Long-term snow, climate, and streamflow trends at the Reynolds Creek Experimental Watershed, Owyhee Mountains, Idaho, United States

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[1] Forty-five water years (1962–2006) of carefully measured temperature, precipitation, snow, and streamflow data for valley bottom, midelevation, and high-elevation sites within the Reynolds Creek Experimental Watershed, located in the state of Idaho, United States, were analyzed to evaluate the extent and magnitude of the impact of climate warming on the hydrology and related resources in the interior northwestern United States. This analysis shows significant trends of increasing temperature at all elevations, with larger increases in daily minimum than daily maximum. The proportion of snow to rain has decreased at all elevations, with the largest and most significant decreases at midelevations and low elevations. Maximum seasonal snow water equivalent has decreased at all elevations, again with the most significant decreases at lower elevations, where the length of the snow season has decreased by nearly a month. All trends show a significant elevation gradient in either timing or magnitude. Though interannual variability is large, there has been no significant change in water year total precipitation or streamflow. Streamflow shows a seasonal shift, stronger at high elevations and delayed at lower elevations, to larger winter and early spring flows and reduced late spring and summer flows.

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1. Introduction

[2] In western North America, seasonal mountain snow cover provides both the primary supply and storage reservoir for water. Water management in this region is largely based on empirical relationships between point measurements of snow water equivalent (SWE) at selected sites and subsequent stream discharge. The assumption is that this relationship is representative of a more comprehensive basin-wide or regional relationship between snow deposition, the timing and magnitude of melt from the snow cover, and the translation of that melt to discharge. It is further assumed that this relationship is robust and stable over time. If climate warming results in trends toward more rain and less snow, this will alter patterns of snow deposition, the timing of melt, and the delivery of meltwater to soils, with consequent changes of flow in streams and rivers. As a result, the empirical relationship between measured SWE and stream discharge will become unstable and unreliable with a substantial impact on water resource management in the region.

[3] Since the beginning of the 20th century, the Earth's mean surface temperature has increased by about 1°C [*Intergovernmental Panel on Climate Change* (*IPCC*), 2007; *Trenberth et al.*, 2007], with greater temperature increases in mountainous regions and strong effects on the seasonal snow cover [*Lemke et al.*, 2007]. Observations and model predictions indicate that persistent warming will substantially alter the hydroclimate, both at global and regional scales [*Leung and Ghan*, 1999; *Leipprand and Gerten*, 2006; *Manabe et al.*, 2004; *Stewart et al.*, 2005; *Randall et al.*, 2007]. In western North America, mean surface temperature has increased by 1–3°C over the last 50 years, with larger increases during winter [*Trenberth et al.*, 2007] and in minimum daily temperature [*Karl et al.*, 1984, 1993; *Quintana-Gomez*, 1999; *Brunetti et al.*, 2000; *Trenberth et al.*, 2007].

[4] The effect of climate warming on the amount of precipitation is not definitive, particularly in mountainous regions of western North America. In the United States and Canada it is reported that total annual precipitation has increased slightly [*Groisman and Easterling*, 1994; *Karl and Knight*, 1998; *Hamlet et al.*, 2005; *Trenberth et al.*, 2007], though some regions have reported persistent drought. The studies cited above are generalizations, extending over hemispheric or continental regions. Extension to mountain environments is complicated by relatively low density of measurement stations, the difficulty of modeling

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snow deposition and melt across mountainous areas, and uncertainty in the change in phase from rain to snow as storms progress from valley bottoms to higher elevations. Precipitation intensity and volume over the western mountains are strongly influenced by storm track and air mass characteristics associated with ocean circulation features such as the El Niño–Southern Oscillation (ENSO) and the Pacific Decadal Oscillation (PDO) [*Hurrell*, 1995; *Hurrell and Van Loon*, 1997; *Dettinger et al.*, 1998]. However, the dynamics of ocean circulation and temperature are poorly monitored [*Bindoff et al.*, 2007], and subsequent linkages between atmosphere-ocean interactions and precipitation in mountainous regions of the western United States and Canada are not well understood [*Trenberth et al.*, 2007]. *Coquard et al.* [2004] compared 15 climate models finding that their ability to simulate precipitation for current and doubled CO_2 conditions over the mountainous western United States was weak. Under doubled CO_2, these models generally agreed that temperature would increase, but showed no consistency for precipitation with some models predicting an increase and others a decrease.

[5] The timing and magnitude of snowmelt in the mountains of the western United States are very sensitive to climate conditions. Along with warmer temperatures, a number of studies have shown substantial changes in snow deposition and melt patterns, a reduced fraction of precipitation that falls as snow over western North America, and a shift in the timing of snowmelt runoff toward earlier in the year [*Aguado et al.*, 1992; *Dettinger and Cayan*, 1995; *Huntington and Hodgkins*, 2004; *Regonda et al.*, 2005; *Stewart et al.*, 2004, 2005; *Knowles et al.*, 2006]. Indications of this shift have been earlier timing of the initial pulse of snowmelt runoff [*Cayan et al.*, 2001; *Stewart et al.*, 2004, 2005], declines in snow cover and spring snow water equivalent [*Mote*, 2003a, 2003b, 2006; *Mote et al.*, 2005], earlier timing of the peak streamflow, and a redistribution of the mean monthly or seasonal fractional streamflow [*Aguado et al.*, 1992; *Cayan et al.*, 2001; *McCabe and Clark*, 2005; *Regonda et al.*, 2005].

[6] Whereas previous studies have demonstrated the general sensitivity of seasonal snowmelt runoff to climate warming, they provide little insight into the dependence of that sensitivity on elevation, site or climate conditions other than temperature. One exception is the work by *Howat and Tulaczyk* [2005] who used over 50 years of 1 April snow course data from 55 stations in the Sierra Nevada. Using a covariance model to interpolate the spatial distribution of SWE in an effort to evaluate the possible effects of climate warming on the spring snow cover in the Sierra Nevada, they found that while 1 April SWE was decreasing at lower elevations it was unchanged and even increasing at higher elevations. While this may limit the impact of warming on SWE volume in the Sierra Nevada, the authors point out that the same warming trends in mountains with a more limited elevation range, such as the Cascades of Washington and Oregon, show a decrease in 1 April SWE at all elevations.

[7] We expect the seasonal snow cover at elevations near the "rain-snow transition zone" to be more affected by climate warming than the snow cover at higher elevations, but few data are available to test this hypothesis, and most of the conclusions reached are circumstantial or anecdotal. *McCabe and Clark* [2005] evaluated streamflow timing for 84 rivers in the western United States and showed a systematic shift toward earlier flows in all regions, with the strongest trend in the Pacific Northwest. Trends were more significant for lower-elevation rivers, but the comparison is limited by the data used for the analysis. Many of the highly significant trends are for low-elevation sites located in the Cascade Mountains of the Pacific Northwest, with weaker trends for higher-elevation sites in the Upper Colorado, Great Basin and California, which include the Rocky Mountains, Wasatch, and Sierra Nevada. While their analysis does indicate that the Pacific Northwest may be more sensitive to climate warming than the other regions studied, the effect of elevation on the trend is not direct because the low-elevation stream gauges are geographically removed from the high-elevation stream gauges.

[8] *Hamlet et al.* [2005] and *Mote et al.* [2005] used a modeling approach to evaluate the sensitivity of trends in snow water equivalent (SWE) across the western United States to changes in temperature and precipitation. Both compared simulated SWE to measured values, and both attempted to separate the effects of changes in temperature from changes in precipitation. These studies show that seasonal maximum SWE is decreasing and occurring earlier, and these trends, throughout this domain, are not the result of changes in precipitation but are strongly correlated to increases in temperature. The decrease in SWE was again greatest in the Pacific Northwest. However, the lack of climate and snow data and a limited range of elevations where data were collected within the same geographic area limited both studies. Though the western United States has an extensive mountain snow measurement program, the NRCS SNOTEL system [*Serreze et al.*, 1999; http://www.wcc.nrcs.usda.gov], the sites are generally located in protected, midelevation locations. The analysis presented by *Hamlet et al.* [2005] and *Mote et al.* [2005], required extrapolation of data across a range of elevations that extended far beyond that of the measurement sites. Temperature measurement sites were typically in valley bottoms and may have been at elevations a km or more below the snow and precipitation sites, so that estimates of precipitation at higher elevations were based on estimated lapse rates that could not be validated.

[9] It is clear that temperatures across western North America have increased and that the seasonal snow cover has been affected. Most research, however, has focused on large-scale analysis, leaving many questions about specific impacts of climate warming, particularly how the mountain snow cover across a range of elevations within a mountain basin may be differently affected. Few locations exist where climate, precipitation, and snow measurement sites are colocated, and fewer still where these are located along a range of elevations spanning the rain/snow transition from the valley bottom to headwaters in a mountain basin. In mountain basins, the distribution of the snow cover and snowmelt and the generation of runoff are heterogeneously distributed across the landscape as a function of terrain structure (elevation, slope, and aspect), wind exposure, and land cover [*Marks and Dozier*, 1992; *Marks and Winstral*, 2001; *Winstral and Marks*, 2002; *Marks et al.*, 1999, 2002; *Garen and Marks*, 2005]. To better understand how climate warming has affected, and may further affect mountain snow cover, snowmelt, streamflow, and catchment hydrology in the western United States, coherent long-term data from a range of elevations within a mountain basin on

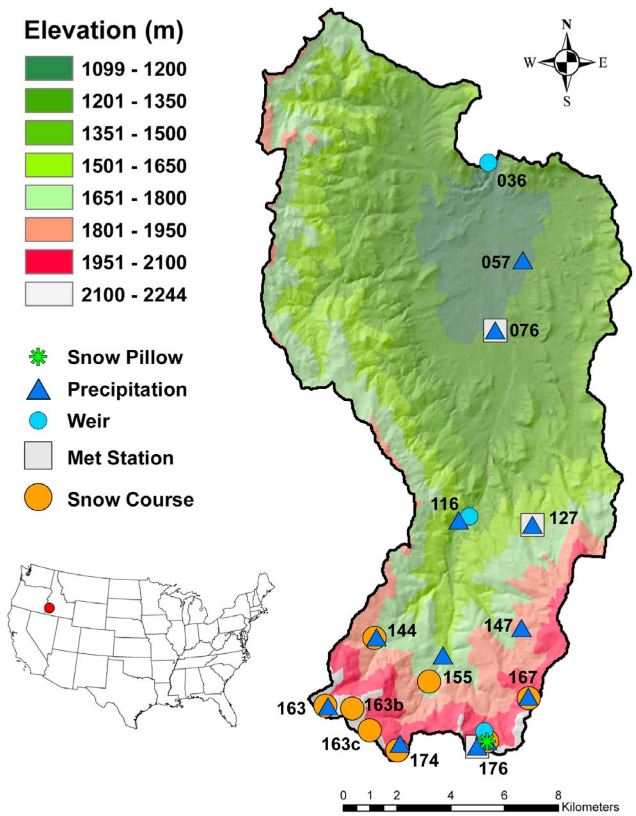

Figure 1. Topographic map of RCEW with long-term climate stations, precipitation gauges, snow courses, snow pillow, and weirs. Site numerical identifiers as presented in Table 1 are shown. Site 076 is located at 116°44′57″W, 43°12′18″N.

snow, precipitation, temperature, humidity, streamflow, and related parameters are required.

[10] Detailed and carefully collected data from 45 years of monitoring (1962–2006) within the Reynolds Creek Experimental Watershed (RCEW), a U.S. Department of Agriculture Agricultural Research Service watershed in the Owyhee Mountains of Idaho [*Robins et al.*, 1965; *Marks*, 2001; *Flerchinger et al.*, 2007; *Marks et al.*, 2007], are analyzed in this paper. Temporal trends in temperature and precipitation are analyzed by year and season, followed by seasonal analysis of precipitation phase (snow or rain), peak SWE, 1 April and 1 May SWE, snow cover initiation and melt out dates, and streamflow for sites across the full range of elevations and site conditions found within the RCEW. All data used for these analyses are available in the auxiliary material.[1]

2. Reynolds Creek Experimental Watershed

[11] The Reynolds Creek Experimental Watershed is a 238 km^2 drainage with an elevation range of 1145 m (1099–2244 m AMSL) located in the Owyhee Mountains near Boise, Idaho (Figure 1). The ecology and hydroclimatology of RCEW are representative of much of the interior mountain west and Great Basin, including areas of Idaho, eastern Oregon, Utah, and Nevada. While there are several hundred snow and precipitation measurement sites across the western United States, *Bales et al.* [2006] point out that there are only 6 comprehensive energy balance sites and that the RCEW is the most detailed and longest studied of those. In this analysis, data collected at 3 meteorological stations, 12 precipitation stations, 8 snow courses, 1 snow pillow, and 3 weirs were selected based on length and continuity of data records to analyze the impact of climate warming on hydrology and related resources at the RECW.

2.1. Meteorological Measurement Sites

[12] The three meteorological measurement sites were initially established to monitor daily climate conditions (maximum/minimum temperature, relative humidity, and pan evaporation during summer) at a range of elevations within the RCEW and have been in continuous operation since the 1960s [*Hanson et al.*, 2001]. As digital data loggers became available in the 1970s and early 1980s, these locations were converted to full micrometeorological measurement stations, providing hourly records of temperature, humidity, wind, and solar radiation. Complementary precipitation measurement stations are also located at each site (Table 1). The low-elevation (076_met) site is located in a relatively broad, flat valley bottom at an elevation of 1207 m, only 108 m above, but nearly 10 km distant from the RCEW outlet weir (Figure 1). Site vegetation is Wyoming big sagebrush and is typical of low elevations in the RCEW. The midelevation site (127_met) is located on the eastern side of the basin near the midpoint elevation (1718 m) of the watershed. Site vegetation is dominated by low sagebrush and is typical of midelevation vegetation on the eastern side of the RCEW. The high-elevation site (176_met) is near the southern rim of the RCEW at an elevation of 2093 m in an exposed area where a few trees and larger shrubs offer limited wind shelter. Site vegetation is a mix of shrubs including mountain big sagebrush, snowberry and buckbrush. Adjacent to the site are Douglas fir and a few Aspen trees. This heterogeneous mix of vegetation is characteristic of higher elevations within southeastern regions of the RCEW.

2.2. Precipitation Measurement Sites

[13] Of the 26 active precipitation measurement sites in the RCEW, data from 12 were selected for this study because they had a long period of record (45+ water years) and represented a range of elevations, site, and wind exposure conditions (Table 1). Two of the precipitation sites are located within a few hundred meters of each other. Site 176_ppt (2096 m) is colocated with the high-elevation met station (176_met) on a very wind exposed ridge near the crest of the RCEW, while site 176e_ppt (2056 m) colocated with the snow pillow (176e_sp) and snow course (176e_sc) in a wind-protected Aspen grove clearing. Data from site 176_ppt were used to determine precipitation events for the precipitation phase analysis, while data from site 176e_ppt was used to determine trends in precipitation volume over the period of record. Because much of the precipitation across the RCEW falls as snow, a significant wind-induced under catch occurs and wind correction of the recorded gauge catch is critical. Each precipitation measurement site consists of a pair of 30.5 cm orifice (orifice height is 3 m)

[1]Auxiliary materials are available at ftp://ftp.agu.org/apend/wr/2008wr007525.

Table 1. Long-Term Climate, Precipitation, Snow Measurement, and Stream Gauge Sites at the Reynolds Creek Experimental Watershed[a]

Site ID	Elevation (m)	Location[b] Easting (m)	Northing (m)	Data Records Type	Start
163_ppt	2169	514,134	4,764,428	Precipitation	1962
163_sc	2162	514,042	4,769,428	Snow course	1961
163b_sc	2147	515,042	4,769,342	Snow course	1961
163c_sc	2125	515,687	4,768,520	Snow course	1961
176_met	2093	519,690	4,767,928	Daily T_a, RH	1967
				Hourly met	1984
174_ppt	2074	516,813	4,768,022	Precipitation	1962
174_sc	2073	516,731	4,767,765	Snow course	1961
176e_ppt	2056	520,055	4,768,122	Precipitation	1962
176e_sc	2056	520,055	4,768,117	Snow course	1961
176e_sp	2056	520,055	4,768,122	Snow pillow	1982
166_sf	2022	519,952	4,768,494	Streamflow	1963
167_ppt	2002	521,601	4,769,781	Precipitation	1962
147_ppt	1872	521,340	4,772,331	Precipitation	1962
144_ppt	1815	515,949	4,771,988	Precipitation	1962
144_sc	1815	515,862	4,771,963	Snow course	1961
155_ppt	1654	518,426	4,771,315	Precipitation	1962
155_sc	1743	517,892	4,770,341	Snow course	1961
127_ppt	1649	521,745	4,776,189	Precipitation	1962
127_met	1652	521,745	4,776,195	Daily T_a, RH	1967
				Hourly met	1984
116_ppt	1460	519,006	4,776,344	Precipitation	1962
116_sf	1404	519,392	4,776,492	Streamflow	1966
076_ppt	1200	520,367	4,783,418	Precipitation	1962
076_met	1200	520,365	4,783,423	Daily T_a, RH	1964
				Hourly Met	1981
057_ppt	1184	521,393	4,786,030	Precipitation	1962
036_sf	1099	520,109	4,789,673	Streamflow	1963

[a]Listed in elevation order, site IDs are related to the historical database identifier. Numerical identifiers are shown on the map in Figure 1. After conversion to hourly data recording, meteorological measurement sites (176_met, 127_met, and 076_met) include hourly measurements of long-term climate (T_a, RH), solar and thermal radiation, wind speed and direction, soil temperature, and moisture. All listed sites are still operating.

[b]Location refers to UTM zone 11.

weighing-recording gauges, one unshielded and the other with an Alter-type wind shield, with baffles individually constrained at 30° from vertical. This system supports the dual-gauge wind correction technique developed by *Hamon* [1973] [see also *Hanson et al.*, 1999]. The dual gauge method uses the ratio of shielded to unshielded catch as an indication of wind-induced under catch. The method was evaluated as part of the World Meteorological Organization (WMO) solid precipitation measurement experiment [*Yang et al.*, 1999, 2001; *Hanson et al.*, 1999, 2004] and was found to be comparable to the standard WMO wind correction and to reliably reproduce reference values based on the Wyoming shield over a wide range of wind, temperature, and precipitation intensity conditions. Shielded, unshielded and wind-corrected data from the period of record for all 12 sites were processed and integrated to an hourly time step, with noise and discontinuities removed using a utility developed by *Nayak et al.* [2008]. Only wind-corrected values were used in the analysis presented in this paper.

2.3. Snow Measurement Sites

[14] Eight biweekly snow courses have been operated at the RCEW since the early 1960s, with one additional snow course added in 1970. These represent the upper 25% of the RCEW with elevations of 1743–2162 m (Table 1). Only during cold, wet years does a continuous seasonal snow cover develop in RCEW between 1500 and 1700 m and hardly ever below that level. All sites are classical snow courses, with a fixed number of samples along a predetermined track. Samples of depth and mass are taken with a Rosen Sampler, which tends to show improved sampling when compared to the Standard Federal Sampler [see *Marks et al.*, 2001a]. Twelve depth and mass samples are taken at each snow course. Prior to 1970, surveys did not always occur at precise 2-week intervals, but thereafter the biweekly interval was maintained between 1 December and 15 May. The annual snow survey period may be extended, depending on snow cover extent. If the ground is bare during a sample visit, zero values are recorded. While all sites have been carefully maintained over the period of record to avoid the effects of site disturbance, vegetation removal, or overgrowth, a land cover change altered the record at one snow course site. For this reason, only seven of the snow courses were used for the analysis in this paper. In addition, a snow pillow has been maintained at the 2056 m elevation (site 176e_sp) just adjacent to snow course 176e_sc and next to precipitation measurement site 176e_ppt since 1982. This site is a wind-protected clearing in an Aspen grove, similar to NRCS SNOTEL sites across the western United States [*Serreze et al.*, 1999]. The snow pillow is a 3 m diameter butyl-rubber device filled with a mix of antifreeze, alcohol, and water which continuously measures snow cover mass. Hourly values of SWE are recorded.

2.4. Determination of Precipitation Phase

[15] *Hanson* [2001] reported that, in the RCEW, the proportion of snowfall in the annual precipitation total varied from around 20% at the lowest elevations to more than 75% at the highest. This estimate was based on the ad hoc assumption by *Cooley et al.* [1988] that precipitation in any month in which the mean temperature was ≤1°C would be considered snow. Similar air temperature–based methods have been used in many investigations [e.g., *Hanson et al.*, 1979; *Lapp et al.*, 2005; *Hamlet et al.*, 2005], but the variable temperature thresholds used are site- and season-dependent. *Knowles et al.* [2006] used daily measurements of increased snow depth as an indicator of snowfall to determine phase. However, their approach was limited to a daily assessment of snow or rain and cannot be used to determine phase for individual storms, or for mixed-phase events, which may begin as rain and end as snow.

[16] Actual observations of precipitation phase are rare, but since 2001, phase determination has been available at many of the precipitation sites within the RCEW. The approach is based on concurrent measurement of precipitation, snow depth, and dew point temperature during precipitation events. *Olsen* [2003] recommends wet bulb temperature as a determinant of air mass potential for rain or snow but suggests near-surface dew point temperature as the most reliable predictor of precipitation phase for a specific location. This approach has been used reliably for analysis of rain-on-snow events [*Marks et al.*, 1998, 2001b] and for time series simulations of the seasonal snow cover at a variety of scales and a number of locations across the western United States [*Marks et al.*, 1999, 2001b, 2002; *Marks and Winstral*, 2001; *Garen and Marks*, 2005]. If the

dew point temperature is above 0°C, precipitation is assumed to be rain, if it is <0°C, snow, and if close to 0°C, mixed rain and snow.

[17] For the analysis presented here, concurrent measurements of humidity (dew point temperature) and precipitation were available from the three long-term meteorological sites back to the early 1980s, so dew point temperature is known for every storm from then to the present. Daily maximum and minimum temperature data were used to estimate storm dew point temperatures for the early part of the RCEW data record. To do this, hourly temperature estimates for the prehourly data period were simulated for the three long-term meteorological stations (176_met, 127_met, 076_met). A sinusoidal diurnal cycle between maximum and minimum temperature with a temperature maximum offset from noon was derived for each month from the 1986–1995 hourly temperature data. These monthly cycles were then used to derive hourly temperatures for the early period of record. Monthly storm period dew point temperature deficit (air temperature minus dew point temperature) was tabulated for events >1 mm of precipitation from the 1986–1995 hourly temperature and humidity data. These were then used to estimate storm dew point temperatures for the 1962–1982 portion of the RCEW precipitation data set. Because it best matched recent observations, and to avoid the "mixed phase" issue, it was assumed that if the dew point temperature was ≤0°C, the precipitation fell as snow.

2.5. Streamflow Measurement Sites

[18] Of the 9 active weirs within the RCEW, data from three were selected for location and the availability of hourly streamflow records that begin in the early to mid-1960s (Table 1). These weirs include the Reynolds Mtn. East weir (166_sf), which drains the 0.39 km^2 headwater (2024–2139 m) catchment, the Tollgate weir (116_sf), which drains the 55 km^2 Tollgate subbasin (1398–2244 m), and the Outlet weir (036_sf), which drains the 238 km^2 RCEW basin (1099–2244 m).

2.6. Data Availability

[19] All data used in these analyses are available as auxiliary material. Data, as described in the auxiliary material, are provided as water year or seasonal totals, peak values or averages, or as values on specific dates, as they are presented in Tables 1–8 and Figures 3–5. More comprehensive hourly and 15 min data can be obtained on request from the Northwest Watershed Research Center in Boise, Idaho, United States (contact, Danny Marks, ars.danny@gmail.com).

3. Trend Analysis

[20] Data from the RCEW were analyzed for trends over the period of record. To capture the annual hydrologic cycle, which begins in fall and ends in summer, annual trend analysis was conducted over water year (WY) intervals (October–September). Greater sensitivity to cold season processes was achieved through seasonal analyses, in which the water year is divided into four seasons, fall (October–December), winter (January–March), spring (April–June), and summer (July–September). Trends in air temperature, snow, and streamflow data were computed using two methods: (1) least square (LS) linear regression and (2) Sen's slope (SS) estimator. Although the LS method is in common use, the slope of the regression can be sensitive to autocorrelation and extreme values. To address this problem, *Hirsch et al.* [1982, 1991] proposed Sen's slope estimator (SS) [*Sen*, 1968], a nonparametric method to detect and estimate the magnitude of temporal trends in hydrologic data. This method computes slopes between all data pairs and estimates the overall representative slope as the median value among all possible slope values. Upper and lower confidence intervals for SS were obtained for $\alpha = 0.10$ from a rank order of all calculated slopes [*Kendall*, 1975; *Gilbert*, 1987].

[21] Significance of trends is evaluated using the nonparametric Mann-Kendall statistic at $\alpha = 0.10$ and 0.05 [*Hirsch and Slack*, 1984; *Lettenmaier et al.*, 1994; *Yue et al.*, 2002a]. The advantage of this statistic is that it tests for consistency in the direction of change for temporally ordered data and is unbiased by the magnitude of change. Two methods were applied to reduce the influence of autocorrelation on statistical significance of trends in time series data: First the Mann-Kendall test with prewhitening (MK-PW), as suggested by *Zhang et al.* [2001], was applied to eliminate the effects of serial correlation in the Mann-Kendall test. Second, we applied the trend-free prewhitening (MK-TFPW) approach [*Yue et al.*, 2002b] to minimize the effect of the MK-PW approach on the magnitude of slope and significance of trend present in original data series.

[22] The assumptions underlying these statistical methods may be invalidated in the presence of long-term persistence (LTP) [*Koutsoyiannis and Montanari*, 2007]. Consideration of LTP is clearly important for developing long-term projections of low flows, which can have long residence times [*Hosking*, 1984, *Khaliq et al.*, 2008]. However, our primary purpose is to relate relatively recent contemporaneous changes in hydrometeorology, snow dynamics and streamflow at the RCEW, which, for headwater catchments, has residence times on the order of 1–2 years. As such the hydrology of the RCEW is strongly tied to seasonal and annual meteorological conditions. Although the period of the RCEW record is sufficient to resolve recent trends in hydrometeorological parameters, a model forecast based on these trends would be subject to falsification if the assumption of no LTP were found to be invalid.

4. Results and Discussion

[23] First, an overview of the statistical results from several methods is presented to give context for the results for each type of data analyzed. This is followed by analysis of temperature, precipitation depth and precipitation phase over the range of elevations within RCEW. Snow course and snow pillow data are then presented for upper elevations, where accumulation is greatest. Finally, streamflow data are presented for nested catchments representing headwater, midelevation and watershed level response.

4.1. Choice of Statistical Method

[24] We found little difference in trend slope values determined by LS or SS for most of our regressions (Figure 2). Trend values are shown per decade for all measurements except melt dates. To fit trends in all parameters on a single set of axes required scaling of snow

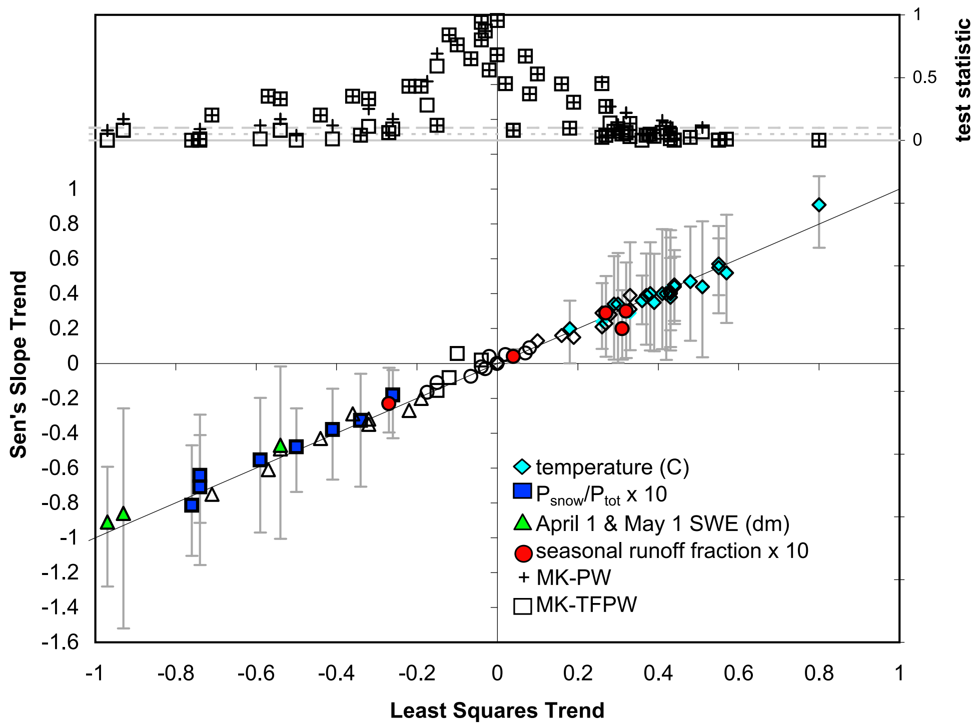

Figure 2. Trend slope values determined by LS with Sen's slope (SS) compared to a 1:1 line for all data except timing of SWE trends. Some data have been scaled as indicated in the legend to fit compactly on the axes. For significant trends ($\alpha < 0.10$), data makers are solid and SS upper and lower limits are shown as error bars. Data with insignificant trends are shown as open data markers. At the top of the chart, Mann-Kendall with prewhitening (MK-PW) and Mann-Kendall with trend-free prewhitening (MKTFPW) test statistics for all trends are compared to dashed lines at $\alpha = 0.05$ and 0.10.

fraction precipitation (10x), seasonal runoff fraction (10x) and SWE (decimeters). The near-zero (insignificant, open data markers on Sen's slope axis) trends, generally have very strong agreement in test statistic values greater than 0.10 as shown near the top of Figure 2. As expected, determination of significance by the MK-PW statistic is slightly more conservative than by MK-TFPW for trends of greater absolute magnitude. The similarity in the results supports use of either regression or a test statistic method. Given the current variance in opinion over the bias introduced or removed by prewhitening [*Bayazit and Önöz*, 2007; *Hamed*, 2008] for trend error estimation we also present Sen's slope upper and lower limits, shown for trends with test statistic values between 0 and 0.10. We found the SS confidence intervals agreed with the MK-PW and MK-TFPW test statistic much better than LS, which has standard error values (not shown) up to two times greater than the SS uncertainty. For the remainder of this paper, we report trends as SS in units per decade and significance tested by MK-TFPW. Since the difference from zero trend for SS upper and lower limits is generally supported by the MK-TFPW test statistic at α = 0.10, these limits are not presented beyond Figure 2.

4.2. Air Temperature

[25] A systematic increase in annual mean temperature over the period 1965–2006 is indicated at all three elevations ($\alpha = 0.05$). Annual mean daily minimum and maximum temperatures increase by 1.4–2.3°C and 0.8–1.4°C, respectively, over the 40 year period of record, although the amount of change depends on elevation (Figure 3). Water year mean daily minimum temperatures increase about 30% more than mean daily maximum temperatures, and temperature increases at low elevations (076_met) are 60–80% that of midelevations and high elevations (127_met, 176_met). The greatest increase in annual mean daily minimum temperature (0.57°C per decade) is found at the midelevation site (Table 2). However, the most important differences in hydroclimatology by elevation within the watershed are observed at the seasonal time scale.

[26] As with water year trends, all seasons show warming, and minimum temperatures generally increase the most (Table 2). In fall, mean daily minimum temperatures increase at the upper elevation ($\alpha = 0.10$) and midelevation ($\alpha = 0.10$) sites, but mean daily maximum temperature increases only at the upper site ($\alpha = 0.05$). This pattern persists into winter, with the addition of increasing daily minimum temperature at the low-elevation site ($\alpha = 0.05$), greater significance ($\alpha = 0.05$) at the midelevation site and lesser significance in the change in mean daily minimum and maximum temperatures at the upper site ($\alpha = 0.10$). In spring, mean daily minimum temperature increases at all elevations ($\alpha = 0.05$), but no significant trend was found for mean daily maximum temperature. In summer both mean daily minimum and mean daily maximum temperature increases at all sites, but the level of significance in the temporal trend is greater at high elevation than low elevation. Following these patterns of statistical significance, the increases in seasonal mean daily minimum temperatures tend to be greatest at the midelevation sites and the increases in seasonal mean daily

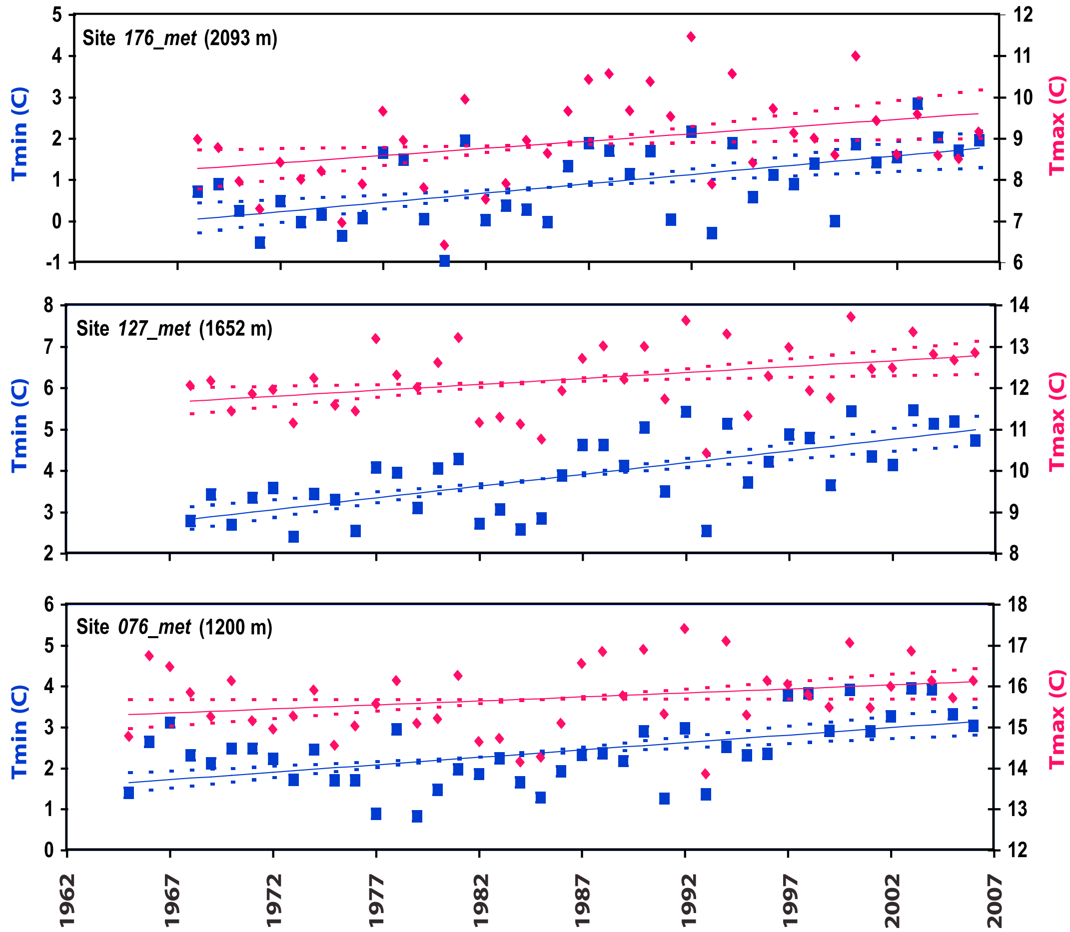

Figure 3. Water year maximum and minimum temperatures for high-elevation (site 176_met, 2093 m), midelevation (site 127_met, 1652 m), and low-elevation (site 076_met, 1200 m) meteorological sites. Red diamonds show maximum temperature, and blue squares show minimum temperature. Trend lines, with upper and lower limits (from the SS estimator) are indicated for each site, showing an increase in both minimum and maximum temperature at all sites.

maximum temperature tend to be greatest at the upper elevation site.

[27] During fall and winter, seasonal mean minimum temperatures are below 0°C and seasonal mean maximum temperatures are above 0°C at all elevations. This corresponds to observations of a diurnal freeze thaw cycle, which prevails throughout the region for much of the cold season. Whereas the range of the diurnal temperature cycle is critical to snow cover status and duration, we define 0°C as a critical threshold for daily mean minimum and maximum temperatures. In this scheme, a cold snow cover is defined by seasonal mean maximum air temperature below zero, a transitional snow cover is defined by a diurnal freeze thaw cycle and an ephemeral snow cover is defined by a seasonal mean daily minimum air temperature above zero.

[28] The fall mean daily minimum temperature at midelevation for the period of record is −0.2°C, with an increasing trend (α = 0.05) of 0.4°C per decade (Table 2). This suggests that at the beginning of the period of record, fall mean daily minimum temperature was about −1°C and by the end of the period of record it was close to 0.6°C. This indicates a fall shift at midelevation from a freeze-thaw diurnal cycle to an above-freezing diurnal cycle.

[29] The winter mean daily maximum temperature at high elevation for the period of record is 0.3°C, with an increasing trend (α = 0.10) of 0.44°C per decade (Table 2). This suggests that at the beginning of the period of record, the winter mean daily maximum temperature was close to −0.6°C, and that by the end it is around 1.2°C. The increase in winter mean daily maximum temperature suggests a shift from winter conditions where the entire diurnal cycle is subfreezing to one that generally has a diurnal freeze-thaw.

[30] These trends indicate that freezing temperatures at midelevations are now limited to winter, and temperatures at high elevations are becoming more like those at midelevation at the beginning of the period of record. Like much of the interior northwestern United States, elevations within the RCEW do not extend much above the high-elevation measurement site (2170 m). In more alpine regions, colder conditions still persist at higher elevations, while in the RCEW there is no higher-elevation land area, so there is no longer a season where the entire diurnal cycle is below freezing.

[31] A complication to elevation-related temperature gradients in mountain valleys is subsidence of cold air. Fall, winter, and spring mean daily minimum temperatures are cooler at the low-elevation site than at the midelevation site,

Table 2. Mean and Standard Deviation in Maximum and Minimum Temperature and Trends Based on Sen's Slope Estimator[a]

	T_{min}			T_{max}		
Site and Elevation (m)	Mean (°C)	SD (°C)	Sen's Slope (°C/decade)	Mean (°C)	SD (°C)	Sen's Slope (°C/decade)
			Annual Water Year			
176_met, 2093	0.9	0.9	0.45**	8.9	1.1	0.35**
127_met, 1652	3.9	0.9	0.57**	12.2	0.8	0.29**
076_met, 1200	2.4	0.8	0.36**	15.7	0.9	0.20*
			Fall (October–December)			
176_met, 2093	−2.9	1.3	0.34*	3.8	1.6	0.38**
127_met, 1652	**−0.2**	1.4	**0.40***	6.5	1.5	0.28
076_met, 1200	−2.1	1.4	0.15	9.4	1.4	0.00
			Winter (January–March)			
176_met, 2093	−5.8	1.6	0.40*	**0.3**	1.7	**0.44***
127_met, 1652	−3	1.4	0.43**	3.4	1.3	0.31
076_met, 1200	−3.6	1.5	0.40**	6.4	1.6	0.23
			Spring (April–June)			
176_met, 2093	2.1	1.4	0.47**	11.1	1.8	0.21
127_met, 1652	5.5	1.4	0.52**	15.1	1.5	0.13
076_met, 1200	4.8	1	0.44**	19.3	1.5	0.16
			Summer (July–August)			
176_met, 2093	10.2	1.3	0.55**	20.3	1.2	0.39**
127_met, 1652	13.2	1.4	0.91**	23.6	1.3	0.39**
076_met, 1200	10.4	1.2	0.40**	27.5	1.3	0.34*

[a]Bold numbers indicate shifts in seasonal temperatures critical to snow cover status caused by trends. Mann-Kendall trend-free prewhitening (MK-TFPW) test significance is indicated for $\alpha = 0.10$ with one asterisk and for $\alpha = 0.05$ with two asterisks.

primarily because strong temperature inversions occur in the valley bottom during the snow season. Generally, the mid-elevation site is just above the inversion level and does not experience this effect.

4.3. Precipitation

[32] Although there is a general increase in water year precipitation with elevation (869 mm km^{-1}, a nearly four-fold increase basin wide), there is an embedded spatial bias evident in the precipitation lapse rate. Sites on the wetter, western side of the RCEW (116_ppt, 155_ppt, 144_ppt, 176e_ppt, 174_ppt, 163_ppt), ranging in elevation from 1491 to 2170 m, show a wind-corrected precipitation range of 474–1124 mm and a precipitation lapse rate of 971 mm km^{-1} but a relative increase by a factor of only 2.4 (1124/474) over the considered gauges (Table 3). Sites on the drier, rain-shadowed eastern side (057_ppt, 076_ppt, 127_ppt, 147_ppt, 167_ppt) ranging in elevation from 1188 to 2003 m, receive 239–806 mm of precipitation, have a smaller lapse rate of 696 mm km^{-1} but a larger relative increase, by a factor of 3.4 (806/239), over the considered gauges (Table 3). Furthermore, as pointed out by *Marks et al.* [2002], *Winstral and Marks* [2002], and *Winstral et al.* [2009], the locations of greatest snow deposition are generally not at the mountain crest, but somewhere just below in wind protected locations. This analysis is presented to illustrate the complexity of precipitation patterns in mountain basins and the limitations of attempting to estimate basin-wide precipitation by establishing a precipitation lapse rate over mountainous areas using limited precipitation data.

[33] Mean annual and seasonal precipitation are presented in Table 3 for locations spanning the range of elevations at RCEW. Water year precipitation data are subject to considerable interannual variability, as evidenced by the high standard deviations in the annual values. However, the division of annual precipitation by season is highly stationary in space and time, as shown by the low standard deviations for seasonal division of precipitation (Table 3). As a consequence, seasonal precipitation at high-elevation,

Table 3. Mean Annual Precipitation and Standard Deviation and Seasonal Mean and Standard Deviation as Percent of Annual Precipitation at 11 Gauges in RCEW for the Period 1962–2006[a]

		Annual		Fall		Winter		Spring		Summer	
Site ID	Elevation (m)	Mean (mm)	SD (mm)	Mean (%)	SD (%)	Mean (%)	SD (%)	Mean (%)	SD (%)	Mean (%)	SD (%)
163_ppt	2169	1124	263	32.9	9.1	38.9	7.7	21.9	7.0	6.3*	3.7
174_ppt	2074	966	230	32.4	9.0	38.2	7.9	22.5	7.2	6.9	3.9
176e_ppt	2056	996	270	33.2	9.5	39.3	8.8	21.0	7.3	6.5*	3.5
167_ppt	**2002**	806	213	32.6	9.9	37.7	9.2	22.3	8.1	7.4	4.0
147_ppt	**1872**	518	148	31.0	9.7	31.9	8.2	27.5	9.8	9.6*	5.6
144_ppt	1815	879	226	32.9	8.7	36.5	8.1	23.3	7.7	7.3	4.4
155_ppt	1654	704	193	32.6	9.4	36.0	8.8	23.2	7.9	8.2*	4.7
127_ppt	**1649**	351	98	28.4	9.6	29.4	8.3	31.1	10.7	11.1*	5.5
116_ppt	1460	474	129	31.3	9.3	32.6	8.4	27.1	8.9	9.0	4.8
076_ppt	**1200**	279	77	28.7	9.5	29.6	9.8	30.1	11.5	11.6	7.2
057_ppt	1**184**	239	68	27.2	9.8	27.4	9.7	32.1	12.3	13.3	8.3

[a]Bold elevations indicate measurement sites on the dry, rain shadow side of the RCEW. Few statistically significant temporal trends ($\alpha = 0.10$, indicated by one asterisk) were found, and these were only for summer precipitation. Note that summer represents a small fraction of annual precipitation, and the modest changes in summer precipitation did not affect the other seasonal percentages.

Table 4. Mean and Standard Deviation and Trends in Percent of Precipitation Falling as Snow Based on SS Estimator[a]

Site and Elevation (m)	Mean (%)	SD (%)	Sen's Slope (%/decade)
	Annual (Water Year)		
176e_ppt, 2056	67.4	10.3	−1.5
127_ppt, 1649	**45.1**	10.8	**−3.8****
076_ppt, 1200	*31.7*	10.5	*−4.8***
	Fall (October–December)		
176e_ppt, 2056	67.4	10.3	−1.5
127_ppt, 1649	**45.1**	10.8	**−3.8****
076_ppt, 1200	*31.7*	10.5	*−4.8***
	Winter (January–March)		
176e_ppt, 2056	83.5	13.2	−6.4**
127_ppt, 1649	**56.1**	16.2	**−5.5****
076_ppt, 1200	**41.6**	17.4	**−8.1****
	Spring (April–June)		
176e_ppt, 2056	87.3	11.6	−0.6
127_ppt, 1649	67.7	17.2	−0.2
076_ppt, 1200	**53.3**	20.7	**−7.1****

[a]Mann-Kendall trend-free prewhitening test significance is indicated for $\alpha = 0.10$ with one asterisk and for $\alpha = 0.05$ with two asterisks. Bold values indicate a shift from snow-dominated to rain-dominated mixed precipitation. Italic values indicate a shift from mixed precipitation to strongly rain dominated precipitation.

midelevation, and low-elevation sites also shows great interannual variability. Slight declines (1% per decade) in the fraction of annual precipitation that occurred in summer were the only significant ($\alpha = 0.10$) findings for trend analysis of precipitation. Though the shift in other seasons is not significant at the 90% level, the decrease in summer precipitation represents a redistribution of water year precipitation to fall and spring at midelevations to high elevations, and to winter and spring at lower elevations.

[34] It is noteworthy that the only statistically significant temporal trend in water year precipitation data is a decrease at the 1815 m site (144_ppt) ($\alpha = 0.10$; not indicated in Table 3) of −45 mm per decade, which is attributed to logging in the mid-1990s. We suggest that controlling for land cover changes in long-term analyses of climate is extremely important, especially for sites where precipitation is dominated by snow.

4.4. Precipitation Phase

[35] At RCEW, 60–75% of the water year precipitation occurs during fall and winter, regardless of elevation (Table 3). Historically, most of that precipitation has fallen

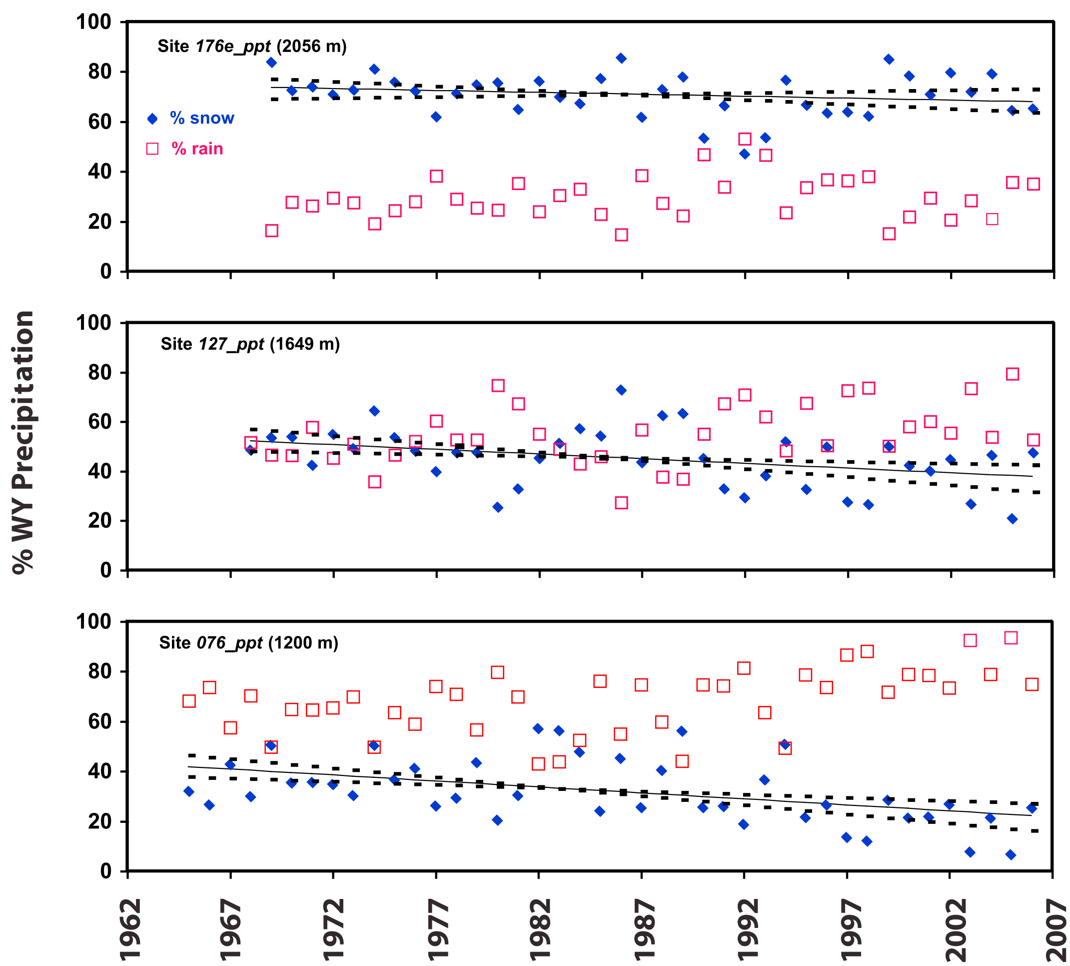

Figure 4. Rain and snow percentage of water year total precipitation at high-elevation (176e_ppt, 2056 m), midelevation (127_ppt, 1649 m), and low-elevation (076_ppt, 1200 m) measurement sites. Blue diamonds show snow, and red squares show rain. Trend lines, with upper and lower limits (from the SS estimator) indicated for each over the period of record show a decrease in snow and an increase in rain percentages at all elevations.

Table 5. Mean and Standard Deviation and Trends in April 1 and 1 May SWE Based on SS Estimator for the Water Year Period 1964–2006[a]

Site and Elevation (m)	1 April Mean (mm)	1 April SD (mm)	1 April Sen's Slope (mm/decade)	1 May Mean (mm)	1 May SD (mm)	1 May Sen's Slope (mm/decade)
163_sc, 2162	701	200	−27	686	241	−20
163b_sc, 2147	640	210	−29	502	288	−61
163c_sc, 2125	663	219	−35	533	289	−49
174_sc, 2073	637	211	−43	491	284	−75*
176e_sc, 2056	542	203	−47*	377	266	−86**
144_sc, 1815	205	163	−91**	42	112	−
155_sc, 1743	163	117	−32	13	55	−

[a]Mann-Kendall trend-free prewhitening test significance is indicated for α = 0.10 with one asterisk and for α = 0.05 with two asterisks. The large 1 April change at the 1815 m elevation (site 144_sc) is attributed to deforestation, which occurred in the mid-1990s. Trends and significance for 1815 and 1743 m elevations (sites 144_sc and 155_sc) are not reported for 1 May because there were numerous null (zero SWE) values, which would violate the assumption of normally distributed residuals used in significance analysis.

as snow, melted during spring, and provided water for streamflow during spring and summer. The high-elevation site, at 2093 m (176e_ppt), shows little change in precipitation phase, with a relatively constant mean water year value of 67% snow over the period of record. However, temporal trend analysis found the percentage of annual precipitation as snow has decreased significantly (α = 0.05) at representative midelevation and low-elevation sites over the past 40 years (Table 4).

[36] Whereas the extent of decrease in snow as a percent of total water year precipitation is similar for midelevation and low-elevation sites, the implication for the snow cover is not. At midelevation, water year precipitation as snow has decreased about 4% per decade over the study period, from 53% to 38% (Table 4). This change indicates that midelevations (e.g., site 127_ppt) have shifted from a hydroclimatic regime of snow dominated mixed snow and rain to a rain dominated mixture (Figure 4). A similar rate of change (5% per decade) at low elevation has reduced water year precipitation as snow, from 41% to 22% and shifted the hydroclimatic regime to a strongly rain dominated mixture (Figure 4).

[37] Seasonal analysis of long-term trends for fall, winter, and spring seasons supports the findings from the analysis of annual totals, and further clarifies the control of elevation on the seasonal transition from snow to rain (Table 4). During fall, there has been a large decrease (α = 0.05) in the percent of precipitation that is snow at all elevations (Table 4). At high elevation the decline from 97% to 70% snow indicates that during fall high elevations have shifted from a totally snow dominated system to one that, while still snow dominated, has a significant rain component. At midelevations and low elevations the declines from 67% to 45% snow and 58% to 25% snow, respectively, indicate that both sites have shifted from a snow-dominated rain-snow mix to one that is rain dominated. During winter, the only significant (α = 0.05) decreasing trend in the percent of precipitation that is snow occurred at low elevation, where snow decreased from 67% to 39% of seasonal precipitation (Table 4). This indicates that, even during winter at low elevations, there has been a shift from a snow-dominated to a rain-dominated system. In spring, significant decreasing trends occur at both midelevations (α = 0.05) and low elevations (α = 0.10). Again, the elevational trend is evident, as the decrease in snow fraction is greatest at lowest elevation and less at midelevation and high elevation.

4.5. Snow Water Storage

[38] Snow water equivalent (SWE) integrates precipitation and energy balance and is thus affected by changes in air temperature and precipitation phase, both of which vary with elevation in mountainous regions. The clear relation of mean 1 April and 1 May SWE (Table 5) to elevation supports the previous finding of *Marks et al.* [2001a] of strong differences in SWE timing and depth between the high-elevation and the low-elevation and midelevation snow courses. These differences persist in the temporal trends in four measures of snow water storage: depth of 1 April and 1 May SWE (Table 5) and timing and depth of peak SWE (Table 6).

[39] While data from all snow courses indicate a decrease in both 1 April and 1 May SWE, only trends for the midelevation to lower-elevation sites are statistically significant (Table 5). Site 176e_sc (2056 m) shows a decreasing trend in 1 April SWE (α = 0.10) of nearly 32% over the period of record, while site 144_sc (1815 m) shows an even larger decreasing trend over the period of record. However, a portion of the very large decreasing trend at course 144_sc may be due, in part, to deforestation that occurred near the site in the mid-1990s. Data for 1 May SWE show statistically significant trends at a slightly higher elevation band, with course 174_sc (2073) showing a 49% decrease (α = 0.10) and course 176e_sc (2056 m) a 66% decrease (α = 0.05) over the period of record. The data indicate that the decrease in SWE and shift to an earlier peak SWE are much more substantial at the low-elevation and midelevation courses than at the high-elevation courses.

[40] Figure 5 presents example time series for 1 May SWE for high-elevation (163c_sc, 2125 m), midelevation (176e_sc, 2056 m) and low-elevation (155_sc, 1743 m) snow courses. At the low-elevation course (155_sc), snow cover persisted past 1 May for about half the years prior to 1984 and no snow has been measured on 1 May since that year (Figure 5). Trend analysis for the (155_sc, 1743 m) is therefore limited for 1 April SWE, and not possible for

Table 6. Mean and Standard Deviation and Trends in Timing and Depth of Peak SWE Accumulation Based on SS Estimator[a]

Site and Elevation (m)	Timing of Peak SWE Mean Date	Timing SD (days)	Timing Sen's Slope (d/decade)	Peak SWE Mean (mm)	Peak SWE SD (mm)	Peak SWE Sen's Slope (mm/decade)
163_sc, 2162	15 Apr	15	−1	757	211	−21
163b_sc, 2147	7 Apr	15	−1	681	222	−35
163c_sc, 2125	7 Apr	16	−1	695	225	−30
174_sc, 2073	6 Apr	15	−1	665	218	−53
176e_sc, 2056	31 Mar	15	−2	573	210	−58
144_sc, 1815	11 Mar	24	−9**	272	147	−90**
155_sc, 1743	28 Feb	21	−6*	254	99	−9

[a]Mann-Kendall trend-free prewhitening test significance is indicated with one asterisk for α = 0.10 and for α = 0.05 with two asterisks. As in Table 5, the very large change in both the timing and volume of peak SWE at the 1815 m elevation (site 144_sc) is attributed to deforestation, which occurred in the mid-1990s.

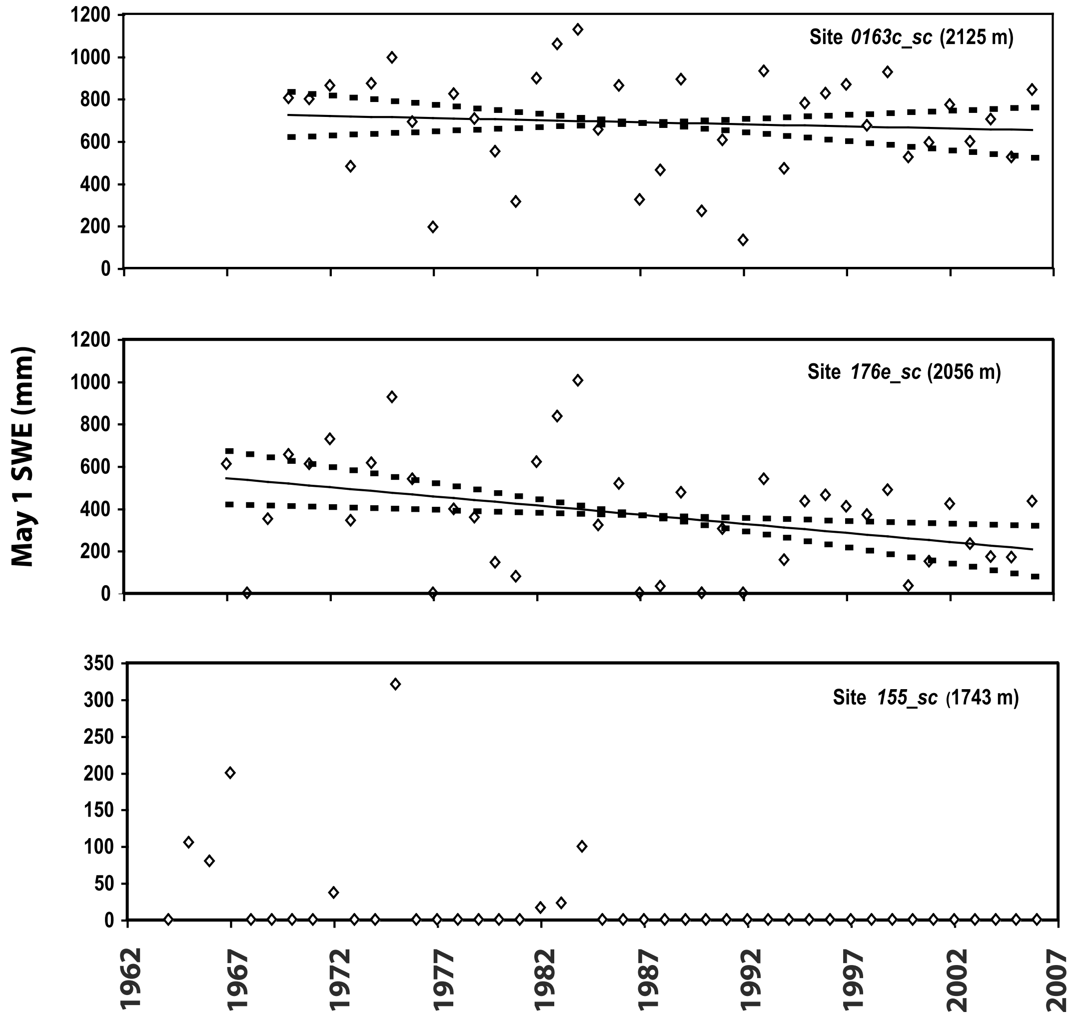

Figure 5. SWE from 1 May (mm) for high-elevation (163c_sc, 2125 m), midelevation (176e_sc, 2056 m), and low-elevation (155_sc, 1743 m) snow courses. Trend lines, with upper and lower limits (from the SS estimator), are indicated for the period of record, showing a consistent decrease at the 2125 and 2056 m snow courses. Trend lines are not shown for the 1743 m snow course (155_sc) because zero values dominate the later part of the record.

1 May SWE for the two low-elevation courses (144_sc, 1815 m; 155_sc, 1743 m) due to the large number of zero values in the later part of the record (Table 5). Melt-out at the low-elevation snow course now consistently occurs in late March or early April.

[41] However, the two low-elevation snow courses (144_sc (1815 m), 155_sc (1743 m)) show a strong temporal response to warming. Site 144_sc shows statistically significant trends (α = 0.05) toward both earlier peak SWE and a reduction in peak SWE, while site 155_sc shows a statistically significant trend (α = 0.10) toward earlier peak SWE (Table 6). The low-elevation snow courses (144_sc, 1815 m; 155_sc, 1743 m) show a shift of 6–9 days earlier peak SWE per decade or 26–39 days earlier over the period of record. At the end of the record, SWE at the low-elevation snow courses tends to peak in early to mid-February, indicating that whereas these two snow courses initially represented different snow environments, they are now similar in that they essentially represent the lower extent of the seasonal snow cover within the RCEW. The lowest-elevation course (155_sc, 1743 m) shows a similar trend in depth of peak SWE, but as with 1 May SWE, the persistence in the trend is not significant, due to the absence of late spring snow cover at this site after 1984 (Figure 5).

[42] The three highest courses (163_sc, 2162 m; 163b_sc, 2147 m; 163c_sc, 2125 m) show an average 18% decrease in 1 April SWE, a 28% decrease in 1 May SWE, a shift in the date of peak SWE to about 5 days earlier, and a 16% reduction in peak SWE. The three lowest courses (176e_sc, 2056 m; 144_sc, 1815 m; 155_sc, 1743 m), however, show a 57% decline in 1 April SWE, a 66% decline in 1 May SWE, a 24 day shift to earlier peak SWE, and a 47% reduction in peak SWE. Though the elevation difference between the high- and low-elevation snow courses is only around 300 m, the impact of climate warming on the seasonal snow cover is more than twice as large for low-elevation courses than for high-elevation courses.

[43] Though the record is shorter (1983–2006, 24 water years), the hourly snow pillow data provide a more detailed perspective on how climate warming has affected the seasonal snow cover at a specific site. Data from the snow pillow on the date and depth of peak SWE are similar to the

Table 7. Mean and Standard Deviation and Trends per Decade for the Hourly Snow Pillow Data Based on Sen's Slope Estimator[a]

	Mean	SD	Sen's Slope
Snow cover initiation date	3 Nov	14 days	6.4 d/decade*
Date of peak SWE	28 Mar	17 days	0.0 d/decade
Peak SWE depth	564 mm	232 mm	−58.5 mm/decade
Melt-out date	15 May	16 days	−4.0 d/decade

[a]Data are for site 176e_sp, 2056 m, 1983–2006. Only the snow cover initiation date shows a statistically significant trend (α = 0.10, indicated by an asterisk), toward a later date. The data indicate that the date of peak SWE is unchanged, the volume of peak SWE has decreased, and the melt-out date is earlier, though these are not statistically significant.

176e_sc snow course located adjacent to the pillow. The snow pillow provides detailed information on the specific dates of snow cover initiation, timing of peak SWE and melt-out, which can only be implied from the biweekly snow course data. Table 7 shows a significant (α = 0.10) trend at that site for snow cover initiation that is nearly a week later per decade, or more than 2 weeks later over the 24 year period of the snow pillow record. Though they are not statistically significant, the data indicate the date of peak SWE has not changed, but a 22% reduction in peak SWE and melt-out about 10 days earlier has occurred. These results are consistent with snow course data presented above, and are useful in illustrating how change has occurred over the 45 year period of record at RCEW. First, the SS value of −58 mm per decade for decrease in depth of peak SWE is the same for snow course 176e_sc (2056 m) with 44 years of data, as for the snow pillow (176e_sp, 2056 m) with only 24 years of data. The same is generally true for the snow pillow and snow course mean peak SWE (564 versus 573 mm), as well as the timing of peak SWE (28 March versus 31 March). Though the snow cover has not been as strongly affected by climate warming at this elevation as at lower elevations (Table 6), data from both the snow pillow and snow course indicate a snow season that is about 9 days shorter now than it was at the beginning of the record. This suggests that in the coming years the snow cover at this site may undergo changes similar to those we have observed at the midelevation site. It is clear that in the future we can expect a limited snow cover at elevations below 2000 m, and a shorter snow season with less snow above 2000 m in RCEW and the similar areas across the Great Basin in the western United States.

4.6. Streamflow

[44] Streamflow data from three weirs, Reynolds Mtn. East (RME), Tollgate (TG), and the RCEW basin outlet were analyzed by water year and by season over the 1964–2006 period of record. At all three weirs, most of the water year flow occurs from March to June. During these months, nearly 90% of annual streamflow occurs at the RME headwater weir (166_sf), 82% at the midelevation TG weir (116_sf), and 70% at the RCEW outlet weir (036_sf). Water year runoff volume variability is great, with standard deviations of at least 50% of the annual mean at all sites. While this is partially due to the interannual variability in precipitation, the standard deviations of water year precipitation ranged from 20% to 30% of the annual mean, which is less than half of the variability in annual runoff. Though the SS values are decreasing for all three weirs, as with precipitation, no significant temporal trends in annual runoff volume emerged for the period of record (Table 8).

[45] Whereas mean water year streamflow, like precipitation, shows no statistically significant trends and is assumed unchanged over the period of record, seasonal streamflow, like snowfall and snowmelt, has shifted toward earlier flow. Trend analysis of monthly streamflow as a fraction of annual total indicates a shift to late winter and early spring flows at RME (166_sf) and TG (116_sf) weirs, with increases in March and April and decreases in May and June flows (Table 8). Temporal trends at RCEW outlet (036_sf) weir are less conclusive, with an increase in May flows as the only significant change. Spring and summer diversions to irrigation below 116_sf probably confound trend analysis at the RCEW outlet weir. There is a strong elevational gradient to this shift. The high-elevation, headwater weir shows significant (α = 0.10) March and April flow increases, while the midelevation weir shows significant (α = 0.05) April flow increases. The outlet weir, however, shows a significant (α = 0.05) May flow increase. The only significant (α = 0.10) change in June flow is a decrease at the high-elevation weir.

5. Conclusions

[46] This analysis presents a comprehensive and coherent assessment of trends in precipitation, snow, climate, and streamflow along a transect of elevations within the RCEW.

Table 8. Mean and Standard Deviation and Trends in Streamflow Based on SS Estimator in Monthly Distribution of Streamflow as Percent of Total Water Year Streamflow During March–June for the Period of Record[a]

	Water Year Stream Discharge		
Weir and Elevation (m)	Mean (10^6 m^3/WY)	SD	Sen's Slope (10^6 m^3/decade)
166_sf, 2022	0.21	0.10	−0.008
116_sf, 1404	13.4	7.52	−0.75
036_sf, 1099	17.1	12.4	−1.66
	Monthly Distribution of Streamflow		
Weir and Elevation (m)	Mean (%)	SD	Sen's Slope (%/decade)
	March		
166_sf, 2022	4.1	4.2	0.4*
116_sf, 1404	11.0	5.0	0.5
036_sf, 1099	14.9	7.5	−0.2
	April		
166_sf, 2022	22.8	13.5	3.0*
116_sf, 1404	24.3	9.4	2.9**
036_sf, 1099	19.6	7.7	0.9
	May		
166_sf, 2022	45.6	11.3	−0.3
116_sf, 1404	33.9	7.9	0.6
036_sf, 1099	24.0	10.4	2.0**
	June		
166_sf, 2022	16.6	10.5	−2.3*
116_sf, 1404	12.5	6.2	−1.1
036_sf, 1099	10.3	6.0	−0.4

[a]Mann-Kendall test significance is indicated for α = 0.10 with one asterisk and α = 0.05 with two asterisks. The water year (1 October to 30 September) is used because in the western United States, the annual water cycle spans the calendar year, beginning in the fall and ending during the dry summer period. WY, water year.

In general, it shows similar hydroclimatic trends to those reported in the literature but provides greater detail into how these trends vary with elevation and location within the mountain basin. The more complete meteorological record and data from a range of elevations and site conditions within the RCEW indicate there are elevational gradients and seasonal differences to climate warming and its effects that may have significant hydrologic impacts in the region.

[47] In agreement with other studies of temperature trends in the western United States and Canada [e.g., *Trenberth et al.*, 2007], temperatures have significantly warmed at all elevations within the RCEW, with trends indicating minimum temperature warming is greater (+0.36 to +0.57°C per decade) than maximum temperatures (+0.2 to +0.35°C per decade). Trends in the data indicate that important thermal thresholds have been crossed during the period of record.

[48] At high elevation, the mean water year diurnal cycle is shifting from freeze-thaw to above freezing and the winter diurnal cycle from below freezing to a daily freeze-thaw. Although this change at high elevation has not been accompanied by statistically significant changes in phase of precipitation or depth and timing of peak SWE, snow pillow data indicate that initiation of the seasonal snow cover occurs later, and melt-out occurs earlier. This results in a snow season that is at least a month shorter than it was in the mid-1960s.

[49] Precipitation phase (snow versus rain) is strongly affected by warming climate. The rain/snow proportions of water year precipitation are critical to the timing of streamflow in the RCEW. While the ratio of rain to snow has increased at all elevations, at midelevation precipitation has moved from a snow to rain dominated mix and at low elevation to a completely rain dominated system. The data also indicate a decrease in SWE at all elevations, with the largest and most significant decreases at midelevations and low elevations. This elevational gradient is important to the hydroclimatology of the RCEW because much more area exists at midelevations and low elevations than at high elevations. Dates of peak SWE occur earlier at all elevations. At low elevations, the date of peak SWE has shifted from mid-March to mid-February.

[50] The data indicate that, while there is large year-to-year variability, water year (annual) streamflow and precipitation have not changed over the period of record. However, as other studies have shown [e.g., *Regonda et al.*, 2005; *Mote et al.*, 2005], there has been a seasonal shift in streamflow, with increases in winter and early spring and decreases in late spring and summer. This shift is stronger at high elevations and delayed at midelevations and low elevations. Fall, winter, and spring precipitation depth has not changed, but the proportion of snow has significantly declined at all elevations.

[51] The hydrologic importance of these trends is in the hypsometry of mountain basins like the RCEW, where only 2% or about 5 km^2 of the land area is above 2060 m, while 24% or about 57 km^2 of the RCEW is between 1700 and 2060 m. Though at higher elevations it is colder and the seasonal snow cover is less affected by climate warming, its hydrologic impact on the RCEW is not as important as the midelevation to low-elevation snow cover because within the RCEW it represents a small area. The loss or redistribution of 100 mm of high-elevation (above 2060 m) SWE represents less than 1% of the mean water year streamflow from the RCEW, while the same 100 mm loss or seasonal redistribution of SWE from 1700–2060 m seasonal snow cover represents more than 14% of water year streamflow from RCEW.

[52] Together these results indicate that the hydroclimatology of the RCEW and similar regions of the interior northwestern United States have already been affected by climate warming. Snowfall and the seasonal snow cover have been significantly affected. Changes in snow deposition and melt have altered patterns of stream flow. In the short term, these trends will likely have a significant impact on land and water management practices. If they were to continue for the next 50–100 years, as suggested by the IPCC report [*IPCC*, 2007], the RCEW and similar areas within the interior Northwest and Great Basin of the United States will be very different hydroclimatically and ecohydrologically.

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