Trends in snow water equivalent in the Pacific Northwest and their climatic causes

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[1] Observations of snow water equivalent (SWE) in the Pacific Northwest are examined and compared with variability and trends in temperature and precipitation at nearby climate stations. At most locations, especially below about 1800 m, substantial declines in SWE coincide with significant increases in temperature, and occur in spite of increases in precipitation. *INDEX TERMS:* 1655 Global Change: Water cycles (1836); 1863 Hydrology: Snow and ice (1827); 1630 Global Change: Impact phenomena. **Citation:** Mote, P. W., Trends in snow water equivalent in the Pacific Northwest and their climatic causes, *Geophys. Res. Lett.*, *30*(12), 1601, doi:10.1029/2003GL017258, 2003.

1. Introduction

[2] Several studies have noted recent declines in snow in various parts of the world. Satellite measurements indicate a decline in snow cover extent since 1966 [*Robinson*, 1999]. Surface observations of snow cover data for the northern hemisphere (1915–1992) indicate an increase in snow extent in winter and a decrease in spring [*Brown*, 2000], especially for the period since the 1950s. These changes have usually occurred in concert with gradual increases in temperature.

[3] This work examines variability and trends in snow course data in the Pacific Northwest (defined here as the states of Idaho, Oregon, and Washington in their entirety, southern British Columbia, and Montana west of the continental divide) and compares the snow data with variability and trends in temperature and precipitation at nearby climate stations. The Pacific Northwest relies heavily on melting snow to provide water for many uses during the summer, when relatively little precipitation falls. Persistent regional warming, therefore, would threaten an important resource, and resource managers need to know whether there is already evidence of a decline in snowpack and, if so, whether it is a harbinger of anthropogenic climate change or a natural cycle that will likely reverse.

2. Data and Methods

2.1. Snow Course Data

[4] Measurements of snow water equivalent (SWE) for the Northwest were described by *Clark et al.* [2001] and were used by *McCabe and Dettinger* [2002] to demonstrate an improved capability for seasonal streamflow forecasting. Data through 2002 were downloaded from the NRCS Water and Climate Center web site (www.wcc.nrcs.usda.gov/snow/snowhist.html) for the U.S. and, for British Columbia, from

the Ministry of Sustainable Resource Management web site (srmwww.gov.bc.ca/aib/wat/rfc/archive/historic.html).

[5] The earliest measurements date back to 1915 at a location in south central Washington, but very few sites were observed routinely before 1935. Nearly all of the early observations were taken only yearly on April 1, near the peak of the snow accumulation season, but in subsequent years other routine observation dates were added.

[6] For inclusion in this study, a record must extend at least from 1950 to 2000, chosen to maximize total data (number of years × number of stations). Records missing more than 25% of the values were excluded, leaving 230 records. A subset of this data set was used to construct Figure 5.

[7] Most locations of snow course data are at high altitudes (1200–2300 m); the lowest used here (576 m) lies in the Washington Cascades, and the highest (2703 m) in south central Idaho.

2.2. Climate Data

[8] The climate records used here were described and analyzed by *Mote* [2003a, 2003b] and include data from the USA and Canada. Canadian data come from the Historical Canadian Climate Database (HCCD [*Vincent and Gullett*, 1999]), which has 32 stations in the study area with temperature records and 81 stations with precipitation records. For the USA I use the Historical Climate Network (USHCN [*Karl et al.*, 1990]), which has 122 stations in our study area with temperature records and 84 stations with precipitation records. In both data sets a majority of stations have periods of record beginning by 1920, and nearly all by 1940.

[9] For each snow record, comparisons with temperature and precipitation were performed as follows. First, I located the nearest five climate stations with complete records for 1950–2000, typically 20–100 km away and at lower elevation than the snow course. The climate anomalies for these stations were averaged for November–March (NDJFM). In this season (but not in summer) precipitation and temperature are themselves nearly uncorrelated at most locations. Correlations of April 1 SWE with NDJFM temperature ranged from -0.71 to $+0.09$ with a mean of -0.29, and correlations with NDJFM precipitation ranged from -0.18 to 0.89 with a mean of 0.55. Correlations between SWE and temperature are strong (<-0.5) in milder climates, notably in the Cascades below 1800m in Oregon and below 1500m in Washington, and correlations are weak (>-0.3) in colder climates, viz., most of BC, Idaho, and Montana. Correlations with precipitation are usually stronger than with temperature, and are especially strong in the Rockies (where temperature plays little role) and especially

weak in the Wallowa Mountains of northeastern Oregon, where *McCabe and Dettinger* [2002] found that SWE varied fairly independently of the rest of the region. The correlations are insensitive to the number of climate stations used in the analysis; in fact, using a regionally averaged reference time series yields similar results. Such regional coherence of snow and climate, and their response to El Niño-Southern Oscillation and the Pacific Decadal Oscillation, hold promise for improving predictions of seasonal streamflow [*McCabe and Dettinger*, 2002].

2.3. Trend Analysis and Comparison

[10] To elucidate the roles of NDJFM temperature and precipitation in producing variations and trends in April 1 SWE, multiple linear regression is performed on the snow data $S(t)$ using reference time series for temperature $T(t)$ and precipitation $p(t)$, yielding coefficients of regression a_T and a_P. The component of any trend $\langle S \rangle$ in the snow data that can be attributed to the temperature trend $\langle T \rangle$ is given by

$$\langle S \rangle_T = a_T \langle T \rangle; \qquad (1)$$

likewise for precipitation. Weighting the climate trend by its interannual regression assumes that if the connection between S and T is strong on the interannual timescale, it is strong also on longer timescales.

3. Results

[11] At nearly all snow course sites, linear trends over the period of record are negative, some substantially so (Figure 1). Decreases are generally largest in the Cascades, many in excess of 40%; this is true also for other periods of record.
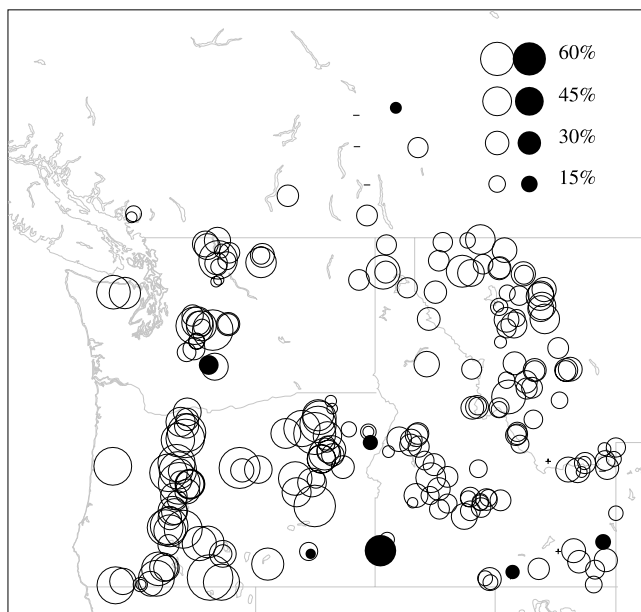

Figure 1. Linear trends, relative to starting value, in snow water equivalent (SWE) on April 1 over the period of record 1950–2000. Negative trends are shown as open circles, positive trends as solid circles; the magnitude of the trend is indicated by the area of the circle according to the legend. Trends less than 5% in absolute value are indicated by a + or − symbol.
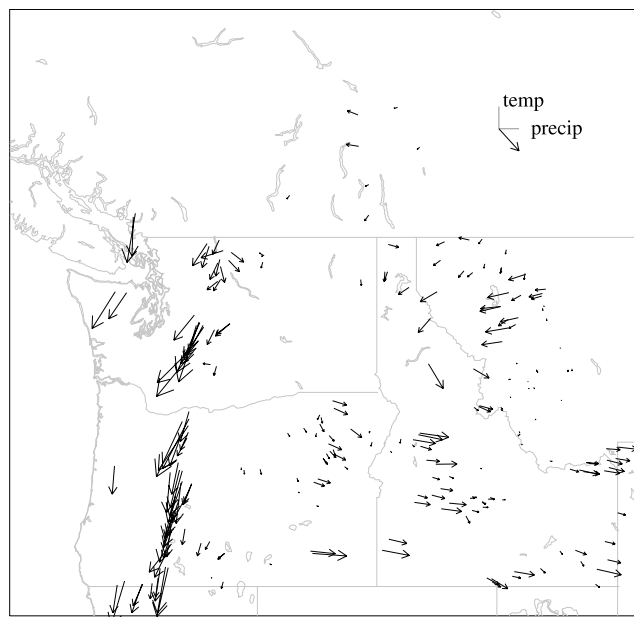

Figure 2. For each location, the arrow shows the trend in April 1 SWE attributed by multiple linear regression to precipitation $\langle S \rangle_P$ (x-direction) and temperature $\langle S \rangle_T$ (y-direction). The reference arrow in the legend has a length of 10 cm of SWE.

[12] Several important questions are raised by the data presented in Figure 1. Can we relate these changes in SWE to changes in temperature and precipitation? How do these changes vary with season and for different periods of record? How well do linear trends capture the important longer-term variations?

[13] To address the first question, Figure 2 shows $\langle S \rangle_P$ and $\langle S \rangle_T$ (see section 2.3). Several regional patterns emerge. In the Cascades and coast ranges, where declines in SWE have been largest, and at some locations in the northern Rockies, both temperature and precipitation have acted to decrease SWE as indicated by arrow directions ranging from south to southwest. In much of eastern Oregon and southern Idaho, and it is surprising how many sites show this pattern, the decreases in SWE have occurred in spite of an increase in precipitation (arrow directions ranging from south to east). For locations with large absolute trends, the sum $\langle S \rangle_P + \langle S \rangle_T$ is typically 40–80% of the observed trend, but in many locations with small trends the fit is poor, as for example in the Wallowa Mountains of northeastern Oregon (see Section 2.2).

[14] Another, more qualitative way to compare the roles of temperature and precipitation is to plot the snow trends as a scatterplot in a space spanned by precipitation trends and temperature trends (Figure 3). Nearly all the snow course data with near-zero or negative precipitation changes (upper left part of the diagram) have experienced decreases, regardless of temperature trend, but in addition nearly all of the points with positive precipitation changes have nonetheless also experienced decreases in SWE.

[15] Another approach to separating the temperature and precipitation influences is to examine the trends as a function of elevation. Relative changes in SWE should be nearly uniform with altitude for changes in precipitation, but
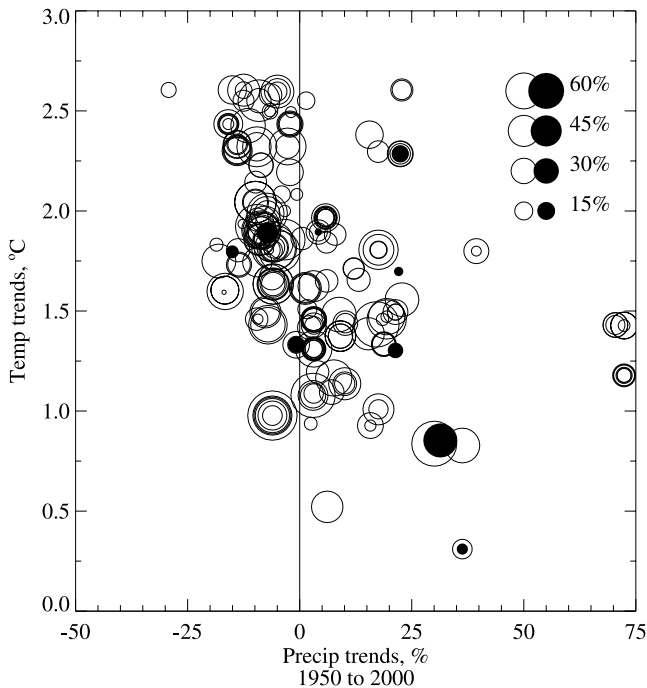
Figure 3. The same data as in Figure 1 are plotted against trends in NDJFM precipitation and temperature at nearby climate stations.

much greater at lower elevations for changes in temperature, since at elevations nearer the mean freezing level a moderate change in temperature can dramatically change the fraction of precipitation that falls as snow. To test this hypothesis, we plot in Figure 4 the trend in SWE as a function of snow course elevation. As expected, the data show a decrease in magnitude of trend with elevation, up to about 1800 m; the correlation of trend with elevation is 0.51. This result is even more pronounced when considering only the snow course records in the Cascades and Olympics (not shown), where freezing levels tend to be higher owing to the marine influence. Thus the elevational dependence of the trends confirms the dominant role of temperature increases in driving the trends.

[16] In order to examine changes in SWE by month and year, we combine the SWE data for those sites with long records. For this part of the analysis we use data from 1925 to 2002 and for each month we include all records that are 75% complete for this interval. Following the approach of *Clark et al.* [2001], the data at different locations are combined by first converting each time series of SWE to a time series of z-scores by subtracting the mean and normalizing by the standard deviation. The individual time series are then averaged and converted back to SWE using the mean and standard deviation averaged over all the time series. With this approach, the time series for each month is less sensitive to the particular combination of snow course records reporting in a given year, but it is inadequate to show the seasonal cycle since a different set of snow course records is used for each month. In fact, there is such a strong reporting bias in May and June toward sites with high SWE that it skews the mean seasonal cycle. For comparison, the mean value of the time series for May has been set to be

79% of the mean April value, using the basin average May/April ratio derived from SNOTEL data [*Clark et al.*, 2001].

[17] The temporal behavior of regional mean SWE thus derived (Figure 5) shows interdecadal variability with generally low values from 1925 to 1945 and after 1975, and high values in the 1950s and 1970s. In this analysis, the mean SWE in the 1990s was the lowest of any 10-year period in the record, but the difference is not significant because so few observations were made in the 1920s. However, regionally averaged USHCN climate data show that for no 10-year period in the 1920s and 1930s was NDJFM precipitation or temperature as high as in the 1990s.

4. Discussion and Conclusions

[18] The declines in Northwest spring snowpack presented here provide further evidence of regional increases in temperature and are qualitatively consistent with observed trends in temperature and precipitation at nearby stations. Both the dependence on elevation and the regression analysis confirm the role of temperature in reducing snowpack since the mid-20th century. Relative declines are most pronounced in the Cascades, where warming, moderate elevation, and declines in precipitation have all contributed to declines in SWE. In the upper Columbia River basin, moderate increases in precipitation in mid-century offset the warming to produce slight increases in SWE since 1940, though these change sign when considering only the period since 1950.

[19] These changes in recent decades are broadly consistent with those derived from lower-elevation weather stations [*Groisman et al.*, 1994; *Brown*, 2000], with little change in winter but with decreases in spring. Positive snow-albedo feedback likely contributes to the springtime
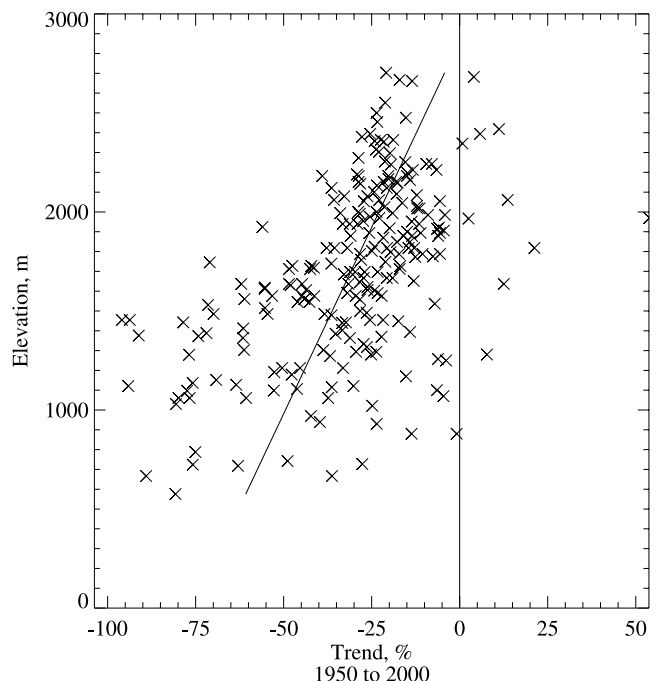

Figure 4. Relative trend in April 1 SWE plotted against snow course elevation (crosses) along with a linear fit.
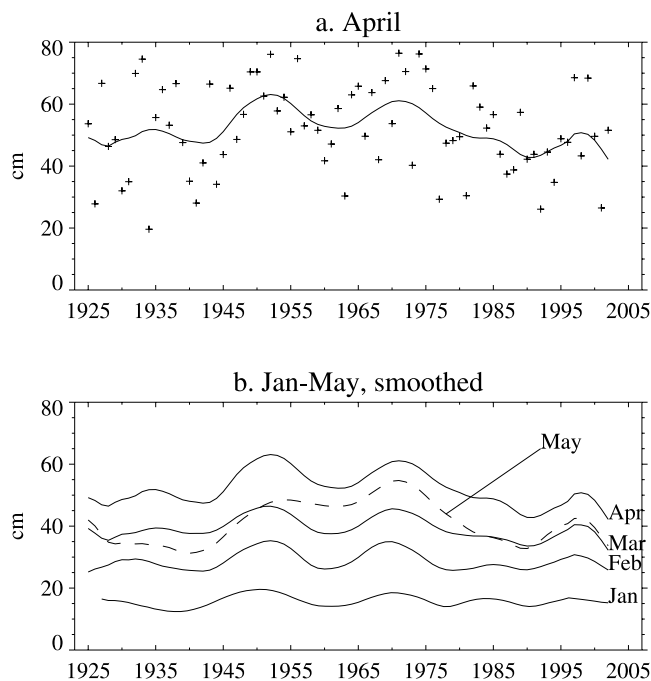
a. April

b. Jan-May, smoothed

Figure 5. Regionally averaged SWE (a) for April 1 of each year, with a smooth curve overlaid, and (b) the smoothed annual values for each month January–May. The May curve has been adjusted for observational bias (see text). Smoothing is performed using locally weighted regression [*Cleveland*, 1993] with parameters chosen to emphasize timescales greater than about10 years.

trends in temperature and snow cover both at low elevations and in the mountains, especially in the forested zone.

[20] An issue of some concern in relating trends in snow course data to trends in climate data is whether trends at the lowland climate stations are representative of trends at the higher altitudes of the snow course data. The results in Figures 2 through 5 suggest that increases in temperature have overwhelmed changes in precipitation in recent decades, but a variety of factors complicate this interpretation. Interannual correlations between SWE and climate are relatively high but may mask long-term changes, whether they reflect a real climatic trend or are a result of changes in the nearby land cover, wind speed or wind direction, instrumental changes, or even air pollution affecting low-elevation precipitation more than mountain precipitation [*Givati and Rosenfeld*, 2003]. Regional averaging might reveal some such changes and remove others. Hydrological modeling may help answer some of these questions.

[21] Could these changes be a reflection of global warming, or are they a natural fluctuation? The Pacific Decadal Oscillation [*Mantua et al.*, 1997] influences regional climate, and the correlation between PDO and the regionally averaged SWE time series (Figure 5a) is −0.49. However, PDO alone cannot explain the changes in SWE between the period before 1945 and the period since 1976: Temperatures have been substantially higher during the later period for comparable values of PDO. Clearly, regional warming has played a role in the decline in SWE, but regional warming at the spatial scale of the Northwest cannot be attributed statistically to increases in greenhouse gases [*Stott and Tett*,

1998]. However, as greenhouse gases continue to accumulate, regional warming is likely to continue as well, and questions of cause will recede.

[22] These results have significant implications for water resources managers in the Northwest. Since snowmelt provides much of the water used during summer for irrigation, municipal and industrial water supply, flow targets for fish protection, recreation, and other uses, future changes in regional snow cover are of great concern. Simulations of the region's hydrology highlight loss of snowpack as a primary impact of future anthropogenic warming [e.g., *Hamlet and Lettenmaier*, 1999], but such studies have usually focused on time horizons (e.g., the 2050s) beyond those of most planning exercises. The fact that a warming-induced loss of springtime snowpack has already been observed heightens the urgency of developing adaptation strategies for coping with the gradual loss of snowpack, which is clearly already well under way.

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