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Glacier variability in the conterminous United States during the twentieth century

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Abstract Glaciers of the conterminous United States have been receding for the past century. Since 1900 the recession has varied from a 24 % loss in area (Mt. Rainier, Washington) to a 66 % loss in the Lewis Range of Montana. The rates of retreat are generally similar with a rapid loss in the early decades of the 20th century, slowing in the 1950s–1970s, and a resumption of rapid retreat starting in the 1990s. Decadal estimates of changes in glacier area for a subset of 31 glaciers from 1900 to 2000 are used to test a snow water equivalent model that is subsequently employed to examine the effects of temperature and precipitation variability on annual glacier area changes for these glaciers. Model results indicate that both winter precipitation and winter temperature have been important climatic factors affecting the variability of glacier variability during the 20th Century. Most of the glaciers analyzed appear to be more sensitive to temperature variability than to precipitation variability. However, precipitation variability is important, especially for high elevation glaciers. Additionally, glaciers with areas greater than 1 km^2 are highly sensitive to variability in temperature.

1 Introduction

Over the past century, most alpine glaciers and ice caps have been responding to climate warming by shrinking (Dyurgerov and Meier 2000; Kaser et al. 2006). Most studies of glacier response to climate change have utilized estimates of changes in glacier mass from field-based measurements and compared these variations with variations of climate variables such as precipitation and air temperature (Tangborn 1980). This approach is the most direct and provides detailed insight into climatic variability governing glacier mass gain and loss. However, field efforts are expensive and time intensive such that only a few glaciers can be monitored by any one program. Although remotely assessing changes in glacier volume show great promise (Baltsavias et al. 2001; Arendt et al. 2008), these efforts have not yet

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developed the long time series necessary for comparing glacier change with climatic variations.

In this study, we make use of historic photographs and maps that date to about 1900 to develop a time series of glacier area change. While this approach is not as direct as using mass change, due to the dynamic complexities between changes in glacier mass and subsequent changes in glacier geometry (Cuffey and Paterson 2010) we assert that for long time series, changes in glacier area follow changes in glacier mass. By making use of area change data, we can expand the population of study glaciers and address regional variations in glacier area to obtain a more integrated perspective on the important factors driving the observed glacier changes. The above approach is used to examine long-term glacier variability because it is difficult and costly to monitor glaciers thus many long-term measurements of glacier variability do not exist.

For this study the region of interest is the glacierized alpine regions of the conterminous US. This region hosts about 8,300 glaciers and perennial snowfields totaling about 247 square kilometers (km^2) of ice (Fountain et al. 2007a, b). The pattern of glacier change in this region has not been well documented (except for Washington State) until recent years and no synthesis effort has been conducted.

2 Data and methods

Historic glacier areas were primarily derived from photographs and topographic maps. We used georectified aerial photos where possible to provide the most accurate assessment of glacier area. No vertical aerial photographs are available prior to about 1940 so glacier area was derived from ground-based photos. In all cases late-summer season photographs were used to examine minimum seasonal snow cover and to reduce errors in defining glacier boundaries. Historic maps were used when available, however the first mapping of glaciers in the conterminous US during the early 1900s depicted glaciers emblematically rather than representationally making their outlines unusable for change detection. The most extensive collection of glacier outlines is from 1:24,000 topographic maps produced by the US Geological Survey. The map collar, which includes the map identification, latitudes and longitudes, and related information, also includes the dates on which aerial photographs were acquired, thus providing a relatively precise date of glacier depiction. The data sources and methods used for different regions are derived from a number of reports (Sierra Nevada (Basagic and Fountain 2011), Oregon Cascades (Jackson and Fountain 2007), Front Range, Colorado (Hoffman et al. 2007), Mount Baker, Washington (Fountain et al. 2007a), Wind River Range, Wyoming (unpublished)) and methodological details for deriving glacier outlines from the 1:24,000 US Geological Survey maps can be found in Fountain et al. (2007b).

We compiled area estimates for 31 glaciers across the conterminous US for use in our study (Fig. 1a; Table 1). Because the glacier areas are based on episodic photographs and from some topographic maps, the time interval between areas for any given glacier varies. Using a linear regression, glacier area was interpolated to decadal intervals for the period 1900 through 2000. In some cases, we extrapolated back to 1900 to provide the same initial time for all glaciers. Time intervals less than a decade were not warranted given the sparse temporal dataset. We acknowledge that we are implicitly filtering glacier variability by relying on decadal intervals, and thereby removing shorter term variability, but the temporal sparseness of the data limits what can be done and we are interested in the longer-term controls and trends of glacier change. Finally, the glacier areas are converted to fractional area (0–1) relative to the original area in 1900.

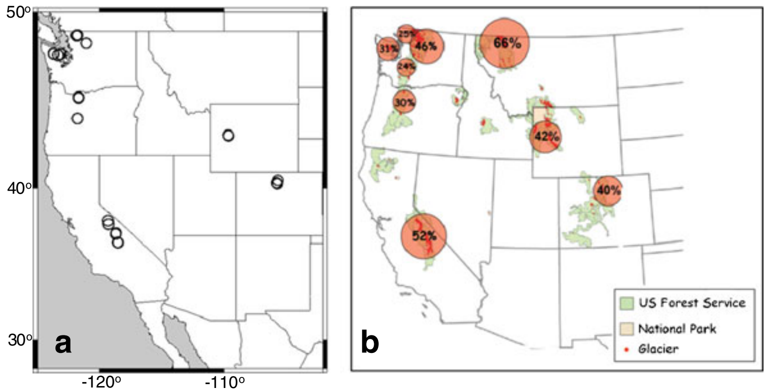


Fig. 1 **a** locations of glaciers analyzed in the study, and **b** fraction of glacier area lost since 1990

Monthly temperature and precipitation data for the period January 1895 through September 2008 were obtained from the Parameter-elevation Regression on Independent Slopes Model (PRISM) dataset (http://www.ocs.orst.edu/prism/). Monthly air temperature and precipitation on a 4-kilometer (km) by 4-km grid for the conterminous US (west of 102 ° west longitude, 209584 PRISM grid cells) were used to calculate monthly values of snow accumulation and melt (snow model), which were used to estimate monthly snow water equivalent (SWE) at each grid cell. The snow model estimates accumulated snow in winter and melt in summer. At the highest elevations seasonal winter snow may survive through the summer and contributes to the total snowpack for the following year. Model-estimated SWE for the years 1900 through 2008 were used for analysis. SWE estimates for 1895 through 1899 were discarded to avoid effects of initial model conditions on SWE estimates. Computed March SWE (hereafter referred to as SWE) is the month of greatest snow accumulation at the lower elevations in the alpine environment and was used as a proxy for annual glacier area (McCabe 1996; Serreze et al. 1999; Bohr and Aguado 2001; Clark et al. 2001).

2.1 The snow accumulation and melt model

Hydrologic models have been used in a number of studies to simulate snow cover, depth, and water equivalent (McCabe and Wolock 1999; Hamlet et al. 2005; Mote et al. 2005; McCabe and Wolock 2008; McCabe and Wolock 2011). The snow model used in this study is based on concepts previously used in monthly water balance models (McCabe and Ayers 1989; Tarboton et al. 1991; Tasker et al. 1991; McCabe and Wolock 1999; Wolock and McCabe 1999; McCabe and Wolock 2008; McCabe and Wolock 2011). Inputs to the model are monthly temperature (T) and precipitation (P); the occurrence of snow is computed as:

$$S = \begin{cases} P\, for\, T_a \leq T_{snow} \\ P\left(\frac{T_{rain}-T_a}{T_{rain}-T_{snow}}\right) for\, T_{snow} < T_a < T_{rain} \\ 0\, for\, T_a \geq T_{rain} \end{cases} \quad (1)$$

where S is monthly snow fall in millimeters (mm), P is monthly precipitation in mm, T_a is monthly air temperature in degrees Celsius (°C), T_{rain} is a threshold above which

Table 1 List of glaciers used in this study from north to south and west to east. The asterisk (*) by the glacier name indicates a commonly used informal name where no official name exists. The elevation (Elev) is the mean elevation in meters above sea level (m asl) of the glacier when the most recent topographic maps were made, typically in the 1970s–1980s. The area is the estimated 1980 area used in the analysis

Glacier name	Mountain/Range	State	Latitude	Longitude	Elev (m asl)	Area (km^2)
Rainbow	Mt Baker	WA	48.799	−121.778	1,818	1.92
Roosevelt	Mt Baker	WA	48.792	−121.831	2,241	4.10
Coleman	Mt Baker	WA	48.784	−121.850	2,166	5.02
Boulder	Mt Baker	WA	48.771	−121.792	2,271	4.15
Deming	Mt Baker	WA	48.761	−121.841	2,271	4.80
Easton	Mt Baker	WA	48.750	−121.830	2,176	3.30
So. Cascade	Cascades	WA	48.355	−121.058	1,916	2.26
Blue	Olympic	WA	47.810	−123.696	1,878	3.90
Hoh	Olympic	WA	47.795	−123.677	1,788	4.60
Constance*	Olympic	WA	47.777	−123.128	1,732	0.19
Eel	Olympic	WA	47.727	−123.339	1,853	0.90
Anderson	Olympic	WA	47.713	−123.335	1,690	0.50
Ladd	Mt Hood	OR	45.392	−121.706	2,211	0.90
Coe	Mt Hood	OR	45.388	−121.695	2,402	1.25
Eliot	Mt Hood	OR	45.381	−121.681	2,473	1.79
NewtonClark	Mt Hood	OR	45.368	−121.682	2,665	1.58
White river	Mt Hood	OR	45.353	−121.697	2,353	0.58
Collier	Three Sisters	OR	44.165	−121.785	2,534	0.69
Conness	Sierra Nevada	CA	37.970	−119.319	3,561	0.19
Lyell east*	Sierra Nevada	CA	37.744	−119.264	3,648	0.24
Lyell west*	Sierra Nevada	CA	37.743	−119.272	3,762	0.32
Darwin	Sierra Nevada	CA	37.171	−118.677	3,871	0.12
Goddard*	Sierra Nevada	CA	37.108	−118.703	3,732	0.20
Lilliput	Sierra Nevada	CA	36.572	−118.553	3,361	0.06
Pickett*	Sierra Nevada	CA	36.559	−118.508	3,734	0.10
Grinnell	Lewis Range	MT	48.752	−113.727	2,054	0.99
Sperry	Lewis Range	MT	48.625	−113.757	2,424	1.05
Gannett	Wind River	WY	43.200	−109.656	3,721	3.30
Dinwoody	Wind River	WY	43.174	−109.636	3,633	2.72
Rowe	Front Range	CO	40.488	−105.646	4,022	0.05
Sprague	Front Range	CO	40.342	−105.734	3,644	0.03
Tyndall	Front Range	CO	40.305	−105.690	3,707	0.04
Andrews	Front Range	CO	40.288	−105.684	3,578	0.08

all monthly precipitation is rain, and T_{snow} is a threshold below which all monthly precipitation is snow. When the monthly air temperature is between T_{rain} and T_{snow}, the proportion of precipitation that is snow or rain changes linearly.

Snow that accumulates during the previous month is added to the current snowpack and is subject to melt if the air temperature is sufficiently warm. Thus, for some cases, snow, rain, and snowmelt can occur in the same month.

Snow melt is computed using a degree-day method of the following form:

$$M = \alpha(T_{air} - T_{snow})d, \quad (2)$$

where M is the amount of snow melted in a month, α is a melt rate coefficient, and d is the number of days in a month. This type of snowmelt model has been used in previous research (e.g. Rango and Martinec 1995).

2.2 Snow model calibration

Parameters for the snow accumulation and melt model (T_{rain}, T_{snow}, α) were taken from previous research and application of the snow model (McCabe and Wolock 2009); T_{rain}= 3.0 °C; T_{snow}=−1.0 °C; and α=0.5. The calibration of the snow model parameters involved comparing the model-estimated SWE with measured values for 314 sites across the western US (McCabe and Wolock 2009). These parameter values are similar to values used with similar models in previous studies. For example, Tarboton et al. (1991) reported Train=3.3 °C and Tsnow=−1.1 °C for use with a monthly time step snow model, and McCabe and Wolock (1999) used Train=5.0 °C and Tsnow=0.0 °C to estimate regionally averaged April 1 SWE for the western US. Additionally, the melt rate coefficient of 0.5 is within the range of values reported by Rango and Martinec (1995). Using these parameters, the model was used to estimate March SWE for the years 1900 through 2008 for each of the PRISM grid cells in the conterminous US.

We used March SWE, rather and September SWE as the annual mass change of the glacier. It can be shown from general principles that a change in mass (or glacier volume) is related to changes in length (Jóhannesson et al. 1989; Nye 1951; 1959). To improve the March SWE as a proxy for glacier area change, the time series for each glacier was smoothed using a 10-year backwards moving average. That is, the average of the 10 years replaces the value for the 10th year in the interval. This is an attempt to replicate the time-scale response of glaciers to a change in mass input (Jóhannesson et al. 1989; Schwitter and Raymond 1993). Small glaciers, typical of those in the conterminous US, have response times of 10 years or less (Nylen 2004; Basagic and Fountain 2011). Without the moving average, the glacier would react immediately yielding an unrealistic flashy response to changes in snowpack.

Because the objective of this study is to examine the spatial and temporal variability in glacier area, the variability of each data set through time is the salient feature of interest. To compare the glacier area data with the SWE estimates, the values of fractional glacier area and smoothed estimated SWE were converted to standardized departures, computed as:

$$z_{ij} = \frac{(x_{ij} - ave_j)}{std_j}, \quad (3)$$

where z_{ij} is the standardized departure for year i at location j, x_{ij} is the original data value for year i at location j, ave_j is the average data value at location j, and std_j is the standard deviation of the data at location j. Transformation into standardized departures causes each data time series to have a zero mean and unit variance. For the remainder of the study standardized departures of the glacier area and the estimated SWE are used.

Given the complexity of alpine terrain and the smoothing inherent in the PRISM methodology used to interpolate data to 4-km by 4-km grid cells, the PRISM cells within 30 km of each glacier were searched to identify the grid cell with the highest correlation between glacier area and SWE. The SWE time series for the PRISM grid cell with the highest correlation was selected for analysis. Figure 2 illustrates comparisons of standardized departures of decadal glacier area time series with decadal model-estimated SWE time series for each glacier included in the study. The median correlation (of the best correlating cell) among all 31 glaciers is 0.68, with a 25th percentile of 0.55 and a 75th percentile of 0.78. The correlations indicate that for most of the glaciers the variability of estimated SWE closely follows the variability of the measured glacier data. Based on these results estimated SWE was used as a proxy for annual glacier variability.

The snow model was used to examine the relative contributions of temperature and precipitation on SWE estimates, and therefore glacier area change, for the 31 glaciers through sensitivity experiments. In the first experiment SWE is estimated using the monthly PRISM precipitation data and long-term mean monthly values of temperature. In this case, the monthly temperature is the mean for that month over the entire period of record. Therefore, only precipitation varied over the period of record and affected the variability in SWE (variable-precipitation (varP) model). In the second experiment SWE is estimated using the monthly PRISM temperature data and long-term mean monthly values of

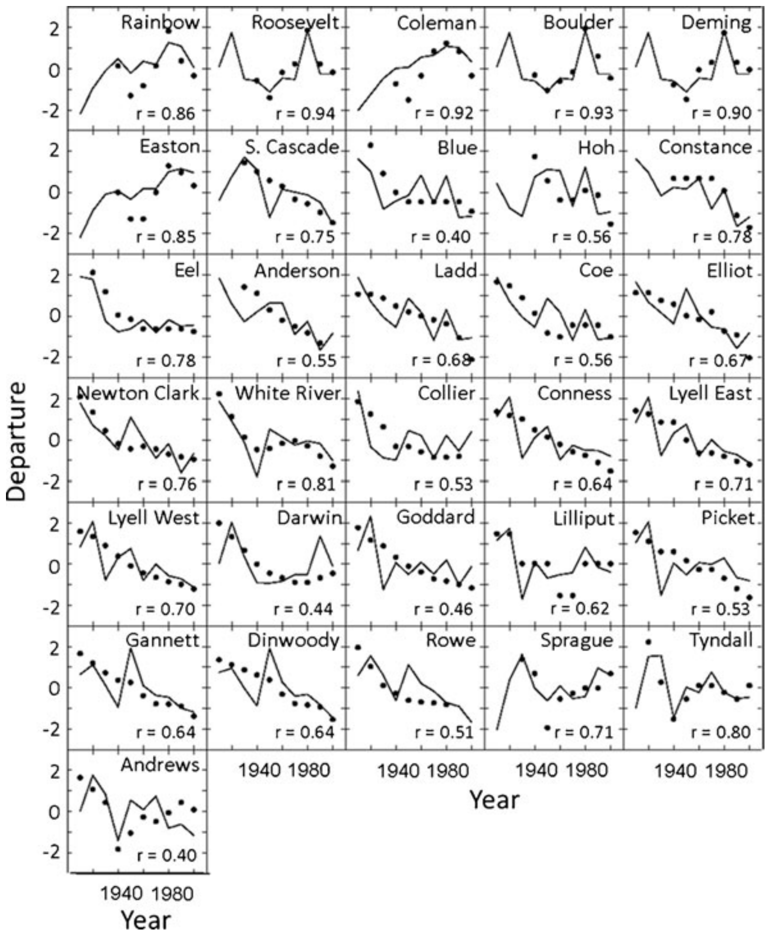


Fig. 2 Comparison of standardized departures of decadal values of measured (*dots*) and snow-model estimated (*lines*) glacier area for 31 sites in the western United States

precipitation (variable-temperature (varT) model). SWE from these experiments was compared with the original SWE computed using the ‘complete model’ that allowed both monthly temperature and precipitation to vary. Just as was done for SWE from the complete model, the SWE time series for each experiment were smoothed with the 10-year backwards moving average and then standardized.

Because of the large temporal autocorrelations in the 10-year moving average SWE time series, a Monte Carlo analysis was used to determine if correlations between time series were statistically significant at a 95 % confidence level ($p<0.05$) (Livezey and Chen 1983; McCabe and Dettinger 1999). The Monte Carlo analysis involved simulating 1,000 100-year long time series of SWE using a stochastic model. The simulated time series preserved the mean, variance, skew, and lag-1 autocorrelations in the SWE time series computed using the complete model for each of the 31 glaciers. The 1,000 simulated SWE time series for each of the 31 glaciers were correlated with SWE time series for each glacier computed using the varP and varT models. From the 1,000 correlations computed for each glacier (and each model) the 5th and 95th percentile correlation coefficients were computed and these values were used to specify the two-tailed 95 % significance level of correlations (for each glacier and each model).

This approach was used rather than simply comparing time series of precipitation and temperature with SWE because the model allows for the interaction of precipitation and temperature. For example, the area of a high-elevation glacier may increase due to positive correlations with increasing winter precipitation and with warming winter air temperatures. However, the correlation with warming winter air temperatures are spurious because the glacier is located at such a high elevation that winter temperatures remain well below freezing thus having little effect on changes in SWE. Interpretations based on simple correlations between time series of glacier area and that of precipitation or temperature can be incomplete.

3 Results and discussion

Obtaining glacier change from historic photographs posed a number of problems. First, photographs had to be taken in late summer to minimize seasonal snow cover and depending on the year late snow cover could obscure glacier boundaries. In addition, rock debris cover and terrain shadows made defining glacier perimeter difficult in numerous instances. We relied on vertical aerial photographs to the fullest extent possible to minimize distortions posed by oblique photographs. Prior to the 1940s all of our data are based on hand-held oblique photographs. We made every attempt to represent the glacier perimeter accurately in a vertical, map-view perspective by noting identical landscape features in the vertical and oblique photographs. From this effort we were able to identify glacier changes for more than several hundred glaciers. However, only a relatively small subset, 50, has data that go back to 1900 (Fig. 1b).

Glacier change since 1900 shows, as expected, a decrease in glacier cover. This is consistent with global trends (Kaser et al. 2006). Perhaps not surprisingly, the change is not spatially uniform over the conterminous US. Extensive shrinkage occurs in the Lewis Range (Glacier National Park) of Montana and the Sierra Nevada of California. The least shrinkage is found in the Pacific Northwest of Oregon and Washington. One has to be careful in drawing too many direct inferences about spatial variations in glacier change, however, due to the differences in topographic setting and local influences. For example, the largest glaciers (~5 km^2) occupy the stratovolcanoes of the Pacific Northwest and descend

2,000 m from elevations of ~4,000 m above sea level (asl). In contrast, the glaciers in the Sierra Nevada are smaller than 0.1 km^2 and descend 300 m from ~3,000 m asl. For the Sierra glaciers, local topographic shading and extra snow accumulation from avalanching play an important role in maintaining the glacier balance (Nylen 2004; Basagic and Fountain 2011), compared to those glaciers on Mt. Rainier in Washington.

To identify temporal changes over the past century, the 31 glaciers had sufficient data to reconstruct decadal-scale trends. Most of the glaciers included in this study show negative long-term trends in SWE during the 20th Century (Fig. 2). Of the 31 glaciers analyzed, 24 indicate long-term decreases in SWE, consistent with the findings of previous glacier area studies (Hoffman et al. 2007; Jackson and Fountain 2007; Sitts et al. 2010; Basagic and Fountain 2011). These results are further supported by snow studies that have shown decreasing winter snow accumulations in the conterminous US (Tangborn et al. 1977; Mote 2003; Mote et al. 2005; McCabe and Wolock 2010); decreases in the ratio of winter snow to winter precipitation (Knowles et al. 2006; Barnett et al. 2008; Bonfils et al. 2008; Pierce et al. 2008); and changes in the occurrence and frequency of rain-on-snow events (McCabe et al. 2007). The decreases in snow accumulations and associated hydroclimatic conditions in the conterminous US primarily have been attributed to increases in winter and summer air temperatures (Knowles et al. 2006; Barnett et al. 2008; Bonfils et al. 2008; Pierce et al. 2008).

To determine whether air temperature or precipitation have the stronger influence on the century-scale change in glacier recession, we compared trends in SWE using the complete model to trends using the varT and varP models. Results showed similar negative trends in SWE using the varT model (Fig. 3) indicating a strong influence of warming air temperatures over the past century in controlling glacier recession. In contrast, trends in SWE using the varP model showed inconsistent results with both positive and negative trends, further underscoring the importance of air temperature driving decreases in SWE and glacier recession during the 20th century. For some glaciers, however, trends in precipitation are also negative indicating that in those settings precipitation also is an important climatic factor associated with decreases in SWE. For a few other glaciers positive trends in SWE (glacier growth) using the complete model are associated with positive trends in varT and varP. These sites include three glaciers on Mount Baker that have experienced a positive trend in winter precipitation over the past few decades, which has buffered glacier mass

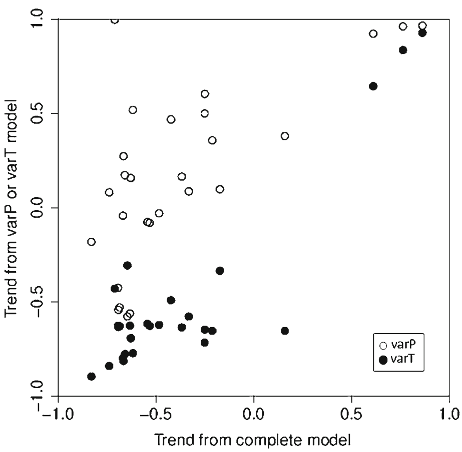


Fig. 3 Comparison of trends in March snow water equivalent (SWE) computed using the complete model compared with trends in SWE computed using the variable precipitation (varP) and variable temperature (varT) models

losses and compensated for reduced snow accumulation due to increasing rain events rather than snow, and due to increased summer melting. This buffering ended during the last decade due to a rapid rise in air temperature without a concomitant increase in winter precipitation.

To examine the controls on decadal-scale glacier variability, the time series of smoothed and standardized SWE for the 31 sites for the varP and varT models, were correlated with the smoothed and standardized SWE time series of the complete model. The results indicate substantial positive correlations for many of the 31 sites for both the varP and varT models. For the varP model half of the sites indicate correlations greater than 0.50, and for the varT model half of the sites indicate a correlation of 0.83, and 75 % indicate a correlation greater than 0.60. These results indicate that both precipitation and temperature have important influences on glacier variability at decadal time scales, with temperature having a greater influence on glacier variability for most glaciers.

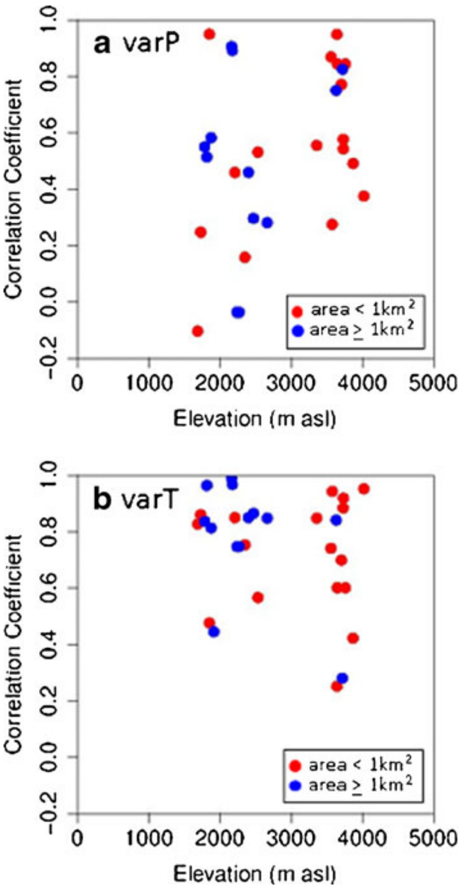
There is a weak but positive correlation ($r=0.42$, $p<0.05$) between elevation and the sensitivity of glaciers to precipitation variability (precipitation sensitivity is expressed as the correlation between SWE computed using the complete model and SWE computed using the varP model) (Fig. 4a). The glaciers most sensitive to precipitation variability are located at high elevations and are located in the interior west and at high elevations of the west coast in the Sierra Nevada of California. In these environments winters are sufficiently cold that historic warming of winter air temperature does not affect the phase of the precipitation. Summer air temperatures, although warming with time, do not influence the decadal variability as much as winter precipitation at high elevation sites. For example, in a region in the northwest dominated by the effects of air temperature, variations in snowfall can be magnified locally by avalanching from the surrounding terrain reducing the glacier's responsiveness to temperature alone.

The correlation between elevation and the sensitivity of glaciers to temperature (temperature sensitivity is expressed as a correlation between SWE computed using the complete model and SWE computed using the varT model) is negative but non-significant ($r=-0.23$) (Fig. 4b) The negative correlation between elevation and glacier temperature sensitivity indicates that glaciers at low elevations are slighty more sensitive to temperature changes than are high elevation glaciers. This finding is consistent with the snow studies in the Pacific Northwest. Relatively warm winters often hover near 0 °C such that a small change in air temperature changes the phase of precipitation (i.e. rain versus snow). Decreasing fractions of snow to total winter precipitation, due to climate warming since 1950, have reduced the water equivalent content of spring snowpacks for low elevation glaciers (Mote et al. 2005; McCabe et al. 2007). Reduced winter snow accumulation starves the glacier of mass and the glacier shrinks, accordingly.

Additional results indicate that glacier size does not appear to have much of an effect on glacier sensitivity to precipitation variability (Fig. 4a). In contrast, all but two of the largest glaciers are highly sensitive to temperature variability (Fig. 4b). Our results for glacier recession is similar to results elsewhere. In southern British Columbia, glacier shrinkage between about 1900 and 2005 was 60 % (Koch et al. 2009) with century-scale recession governed by warming air temperatures and decadal variations influenced by air temperature and precipitation. These results also mirror global variation of glaciers (Kaser et al. 2006).

To determine if winter or summer temperature and precipitation had the greatest effect on glacier area over the 20th century, other model experiments were performed in which winter or summer temperature and precipitation was allowed to vary according to the data record while using the monthly average over the entire record for the other variables. This resulted in four more experiments (variable winter (October through March) precipitation ($\mathrm{varP_{win}}$),

Fig. 4 Comparison of elevation (in meters above sea level) with correlations of March snow water equivalent (SWE) computed using the complete model with SWE computed using the **a** variable-precipitation (varP) model, and **b** variable-temperature (varT) model. *Red dots* indicate glaciers with areas less than 1 square kilometer (km^2), and *blue dots* indicate glaciers with areas ≥1 km^2



variable summer (April through September) precipitation ($varP_{sum}$), variable winter temperature ($varT_{win}$), and variable summer temperature ($varT_{sum}$)). Results from these model experiments were also smoothed with a 10-year backwards moving average and standardized (Tangborn 1980).

Correlations between SWE using the complete model and SWE determined from the seasonal models ($varP_{win}$, $varP_{sum}$, $varT_{win}$, and $varT_{sum}$) indicate statistically significant positive correlations (at $p<0.05$) for most of the glaciers for the $varP_{win}$ and $varT_{win}$ models (Fig. 5). SWE simulated using the $varP_{sum}$ model is not well correlated, as expected, except for Mt. Baker in the far Northwest, and may be associated with extensive cloudiness reducing summer loss of snow and ice. The poor correlations of SWE computed using the $varT_{sum}$ model, with the exception of Mount Baker, is a bit surprising and contrary to local studies that have compared glacier area change with trends in precipitation and temperature. In these studies, typically long-term trends in climate variables are compared with trends in glacier mass or area, whereas decadal glacier variability is examined here. For reasons that are unclear the only glaciers with substantial correlations with summer temperature are the most northerly ones. These results suggest that changes in winter climate have had a larger effect on glacier variability than have changes in summer climate. These results support

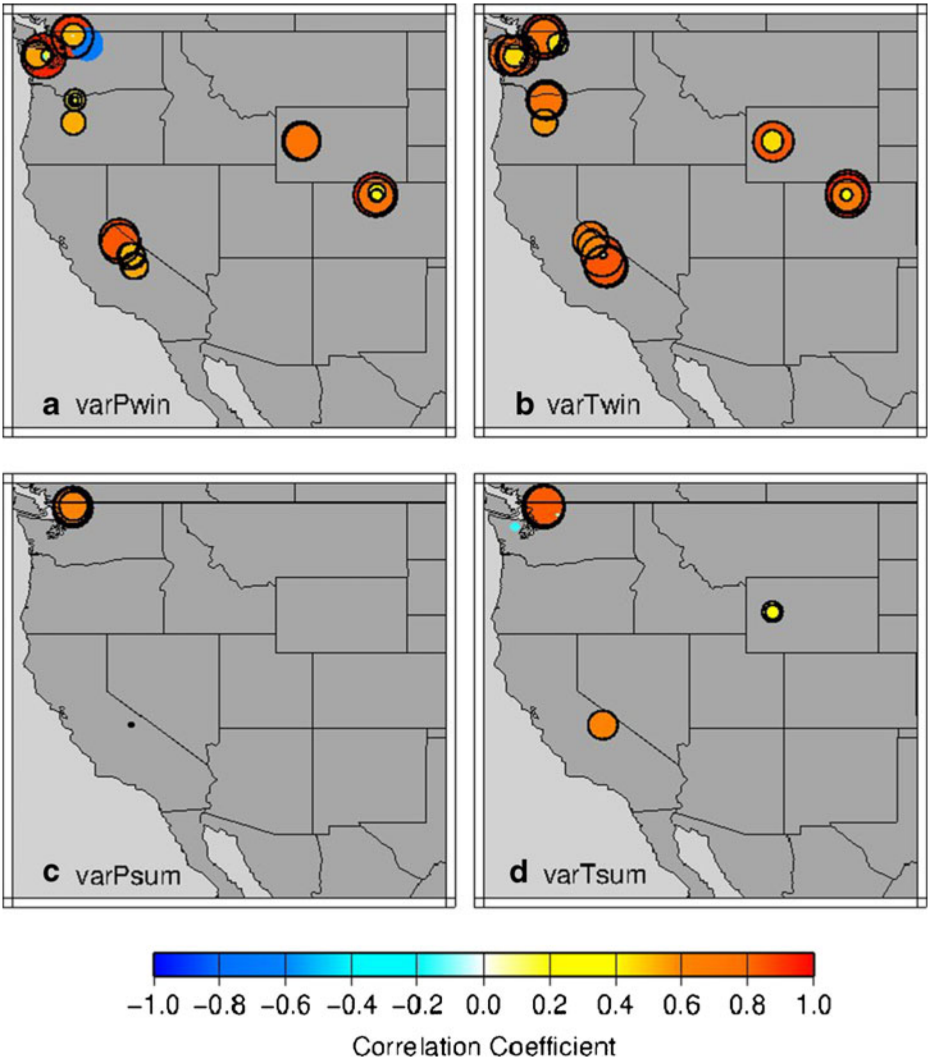


Fig. 5 Correlations between March snow water equivalent (SWE) computed using the complete model and SWE computed using the **a** variable winter precipitation (varPwin) model, **b** variable winter temperature (varTwin) model, **c** variable summer precipitation (varPsum) model, and **d** variable summer temperature (varTsum) model. The size of the circles is indicative of the magnitude of the absolute value of the correlation, and those outlined in black indicate correlations significant at a 95 % confidence level

previous findings (McCabe and Fountain 1995) that changes in winter snow accumulation were most correlated with changes in mass balance at South Cascade Glacier and that the reduction in glacier mass since the winter of 1986/87 were associated with the reduction of winter snowfall. Decreasing winter snow accumulation presents a doubly negative effect on glacier mass by reducing the mass contribution to the glacier from snowfall and the thin seasonal snowpack exposes the glacier ice earlier in spring to melting and mass loss.

4 Conclusions

The temporal and spatial variability of 31 glaciers in the conterminous US during 1900 to 2000 was examined using modeled estimates of snow water equivalent (SWE) as a proxy. Our simple model of SWE is well correlated with glacier area changes providing confidence in the results of our modeling experiments. Glacier shrinkage over the century time scale is largely controlled by warming air temperatures. No century-scale trend exists for precipitation. At decadal time scales changes in glacier area are largely controlled by temperature for most glaciers, however there is indication that precipitation variability also is important, expecially for high elevation glaciers.

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