ASSESSING THE IMPACTS OF GLOBAL WARMING ON SNOWPACK IN THE WASHINGTON CASCADES

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ABSTRACT

The decrease in mountain snowpack attributable to global warming is difficult to estimate in the presence of the large year-to-year natural variability in observations of snow water equivalent (SWE). A more robust approach for inferring the impacts of global warming is to estimate temperature sensitivity of spring snowpack and multiply it by putative past and future temperature rises attributable to global warming.

Estimates of sensitivity can be obtained from (*a*) geometric considerations regarding the change in the climatological snow line resulting from warming, (*b*) regression of historical April 1 SWE measurements upon mean winter temperatures, (*c*) a hydrological model forced by daily temperature and precipitation observations, and (*d*) the distribution of precipitation versus temperature at SNOTEL stations. All four methods yield an estimated 20% loss of spring snowpack for 1°C warming; considering warming-induced precipitation increases, the sensitivity would decrease to 16%.

Using various rates of temperature rise over the Northern Hemisphere, it is estimated that spring snow water equivalent in the Cascades portion of the Puget Sound drainage basin should have declined by 8-16% over the past 30 years due to global warming and it can be expected to decline by another 11-21% by 2050.

INTRODUCTION

Recent investigations of Cayan et al. (2001), Groisman et al. (2004), Regonda et al. (2005), Stewart et al. (2005), Hamlet et al. (2005), Knowles et al. (2006), Mote (2006), Mote et al. (2008), and Barnett et al. (2008) have all found evidence of hydrological impacts of global warming over parts of the western United States since the mid-20th century. Quantitative assessments of the extent of these effects are subject to large uncertainties because hydrological variables like snowpack exhibit large year-to-year and decade-to-decade variability in association with changes in the atmospheric circulation that affect the distribution of precipitation (Cayan, 2001). In the presence of this presumably natural background variability, the magnitude and sometimes even the sign of trends may be dependent on the choice of end points used in the calculations, in which case, different choices made by different analysts can yield conflicting impressions of the significance of the impacts of global warming. The sensitivity of the trends to the choice of period of record is underscored by sharply contrasting results of Mote et al. (2008) who reported losses of springtime snow water equivalent (SWE) at stations in the Pacific Northwest since the mid-20th century at a large span of elevations (Figure 1), and the lack of significant trends in time series of SWE at some snow courses in the same region from 1977 to 2006 (Figure 2).

REGION OF STUDY

This analysis is specifically applicable to the Cascades portion of the Puget Sound drainage basin shown in Figure 3, comprising most of the west-facing slope of the Cascades range. This area is of interest due to the media coverage regarding the trends in snowpack in the basin (Cornwall, 2007) in addition to the recent work of Mote et al. (2008).

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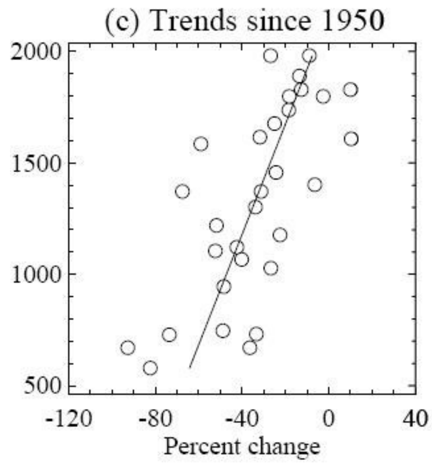

Figure 1. Trends in Cascades April 1 SWE versus elevation, 1950-2006 as observed at a variety of snow courses. Taken directly from Mote et al. (2008), Figure4c.

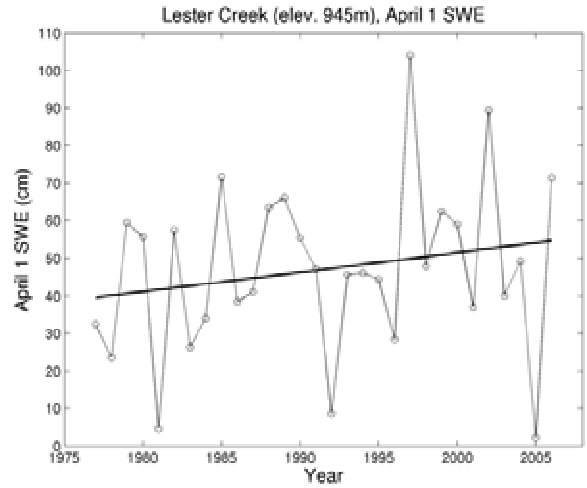

Figure 2. April 1 SWE at Lester Creek (Elevation 945m) for the period 1977-2006.

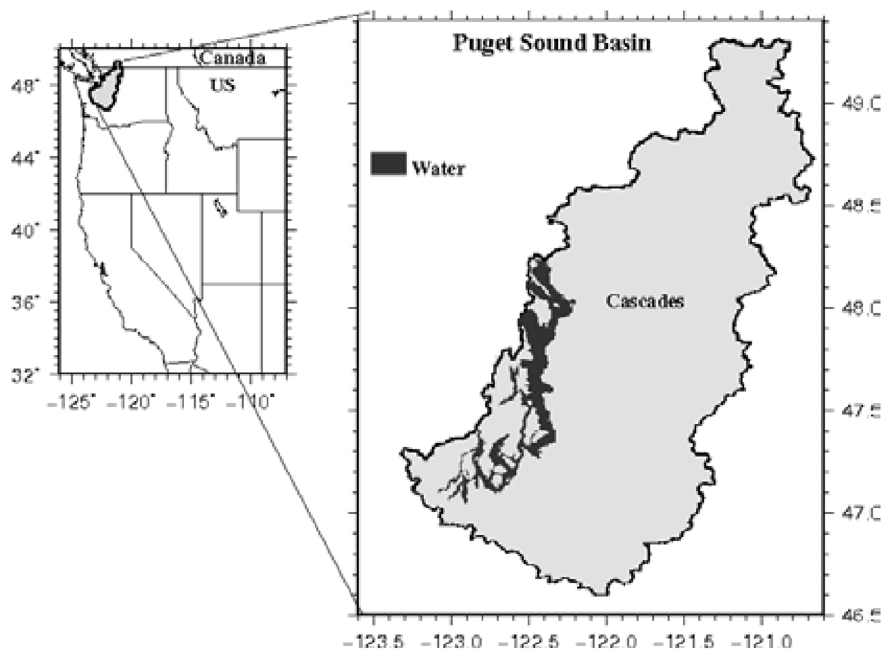

Figure 3 The Cascades portion of the Puget Sound drainage basin.

DEFINING SENSITIVITY

In order to isolate the impact of warming on the snowpack from impacts related to natural variations in precipitation, the sensitivity (λ) is defined as follows:

$$\delta SWE = \lambda(\delta T) \qquad (1)$$

where δSWE represents the change in April 1 SWE following a change in local temperature attributable to global warming, δT. The sensitivity is essentially a constant of proportionality between SWE and temperature. The sensitivity calculated here will be expressed as the percentage loss of mean April 1 SWE in the Cascades for 1°C warming.

ESTIMATING SENSITIVITY

The sensitivity is calculated via four methods: (a) Considerations of the basin geometry, specifically with regard to the effect of warming on the vertical profile of SWE and the distribution of area with elevation; (b) Regression of basin-integrated April 1 SWE values upon observations of mean wintertime temperature. April 1 SWE measurements at various snow courses have been used to estimate the basin-integrated SWE; (c) Comparison of hydrological model output for a control climate and a climate that is 1°C warmer; and, (d) Examination of the distribution of precipitation with respect to daily-mean temperature as recorded at two SNOTEL stations.

Basin Geometry Considerations

Making some simple assumptions about the vertical profile of SWE, the distribution of area with elevation, and the influence of warming on the location of the base of the snowpack, the sensitivity can be calculated.

We can define the basin-integrated April 1 SWE for a narrow band of elevation as being the product of the mean SWE at that elevation and the amount of area in the basin at that elevation. Integrating from the base of the snowpack to the maximum elevation in the basin would be equal to the total basin-integrated April 1 SWE. Equation 2 expresses this relationship, where *S* represents mean SWE as a function of elevation (*z*), and *A* represents the proportion of area in the basin as a function of elevation.

$$\text{Basin-Integrated SWE} = \int_{BASE}^{TOP} S(z)A(z)\,dz \qquad (2)$$

The difference in basin-integrated April 1 SWE between the current climate and a climate that experiences 1°C warming can be calculated using Eqn. 2 and the following assumptions:

- SWE increases linearly with elevation above 600m (e.g., *S(z)* increases linearly with increasing *z*).
- Warming raises base of the snowpack and the entire vertical profile of SWE by 153m. This assumes that the moist adiabatic lapse rate (-6.5°C/km) is a good estimate for the rate of cooling with elevation in the atmosphere. For 1°C warming, 1°C x $(6.5°\text{C/km})^{-1}$ = 153m.
- The shape of the vertical profile of SWE is unchanged following warming (i.e., SWE still increases linearly with elevation; *S(z)* has the same shape following warming).
- The mean base of the April 1 snowpack is located at 600m.

To make the calculation, the function for the distribution of area with elevation (*A(z)*) in the Cascades was taken from the topographic data in the Distributed Hydrology Soil Vegetation Model (DHSVM; Wigmosta et al., 1994, 2002; Figure 4), which is discussed further in the Hydrological Modeling subsection. Figure 5 shows the SWE volume as a function of elevation for the current climate (outer curve) and a climate that experiences 1°C warming (inner curve). These curves indicate that a 1°C warming would lead to a 23% reduction in April 1 SWE in the Cascades.

Two aspects of the calculation should be noted: first, the precise amount of SWE at each elevation is arbitrary (hence there are no units on the *x*-axis in Figure 5.) since the vertical profile of SWE has merely been shifted upward with warming. In other words, the *relative* difference between the basin-integrated SWE is independent of how quickly SWE increases with elevation. Second, altering the assumed location of the base of the snowpack does not change the sensitivity value markedly. Using a base of 400m yields a sensitivity of 20% while a base of 800m yields an estimate of 26%.

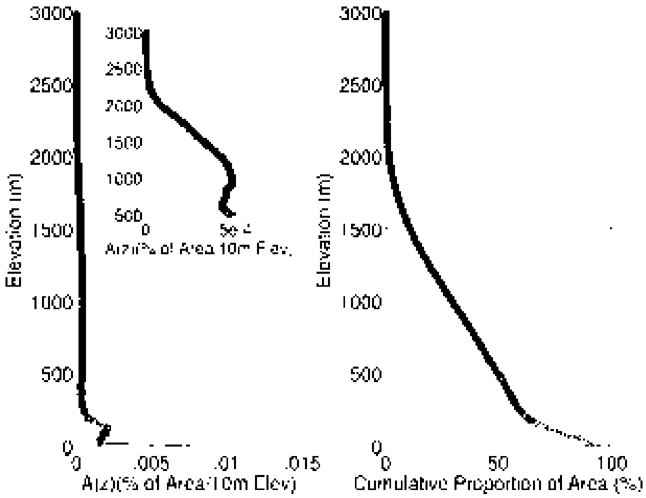

Figure 4. Left panel: *A(z)* for the Cascades with an inset showing an expanded plot for elevations above 500 m. Right panel: the corresponding hypsometric curve.

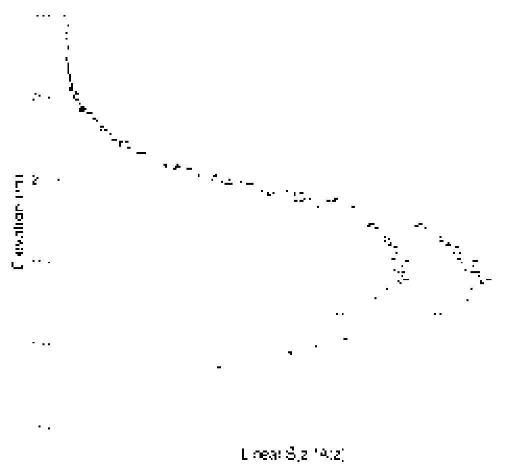

Figure 5. Idealized illustration of SWE loss in the Cascades assuming a linearly increasing profile for *S(z)*. Outer curve corresponds to the original climatology; inner curve corresponds to a 1°C warming and a lifting of the *S(z)* profile by ~ 150m.

<u>Regression between April 1 SWE and Winter Temperature</u>

In this approach, naturally occurring year-to-year variations in temperature in the past record are used as an analog for global warming. We apply this approach to historical measurements of April 1 SWE and mean winter temperature in the Cascades as a means of estimating the temperature sensitivity of the snowpack.

A time series of historical basin-integrated April 1 SWE values was constructed in order to perform the regression of SWE upon temperature. For each year during the 1970-2006 period, the April 1 SWE measured at 24 snow courses in the Cascades was regressed upon the respective elevation of each of the snow courses; the resulting best-fit regression line is analogous to *S(z)* but based on the data for just one year. Then, each year's best-fit line was multiplied by the same *A(z)* function from the DHSVM and integrated with respect to *z*, yielding an estimate for the basin-integrated April 1 SWE for that year.

The regression of the basin-integrated April 1 SWE values upon the wintertime mean temperature as observed in Washington's Climate Division 4, which represents the east slope of the Olympics and the foothills of the Cascades, can be seen in Figure 6; the slope of the best-fit regression line yields a sensitivity of 27% of mean April 1 SWE for 1°C. However, it is clear that the fit of the regression is poor (r^2 value is only 0.28) and subsequently the 95% confidence limits on the sensitivity estimate range from 12% to 42%. Using other adjacent Climate Divisions or averages of nearby Historical Climate Network (HCN) stations yields sensitivity estimates that range from near 0 to over 40% (Table 1.).

The sensitivity estimates from the regression method are generally consistent with the sensitivity calculated using geometric considerations, in the sense that all regression-derived estimates include 20%, regardless of the temperature data set used to derive them. However, the large uncertainty associated with the regression-derived sensitivity estimates undercuts their utility in making a quantitative estimate of the impact of global warming on the Cascades snowpack.

The wide range of the estimates reflects uncertainty arising from several sources. First, the use of seasonal-mean temperature statistics fails to capture the daily covariability between temperature and precipitation that plays an important role in determining snow accumulation. A more precise estimate could be made using a regression between temperatures on precipitating days and observations of daily snow accumulation, if such data were available at snow courses. Second, the paucity of stations at relatively low elevations (around 500 m, near the typical base of the April 1 snowpack) leads to uncertainty in the estimate of the basin-integrated SWE because a significant proportion of the basin area is located at those low elevations. Third, it is unclear which of the existing temperature records best represents conditions within the zone of snow accumulation. Most stations are located at lower elevations to the west or east of the Cascade crest, not in the area of the snowpack itself.

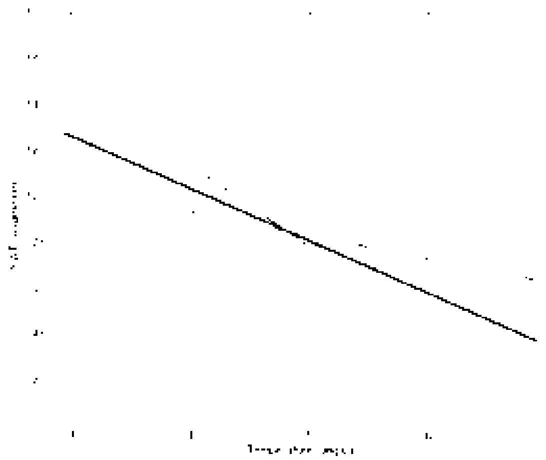

Figure 6. Regression of Cascades basin-integrated April 1 SWE upon mean winter (Nov.-Mar.) temperature for Washington's Climate Division 4, 1970-2006. The slope of the best-fit regression line (thick line) yields the sensitivity of the Cascade snowpack to warming. The thin lines represent the 95% confidence limits associated with the slope estimate.

Temperature Data Used	**Sensitivity (Range)**	**r^2**	**Temperature Variance (° C^2)**
Climate Division 4	27% (12-42%)	0.28	0.68
Climate Division 5	21% (9-33%)	0.28	1.14
Nearby US HCN stations	18% (3-33%)	0.15	0.88
Nearby US HCN stations, West	25% (8-39%)	0.23	0.68
Nearby US HCN stations, East	10% (+2-22%)	0.07	1.42

Table 1. Sensitivity estimates derived from regression of basin-integrated April 1 SWE upon seasonal mean temperature. Climate Division 4 includes the east slope of the Olympic Mountains and the Cascade Foothills; Climate Division 5 includes the west slope of the Cascades. Sixteen Historical Climate Network (HCN) stations that straddle the Cascades have been used for the HCN estimates; 10 are located west of the crest of the Cascades and 6 are located east of the crest.

Hydrological Modeling

Here we input historical temperature and precipitation data into the DHVSM to estimate the April 1 SWE in the Cascades. The temperature sensitivity is estimated by comparing the climatological-mean, basin-integrated April 1 SWE derived from a control run of the model with that derived from a perturbed run in which all the temperatures are raised by 1ºC.

The DHSVM is a spatially distributed hydrology model that represents the water and energy balance of the land surface, and resulting runoff production and streamflow, at spatial resolutions that typically range from 30 to

200 m. DHSVM includes a snow accumulation and ablation model that represents snow either in the presence or absence of forest canopies, and the interaction of the vegetation canopy with the snowpack energy budget (e.g., through differential accumulation and melt processes in, and under, forest canopies). DHSVM represents explicitly the effects of topography on the surface energy balance, most importantly, the role of slope and aspect on incident and reflected solar radiation. Although DHSVM represents a range of surface and subsurface processes related to the production of runoff and streamflow, in this study we utilized only the snowpack model. This model has been tested and evaluated in comparison with observations (see e.g. Nijssen et al., 2003), and is generally able to reproduce observed SWE at sites where high quality forcings (especially precipitation) are available.

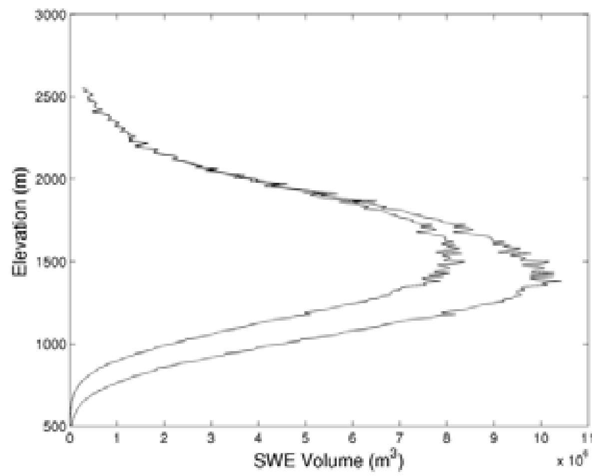

Figure 7. DHSVM estimates of basin-integrates April 1 SWE in the Cascades for the 1970-2000 climatology (control climate; outer curve) and a 1°C warmer climate (inner curve).

DHSVM was run for the Cascades portion of the Puget Sound drainage basin using temperature and precipitation observations for the water years 1916-2002 (Oct. 1915-Sept. 2002). In order to calculate the temperature sensitivity, we focused on the more recent period October 1970 through September 2000. Using this period as the control climatology, a perturbed run was created by increasing the temperature on all days by 1°C. Figure 7 shows April 1 SWE in the Cascades versus elevation in the control run and the perturbed run. In the perturbed run the decrease in SWE in response to the 1°C temperature rise is greatest at elevations ranging from 1000 to 1500 m, where it is roughly equivalent to a 150 m rise of the SWE profile. Averaged over the Cascades, 22% of the April 1 SWE is lost due to a 1°C warming, consistent with the estimate based on geometric considerations

Daily Distribution of Precipitation as a Function of Temperature

By examining the distribution of precipitation with temperature as observed at SNOTEL stations, we can also estimate the temperature sensitivity of the Cascades snowpack. By dividing the winter precipitation into 1°C class intervals based on the daily-mean temperature on which it fell (Figure 8), the amount of snow occurring between -1°C and 0°C can be estimated. The snow falling in this temperature range in the present climate would presumably fall as rain in a 1°C warmer climate. By dividing the precipitation amount in each class interval by the sum of precipitation falling below 0°C, the values for each class interval are expressed as a percentage of total snowfall.

In Figure 8, a Gaussian curve has been fit to the data to eliminate the spike in the distribution near 0°C. Using the values of this curve, the temperature sensitivity for 1°C warming at Corral Pass (elevation 1828m) is about 11%, and at Olallie Meadows (elevation 1128m) it is about 20%. These estimates are in accord with our expectation that sensitivity should gradually decrease with elevation above the base of the snowpack.

Unfortunately, these SNOTEL-derived sensitivity estimates are applicable for one point, and do not necessarily represent the entire basin. It would have been ideal to calculate sensitivities for SNOTEL stations for a variety of elevations throughout the Cascades and compare them to the vertical profiles of SWE loss derived in the previous subsections. However, the paucity of SNOTEL stations below 1000m prevents such a comparison, especially near the base of the snowpack. Given that the elevation of Olallie Meadows lies near the centroid of the snowpack in the Cascades basin, the point-estimate of sensitivity at Olallie Meadows is consistent with estimates based on simple geometric considerations and the DHVSM.

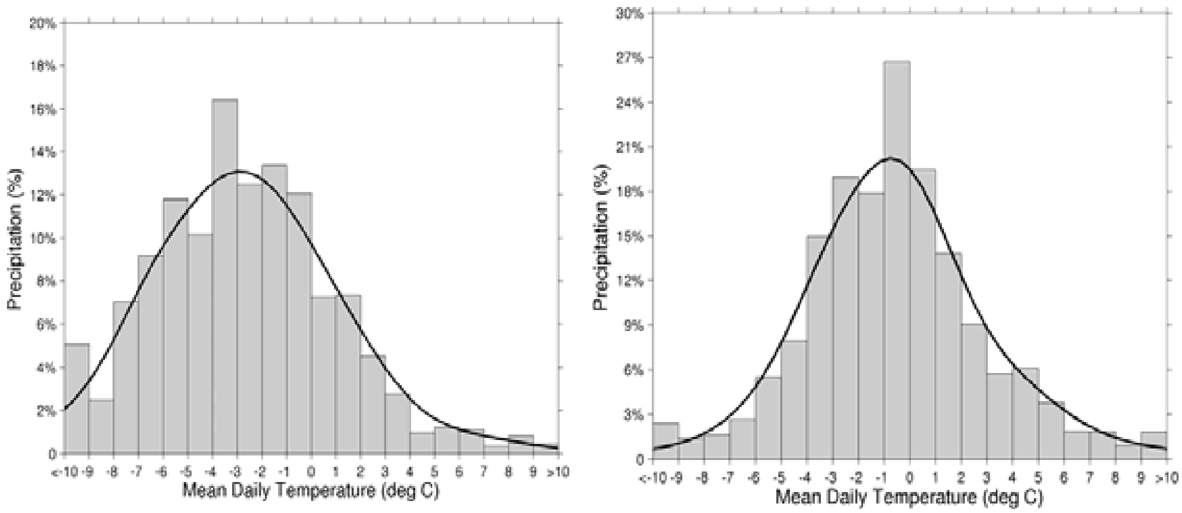

Figure. 8. Contribution of days with daily mean temperatures in various ranges to the total precipitation at Corral Pass (left, elevation 1828m) and Olallie Meadows (right, elevation 1128m), based on November-March data for the water years 1991-2007. The boundaries between class intervals correspond to integral values of the temperature in °C. Precipitation in each bin is divided by the sum of precipitation in all the bins below 0°C. A Gaussian curve (black line) has been fitted to the data.

Temperature-Induced Increases in Precipitation

The Clausius-Clapeyron equation indicates that the increase in the rate of precipitation could be as high as 7% for a 1°C warming, solely from an increase in the specific humidity of the atmosphere (i.e., ignoring any precipitation increases resulting from changes in circulation patterns). However, several studies (e.g., Lorenz and DeWeaver, 2007) argue that while the Clausius-Clapeyron relationship provides a good prediction for the increase in atmospheric water vapor, the increases in precipitation are actually smaller than 7% for 1°C warming. Additionally, since there will be an overall decrease in the days where it is sufficiently cold to accumulate snow at many elevations in the Cascades, only a fraction of the increase in precipitation will be realized as an increase in snowfall. For example, for Olallie Meadows (Figure 8), approximately 20% of the days in today's climate receive snowfall at just between -1°C and 0°C; thus, the incremental change in snowfall would be equal to 80% of the incremental change in overall precipitation in a warmer climate. For purposes of discussion we will assume that the increase in snowfall due to an increase precipitation is 4%, and that the temperature sensitivity, is 20% – 4% = 16%.

ESTIMATING PAST AND FUTURE WARMING

In this section we will use the temperature sensitivity of snowpack (16% for 1°C warming), in conjunction with various estimates of δT to assess the cumulative loss of snowpack in the Cascades attributable to global warming over the past 30 years, and the additional losses that can be expected between now and the year 2050 if the warming continues at the same rate.

A theoretical framework for interpreting the simulation of circulation features on the scale that governs winter temperature and precipitation over regions such as the Cascades does not yet exist. In this study we will not attempt to infer how much winter temperatures over the Cascades have risen in response to global warming. As in the adage, "a rising tide lifts all ships", we simply assume that the contribution of global warming to the rise in winter temperatures over the Cascades is the same as the observed rise in temperature averaged over the Northern Hemisphere as a whole.

Most previous studies examining loss of snowpack have emphasized trends in temperature over land. Since most of the winter precipitation occurs when marine air masses from the North Pacific are swept ashore, it could be argued that changes in sea surface temperature, rather than changes in land temperature, provide a more accurate basis for estimating δT affecting the snowpack. Here we will consider both land and ocean temperatures.

Table 2 shows various estimates of the linear trend in temperature at the Earth's surface over the Northern Hemisphere over the past 30 years (1977 through 2006). The rate of warming has been roughly twice as large over the continents as over the oceans. Distinctions between the trends based on various regional and seasonal breakdowns of the data are less pronounced.

Domain	**Land**	**Ocean**
Northern Hemisphere Annual	0.93	0.52
Northern Hemisphere Winter (NDJFM)	1.04	0.46
45°N to 50°N Winter (NDJFM)	1.28	0.55

Table 2. Linear trends in surface air temperature in °C per 30 years over land and sea surface temperature over various domains in the Northern Hemisphere for the period of record 1977-2006. Land data is based on the CRUTEM 3 data set and ocean data is based on the HadSST2 data set, both from the Climate Research Unit at the University of East Anglia (Brohan et al., 2006).

A reasonable upper bound of δT over the 30-year period is 1°C, a value representative of the zonal-mean land temperature trends (left hand column of Table 2), a lower bound of δT over the 30-year period is 0.5°C, a value representative of the zonal-mean ocean temperature trends (top three estimates in the right hand column of Table 2). Combining these estimates with the sensitivity estimate of 16% for 1°C warming for the Cascades, we estimate that the incremental loss of snowpack that can be attributed to global warming over the past 30 years ranges from 8% to 16%. If land and ocean temperatures continue to rise at the same rate (0.33°C/decade for the land; 0.17°C/decade for the ocean) over the next 40 years, consistent with IPCC projections (2007) and regional climate modeling studies (Salathé et al. 2007), it will result in a further 11-21% decrease by 2050, bringing the cumulative loss since the 1970s up to 19-37%.

ISSUES INVOLVING TREND ANLAYSIS

Historical time series of April 1 SWE in the Cascades are subject to relatively large variability, much of it driven by variations in year-to-year and decade-to-decade precipitation. Hence, the influence of temperature on the Cascade snowpack will tend to be masked by large variations in precipitation, especially for records of only a few decades.

A measure of the detectability of a trend in the presence of background noise can be represented by the student *t*-statistic:

$$t = \frac{\beta N}{\sigma}\sqrt{\frac{N-2}{12}} \tag{3}$$

where β is the slope of the least squares best-fit trend line, N is the number of years in the time series, and σ is the standard deviation of the residual time series after the removal of the trend. The factor βN in equation 3 can be interpreted as the cumulative change in the variable in question attributable to the linear trend over the length of the time series. In essence, it represents the "signal." The denominator (σ) represents the variance that is not explained by the trend; it constitutes the "noise." In this expression, the terms can be considered as percentages relative to a mean value (e.g., the numerator would be the percentage change in April 1 SWE over N years; σ would be the coefficient of variation of the de-trended data). The term ($N - 2$) represents the number of statistical degrees of freedom. It indicates that the longer the time series, the greater the chance of detecting a trend of a given magnitude.

As an example of the noise level inherent in the SWE time series, Table 3 shows means and standard deviations of April 1 SWE for a set of representative stations in western Washington, together with the corresponding time series for the Cascades as derived from the DHSVM. The coefficients of variation of are on the order of 50% for the individual snow course time series and 35% for the basin mean.

Given σ = 35% for the basin, the minimum detectable trend at the 95% confidence level for a one-sided *t*-test for 30 years of data (t~1.7) would be equivalent to a 40% loss of mean of April 1 SWE. Any trend smaller than that would not be distinguishable from the lack of any trend, statistically speaking. For individual station records, the magnitude of the minimum detectable trend would be even greater. In both cases, the trends required for detection are far greater than the 8-16% loss estimated in the analysis in the previous section. Hence, the lack of a downward trend in Cascades snowpack during the past 30 years is not necessarily inconsistent with findings of a statistically significant downward trend from the mid-20th century onward or with the attribution of that downward trend to global warming.

Station/Model	**Station ID**	**Elevation (m)**	**Record**	**Inferred Mean (cm)**	**St. Dev. (cm)**	**Coeff. of Variation (%)**
Freezeout Creek Trail	20A01	1067	1944-2006	38	14	37
Beaver Pass	21A01	1122	1944-2006	97	32	33
Beaver Creek Trail	21A04	671	1944-2006	37	22	59
Thunder Basin	20A07	732	1948-2006	72	21	29
Mt. Gardner	21B21	1006	1959-2006[1]	41	29	71
Cascades DHSVM run			1916-2002	35	12	34

[1]1967 and 2004 are missing.

Table 3. Selected statistics for time series of snow course stations in the Cascades. The Inferred Mean is equal to the SWE value of the fitted trend line at the beginning of the record; the Standard Deviation has been calculated for the residual, detrended time series. The bottom row shows basin-integrated April 1 SWE from the DHSVM.

SUMMARY AND CONCLUSIONS

- The temperature sensitivity (λ) of snowpack in the Cascades, as estimated from all four methods listed in Table 4 are on the order of 20% of mean April 1 SWE for 1°C warming in the absence of indirect effects, and 16% taking the warming-induced increase in precipitation into account.

Estimation Method	**Basin-Integrated Sensitivity (λ) (expressed as loss of April 1 SWE for 1°C warming)**	**Comment**
Geometric Considerations	23%	-
Regression	10-27%	Large uncertainty; 95% confidence limits range from 0-40%
Hydrological Model Output	22%	-
Distribution of Precip vs. Temperature from SNOTEL	~20%	Extrapolated from point estimate for Olallie Meadows

Table 4. Summary of results from the four methods of estimating sensitivity.

- Approaches (*a*), (*c*), and (*d*) for estimating sensitivity emphasize different controls on snowpack: approach (*a*) emphasizes the basin geometry and the vertical profile of SWE, while approaches (*c*) and (*d*) emphasize the mean temperature and range of temperatures observed during winter snowfall events at various elevations within the basin.

- The large uncertainty associated with the sensitivity estimate from regression (*b*), which ranges from near 0 to over 40% of mean April 1 SWE, limits the value of the method.

- IPCC (2007) concludes that most of the observed increase in globally-average temperatures since the mid-20th century is very likely due to the observed increase in anthropogenic greenhouse gas concentration. Most of this warming has occurred during the past 30 years. Using the sensitivity estimate derived in this study, we estimate that in the absence of sampling fluctuations, global warming would have produced an 8-16% decrease in snowpack in the Cascades.

- Sensitivity-based assessments of the impacts of global warming on snowpack can provide useful information for water managers who need to make long range planning decisions between now and the time that the impacts can be confirmed on the basis of direct observations of trends in hydrological variables. In view of the large background variability, assessments of this kind are likely to be most useful if they are expressed in probabilistic terms rather than as decade-by-decade forecasts of the mean snowpack.

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