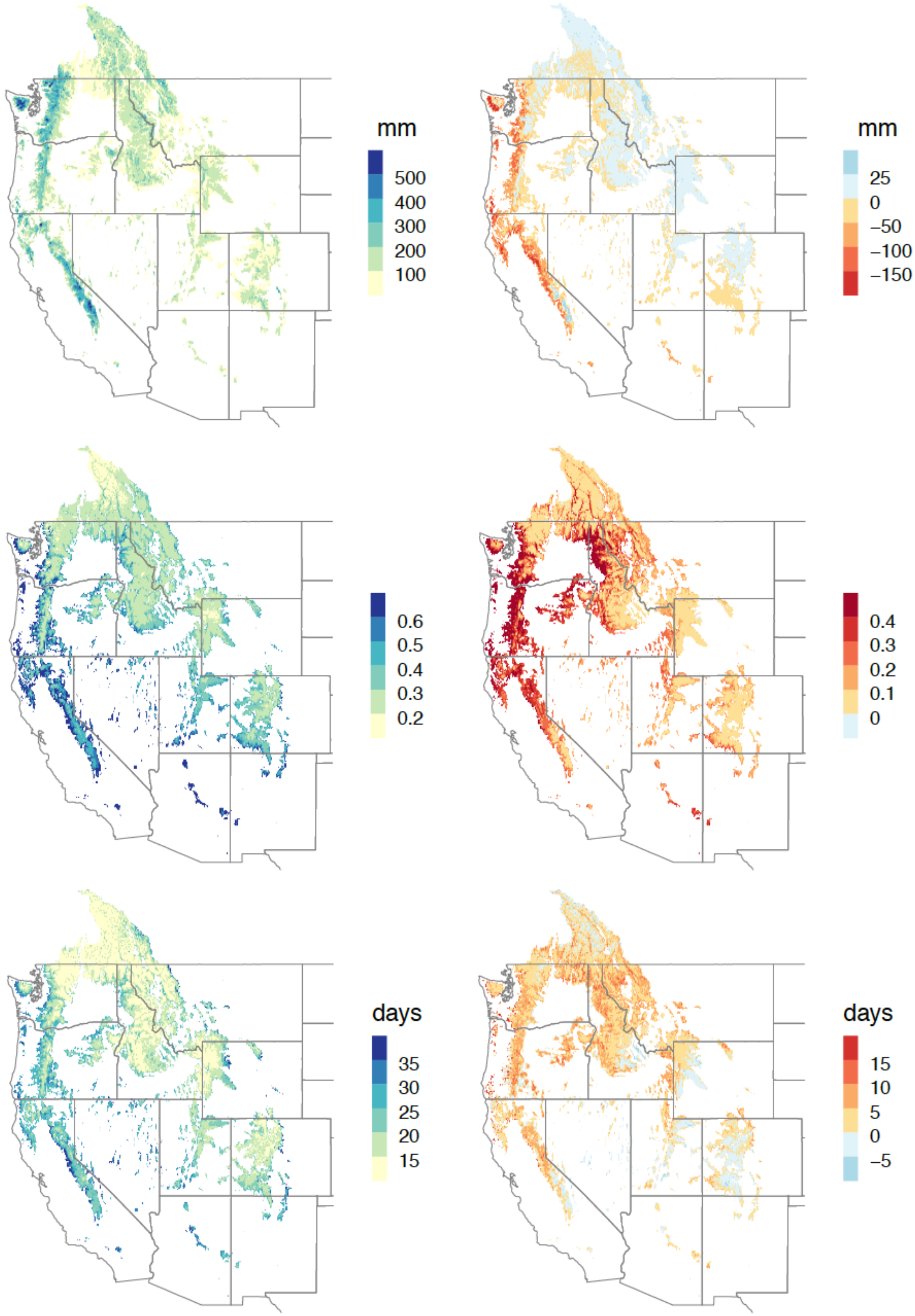


Figure B.11 Change in SWE_{max} and DOMS IQR as a function of change in mean winter precipitation IQR. The correlation coefficient suggests a slight negative relationship between change in SWE_{max} IQR and change in winter precipitation IQR.


mm
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−50
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0.3
0.2
0.1
0
days
35
30
25
20
15
days
15
10
5
0
−5

Figure B.12 Additional measures of snowpack variability: (a) Historical SWE_{max} standard deviation (SD), (b) change in SWE_{max} SD, (c) historical SWE_{max} coefficient of variation, (d) change in SWE_{max} coefficient of variation, (e) historical DOMS SD, (f) change in DOMS SD. For both DOMS and SWE_{max} historical patterns and spatial patterns of change in standard deviation are qualitatively similar to those for IQR. SWE_{max} CV shows large increases in the snow-to-rain transition zone due to decreasing means; this metric is likely more reflective of changing means than variability.

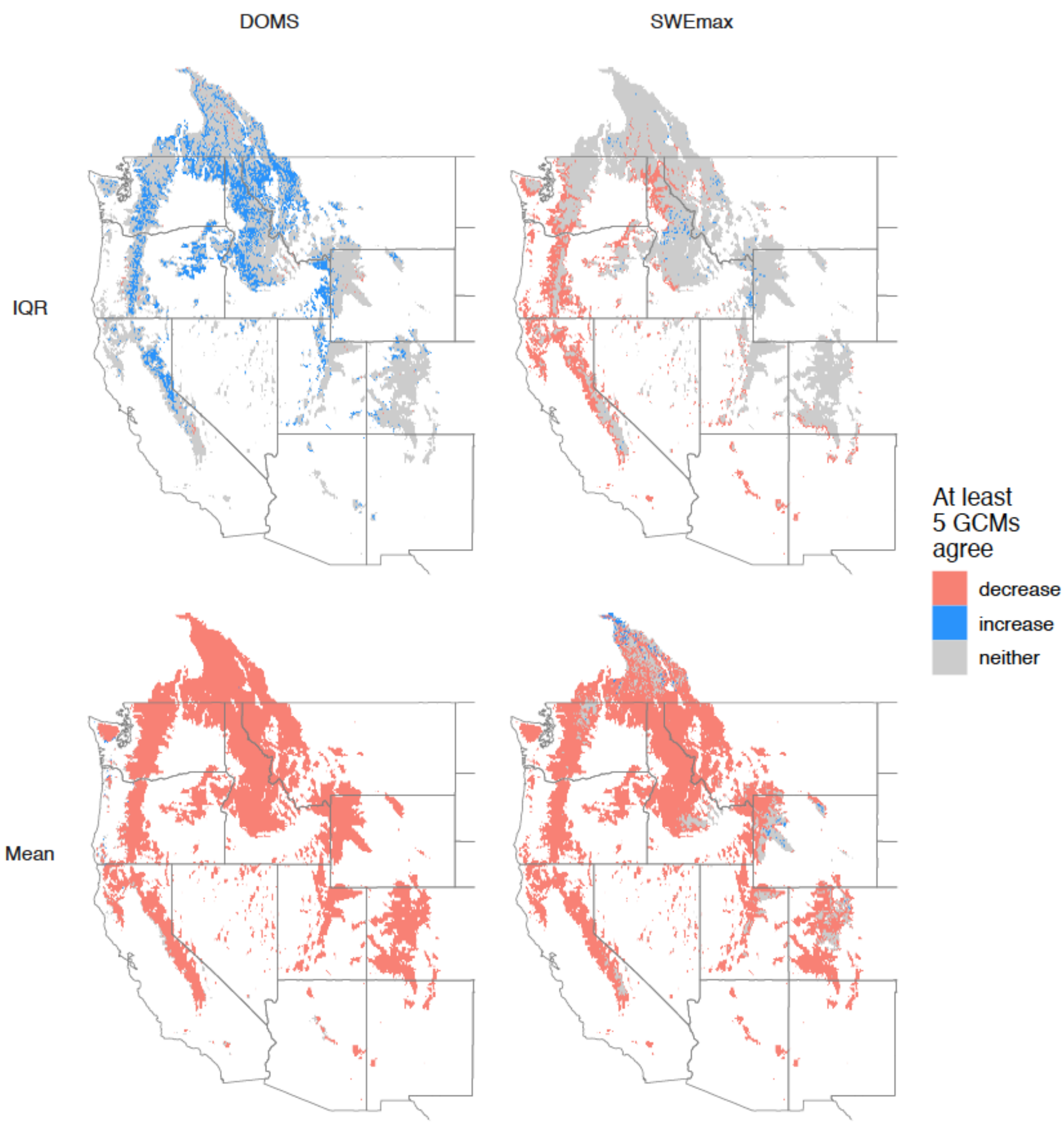

Figure B.13 Summary of model agreement using Monte Carlo bootstrapping approach. Model agreement suggests that at least 5 GCMs agree on a significant change in the same direction (increasing or decreasing). Significance is defined such that the future value falls outside the historic 5th-95th empirical percentiles.

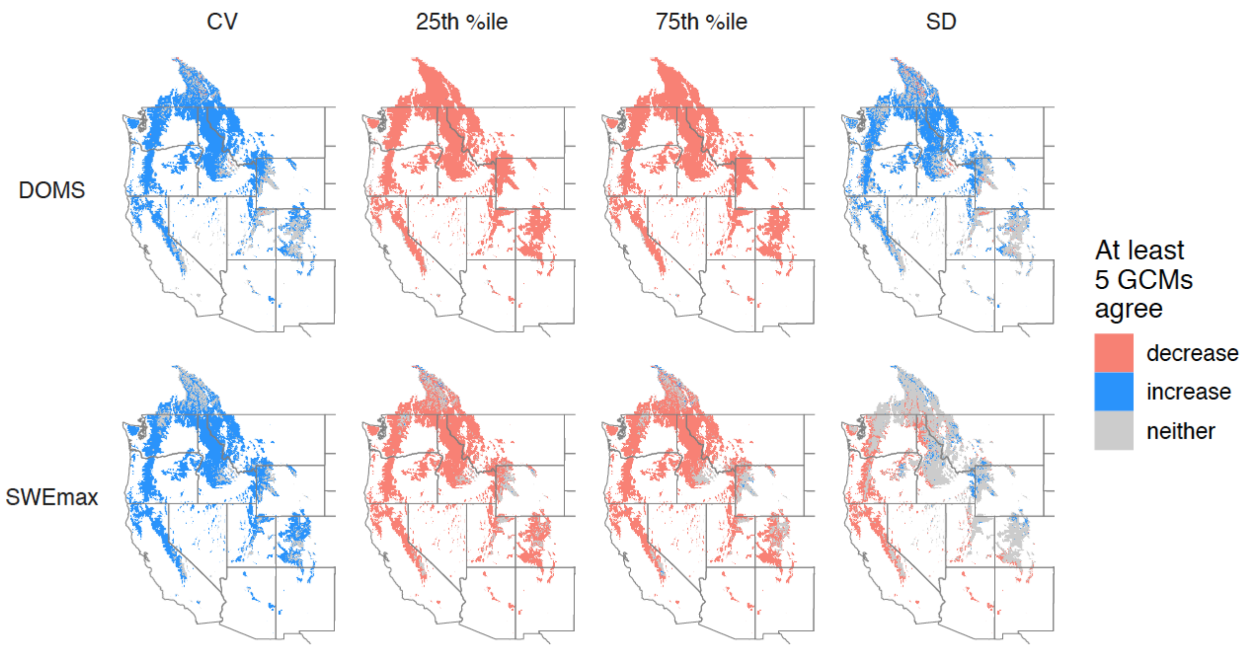

Figure B.14 As in Figure S13, but for variables that are not focused on in the text (from left to right: coefficient of variation, 25th percentile, 75th percentile, and standard deviation).

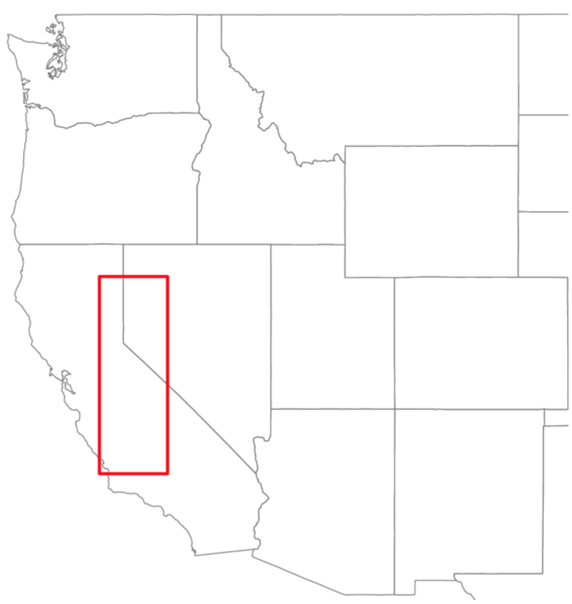

Figure B.15 Red box shows the area inset in Figure 2 in the main text.

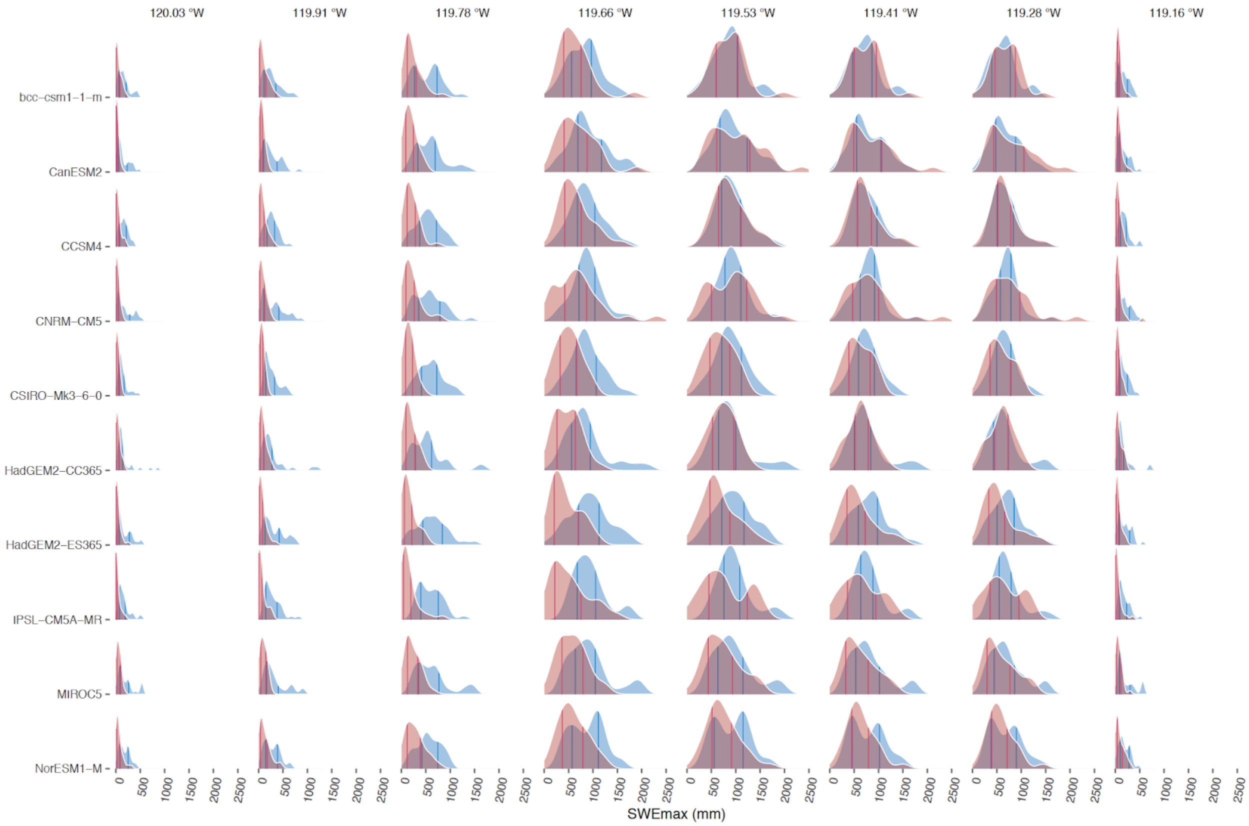

Figure B.16 Values of SWE_{max} along transect as in Figure 2 for each of 10 GCMs.

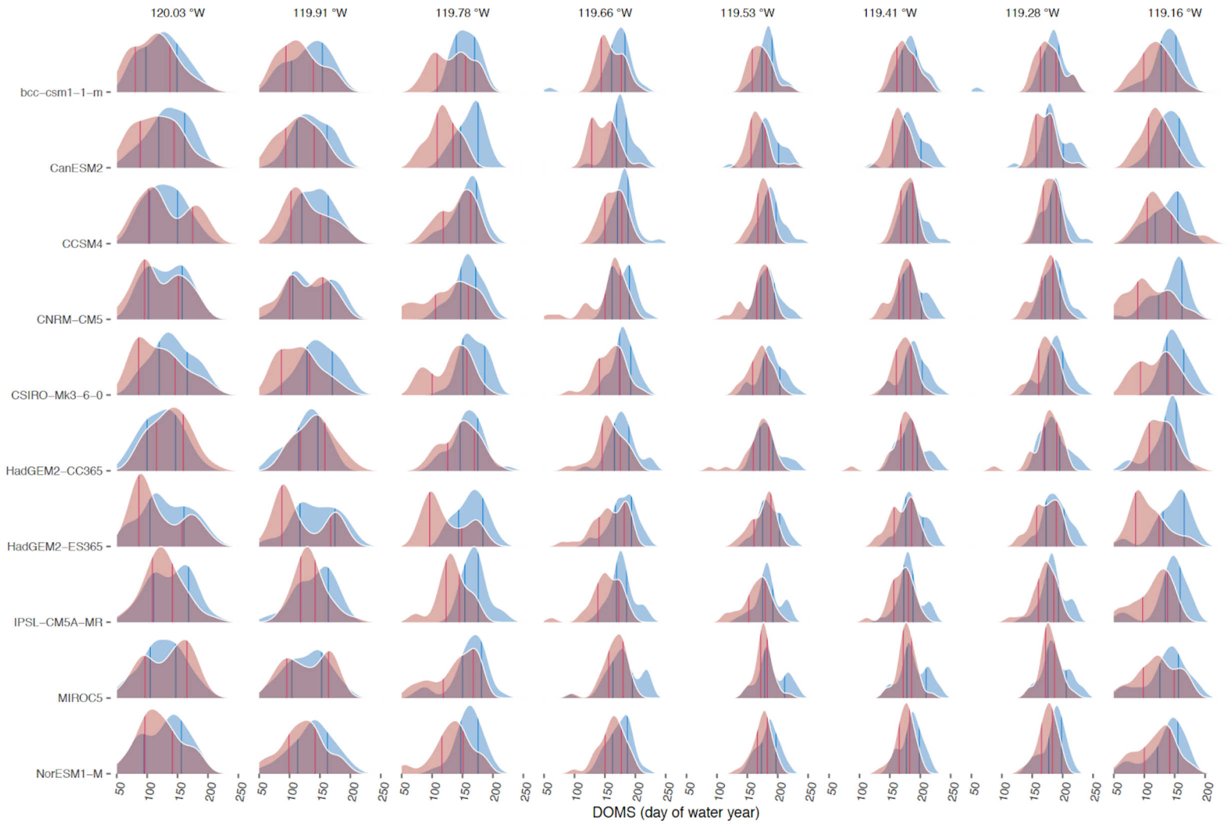

Figure B.17 Values of DOMS along transect for each of 10 GCMs as shown in Figure 2.

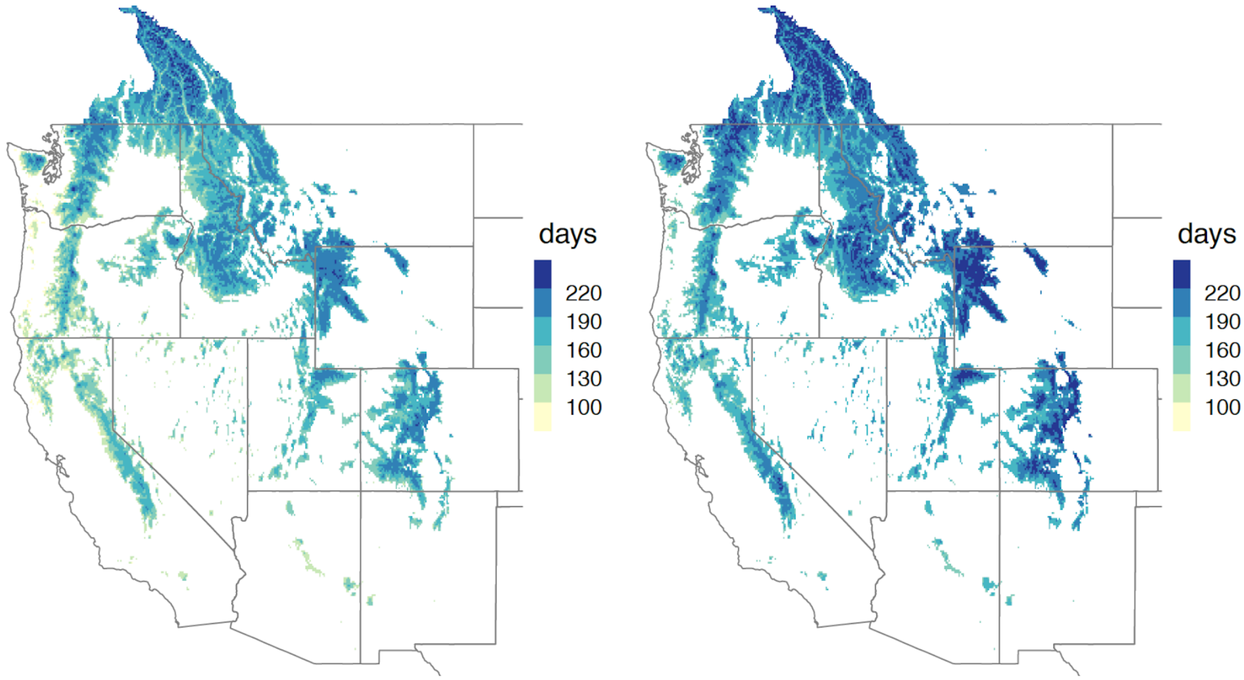

Figure B.18 Historical values of (left) 25th and (right) 75th percentiles of DMS.

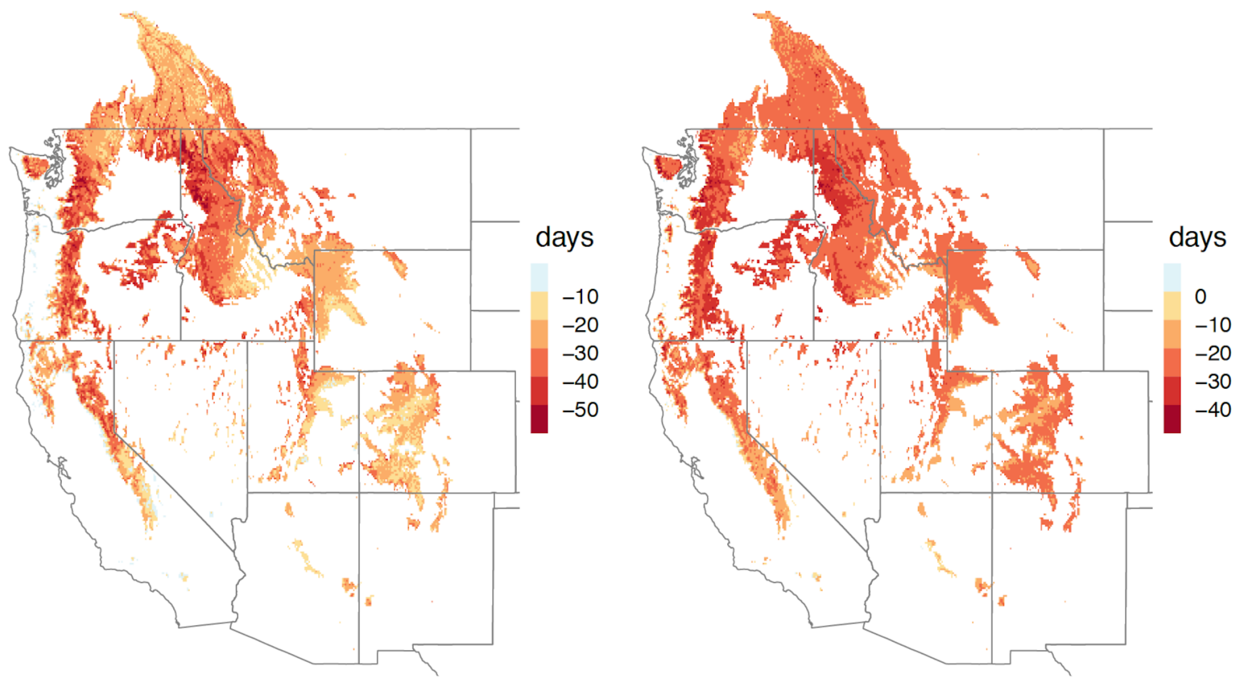

Figure B.19 Changes in values of (left) 25th and (right) 75th percentiles of DMS.

Appendix C – Supplemental Information for Chapter 4

The figures and tables presented in this supplementary information display model diagnostics not presented in the primary article text (Figures C1-C4), additional details of model results (Table C1), or display results from individual GCMs that have been averaged over many GCMs in the main article (Figures C5-C9).

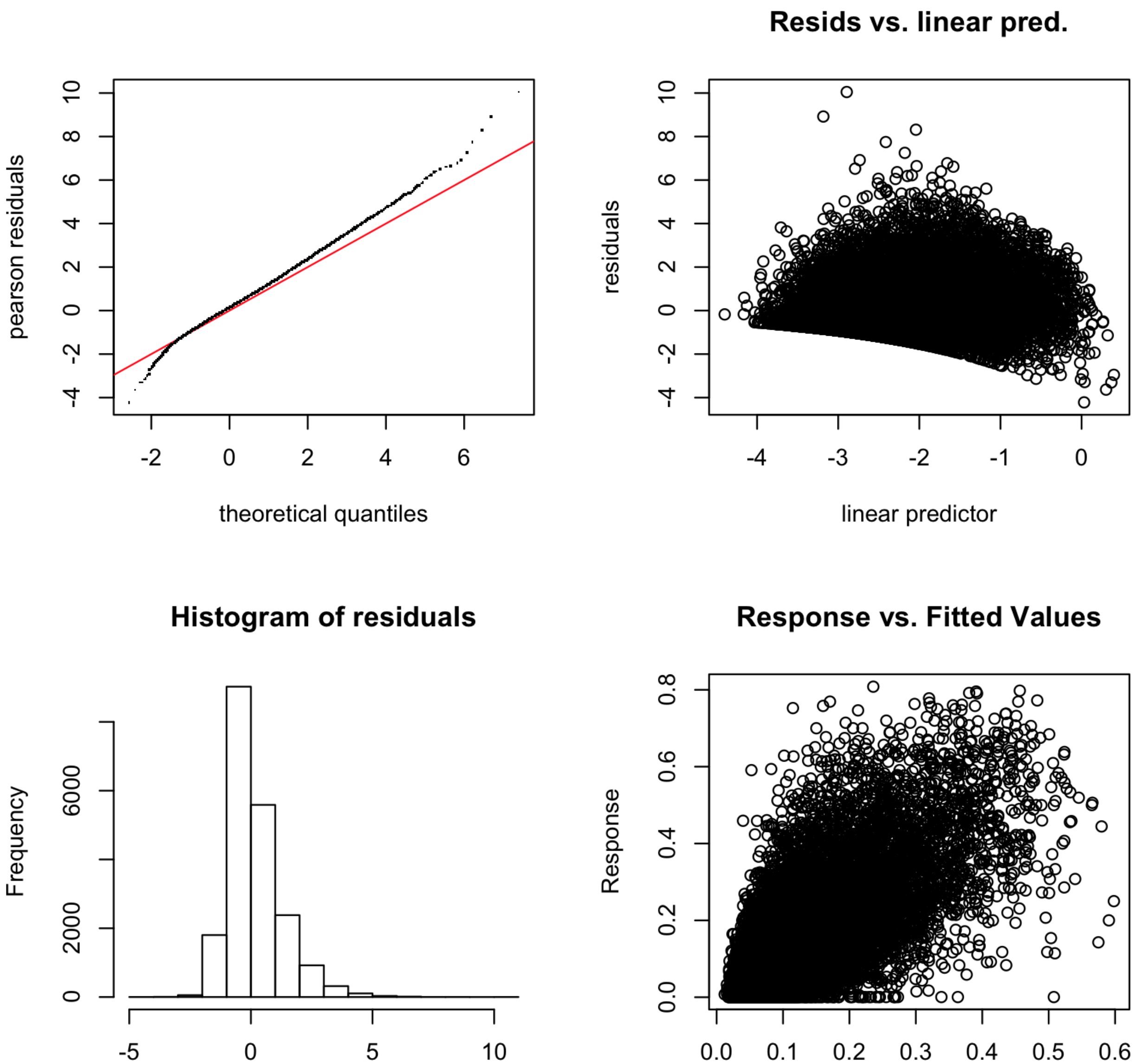

Fig C.1 Pearson residuals for SNOTEL model. Figure shows a Q-Q plot, residuals versus linear predictors, histogram of residuals, and responses versus fitted values. Despite slight non-normality of residuals, these diagnostic plots generally suggest that the model assumptions are adequately met.

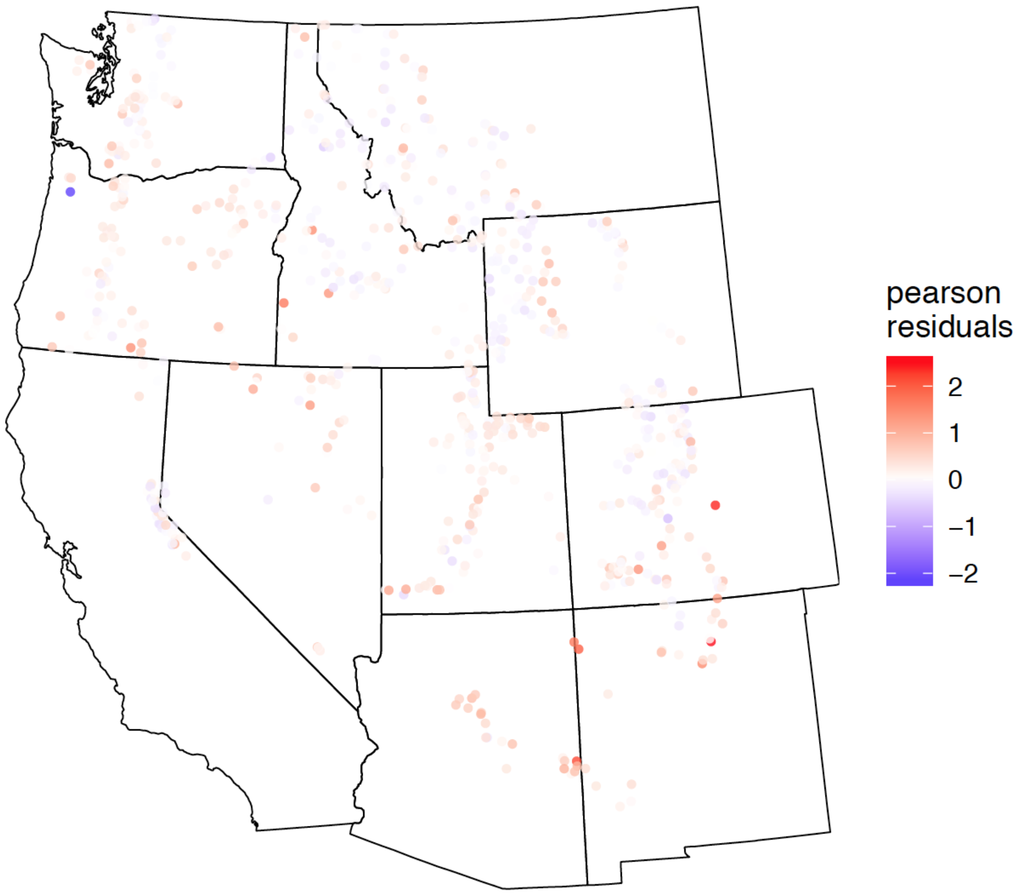

Figure C.2 Mapped pearson residuals (unitless) averaged over time for each SNOTEL site. The mapped residuals generally do not show significant spatial patterning, suggesting that the model is meeting the assumption that residuals are not spatially correlated.

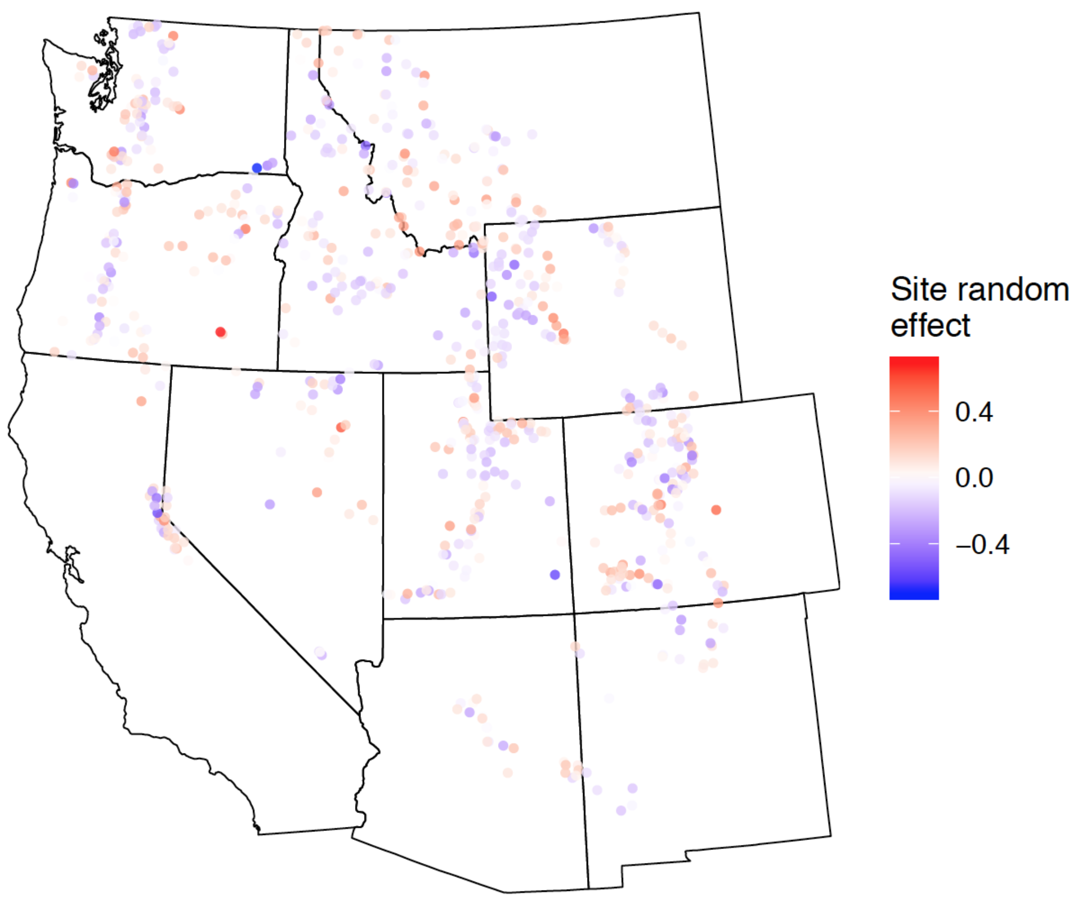

Figure C.3 Map of site random effects for SNOTEL model.

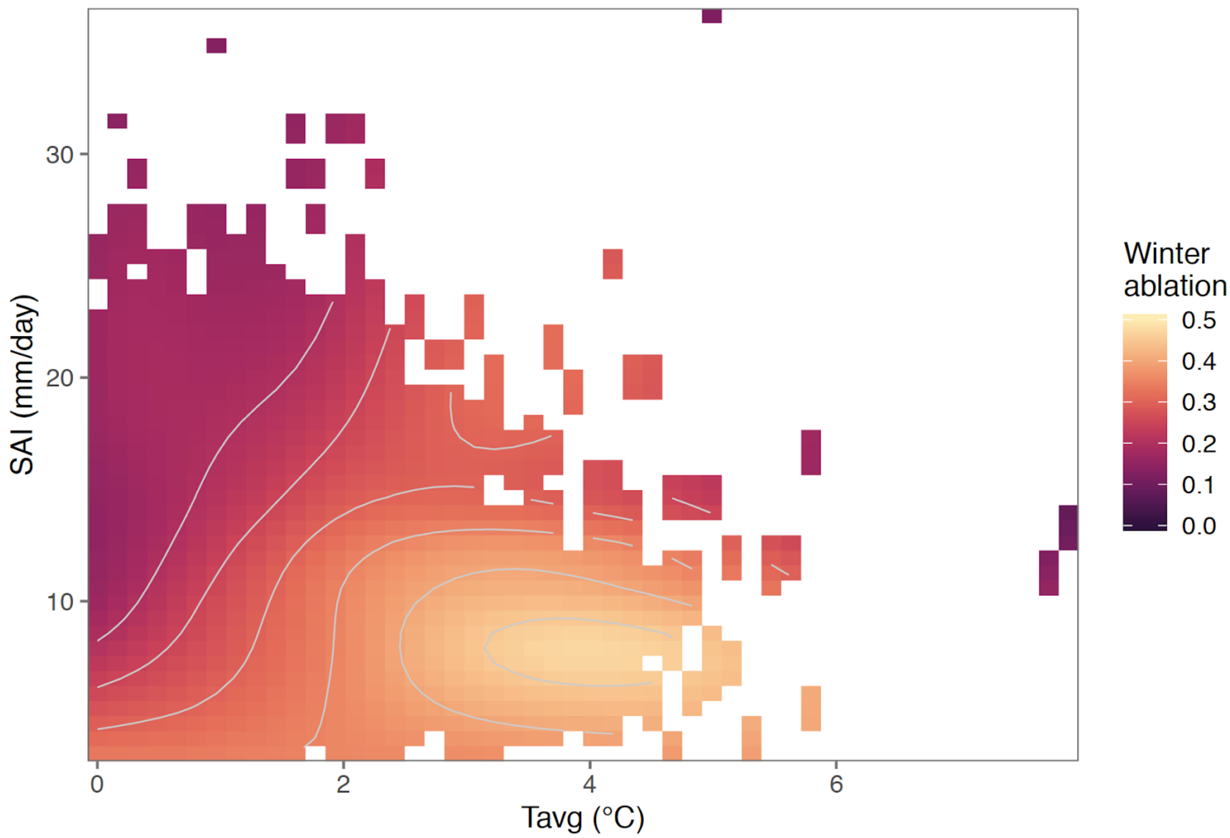

Figure C.4 Contour plot of statistical model projections using only SNOTEL data with winter T_{avg} > 0°C.

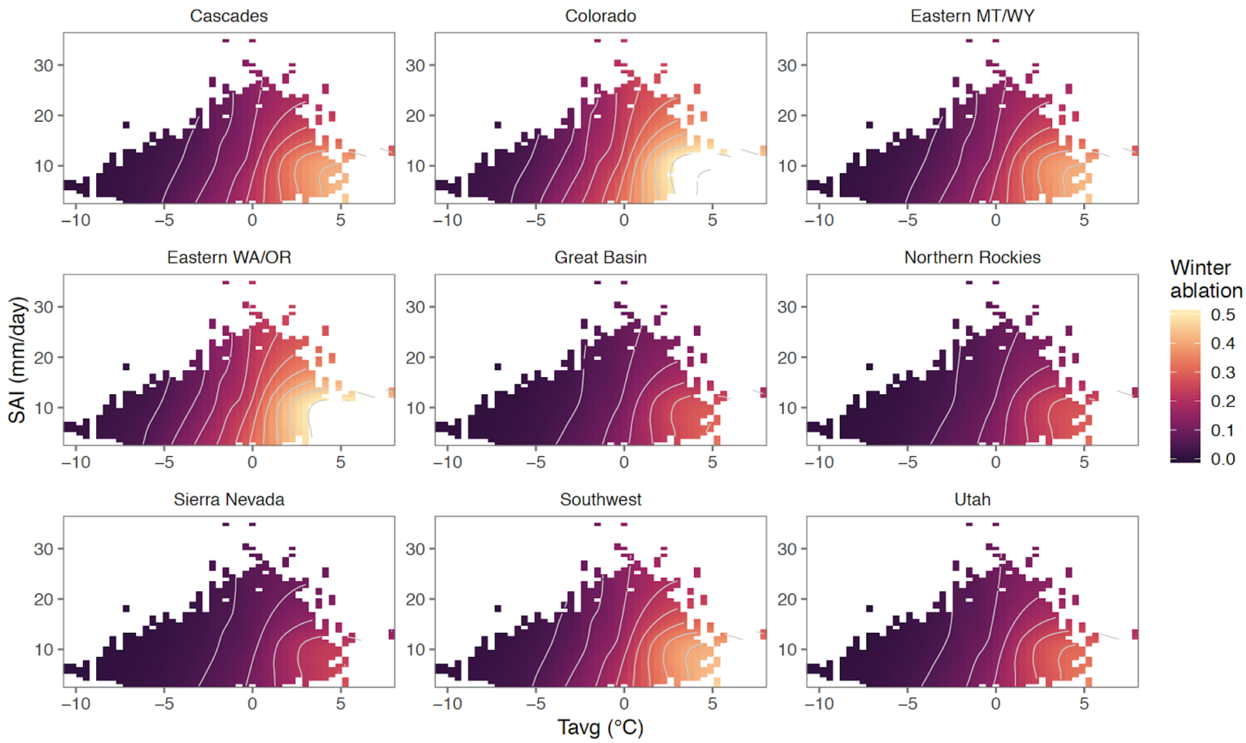

Figure C.5 Contour plots of winter ablation for SNOTEL model at each Serreze region. For each region, the median site random effect value is used for the model. Different regions show different values of winter ablation, but the contour line slopes are approximately the same in all cases. The same is true if values other than the median (e.g., 10th or 90th percentile) are used (not shown).

Table C.1 AIC, R^2 and deviance explained for SNOTEL and each VIC model, for both the full and null models. For all diagnostic variables and all models, the results suggest that the full model, including SAI, is preferable to the null model.

DATA SOURCE	NUMBER OF SITE-YEARS	AIC ($X10^3$)			ADJUSTED R^2 (%)			DEVIANCE EXPLAINED (%)		
		Full	Null	Difference	Full	Null	Difference	Full	Null	Difference
SNOTEL	20249	-63.6	-63.5	0.13	48.1	47.4	0.7	51.4	51.0	0.4
SNOTEL (>0°C)	3505	-6.8	-6.7	0.09	42.8	40.5	2.3	50.2	48.5	1.7
BCC-CSM1-1-M	117188	-480.1	-472.5	-7.5	49.7	45.6	4.1	56.2	52.4	3.8
CANESM2	116918	-467.4	-458.2	-9.1	45.0	39.8	5.2	53.5	48.6	4.9
CCSM4	117066	-468.1	-460.1	-8.0	50.1	45.7	4.4	58.3	54.4	3.9
CNRM-CM5	117124	-486.0	-474.8	-11.2	53.9	48.2	5.7	61.8	56.7	5.1
CSIRO-MK3-6-0	117099	-477.6	-467.0	-10.6	43.9	37.6	6.3	53.9	48.0	5.9
HADGEM2-CC365	116928	-493.9	-487.0	-6.9	50.0	46.2	3.8	60.5	57.2	3.3
HADGEM2-ES365	117017	-476.7	-468.9	-7.8	49.8	45.4	4.4	58.1	54.2	3.9
IPSL-CM5A-MR	116950	-447.7	-437.5	-10.1	48.0	53.5	5.6	56.1	50.8	5.3
MIROC5	117233	-477.4	-470.3	-7.0	49.9	45.9	4.0	58.9	55.6	3.3
NORESM1-M	117191	-476.4	-467.8	-8.6	42.6	37.8	4.8	51.7	46.9	4.8

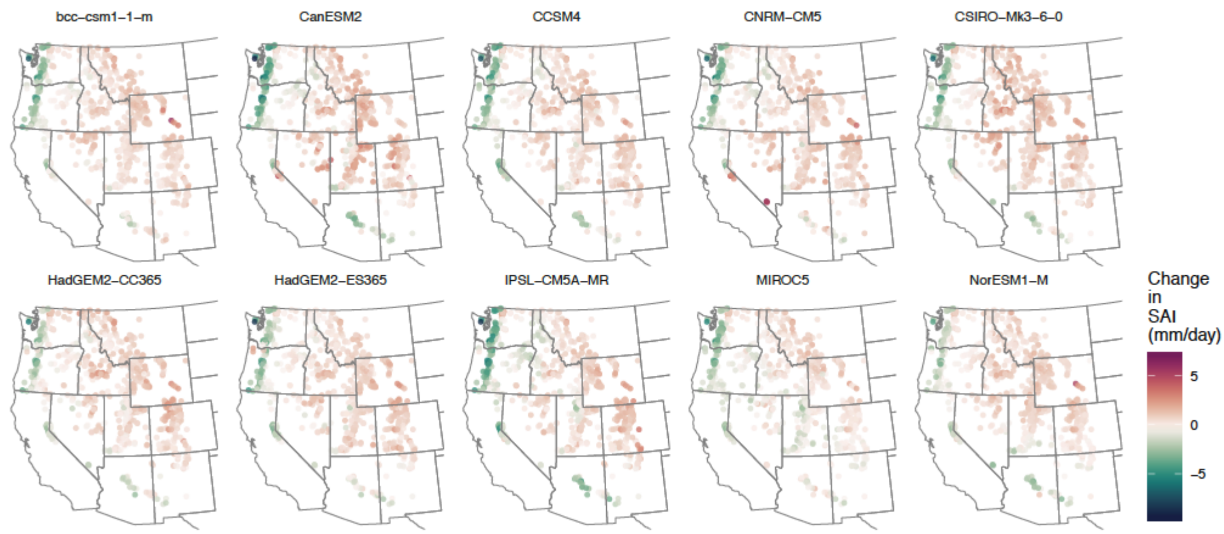

Figure C.6 Change in 30-year average SAI for each GCM from 1970-1999 to 2070-2099. Spatial patterns of trend are similar between models, though magnitudes of change differ.

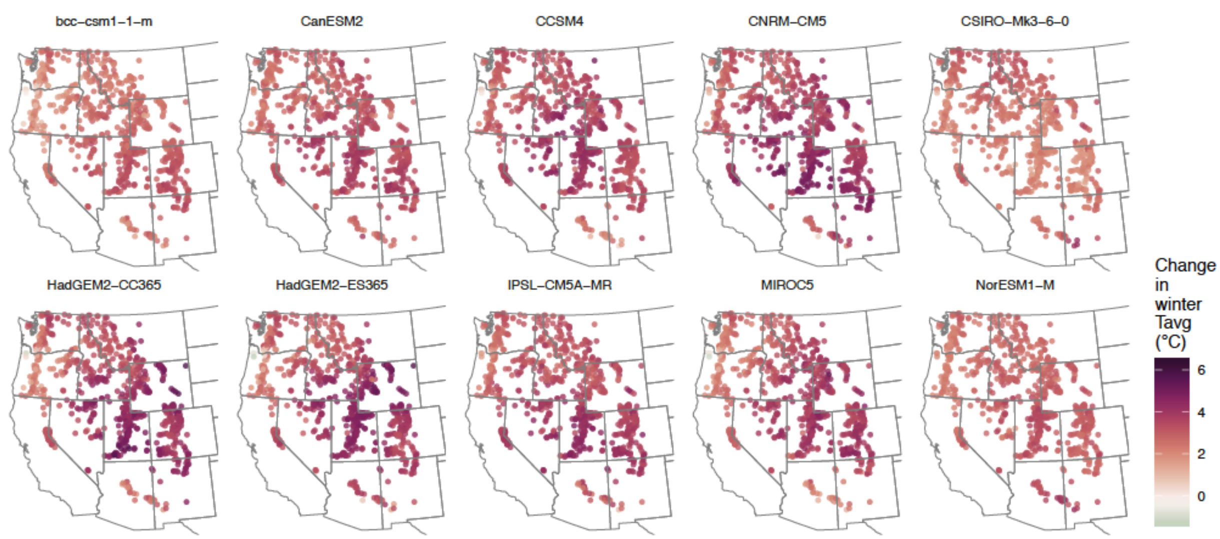

Figure C.7 Change in 30-year average winter T_{avg} for each GCM from 1970-1999 to 2070-2099. Winter T_{avg} consistently increases, with some variability in magnitude between GCMs.

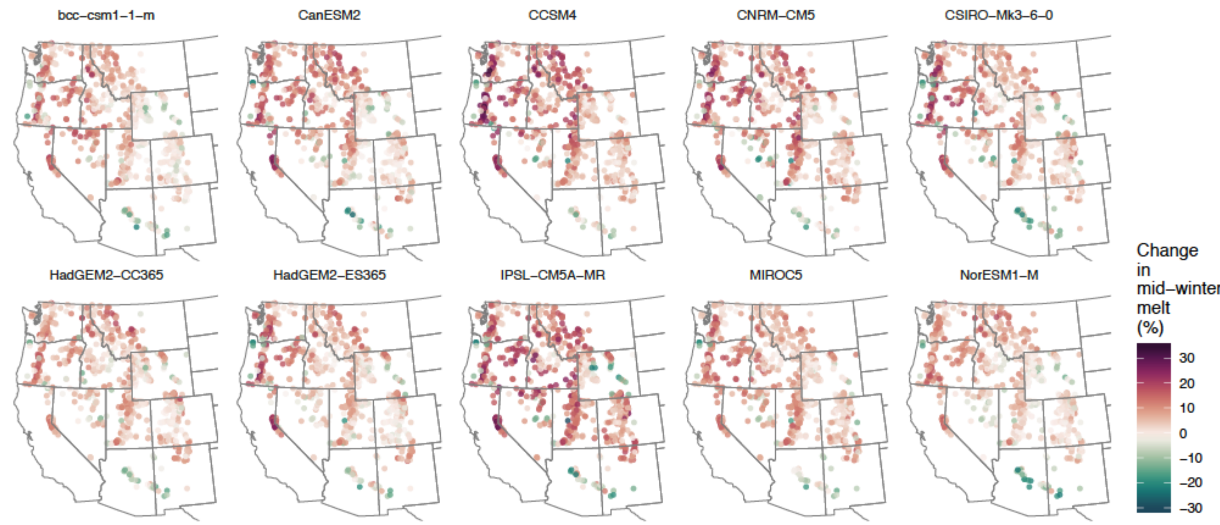

Figure C.8 Change in 30-year average winter ablation for each GCM from 1970-1999 to 2070-2099. Winter ablation fraction predominantly increases, with isolated sites showing decreases in winter ablation, particularly in the mountains of Arizona and New Mexico.

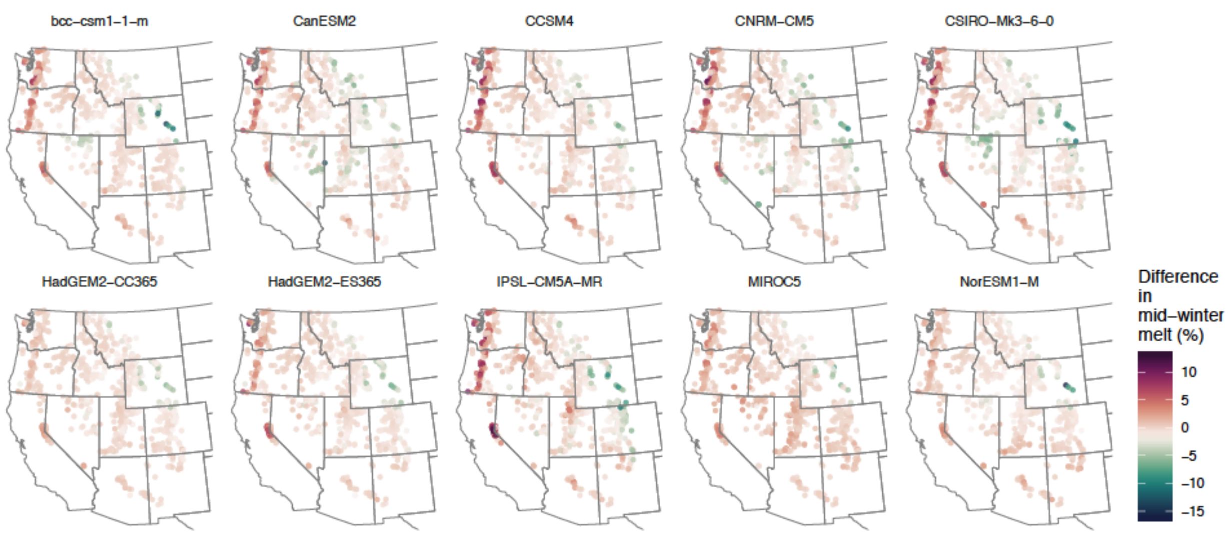

Figure C.9 Difference in winter ablation with detrended versus original data, averaged over water years 2070-2099. Spatial pattern is similar between GCMs, with differences in magnitude of effect.

Appendix D – Supplemental Information for Chapter 5

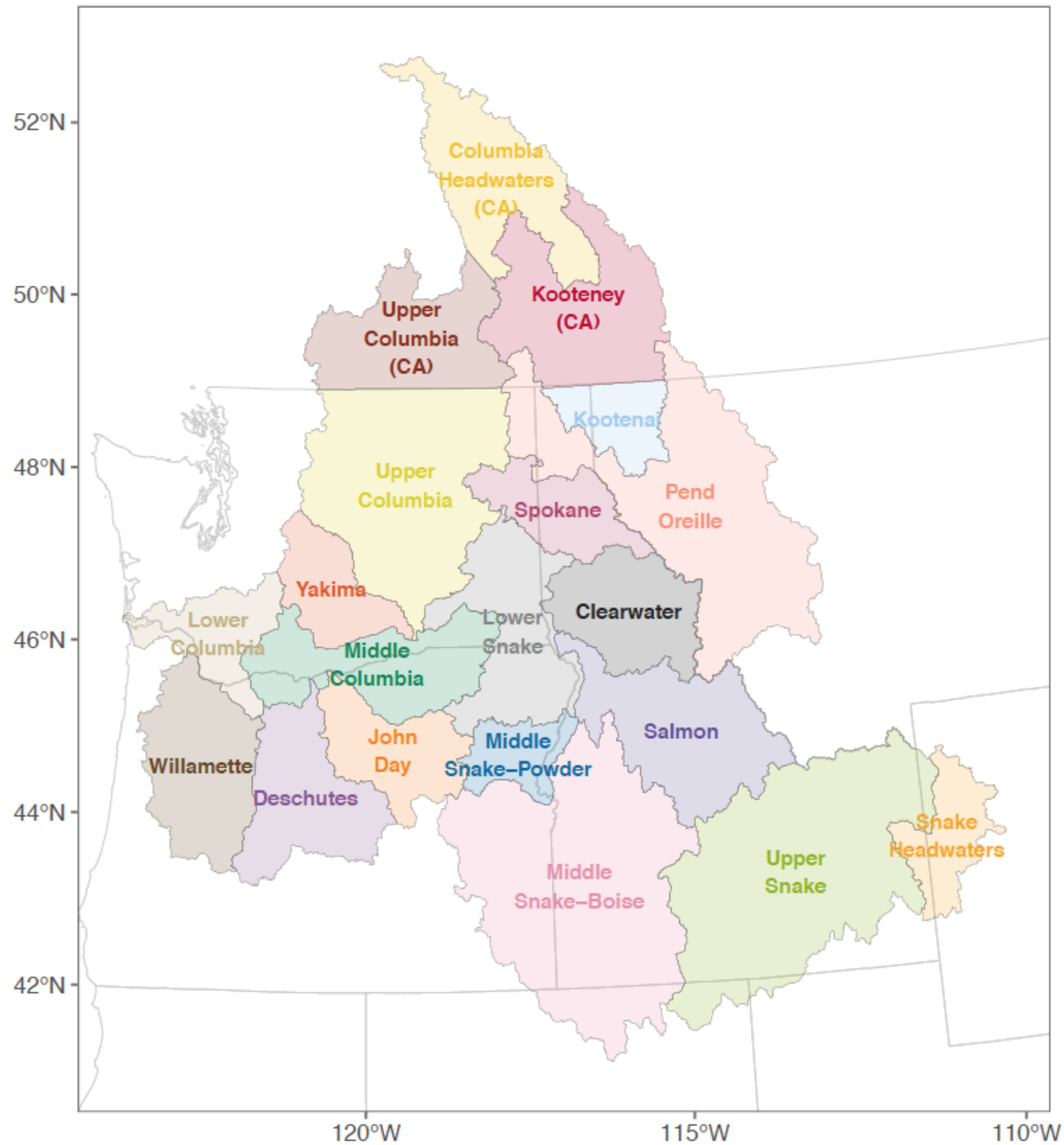

Figure D.1 HUC-6 units with names.

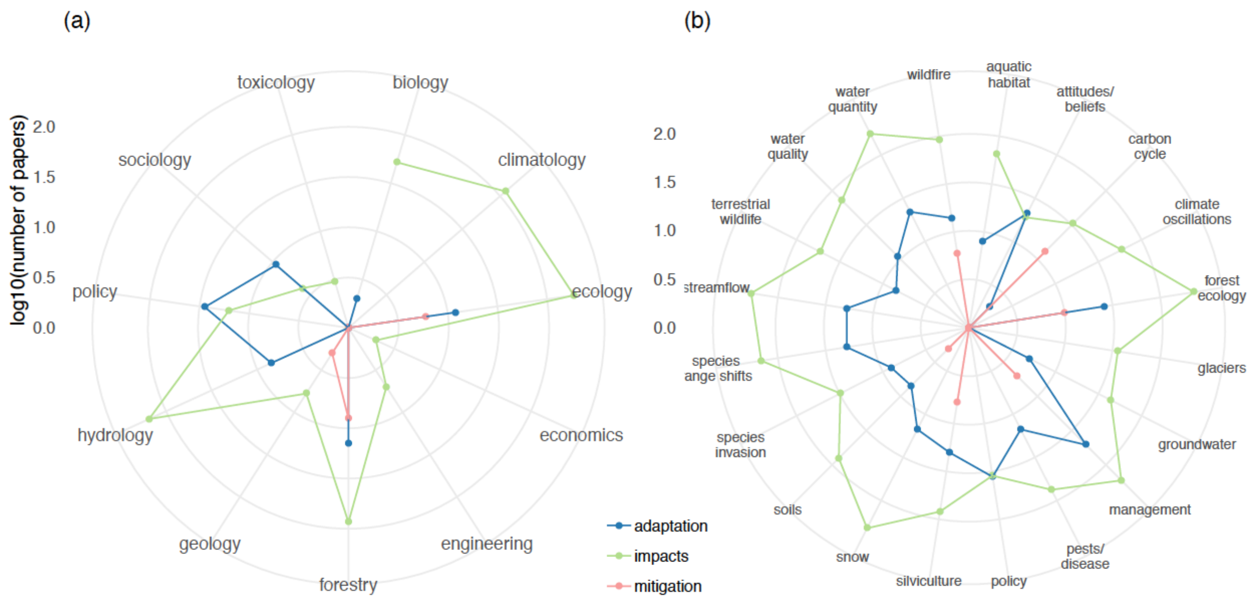

Figure D.2 As Figure 5.6, with number of papers, rather than percentages by category.

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