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Interannual variations in snowpack in the Crown of the Continent Ecosystem

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Abstract:

Ecosystem changes such as glacier recession and alpine treeline advance have been documented over the previous 150 years in the Rocky Mountains of northern Montana and southern British Columbia and Alberta, a region known as the Crown of the Continent Ecosystem (CCE). Such changes are controlled, at least partially, by variations in snowpack. The CCE consists primarily of public lands, the majority of which is undeveloped or wilderness. Consequently, this region is well suited for an examination of long-term snowpack variation and associated ecosystem change. Data from nine SNOTEL sites provide an indication of the daily accumulation and ablation of snowpack over the period 1977–2001, as well as the relationship between precipitation, temperature and snowpack. 1 April data from 21 snow courses indicated the extent of regional snowpack variation and trends over the period 1950–2001, and 1 May data from three snow courses in Glacier National Park allow this record to be extended back to 1922. SNOTEL data suggest CCE snowpacks are larger and more persistent than in most regions of the western USA, and that water year precipitation is the primary control on 1 April snow water equivalent (SWE). Snow course data indicate that variations in both 1 April and 1 May mean SWE are closely tied to the Pacific decadal oscillation, an El Niño–southern oscillation-like interdecadal pattern of Pacific Ocean climate variability. Despite relatively stable snowpacks and summer temperatures since 1922, the glaciers in Glacier National Park have receded steadily during this period, implying a significant climatic shift between their Little Ice Age glacial maxima (*ca* 1860) and 1922. Published in 2002 by John Wiley & Sons, Ltd.

KEY WORDS Crown of the Continent Ecosystem; northern Rocky Mountains; Glacier National Park; snowpack; SNOTEL; Pacific decadal oscillation; glaciers; alpine treeline

INTRODUCTION

Long-term climatic variability across the world's mountains is being studied because mountains cover approximately one-fifth of the terrestrial surface of the globe, provide 50% of the freshwater humans consume, are ranked high in biodiversity and, because of their spatial complexity, present a challenge to our understanding of atmosphere–landscape interactions (Messerli and Ives, 1997). The Rocky Mountains of northern Montana and southern Alberta and British Columbia comprise the Crown of the Continent Ecosystem (CCE), an area that encompasses more than 10 000 km^2 of designated wilderness and national parks (Figure 1). Within this region, numerous ecosystem changes have been documented since the end of the Little Ice Age (*ca* 1850) that are likely driven by changes in climate. In Glacier National Park, Montana, USA, fewer than 37 glaciers remain of the 150 estimated to have been present around 1850 (Carrara, 1989). Key *et al.* (in press) document a reduction in glacial ice and perennial snow cover from 99 to 27 km^2 during the past 150 years and Hall and Fagre (in press) estimate that all glaciers are likely to melt by 2030. Alpine treelines have advanced upward in elevation (Butler and DeChano, 2001), have increased in biomass as spaces between patches have filled in (Klasner and Fagre, 2002), and many trees have changed from the prostrate, krummholz form to begin growing as upright trees (Butler *et al.*, 1994; Klasner and Fagre, 2002). Annual mean precipitation for nearby Kalispell has actually increased during the last century (0·09 cm year^{-1}, $p = 0{\cdot}03$). Because winter

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snow accumulation is a major contributor to glacier mass balances, and the timing and magnitude of seasonal snowpacks are the proximate drivers for tree seedling establishment and growth at upper elevation sites (Peterson, 1998), we hypothesized that long-term decreases in snowpack size and duration had occurred in the CCE despite slight increases in annual precipitation. Furthermore, changes in glacier mass balance and alpine tree establishment and growth have occurred in episodic fashion rather than incrementally, suggesting that snowpack variation may have similar periodicity.

Efforts have been made to examine the spatial and temporal characteristics of mountain snowpack in the western USA, trends in snow cover, and potential future responses to simulated climate change scenarios. Dettinger and Cayan (1995) indicated that snowmelt and runoff are beginning earlier in northern and central California, particularly in mid-elevation basins sensitive to changes in mean winter temperatures. Cayan (1996)

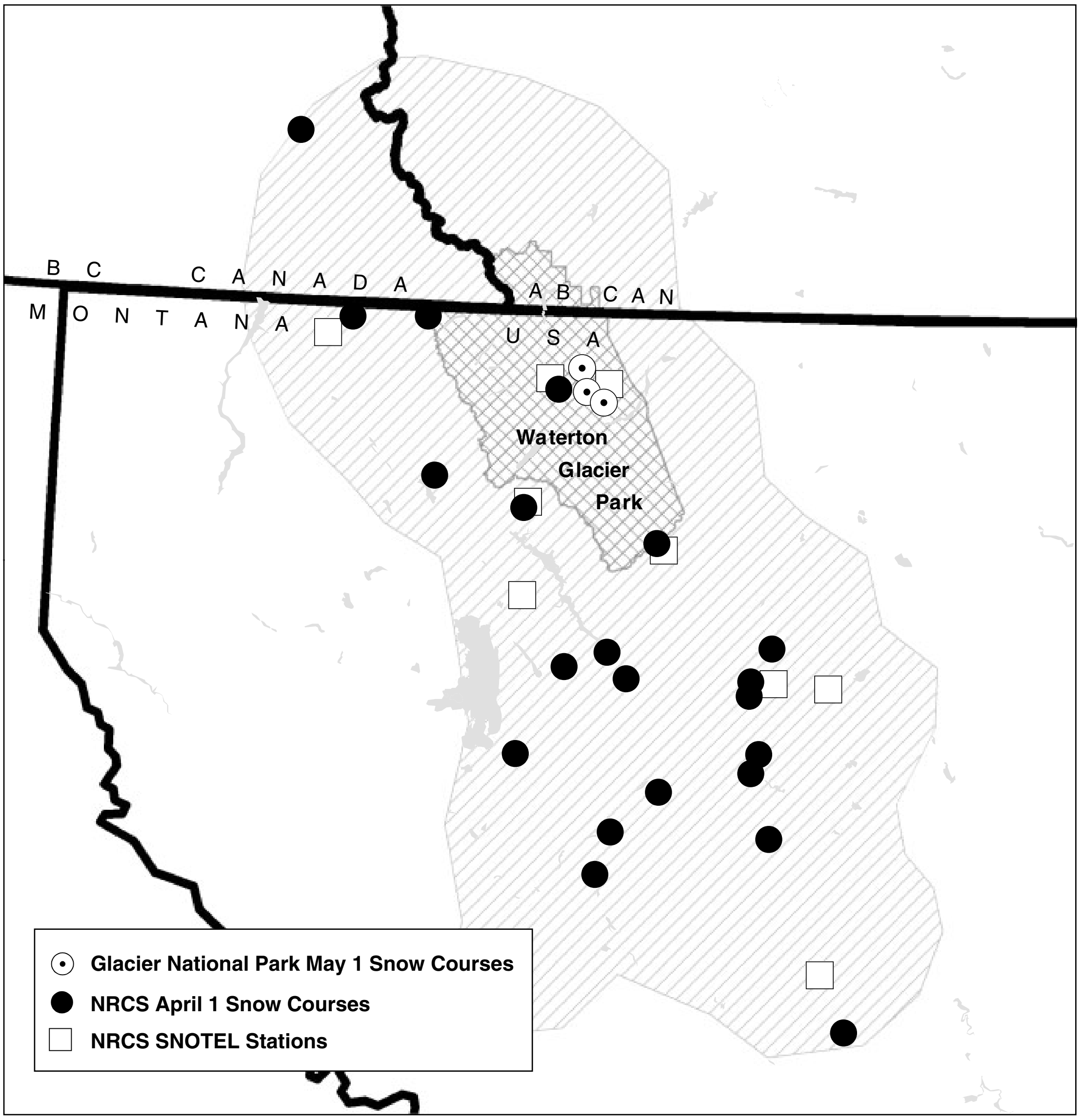

Figure 1. The Crown of the Continent region, with snow courses and SNOTEL stations examined in this study

demonstrated that precipitation anomalies have a greater effect on 1 April snow water equivalent (SWE) than temperature anomalies, excluding low-elevation areas in the Pacific Northwest. Using data from SNOTEL stations, Serreze *et al*. (1999) compared eight regions in the west and found that, although variations in the SWE/precipitation ratio can have a large effect on snowpack variability in the coastal mountain ranges and in the southwest, snowpack variability in the colder interior regions of the west is driven primarily by available precipitation. Based on general circulation model simulations of future climate, McCabe and Wolock (1999) predicted that, despite increases in winter precipitation, increases in temperature over the next century will result in large decreases in 1 April SWE throughout the western USA. McGinnis (1997) estimated that the duration of snow cover on the Colorado Plateau will decrease by an average of 58 days, based on a carbon dioxide doubling ($2 \times CO_2$) scenario. Leung and Wigmosta (1999) estimated that, based on a regional climate model, snowpack will decrease by 60% in a representative watershed in the Pacific Northwest but only by 18% in a representative watershed in the CCE. More recently, multi-decadal patterns in western North American snowpack variation have been linked to Pacific decadal oscillation (PDO) indices (McCabe and Dettinger, 2002).

The CCE is well suited for an examination of long-term trends in snowpack because there are relatively few local influences on climate patterns, such as large urban areas, extensive irrigation, or major changes in land cover. This makes it easier to attribute changes in regional snowpack to long-term climatic variability without the confounding influence of major landscape change. Moreover, the CCE has some of the earliest systematically collected snow data in the western USA (dating back to 1922) with which to examine snowpack trends and relate them to ecosystem responses. As part of a 12 year global change research program, several modelling studies have demonstrated the CCE's potential sensitivity to long-term snowpack changes (White *et al*., 1998; Running and Nemani, 1991). Under one scenario of reduced snowpack, tree growth rates increased at upper elevations but were reduced at lower elevations, streamflows peaked earlier during spring run-off and many streams became ephemeral, and competitive relationships between grassland and forest communities were altered (White *et al*., 1998). Thus, cascading ecological effects of reduced snowpack have the potential to alter the CCE significantly.

Our objectives were to examine CCE snowpacks for spatial and temporal patterns that might explain the observed changes in key ecosystem features. Relevant long-term temporal patterns included trends that might reflect climate change (i.e. global warming), changes in interannual variability, or multi-decadal patterns. The PDO has recently been described by Mantua *et al*. (1997) and Zhang *et al*. (1997) and shown to have a strong influence on the state of natural resources in the Pacific Northwest regions of the USA (e.g. salmon productivity). Peterson and Peterson (2001) have described synchrony between high-elevation tree growth and PDO signals. They found that growth of hemlock (*Tsuga mertensiana*) was negatively correlated with spring snow depth and positively correlated with the winter PDO index. Because similar natural resources have shown distinct changes in the CCE region, we examined our regional snowpack in relation to the PDO.

STUDY SITE

The 'Crown of the Continent' term was first applied to the Rocky Mountains near the USA–Canada border by George Bird Grinnell, an early advocate of establishing national parks in this region. The CCE extends from Fernie, British Columbia, south to just north of Missoula and Helena in Montana. It encompasses Waterton Lakes National Park (Canada) and Glacier National Park (USA), which were designated the Waterton–Glacier International Peace Park in 1932. The Bob Marshall Wilderness Complex, extensive national, state, and provincial forest lands, and the Blackfoot Indian Reservation surround the national parks to form a relatively unaltered landscape when contrasted with other areas of western North America. The CCE is a snow-dominated region, with over 70% of the annual precipitation falling as snow at higher elevations, which remain snow-free for as little as 6 weeks in late summer. The CCE is the headwaters for its region. Elevations range from 800 m in valley bottoms to 3200 m peaks comprised of sedimentary rock up to 1·3 billion years old. The mountain topography was extensively reshaped by glaciation. Expansive conifer forests cover approximately 75% of the area. This region contains relatively intact floral and faunal assemblages. Species distribution and abundance vary along elevational gradients (extending to alpine vegetation) and from west to east (including grassland). Climate is controlled by dominant air masses, with areas west of the Continental Divide receiving a stronger maritime influence from the Pacific Ocean and areas east of the Continental Divide having a distinctly more continental climate. Precipitation varies dramatically between high-elevation sites located near the Continental Divide and lower elevation sites along the plains near the eastern edges of the region. For example, precipitation varies from 350 cm year^{-1} (west side, high elevation) to 40 cm year^{-1} (east side, low elevation). Other factors, including dessicating east-side winds, can enhance smaller differences in precipitation regimes between the east and west sides. This contrast in precipitation and other climatic factors over relatively small distances has a profound impact on microclimate and vegetation distribution (Peterson *et al*., 1997).

DATA AND METHODS

Snow course and SNOTEL data

The data analysed in this study were drawn primarily from two distinct but related snow sampling networks, both created and maintained by the Natural Resources Conservation Service (NRCS). Both the snow course survey network and the more recent SNOTEL network were designed to monitor the water equivalence of mountain snowpacks, which typically provide the majority of available water in the western USA. Both networks contain a large number of sampling points located throughout the mountains of the western states and Alaska.

The NRCS snow course network includes some snow courses that date back to the early part of the 20th century. A large number of snow courses in the network have been monitored since the 1940s. Snow courses generally consist of a line of points in an open, sheltered area below the treeline. Depth and SWE are measured at each point. The average values for these points are then archived as the representative depth and SWE measurements for the snow course. Surveys are conducted on or about the first of each month between January and May, although generally only a few priority sites are surveyed for all of these months. More data are available from 1 April snow surveys than from any other month.

NRCS maintains 34 snow courses in the CCE. These are supplemented by 12 similar snow courses in the Provinces of British Columbia and Alberta maintained by the provincial governments. Twenty-one of these 46 snow courses were selected based on the availability of a continuous or near-continuous time series of 1 April measurements between 1950 and 2001. Twenty of these snow courses are located in Montana; the remaining snow course is located near Fernie, British Columbia. Measurements from the period 1950–2001 for these 21 snow courses will be referred to as the 1 April snow course dataset. Three additional snow courses, each with 80 years of continuous 1 May measurements, were selected to provide a longer but limited dataset. All three of these snow courses are located in the Many Glacier Drainage of Glacier National Park and are conducted jointly by water supply forecasters in Montana and Alberta. Measurements from the period 1922–2001 for these three snow courses will be referred to as the 1 May snow course dataset.

The NRCS SNOTEL network provides daily measurements of SWE and, more recently, other climate parameters, including precipitation and temperature. SNOTEL data allow for a detailed examination of the seasonal evolution of the snowpack. Like snow courses, SNOTEL stations are generally located in sheltered, open areas below the treeline. SNOTEL stations utilize pillows filled with solution and a transducer that converts pressure from the weight of accumulated snow on the pillow into SWE measurements. In recent years, SNOTEL stations have been equipped to measure precipitation and temperature as well. The majority of SNOTEL stations do not currently measure snow depth.

SNOTEL stations are automated and present the possibility of instrument error being introduced into SNOTEL datasets. Serreze *et al*. (1999) indicate that the most significant errors in SNOTEL data measurements likely occur during the summer, when negative SWE values are occasionally recorded. During autumn and spring wet snow events, snow may also stick to the inside of the precipitation gauge and be recorded days later as new precipitation (Serreze *et al*., 1999). Neither of these two potential sources of error should affect this analysis of CCE snowpack, since neither summer SWE values nor daily precipitation values were analysed. The possibility of the formation of snow or ice bridges above the SNOTEL pillow surface, as well as the deposition of foreign material, such as fallen branches, on the pillow represent the most likely sources of error in the SNOTEL data used in this study.

Nineteen SNOTEL stations are located in the Montana portion of the CCE. Nine stations with SWE records extending back to water year 1977 or before were selected to provide daily SWE measurements to augment the monthly snow course survey measurements. In addition, these stations each contain at least 20 years of precipitation measurements and 10 years of temperature measurements, allowing for a limited analysis of the effects of precipitation and temperature on the evolution of seasonal snowpacks.

A small number of missing SWE values in the 1 April snow survey dataset were estimated to allow for the inclusion of four sites that would have otherwise been excluded from a time series analysis. These values were calculated based on correlations with values at nearby sites for more than 50 years of data. Three sites required the calculation of one missing value and one site required the calculation of two missing values.

All of the selected SNOTEL stations and all of the snow courses in the 1 April dataset were divided into subgroups based on elevation, relationship to the Continental Divide, and whether north or south of the Marias Pass/Route 2 corridor (also the southern boundary of Glacier National Park). These classifications were designed to facilitate analysis regarding the spatial distribution of snowpack and the spatial dimensions of changes in snowpack. The elevation classification criteria (Table I) were designed to classify an approximately equal amount of sites in each category; consequently, they differ for SNOTEL and snow courses. After SNOTEL data indicated no significant difference between snowpacks in the high-elevation and mid-elevation categories, the mid-elevation classification was dropped and snow courses were classified as either high elevation or low elevation.

SNOTEL and snow course data for the period 1922–1999 were obtained from the NRCS anonymous ftp server, ftp.wcc.nrcs.usda.gov. Snow course and SNOTEL data for water years 2000 and 2001 were obtained via a remote connection to the NRCS Western Climate Center Centralized Forecast System Database.

Temperature data for Kalispell, Montana

Temperature data from Kalispell, Montana, were also obtained. Kalispell is the closest meteorological station to Glacier National Park, with data extending back to the early 1900s. Mean summer temperature values from Kalispell were compared with mean summer temperature values at the Flattop Mountain SNOTEL site (located within 10 km of several glaciers).

PDO data

The PDO is a pattern of Pacific Ocean climate variability driven by sea surface temperature anomalies, similar to the El Niño–southern oscillation, but played out over a much longer time period (Zhang *et al*., 1997). The PDO index is defined as the first principal component of North Pacific (north of 20 °N latitude) monthly

Table I. Elevation classification scheme for SNOTEL and snow course sites

	SNOTEL	Snow course
Low elevation (m)	<1600	<1650
Mid elevation (m)	1600–1900	
High elevation (m)	>1900	>1650

sea surface temperature variability (Mantua *et al*., 1997). Monthly indices for the PDO were obtained from the University of Washington's Joint Institute for the Study of Atmosphere and Ocean at ftp.atmos.washington.edu.

RESULTS

CCE snow database

All standard snow course records for the defined CCE were compiled and merged into a single database consisting of 6543 individual SWE measurements from 46 snow courses. 1 April SWE measurements from 21 of these snow courses met the 50 year minimum criterion for inclusion in the analysis dataset. Three additional sites with 80 year records were included as a second analysis dataset. Nine SNOTEL sites with 25 years or more of SWE measurements were identified and included as a third analysis dataset. SWE, temperature, and precipitation data from these nine sites were also merged into a single database.

Spatial variations in the seasonal distribution of daily SWE

Data from the nine SNOTEL sites with 25 year SWE records provide an indication of the typical evolution of seasonal snowpacks in the CCE and reveal differences between accumulation–ablation patterns across spatial and elevation gradients (Figure 2, Table II). The accumulation–ablation curve incorporating data from all nine SNOTEL sites indicates a mean annual maximum of 60·6 cm of SWE on 13 April. Differences in the 1 April mean SWE indicate that the 1 April mean SWE is not significantly different at mid- and high-elevation sites. Low-elevation sites, however, recorded significantly less SWE on 1 April and earlier meltout. Differences between sites east and west of the Continental Divide were also very significant, with sites west of the Continental Divide exhibiting higher peak and 1 April SWE, as well as longer persistence. Latitude appears to be a less important, but still significant, factor in determining snow accumulation and ablation in the CCE, with sites in the northern half of the region exhibiting slightly higher peak SWE and 1 April SWE.

Spatial variations in 1 April mean SWE

Snow courses in the 52 year dataset were divided into paired groups based on elevation, relationship to the Continental Divide, and latitude. Mean values from all years and all sites in each paired group were compared using independent samples *t*-tests with equal variances assumed (Table III). Significant differences existed between high- and low-elevation sites, east- and west-side sites, and sites in the northern and southern portions of the Crown of the Continent.

Precipitation and temperature as drivers of SWE

At wind-sheltered SNOTEL sites and snow courses, precipitation and temperature are assumed to be the primary drivers of measured snow accumulation and ablation. Solar radiation and humidity also influence snow ablation, but most SNOTEL sites do not record measurements of these parameters; consequently, we were not able to examine their effect on the snowpack. The 1 April water-year-to-date precipitation in the CCE is highly correlated with 1 April SWE, with an overall correlation coefficient of 0·898 and subset values ranging from 0·6 to 0·972 (Table IV). The 1 April water-year-to-date precipitation is better correlated with SWE at higher elevation sites and sites to the west of the Continental Divide. October–March temperature averages for the entire dataset are not as strongly correlated with 1 April SWE (Table V). Many of the temperature–SWE data subsets (e.g. high elevation), composed of a limited set of observations, are not significantly correlated with SWE. Surprisingly, SWE at low-elevation sites, which are most prone to winter rain and thaw cycles, is the least well correlated with October–March mean temperatures.

SNOTEL versus snow course survey data

The mean 1 April snow course measurements and mean 1 April SNOTEL measurements are highly correlated ($r = 0{\cdot}96$). Consequently, it is reasonable to assume that CCE seasonal snowpack evolution patterns

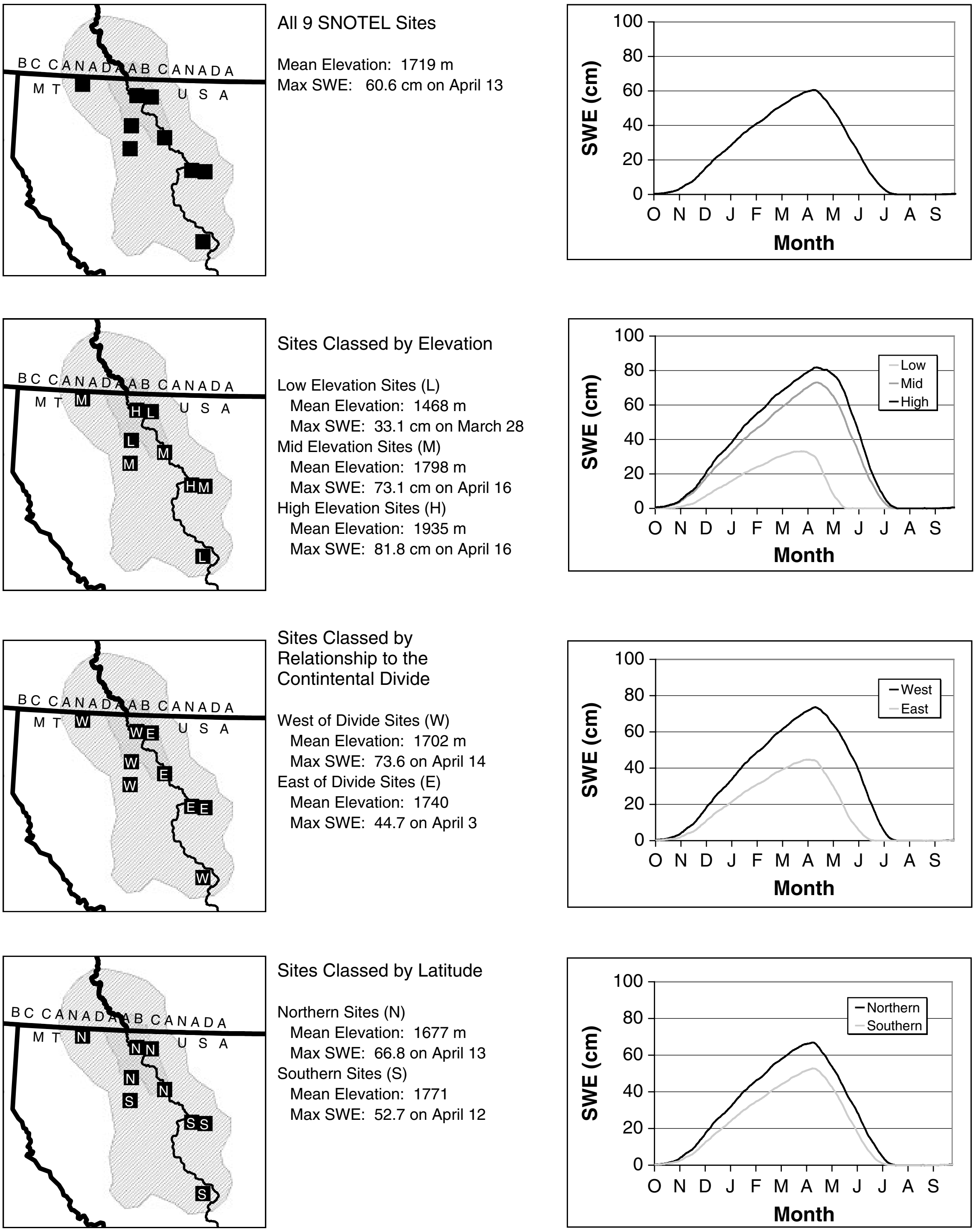

Figure 2. SNOTEL accumulation–ablation curves with site locations

Table II. Results of *t*-tests comparing 1 April mean SWE values for SNOTEL sites in different elevation classes, on opposite sides of the Continental Divide, and in the northern and southern sections of the CCE

	Mean	N	t	Degrees of freedom	Significance
All nine	59·2	225			
Low elevation	33·0	75			
High elevation	78·2	50	−10·7	123	0·000
Low elevation	33·0	75			
Mid elevation	69·3	100	−9·0	173	0·000
Mid elevation	69·3	100			
High elevation	78·2	50	−1·5	148	0·133
East of the Continental Divide	44·5	100			
West of the Continental Divide	70·9	125	−6·3	223	0·000
Northern	63·5	125			
Southern	51·1	100	3·3	223	0·001

Table III. Results from *t*-tests comparing means for groups of snow courses based on elevation, relationship to Continental Divide, and latitude

	Mean	N	t	Degrees of freedom	Significance
All 21	47·7	1092			
Low elevation	30·6	416			
High elevation	58·2	676	−14·0	1090	0·000
East of the Continental Divide	33·2	364			
West of the Continental Divide	55·0	728	−10·3	1090	0·000
Northern	53·0	312			
Southern	45·6	780	3·2	1090	0·001

Table IV. Relationship between 1 April water-year-to-date precipitation and 1 April SWE

	Correlation coefficient	Significance
SWE vs PRCP, all	0·898	0·000
SWE vs PRCP, low	0·600	0·000
SWE vs PRCP, mid	0·972	0·000
SWE vs PRCP, high	0·944	0·000
SWE vs PRCP, east	0·669	0·000
SWE vs PRCP, west	0·955	0·000

depicted by SNOTEL data are representative of the seasonal snowpack evolution patterns for snow courses in the region for the period 1977–2001.

Temporal variations in SWE

Interannual variation in 1 April SWE for different data subsets was compared using the coefficient of variation (100 × standard deviation/mean). Interannual variation in the Crown of the Continent averaged

Table V. Relationship between October–March temperatures and 1 April SWE

	Correlation coefficient	Significance
SWE vs TAVG, all	−0·399	0·000
SWE vs TAVG, low	−0·136	0·473
SWE vs TAVG, mid	−0·284	0·076
SWE vs TAVG, high	−0·425	0·061
SWE vs TAVG, east	−0·350	0·027
SWE vs TAVG, west	−0·368	0·009

Table VI. Means, standard deviations, and coefficients of variation for 1 April SWE by elevation band and geographic zone

	Mean	Standard deviation	Coefficient of variation
All 21	47·7	12·8	26·8
Low elevation	30·6	11·7	38·1
High elevation	58·2	14·2	24·3
East of the Continental Divide	33·2	12·0	36·2
West of the Continental Divide	55·0	13·7	24·9
Northern	53·0	15·0	28·4
Southern	45·6	12·5	27·5

26·8% for all 21 sites and ranged from 24·3% at high-elevation sites to 38·1% at low-elevation sites (Table VI). Variation was also higher for sites east of the Continental Divide (36·2%) than for sites west of the Continental Divide (24·9%). Increased interannual variation was highly correlated with decreased mean 1 April SWE ($r = -0{\cdot}96$).

Residuals from the linear regression of the 1 April SWE time series (1950–2001) and the 1 May SWE time series (1922–2001) were normally distributed and not significantly autocorrelated, initially indicating the dataset was suitable for linear trend analysis. The mean 1 April SWE exhibited a downward trend ($-0{\cdot}38$ cm year^{-1}, $p = 0{\cdot}001$) in the complete 21-site dataset over the period 1950–2001 (Figure 3), whereas the mean 1 May SWE exhibited a slight positive trend that was not significant (0·12 cm year^{-1}, $p = 0{\cdot}421$) over the period 1922–2001 (Figure 4). Further examination, however, revealed a low-frequency (20–30 year) cyclical component in both time series, apparently associated with the PDO. Linear trend analysis is most likely not appropriate for a cyclic time series that includes so few cycles.

Interannual SWE variations and the PDO

Standardized indices calculated for 1 April SWE and October–March PDO index values indicate a strong relationship between the PDO and 1 April SWE in the Crown of the Continent for the period 1950–2001 (Figure 5). Comparing indices for the 1 May SWE time series and October–March PDO also indicates that the relationship existed for the period 1922–2001 (Figure 5). The average of the October–March PDO index explains 26·7% of the variability in mean 1 May SWE values from 1922 to 2001. However, the explanatory ability of the October–March PDO index increases to 61·9% when the 5 year moving averages for October–March PDO and 1 May SWE are compared. In a similar fashion, the October–March PDO index can explain 28·4% of the variability in 1 April SWE measurements for the period 1950–2001; using moving average values increases the explanatory ability to 70·4%.

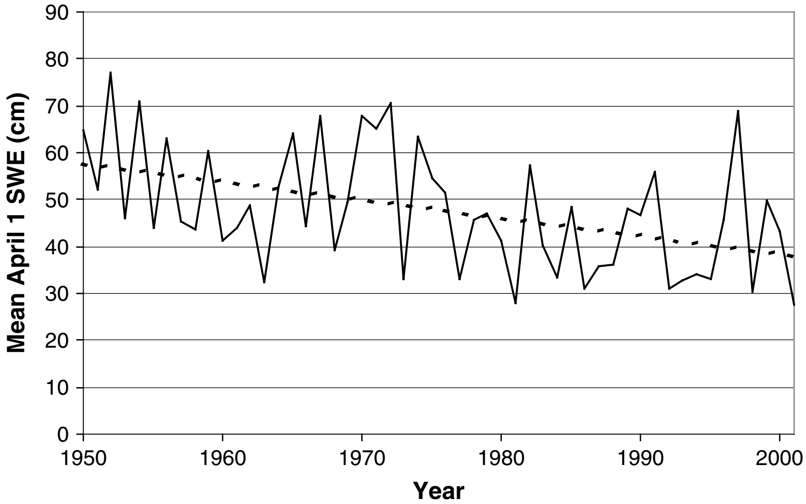

Figure 3. Mean 1 April SWE at 21 snow courses in the Crown of the Continent region, 1950–2001

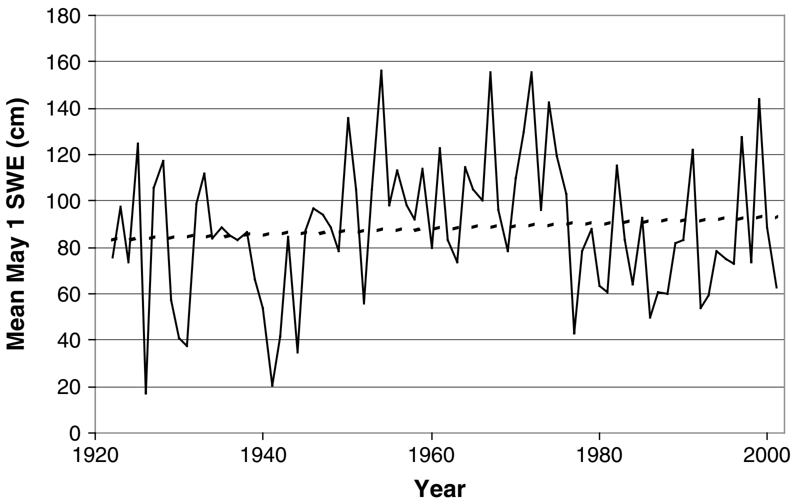

Figure 4. Mean 1 May SWE at three snow courses in the Many Glacier drainage, Glacier National Park, 1922–2001

Summer temperatures

Mean summer (July, August, September) temperatures were calculated for Kalispell, Montana. Located 40 km from West Glacier and approximately 70 km from the Continental Divide (where the majority of Glacier National Park's glaciers are found), Kalispell is the closest meteorological station with a period of record extending back to the early part of the 20th century. Kalispell mean summer temperatures for the period 1983–2000 are well correlated with mean summer temperatures at the Flattop Mountain SNOTEL site (Figure 6; $r = 0{\cdot}79$), located near the Continental Divide within 10 km of numerous glaciers. Since 1922, July–September temperatures at Kalispell have fluctuated but remained relatively stable, with the exception of a large increase in the 1920s and early 1930s followed by a large decrease in the late 1930s and early 1940s (Figure 7).

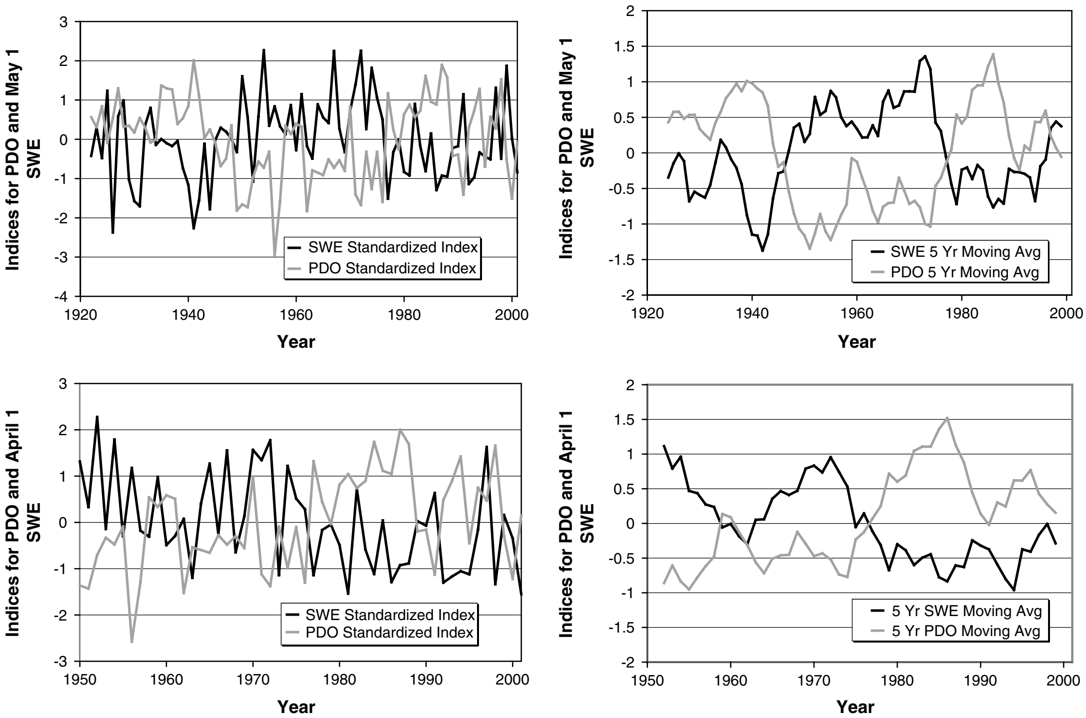

Figure 5. October–March PDO indices and 1 April and 1 May mean SWE. The strength of the relationship between PDO and SWE is far more apparent when 5 year moving averages for SWE and October–March PDO indices are considered

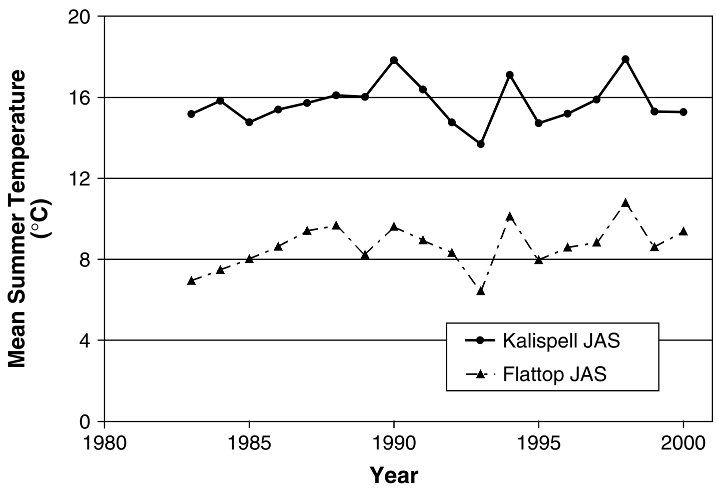

Figure 6. Mean summer (July, August, September) temperature at Kalispell and Flattop Mountain SNOTEL site

DISCUSSION

Prior to discussing spatial patterns of snowpack accumulation and ablation or interannual variations in snowpack in the CCE, it is useful to compare the CCE snowpack with other snowpacks across the western USA. Serreze *et al*. (1999) summarized the basic patterns of snow accumulation and ablation at SNOTEL

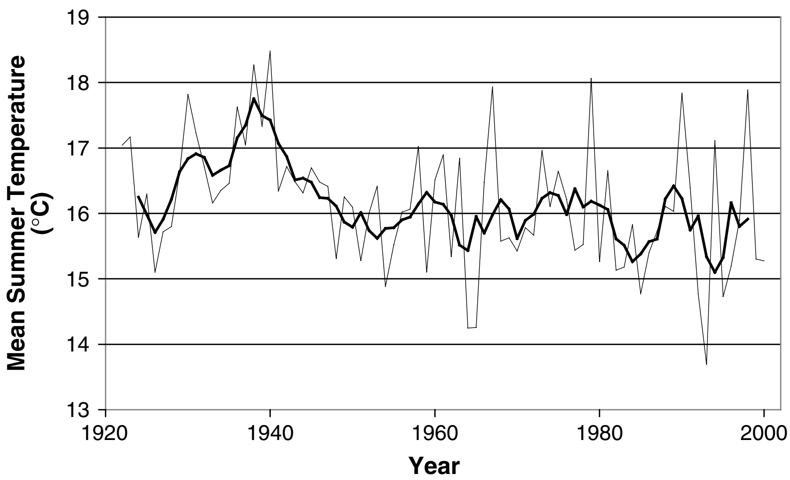

Figure 7. Mean summer temperature at Kalispell, 1922–2000

sites in eight regions they define in the western USA. The accumulation–ablation pattern for the CCE is similar to the pattern they found for the Idaho–western Montana region, with mean maximum SWE in the CCE slightly higher and occurring 1 day later than mean maximum SWE in Idaho–western Montana. Both the mean maximum SWE and the snowpack persistence are greater in the CCE than in all regions defined by Serreze *et al.* (1999), with the exception of the Sierra Nevada and Pacific Northwest.

As expected, higher elevation SNOTEL sites have larger, more persistent snowpacks in the CCE. The similarity between the accumulation–ablation patterns for mid- and high-elevation sites probably reflects the small difference in mean site elevations between the two categories and the small sample size for each category. Consequently, this similarity should not be interpreted as an indication that high-elevation snowpacks are very similar to mid-elevation snowpacks in the region. It is also worth noting that true high-elevation snowpacks are not included in this analysis, as all SNOTEL stations and snow courses in this study are located below the treeline.

The indication that snowpacks at sites to the west of the Continental Divide tend to be larger than snowpacks at sites to the east of the Continental Divide is not surprising, but it does merit some consideration. A precipitation gradient decreasing from west to east could be hypothesized to explain this difference, but Finklin (1986) indicates that, at least for the Waterton–Glacier International Peace Park, precipitation appears to be approximately equal on both sides of the Continental Divide and that major decreases in precipitation only occur along the far eastern edges of the region. The mean 1 April water-year-to-date precipitation values do indicate more precipitation at sites west of the Continental Divide (81·7 cm) than at sites east of the Continental Divide (63·8 cm), but this relatively small difference is insufficient to explain the larger difference in mean maximum SWE between west-side sites (73·6 cm) and east-side sites (44·7 cm). The unexplained difference is most likely the result of the effects of temperature, wind, and humidity. In particular, periodic warm Chinook winds that race downslope on the east side of the Continental Divide may be responsible for rapid snowmelt at times when snowpacks to the west of the Continental Divide remain stable.

The relationships between the 1 April water-year-to-date precipitation and 1 April SWE and between the October–March mean temperatures and 1 April SWE demonstrate that, as previously indicated for the interior northwest by Cayan (1996) and Serreze *et al.* (1999), precipitation is the dominant control of 1 April SWE in the CCE. October–March mean temperatures are much less important. The lack of strong correlations between 1 April SWE and October–March mean temperatures suggests that temperatures in the Crown of the Continent are typically cool enough so that small fluctuations in the seasonal mean do not have a large effect on 1 April SWE. October–March mean temperatures, however, may be too crude a measure to capture some of the important characteristics of the relationship between temperature and 1 April SWE in the CCE. Slight temperature variations on heavy precipitation days can make the difference between rain and snow events. In addition, a single very warm day can be responsible for significant ablation. Neither of these two events are likely to have a large impact on October–March mean temperatures.

Based on 1 April data from 21 snow courses, interannual variation in snowpack averages 26·8%. It appears to be most variable at low-elevation snow courses and snow courses east of the Continental Divide. Snow courses at low elevations and to the east of the Continental Divide also have the lowest mean 1 April SWE values. These two groups with the smallest snowpacks and greatest snowpack variability are also at the greatest risk for being impacted significantly by future changes in climate. Predicted climate warming will likely affect lower elevation snowpacks earlier and more significantly than middle and higher elevation snowpacks. In addition, ecosystem modelling has predicted that, under a variety of climate change scenarios, areas to the east of the Continental Divide will begin to dry out while areas to the west of the Continental Divide will remain more stable (White *et al.*, 1998). This particular combination of snowpack size, variability, and likelihood of future change for low-elevation and east-side snowpacks has important ecological implications, since vegetation in low-elevation areas and areas to the east of the Continental Divide are commonly limited by moisture availability.

Higher variability in mean 1 April SWE at lower elevation sites also suggests that orographic precipitation in the CCE is a more consistent source of snowfall than frontal precipitation. Whereas high-elevation sites consistently receive orographically derived or enhanced precipitation every year even in the absence of major frontal systems, lower elevation sites are likely to remain relatively dry in the absence of frontal systems.

The negative trend in 1 April SWE between 1950 and 2001 is most likely the result of positioning of PDO phases (negative from 1947 to 1977, positive from 1977 to the present) and does not necessarily stand as evidence of a climate-change-related decrease in 1 April SWE in the CCE. As mentioned earlier, linear trend analysis may not be appropriate for a cyclical time series incorporating so few cycles.

Glacier mass balance (and often area) is generally considered to be a function of temperature (most important in summer) and winter precipitation. For glaciers in the Jackson–Blackfoot glacier basin of Glacier National Park, Hall (1994) designed an accumulation–ablation model in which a 1 °C temperature increase had an effect equivalent to a 14% decrease in annual precipitation. The 1 May SWE, however, is likely a more direct driver of glacier response than annual precipitation, given that it is an integrative measure of total winter accumulation that incorporates temperature, precipitation, and any other relevant factors in the winter accumulation process. It is evident that a combination of summer temperature data and winter snow accumulation data is necessary to explain glacier behaviour adequately. Despite the fact that both mean summer temperatures at nearby Kalispell and 1 May SWE in the Many Glacier Basin have remained relatively stable in the period 1922–2001, glaciers throughout Glacier National Park have receded steadily throughout this period (Key *et al.*, in press). The 1922–2001 climate regime is clearly incapable of maintaining steady-state glacier mass balances. Carrara (1989) estimated that glaciers in Glacier National Park reached their Little Ice Age maxima just prior to 1860. Allowing for between 10 and 25 years of demonstrated lag time between climatic change and glacier response (Hall and Fagre, 2002), this implies that major changes in summer temperatures or winter snowpacks likely occurred sometime between 1840 and 1922. Temperature records do indicate warming between 1900 and 1922 for Kalispell, but, unfortunately, no mountain snowpack records exist for the area prior to 1922. The period 1840–1922 is, however, the period when tree establishment above Little Ice Age treeline is first recorded (Bekker *et al.*, 2000), and Peterson (1998) has shown that high-elevation tree establishment patterns are related to variations in snowpack.

The relationship between PDO and snowpack has been suggested previously, most recently by McCabe and Dettinger (2002). They found that 1 April snowpack variation is more strongly linked to the PDO at snow courses in the northwestern USA and southwestern Canada than anywhere else. The fact that 1 April SWE variation is much more closely related to 5 year averages of October–March PDO indices than to single-season PDO indices emphasizes the importance of the PDO phase over the specific strength of the signal for one season. Mantua *et al.* (1997) identify reversals in the prevailing polarity of the PDO in 1947 and 1977, and these are evident in both the 1 April 52 year SWE dataset and the 1 May 80 year SWE dataset. Because PDO is generally slow to evolve (McCabe and Dettinger, 2002), and follows a somewhat regular cycle, it can potentially be incorporated into attempts to explain past changes in physical and biological systems of the CCE, as well as in attempts to model potential changes in these same systems in the future.

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