

Assessing the Impacts of Global Warming on Snowpack in the Washington Cascades*

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ABSTRACT

The decrease in mountain snowpack associated with 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 the temperature sensitivity (λ) of spring snowpack and multiply it by putative past and future temperature rises observed across the Northern Hemisphere.

Estimates of λ can be obtained from (i) simple geometric considerations based on the notion that as the seasonal-mean temperature rises by the amount δT , the freezing level and the entire snowpack should rise by the increment $\delta T/\Gamma$, where Γ is the mean lapse rate; (ii) the regression of 1 April SWE measurements upon mean winter temperatures; (iii) a hydrological model forced by daily temperature and precipitation observations; and (iv) the use of inferred accumulated snowfall derived from daily temperature and precipitation data as a proxy for SWE. All four methods yield an estimated sensitivity of 20% of spring snowpack lost per degree Celsius temperature rise. The increase of precipitation accompanying a 1°C warming can be expected to decrease the sensitivity to 16%.

Considering observations 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 yr resulting from global warming, and it can be expected to decline by another 11%–21% by 2050. These losses would be statistically undetectable from a trend analysis of the region's snowpack over the past 30 yr.

1. 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 the hydrological impacts of global warming over parts of the western United States since the mid-twentieth 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 et al. 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 up to 35% in springtime snow-water equivalent (SWE) at stations in the Pacific Northwest for a variety of periods beginning around the mid-twentieth

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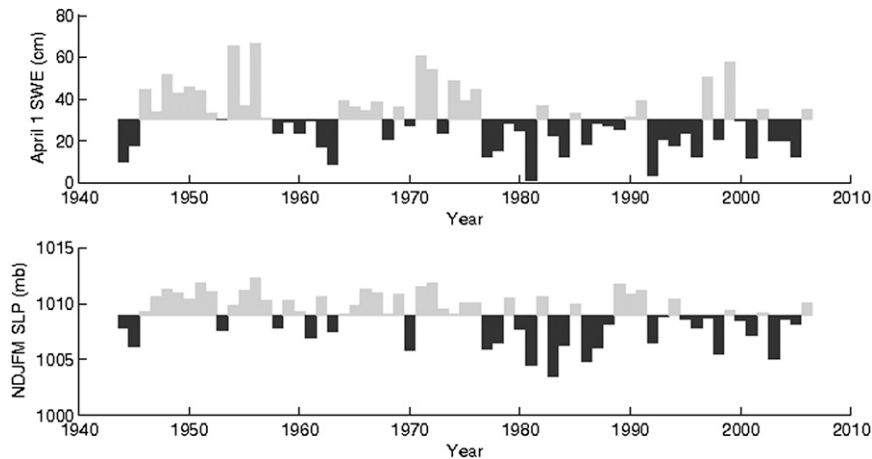


FIG. 1. (top) 1 April SWE at the snow course on the Freezeout Creek Trail. (bottom) The North Pacific Index (NPI), an average of sea level pressure over the area of 30° – 65° N, 160° E– 140° W, as defined in Trenberth and Hurrell (1994) for the winter months of November through March. The NPI is an indicator of the amplitude and polarity of the PDO.

century and ending in 2006, and Stoelinga et al. (2008, manuscript submitted to *J. Climate*), who found little, if any, trend in SWE at stations in the same region from 1977 to 2006.

Mote et al. (2005) noted that part of the decrease in snowpack during the period of record that they examined was attributable to a “regime shift” in the so-called Pacific decadal oscillation (PDO; Mantua et al. 1997), a prominent pattern of natural variability affecting regional temperature and precipitation. Figure 1 shows a time series of wintertime sea level pressure over the Gulf of Alaska, an indicator of the status of the PDO, together with an extended time series of SWE at a snow course on the Freezeout Creek Trail (48.95° N, 120.95° W, station NRCS 20A01, elevation 1067 m) on the western slope of the Cascade Range in western Washington. The extended episode of relatively high snowpack from the late 1940s to the mid-1970s was characterized by above-normal sea level pressure over the Gulf of Alaska, a condition that tends to favor relatively low freezing levels and frequent winter storms in the Pacific Northwest. From 1977 onward, and particularly during the interval from 1977 to 1988, the opposite conditions prevailed and spring snowpack tended to be substantially lower at stations throughout the Pacific Northwest than in the previous decades.

Based on periods of record starting in 1960 and continuing up to the present, approximately 75% of the snow course sites in the West experienced declines in snowpack (Mote 2006), and Knowles et al. (2006) show that about 75% of a set of surface stations across the West experience a decrease in the fraction of precipitation falling as snow. In contrast, a simulation with a

hydrological model forced with observed daily precipitation data, but holding temperature fixed, yields a nearly equal balance of increases and declines in snowpack (Hamlet et al. 2005). Mote (2006) presented further evidence, based on a statistical analysis of snow course records, that warming over the Pacific Northwest has contributed to the loss of snowpack over this extended period of record.

While the above evidence suggests that snowpack over the Pacific Northwest is declining in response to global warming since the midcentury, it should be noted that virtually all of the loss of snowpack over the Pacific Northwest in snow course records and in hydrological simulations from the middle of the twentieth century onward took place before 1977. These time series exhibit little, if any, trend during the past 30 yr, a period of pronounced global warming. This contrast should not be altogether surprising, given the limited geographical extent of the Pacific Northwest. On this relatively small spatial scale, the “signal” related to the impact of global warming on the local snowpack is competing with sizeable “noise” related to the year-to-year and decade-to-decade climate variability that is likely unrelated to global warming. Thus, the historical record of local variations in snowpack, in and of itself, does not provide unequivocal evidence of a causal connection between global warming and the loss of snowpack or a quantitative estimate of the loss of snowpack that is attributable to global warming.

Just as estimates of climate sensitivity (i.e., the change in global-mean surface temperature per unit change in net downward radiation at the top of the atmosphere) provide a range of potential warming caused by the

buildup of greenhouse gases in the atmosphere, a similar sensitivity parameter (λ) can be used to assess the impact of global warming upon a variable x . Let

$$\delta x = \lambda(\delta T), \quad (1)$$

where T is temperature, δT is the local temperature change associated with global warming, and

$$\lambda \equiv \frac{dx}{dT} = \frac{\partial x}{\partial T} + \sum_i \frac{\partial x}{\partial y_i} \frac{dy_i}{dT} \quad (2)$$

is the temperature sensitivity of x . In (2), the summation term represents the changes to the snowpack not directly associated with temperature; y represents an arbitrary variable that is both dependent on T and affects the value of x .

Taking x to be the average 1 April SWE for the region, the terms that constitute the sensitivity in (2) represent the effects of several physical processes that relate temperature to the accumulation and maintenance of the snowpack. The first term ($\partial x/\partial T$) represents the direct effect of warming on the snowpack, which includes the increase in the proportion of precipitation falling as rain instead of snow and any increase in the frequency and/or magnitude of melting events during the season. The summation term includes warming-induced changes in winter precipitation, which could arise from an overall increase in specific humidity or an alteration of the regional circulation. In this study, we will show that the direct effect of warming on the snowpack is primarily due to the shift in precipitation from snow to rain and that, on average, melting plays a less important role in determining the Cascades spring snowpack. In addition, we will provide an estimate of the increase in spring snowpack attributable to the temperature-induced increase in specific humidity, while ignoring the effect of temperature-induced circulation changes. While we acknowledge the possibility that the winter circulation of the North Pacific could be affected by global warming (Gillett et al. 2003; Salathé 2006), we consider the results to date too uncertain to be incorporated into this analysis in a quantitative manner.

The sensitivity-based approach has been used to estimate the influence of temperature and precipitation on the snowpack in the Sierra Nevada mountains (Howat and Tulaczyk 2005), and more generally across the West (Bales et al. 2006), on glacial mass balance (Rasmussen and Conway 2005), and to estimate the potential impact of future warming on agricultural production (Lobell and Asner 2003; Peng et al. 2004). To illustrate how this works, consider how the calendar date of an event such as the spring thaw or the start

of the growing season changes in response to global warming. In this context, the temperature sensitivity is $\lambda = d\tau/dT_c = \partial\tau/\partial T_c$, where T_c is the seasonally varying climatological-mean temperature at the site in question, τ is calendar date, any indirect effects are considered negligible, and the derivative is evaluated at the time of the calendar year when the event occurs. Hence, the temperature sensitivity in this case is simply the inverse of the rate of change of T_c around the time of the event. For example, if T_c rises at a rate of $4^\circ\text{C month}^{-1}$, a 1°C temperature rise would advance the date of occurrence of the event by 1/4 month.

In this paper we will show how the sensitivity-based approach can be used to infer the impact of global warming upon Pacific Northwest snowpack using the Cascades portion of the Puget Sound drainage basin as an example. In the next section we will consider four different approaches to estimating the temperature sensitivity of the snowpack, all of which yield estimates on the order of a 20% decrease per degree Celsius temperature rise in the absence of any increase in precipitation. We will argue that the increase in precipitation resulting from an increase in specific humidity that accompanies a 1°C warming could reduce the sensitivity to about 16%. In section 3 we will consider upper and lower bounds for temperature rise across the Northern Hemisphere over the past 30 yr. While we recognize that it is the local temperature, rather than the hemispheric-mean temperature, that affects the regional snowpack, we will assume that the local, human-induced warming is best reflected in hemispheric or global means, which are less subject to smaller-scale, circulation-related variability. In section 4 we will estimate how large a decrease in snowpack over a 30-yr period would need to occur in order to be detectable in the presence of smaller-scale variability unrelated to global warming. In the final section we will offer some further reflections on the merits of the sensitivity-based approach as applied to snowpack and other hydrological and ecosystem indicators.

2. Estimating the sensitivity of SWE to changes in temperature

In this section we will consider four different ways of estimating the temperature sensitivity of SWE integrated over a prescribed drainage basin to changes in winter temperature under the assumption that precipitation does not change. In particular, the third subsection shows that winter melt events have a relatively small impact on snowpack at the end of the season and, thus, that the changes in SWE can be attributed primarily to a shift in precipitation from snow to rain. In the final subsection we consider the effect upon SWE of

temperature-induced changes in specific humidity that would increase precipitation.

This analysis is specifically applicable to the Cascades portion of the Puget Sound drainage basin¹ shown in Fig. 2, hereafter referred to simply as “the Cascades.” Strictly speaking, it comprises most of the west-facing slope of the Cascades range in Washington State.

a. A simple geometric approach

Expanding on the conceptual approach outlined by Fleagle (1991), the temperature sensitivity of snowpack can be estimated by assuming that as the temperature rises in response to global warming, the freezing level and the entire snowpack will be shifted upward by the increment $\delta z = \delta T/\Gamma$, where Γ is the environmental lapse rate $-\partial T/\partial z$. The volume of water contained in the snowpack, integrated over the total area of the basin, which we define as the snow-water storage (SWS; units of volume, such as m^3), can be expressed as

$$\text{SWS} = \int S dA, \quad (3)$$

where S represents the SWE at a particular location (units of depth, such as cm), and A is the area represented by that location. The SWS can be viewed as consisting of the sum of the contributions from the terrain that lies in various ranges of elevation within the basin, that is,

$$\text{SWS} = \sum \bar{S}_i A_i, \quad (4)$$

where A_i is the area and \bar{S}_i is the average SWE of the i th layer. If the layers are sufficiently thin, the summation can be written as the integral

$$\text{SWS} = \int \bar{S}(z) A(z) dz, \quad (5)$$

where $\bar{S}(z)$ is a function representing the average SWE occurring at elevation z . The differential area that lies between the elevation contours z and $z + dz$ is $A(z)dz$; in hydrology terminology it is the vertical derivative of the so-called “hypsothetic curve” for the drainage basin, in which the fractional area of the basin that lies above a specified elevation is plotted as a function of

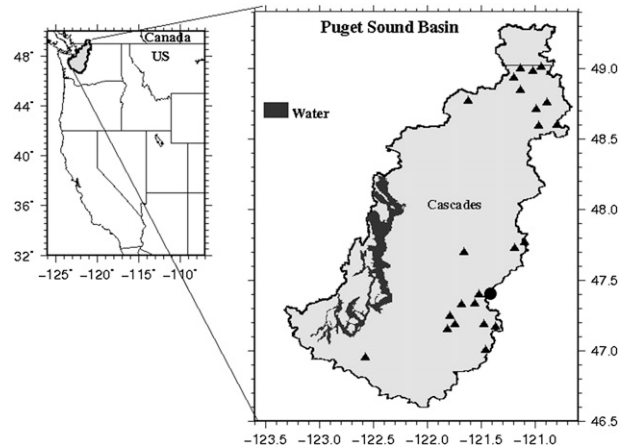


FIG. 2. The Cascades portion of the Puget Sound drainage basin. Locations of snow courses used in the regression analysis (section 2b) are indicated by the triangles. The location of the Olallie Meadows snow course, which was also used in the regression analysis, and SNOTEL station (section 2d) are represented by the circle.

elevation. For example, in the case of conical mountains, the area above a specified elevation contour decreases quadratically with elevation and A decreases linearly with elevation; for a sloping plane surface the area above a specified elevation contour decreases linearly with elevation and A is uniform.

Figure 3 shows the vertical profiles of A for the Cascades, expressed as fractions of the total area of the basin per 10 m of elevation change, together with the hypsothetic curve. The vertical profile of $A(z)$ is derived by differentiating the hypsothetic curve. It is notable that between elevations of 500 and 1200 m, $A(z)$ is relatively uniform.

We can obtain a lower bound on the temperature sensitivity of the snowpack by assuming that $\bar{S}(z)$ is vertically uniform within the snowpack and estimating the fractional area of the snowpack that would be lost if the base of it were to rise by 1°C . The rise in the base of the snowpack for a 1°C warming is the inverse of the lapse rate, here assumed to be $-6.5^\circ\text{C km}^{-1}$, which is close to the moist-adiabatic value. We will assume that the base of the snowpack is at the 600-m level in the Cascades. For this assumed base elevation, the hypsothetic curves indicate that snowpack occupies 46% of the area of the basin. Raising the base elevation by $1^\circ\text{C} \times (6.5^\circ\text{C km}^{-1})^{-1} = 153\text{ m}$ reduces the areal coverage from 46% to 40% of the area of the Cascades, a loss of 12% of the current area of the snow accumulation.

In reality, \bar{S} increases with elevation within the snowpack because of the excursions of the freezing level during the winter season. Hence, a rise in the elevation of the base of the snowpack will result in an additional

¹ The Puget Sound is also fed by snowmelt from the Olympic Mountains, located to the west of the sound. The analysis presented in the following sections was performed for the Olympic basin; however, given the similarity of the results for the Olympics and Cascades and the relatively small size of the Olympics (it is 1/10 the area of the Cascades basin), the results for the Olympics are not shown.

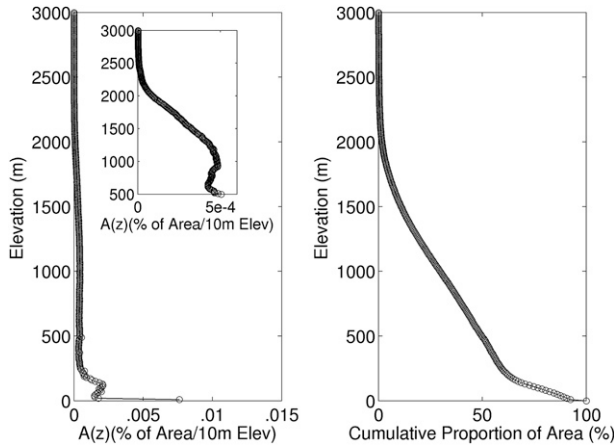


FIG. 3. (left) $A(z)$ for the Cascades with an inset showing an expanded plot for elevations above 500 m. (right) The corresponding hypsometric curve. The $A(z)$ curves have been smoothed with a five-point triangular filter.

loss of SWS by shifting the profile of \bar{S} upward and thereby thinning the snowpack at any given elevation. For example, if \bar{S} increases linearly with elevation [$\bar{S}(z) = mz$; where m is a constant] above the base ($z = z_B$), then

$$\text{SWS} = m \int_{z_B}^{z_T} zA(z)dz, \quad (6)$$

where z_T is the top of the snow accumulation region. If the base rises by the increment Δz , then the loss in SWS can be expressed as

$$\delta\text{SWS} = m \int_{z_B}^{z_B + \Delta z} zA(z)dz + m\Delta z \int_{z_B + \Delta z}^{z_T} A(z)dz, \quad (7)$$

where the first term can be written as $A(z_B)m(\Delta z^2)/2$, because $A(z)$ is nearly constant within the relatively small elevation range of Δz . Thus, the fractional loss of SWS for the case of an \bar{S} profile that increases linearly with elevation is

$$\frac{\delta\text{SWS}}{\text{SWS}} = \frac{\Delta z \left[A(z_B) \frac{\Delta z}{2} + \int_{z_B + \Delta z}^{z_T} A(z)dz \right]}{\int_{z_B}^{z_T} zA(z)dz}. \quad (8)$$

Figure 4 shows the relative estimates of SWS made using (6) for the Cascades for a linearly increasing \bar{S} profile. When the profile of \bar{S} is raised by 153 m, simulating a 1°C warming, the estimated loss of SWS is $23\% \text{ } ^\circ\text{C}^{-1}$ for the Cascades. The percentage loss in SWS is equivalent to the percentage loss of 1 April SWE averaged over the basin, since the area of the basin will be unaffected by warming. Notably, the value of 23% is nearly twice as large as the estimated loss based on

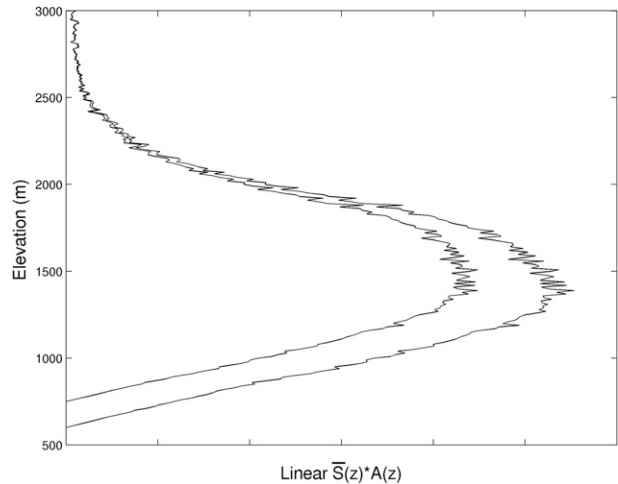


FIG. 4. Idealized illustration of SWE loss in the Cascades, assuming a linearly increasing profile for \bar{S} . Outer curve corresponds to the original climatology; inner curve corresponds to a 1°C warming and a lifting of the \bar{S} profile by ~ 150 m.

the assumption that \bar{S} is independent of height. We will show evidence in the next section that the linear \bar{S} profile is the more realistic one.

The sensitivity estimate (λ) is relatively insensitive to the assumed elevation of the base of the snowpack and lapse rate; using a base of 400 m yields $\lambda \sim 20\% \text{ } ^\circ\text{C}^{-1}$ warming, while a base of 800 m yields $\lambda \sim 26\% \text{ } ^\circ\text{C}^{-1}$ warming. For a lapse rate of $-5.5^\circ\text{C km}^{-1}$, λ is $27\% \text{ } ^\circ\text{C}^{-1}$ warming, and a lapse rate of $-7.5^\circ\text{C km}^{-1}$ yields $\lambda \sim 20\% \text{ } ^\circ\text{C}^{-1}$ warming.

b. Use of seasonal snowpack and temperature data

Regression analysis of seasonal-mean historical data has been used as a basis for estimating the sensitivity of the Sierra Nevada snowpack (Howat and Tulaczyk 2005) and crop yields (Lobell and Asner 2003; Peng et al. 2004) to global warming. In this approach, naturally occurring year-to-year variations in temperature in the past record are used as an analog for global warming. Here we apply this approach to historical measurements of 1 April SWE and mean winter temperature in the Cascades as a means of estimating the temperature sensitivity of the snowpack.

To perform the regression of SWE and temperature, a time series of historical basin-integrated 1 April SWE values was constructed. At each snow course, the snow melts completely in the summer, and thus the values measured in April correspond only to snowfall occurring during the previous winter. For each year during the 1970–2006 period, the 1 April SWE measured at 24 snow courses in the Cascades (see the appendix, Table A1) was regressed upon the respective elevation of each of the snow courses; the resulting best-fit regression line is

analogous to $\bar{S}(z)$, but is based on the data for just 1 yr. Then, each year's best-fit line was multiplied by the $A(z)$ function and integrated with respect to z , yielding an estimate for the volume of water stored by the snow; the volume was then divided by the area of the basin that is greater than 600 m in elevation. The 24 snow courses selected from the 44 that are available in the region are the ones that had measurements for at least 33 of the 37 yr in the 1970–2006 period.

The regression of the basin-integrated 1 April SWE values upon the wintertime mean temperature as observed in Washington's Climate Division 4² can be seen in Fig. 5; the slope of the best-fit regression line yields a sensitivity of 27% of mean 1 April SWE for 1°C. However, it is clear that the fit of the regression is poor (the 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 the nearby Historical Climate Network (HCN; see Karl et al. 1990; see also the appendix, Table A2) stations yields sensitivity estimates that range from near 0% to over 40% (Table 1). Because winter mean precipitation is largely uncorrelated with the winter mean temperature (e.g., for Division 4 for the 1970–2006 period, the correlation coefficient is 0.02), the variability associated with precipitation can be removed from the snowpack time series, and the regression is repeated against the residual snowpack time series. However, using this procedure only narrowed the uncertainty range slightly (i.e., from 12%–42% to 18%–35%), and the central estimate of the sensitivity is virtually unchanged (i.e., from 27% to 26%).

The sensitivity estimates from the regression method are generally consistent with the sensitivity calculated in section 2a, in the sense that all regression-derived estimates include 20%, regardless of the temperature dataset used to derive them. The 20% value is also consistent with the result presented in Mote et al. (2008; Fig. 7, bottom left), where a similar regression analysis was performed between area-weighted observations of SWE and wintertime temperature for the Cascades. 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

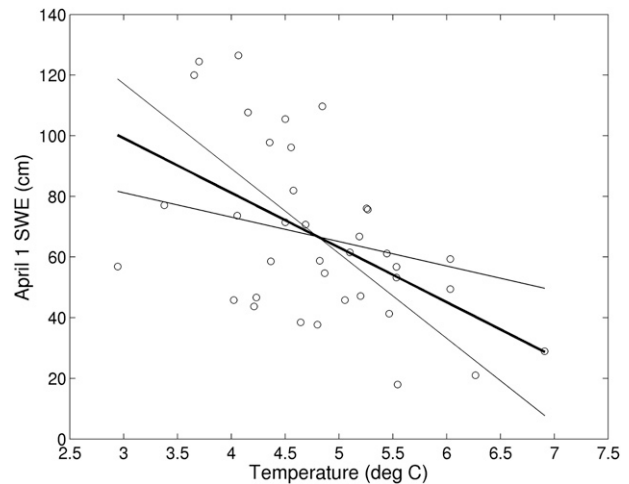


FIG. 5. Regression of Cascades basin-integrated 1 April SWE upon mean winter [November–March (NDJFM)] temperature for Washington's Climate Division 4, 1970–2006. The slope of the best-fit regression line (thick line) yields the sensitivity of the Cascade snowpack to warming. 95% confidence limits associated with the slope estimate (thin line). The basin-integrated 1 Apr SWE has been estimated using SWE measurements from 24 snow courses located in USGS Hydrologic Unit 1711 (see the appendix, Table A1).

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 1 April snowpack) leads to uncertainty in the estimate of the basin-averaged 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.

c. Estimates based on a hydrological model

Here we use a hydrological model forced with historical temperature and precipitation data to estimate the temperature sensitivity of the snowpack in the Cascades. The temperature sensitivity is estimated by comparing the climatological-mean, basin-integrated SWE derived from an extended control run of the model with that derived from a perturbed run forced with the same subdaily "observations" in which all of the temperatures are raised by 1°C.

The Distributed Hydrology Soil Vegetation Model (DHSVM; Wigmosta et al. 1994, 2002) is a spatially

² Climate Division 4 represents the east slope of the Olympics and the foothills of the Cascades.

TABLE 1. Sensitivity estimates derived from regression of basin-integrated 1 April SWE upon seasonal-mean temperature. Basin-integrated SWE has been calculated from snow courses across the western slope of the Cascades; the temperature data used are indicated in each row. 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 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 (see the appendix, Table A2).

Temperature data used	Sensitivity (range) (%)	r^2	Temperature variance ($^{\circ}\text{C}^2$)
Climate Division 4	27 (12–42)	0.28	0.68
Climate Division 5	21 (8–31)	0.28	1.14
Nearby U.S. HCN stations	18 (3–31)	0.15	0.88
Nearby U.S. HCN stations, west	25 (8–39)	0.23	0.68
Nearby U.S. HCN stations, east	10 (+2–22)	0.07	1.42

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 (although both finer and coarser spatial resolutions have been used). 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 explicitly represents the effects of topography on the surface energy balance, most importantly, the role of slope and aspect on incident and reflected solar radiation. While 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, which is essentially identical to the snow model used in the Variable Infiltration Capacity (VIC; Liang et al. 1994) macroscale hydrology model. This model has been tested and evaluated in comparison with observations (see, e.g., Nijssen et al. 2003), and it is generally able to reproduce observed SWE at sites where high-quality forcings (especially precipitation) are available.

Temperature and precipitation forcings were taken from a daily, $1/16^{\circ}$ gridded dataset produced following methods outlined in Maurer et al. (2002) and Hamlet and Lettenmaier (2005), based on National Climatic Data Center (NCDC) Cooperative Observer station data, including U.S. Historical Climate Network station data. The monthly means of the gridded data were adjusted to have the same spatial distribution as the Parameter-Elevation Regression on Independent Slopes Method (PRISM) dataset described by Daly et al.

(1994). The daily temperature and precipitation values were interpolated to a 3-h time step in order to input them into the model. The precipitation data were further adjusted for elevation by interpolating the station data to a $1/16^{\circ}$ grid. This second stage adjustment was tuned to yield a reasonable reproduction of monthly-mean streamflow by the model as observed at long-term U.S. Geological Survey (USGS) stream gauge stations throughout the region.

At some high-elevation grid cells (mostly above 1500-m elevation, and comprising less than 1% of the region) the model did not ablate all of the previous winter's snowpack during the following summer, resulting in long-term accumulation of snow. The pixels where this occurred are in fact in areas where glaciers exist or have existed. DHSVM does not simulate glaciers explicitly, and therefore does not have a mechanism for balancing these accumulations with downslope movement, which occurs in nature. To solve this problem, we removed any snow that remained on 1 August of each year.

DHSVM was run for the Cascades portion of the Puget Sound drainage basin using the $1/16^{\circ}$, 3-hourly dataset for the water years of 1916–2002 (October 1915–September 2002); simulated SWE values were archived for each grid cell for the period of December through June for each year. Comparison between the basin-averaged values of 1 April SWE generated by DHSVM and a variety of snow course records (not shown) indicates the model's ability to reproduce the year-to-year and decade-to-decade variations in SWE for the Puget Sound basin.

To calculate the temperature sensitivity, we focused on the more recent period of 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 6 shows vertical profiles of 1 April SWE in the Cascades in the control run and the perturbed run. It is notable that the profiles exhibit a nearly linear increase of SWE with elevation above the base of the snowpack, consistent with the assumption in section 2a. 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 lifting of the SWE profile. Averaged over the Cascades, 22% of the 1 April SWE is lost due to a 1°C warming, consistent with the estimate based on the linear SWE profile considered in section 2a.

d. Estimates based on inferred accumulated snowfall

In this section we will show that the buildup of snowpack during the winter is more sensitive to the cumulative snowfall in winter storms than to the melting that occurs in between storms, and we will exploit this

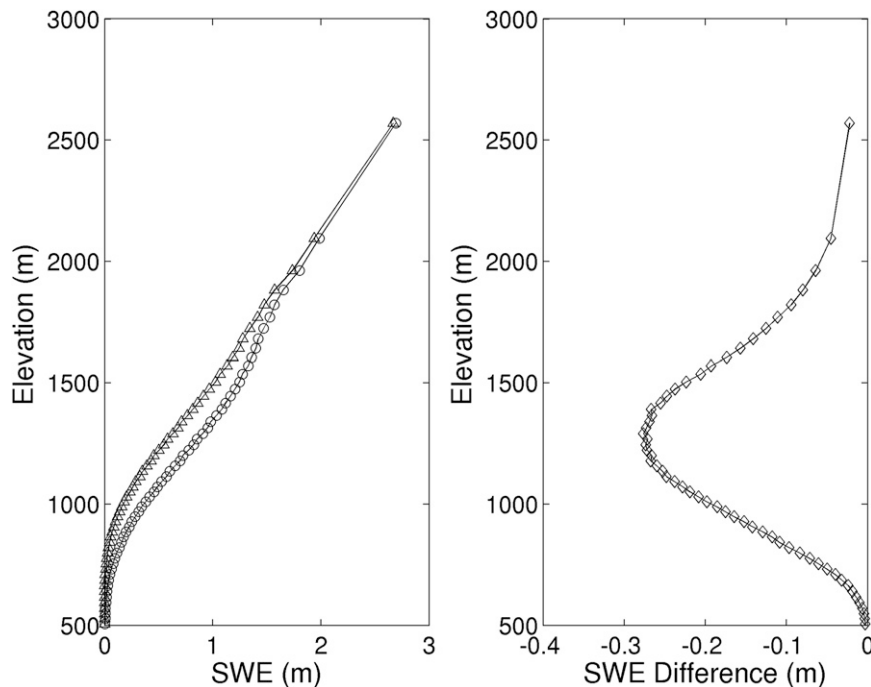


FIG. 6. (left) 1 April SWE averaged over the Cascades as simulated with the DHSVM snow accumulation model plotted as a function of elevation: control run (circles) and a perturbed run (triangles) in which the temperature for all the input data is raised by 1°C . The model was run for the water years 1971–2000 (October 1970–September 2000). (right) The difference between the control and perturbed runs.

finding to estimate the temperature sensitivity of the 1 April snowpack using the accumulated snowfall inferred from daily temperature and precipitation measurements at individual stations.

Figure 7 compares daily observations of SWE at the Olallie Meadows (1128 m) Snowpack Telemetry (SNOTEL) station in the Cascades with the inferred accumulated snowfall (IAS) and rainfall based on collocated observations of daily precipitation and mean temperature. This station was chosen to be representative of the Cascades in general because it lies near the centroid of the snowpack of the Cascades (Mote et al. 2008).

When the daily-mean temperature is above 0°C , all precipitation is considered to be rain and the inferred accumulated snowfall does not increase; when the daily-mean temperature is equal to or below 0°C , all precipitation is considered to be snow and the IAS increases by the amount of the precipitation on that day. As shown in Fig. 7, the buildup of SWE over the winter seasons tracks the IAS remarkably well, especially considering that the daily station data upon which it is based do not fully resolve the large variations in the freezing level and the lapse rate observed in association with the passage of winter storms. For most stations located above 1000 m, the buildup of SWE tracks the

IAS well; exceptions occur at the few relatively low-lying SNOTEL stations where melting is more frequent, causing the IAS to overestimate the SWE. The tendency for melting to be more important at the lower elevations was noted by Mote et al. (2005) in their analysis of the correlation between melt events and 1 April SWE at SNOTEL sites.

Figure 8 shows the winter precipitation divided into 1°C class intervals based on the daily-mean temperature on which it fell. The value in each class interval is divided by the sum of precipitation in all of the class intervals below 0°C . A Gaussian curve has been fit to the data to eliminate the spike in the distribution near 0°C .³

³ A spike in temperature near 0°C is common to the precipitation–temperature histograms of many of the SNOTEL stations, especially those at lower elevations. This feature may be a consequence of the formation of a nearly isothermal (0°C) freezing layer during episodes of heavy precipitation and the cooling effect of melting snow upon air temperatures just above the ground during rain events. These spikes are anchored to 0°C , and would not be expected to shift toward higher temperatures in response to global warming. Hence, in estimating the temperature sensitivity of IAS, we use a smoothed histogram in which the excess frequency of occurrence represented by the spike is redistributed among the neighboring bins.

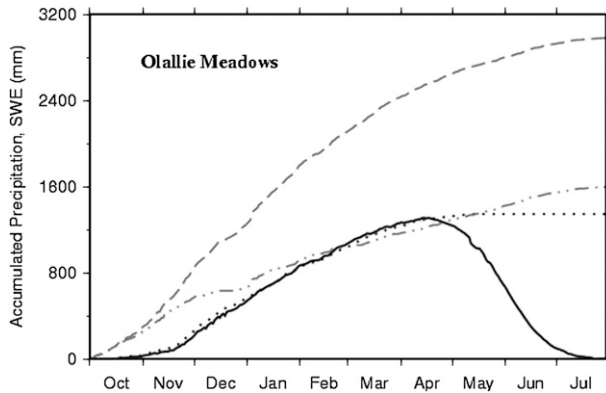


FIG. 7. Observed snow-water equivalent (black solid line) at Olallie Meadows (station NRCS 21B55S; 1128-m elevation) vs estimates of accumulated snowfall (dotted line) and rainfall (dot-dash line) as inferred from daily temperature and precipitation data, averaged for water years 1991–2007 (October 1990–September 2007). The gray dashed line represents accumulated precipitation. The precipitation that falls during each day is classified as snow if the daily-mean temperature is at or below 0°C .

The temperature sensitivity of the IAS, expressed in percentage loss for 1°C temperature rise, is simply the value of the smoothed curve that spans the range from -1° to 0°C . The sensitivity inferred from Fig. 8 is just below $24\% \text{ }^{\circ}\text{C}^{-1}$ warming at Olallie Meadows, which is consistent with the estimates based on the simple geometric considerations (section 2a) and the DHSVM (section 2c). In agreement with our expectation that sensitivity should decrease with elevation, stations located above (below) Olallie Meadows exhibit lower (higher) sensitivities (not shown).

Some caveats with respect to the SNOTEL analysis deserve mention. First, the chosen threshold value (0°C) for partitioning precipitation between rain and snow is lower than used in previous studies, which range up to either 1° or 2°C (Rasmussen and Conway 2005, and references therein). A threshold temperature of 0°C was used in the present study because it yielded a better fit between IAS and SWE in time series such as those shown in Fig. 7. Repeating the analysis with 1°C as the threshold reduced the sensitivity by a few percent.

Second, Serreze et al. (1999) and Pepin et al. (2005) have noted the tendency for SNOTEL temperature sensors to provide spurious data (e.g., missing values, the same temperature can be registered on consecutive days during a malfunction, unrealistically anomalous temperatures). Although we did not apply rigorous quality control procedures to the data, the binning analysis did not reveal any extraordinarily high or low temperatures. The errors associated with days when temperature values were not recorded is small: only 1% of the days in our period of record (1991–2007) and

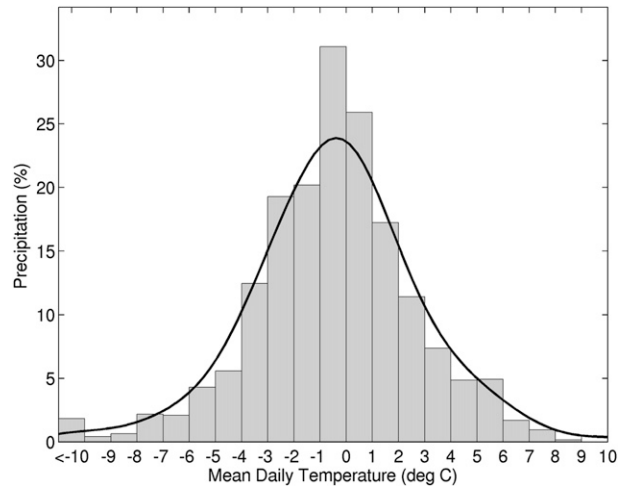


FIG. 8. Contribution of days with daily-mean temperatures in various ranges to the total precipitation at Olallie Meadows, based on November–March data for water years 1991–2007. The boundaries between class intervals correspond to integral values of the temperature ($^{\circ}\text{C}$). Precipitation 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.

about 1% of the total accumulated snowfall were associated with missing temperature values.

e. Temperature-induced increase in precipitation

In this section, we consider the increase in precipitation in the Cascades resulting from a warming-induced increase in the specific humidity. While we acknowledge that precipitation could also be affected by changes in the regional circulation, such as a shift in the location or intensity of the jet stream or the storm tracks in the North Pacific, arriving at quantitative estimates of their effect on the regional snowfall is beyond the scope of this study.

On a global basis, climate models predict that precipitation will increase at a rate of about 1%–2% for 1°C of warming (Vecchi and Soden 2007; Held and Soden 2006), as determined by the time scale of radiative cooling in the downward branch of deep overturning circulations. In the extratropics, models indicate that wintertime precipitation would be more sensitive to global warming ($\sim 3\%$ – 5% for 1°C of warming) if changes in the strength of the zonal-mean circulation and the associated increases in moisture transport are taken into account (Lorenz and DeWeaver 2007). For locations where orographic forcing (see Roe 2005) plays a dominant role in precipitation, it is conceivable that the rate of precipitation increase could be as high as $7\% \text{ }^{\circ}\text{C}^{-1}$ of warming, the limiting value set by the Clausius–Clapeyron equation.

A large proportion of the area of the Cascades is likely similar to Olallie Meadows (Fig. 8), and receives

much of its snow at a temperature close to 0°C. Hence, the increase in snowfall resulting from a 1°C temperature rise is given by

$$\frac{\partial P}{\partial T} \times \frac{\int_{-\infty}^{-1^{\circ}\text{C}} p dT}{\int_{-\infty}^{0^{\circ}\text{C}} p dT},$$

where P represents precipitation, T is temperature, and p is the probability density function of precipitation occurring for a particular range of temperatures. Thus, $\partial P/\partial T$ represents the increase in overall precipitation from warming, while the second term is a correction factor between 0 and 1, indicating what fraction of the increase in precipitation will be in the form of snow. In the smoothed distribution of precipitation amount versus temperature at Olallie Meadows, the correction factor is equal to 0.76.

It follows that the reduction in the sensitivity resulting from increases in precipitation caused by an increase in the specific humidity is unlikely to be more than 5% °C⁻¹ warming and could be substantially less than that. For purposes of discussion in the subsequent sections of this paper, we can conservatively use the value of 20% of 1 April SWE lost per degree Celsius temperature rise as the direct effect of warming. Combining this with an assumed 4% increase in SWE for the indirect effect of increasing precipitation, the total sensitivity is 20%–4% = 16% °C⁻¹ warming.

3. Estimating the local temperature rise associated with global warming

In this section we will use the temperature sensitivity of snowpack, as estimated in section 2, in conjunction with various estimates of δT to assess the cumulative loss of snowpack in the Cascades associated with global warming over the past 30 yr, and the additional losses that can be expected between now and the year 2050 if the warming continues at the same rate. Although many previous studies (Mote et al. 2005, 2008; Mote 2006; Knowles et al. 2006) document the changes in snowpack over the period of relatively abundant data from around 1950 onward, we focus on the shorter and more recent period when there is an apparent contradiction between the relatively rapid rate of global warming and the absence of a downward trend in the snowpack.

The science of modeling the regional impacts of global warming is still in its infancy. A rudimentary set of principles is just beginning to emerge for interpreting how the atmospheric general circulation should evolve in response to global warming (e.g., see Held and Soden 2006, and references therein), but it has thus far been

mainly concerned with planetary-scale features, such as the width of the tropics, the strength of the climatological-mean stationary waves, and the latitude of the storm tracks. A theoretical framework for interpreting the simulation of 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 (as opposed to other regions) 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. Presumably, most of this warming has been caused by the increase in the atmospheric concentration of greenhouse gases (Solomon et al. 2007); however, we will not quantify the portion attributable to natural or anthropogenic sources.

Most previous studies examining losses of snowpack have emphasized trends in temperature over land. Because it could be argued that most of the winter precipitation over the Cascades takes place when marine air masses from the North Pacific are swept ashore, changes in temperature averaged through the depth of the boundary layer are more relevant to the snowpack than changes in local surface air temperature. Because the properties of the boundary layer are influenced by surface properties extending far upstream of the Cascades, it could thus be argued that changes in SST, 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 yr (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.

A reasonable upper bound of δT over the 30-yr period is 1°C, a value that is representative of the zonal-mean land temperature trends (left-hand column of Table 2), a lower bound of δT over the 30-yr 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 $\lambda \sim 16\%$ for 1°C warming for the Cascades, as calculated in section 2, we estimate that the incremental loss of snowpack that is associated with global warming over the past 30 yr ranges from 8% to 16%. If land and ocean temperatures continue to rise

TABLE 2. Linear trends in surface air temperature ($^{\circ}\text{C}$ per 30 yr) over land and sea surface temperature over various domains in the Northern Hemisphere for the period of record 1977–2006. “North Pacific” denotes the average over the box (32.5° – 57.5°N , 142.5°E – 122.5°W). Land data based on the Hadley Climate Research Unit Global Temperature (CRUTEM) 3 dataset and ocean data based on the Hadley Centre SST (HadSST) 2 dataset from the Climate Research Unit at the University of East Anglia (Brohan et al. 2006).

Domain	Land	Ocean
Northern Hemisphere annual	0.93	0.52
Northern Hemisphere winter (NDJFM)	1.04	0.46
45° – 50°N winter (NDJFM)	1.28	0.55
North Pacific winter		0.38

at the same rate (0.33°C decade $^{-1}$ for land and 0.17°C decade $^{-1}$ for ocean) over the next 40 yr, consistent with Intergovernmental Panel on Climate Change (IPCC) projections (Solomon et al. 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%.

If mean temperature of the North Pacific domain in Table 2 was used in place of hemispheric or zonal averages as the lower bound of δT , the temperature rise would have been 0.38°C rather than 0.5°C . Evidently there has been some degree of cancellation between the hemispheric-scale warming over the past 30 yr and the dynamically induced temperature trend over the North Pacific. If this cancellation was understood to be an integral part of the spatial signature of global warming, there would be a basis for expecting it to continue to mitigate the impacts of global warming on the Cascades snowpack. Lacking dynamical support for such an interpretation, we are inclined to regard the smallness of the temperature rise over the North Pacific over the past 30 yr as sampling or regional variability that has no predictive implications.

4. Further consideration of statistical issues

Long-term trends in records of snowpack tend to be obscured by the presence of natural, regional variability, especially related to precipitation. A measure of the detectability of a trend⁴ in the presence of background noise is the Student’s t statistic

$$t = \frac{r}{\sqrt{1-r^2}} \sqrt{N-2}, \quad (9)$$

⁴ Lettenmaier (1976) provides a more in depth and theoretical framework for estimating the probability of detecting a trend of a prescribed magnitude (relative to the background noise) at a prescribed confidence level in a record of a prescribed length.

where r is the correlation coefficient between the time series and the least squares best-fit trend line. For noisy SWE data, the linear trend accounts for only a small fraction of the temporal variance [$(1-r^2) \sim 1$]. Making this simplification in (9) and substituting for r in the numerator from the least squares regression formula and the product moment formula for the correlation coefficient yields

$$t = \frac{\beta N}{\sigma} \sqrt{\frac{N-2}{12}}, \quad (10)$$

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 time series.

The factor $\beta N/\sigma$ in (10) can be interpreted as the cumulative change in the (standardized) variable in question attributable to the linear trend over the length of the time series, expressed in dimensionless units. Because the Student’s t statistic in this case is being computed for a quantity divided by its own mean, it follows that σ is the coefficient of variation (i.e., the standard deviation divided by the mean).

The term $(N-2)$ in (10) represents the number of statistical degrees of freedom. Because 1-yr lag correlations in the SWE time series tend to be on the order of 0.1 or less, the number of degrees of freedom is assumed to be equal to the number of years in the time series being analyzed. If the autocorrelation were significant, the number of statistical degrees of freedom would be less than $N-2$ and the minimum detectable trend would be even higher than indicated in this analysis.

As an indication of the noise level inherent in the SWE time series, Table 3 shows means and standard deviations of 1 April 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 are on the order of 50% for the individual snow course time series and 35% for the basin mean.

For purposes of illustration, let us calculate the percentage decline in snowpack (expressed as a linear trend) over a 30-yr interval that would just meet the requirement for statistical significance. Because we are considering only decreases in snowpack, it is appropriate to use the one-sided test, for which the 95% confidence level is $t \sim 1.7$. Substituting these values into (10) yields a reduction of the basin-mean snowpack of at least 40% that would be required for detection at the 95% confidence level over the 30-yr period. 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. For a

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 bottom row shows basin-integrated SWE from the DHSVM.

Station/model	Station ID	Elevation (m)	Record	Inferred mean (cm)	Std dev (cm)	Coef of variation (%)
Freezeout Creek Trail	20A01	1067	1944–2006	38	15	39
Beaver Pass	21A01	1122	1944–2006	97	34	35
Beaver Creek Trail	21A04	671	1944–2006	37	22	59
Thunder Basin	20A07	732	1948–2006	72	22	31
Mt. Gardner	21B21	1006	1959–2006*	41	29	70
Cascades DHSVM run			1916–2002	35	12	35

* 1967 and 2004 are missing.

50–60-yr-long time series the minimum detectable trend drops to around a 30% loss in snowpack for the basin mean, which is within the range reported in the studies of Mote et al. (2005, 2008). Certainly, the mere presence of a statistically significant trend in the snowpack time series is not, in and of itself, evidence of global warming. Conversely, as shown by this analysis, the *lack* of a significant downward trend in snowpack over the Cascades during the past 30 yr is not necessarily inconsistent with findings of a statistically significant downward trend from the mid-twentieth century onward or with the linking of that downward trend to global warming.

The same background noise that tends to obscure the trends in historical snowpack data makes it difficult to predict the evolution of snowpack over a limited region, such as the Cascades. Even if the projections of global warming and the above estimate of the temperature sensitivity of snowpack are both proven to be accurate, the impacts of global warming will continue to be masked at some times and exacerbated at other times by inherently unpredictable year-to-year and decade-to-decade variations in snowpack. In the face of these large uncertainties in forecasting the snowpack for a given number of years into the future, assessments of the risks of extreme events (e.g., an annual snowpack insufficient to meet demands for water) are likely to be more useful to planners and policymakers than actual year-by-year forecasts.

5. Concluding remarks

We conclude with some comments on the use of temperature sensitivity as a basis for inferring the impacts of global warming.

- The temperature sensitivity λ of snowpack in the Cascades, as estimated from (a) simple geometric considerations, (b) regression of 1 April SWE measurements upon seasonal-mean temperature, (c) a hydrological model forced with historical daily temperature and precipitation data, and (d) a simple analysis of inferred accumulated snowfall would be on the order of 20% of mean 1 April SWE per °C warming in the absence of

indirect effects, and 16%, taking the warming-induced increase in precipitation into account.

- That approaches (a) and (c) to estimating the temperature sensitivity of SWE yield mutually consistent results is understandable, given that the vertical profile of basin-integrated SWE [i.e., $\bar{S}(z)A(z)$] based on the assumed linear profile for \bar{S} closely matches the shape of the corresponding profile generated by the model (Fig. 9).
- Approaches (a) and (d) emphasize different controls on snowpack, respectively, the basin geometry and the vertical profile of SWE, and 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 1 April SWE, limits the value of the method. Removing the variability associated with precipitation from the snowpack record prior to the regression upon temperature does not significantly narrow the ranges of the estimates.
- The 1 April SWE is closely approximated by the inferred accumulated snowfall, except near the base of the snowpack. It follows that sensitivity estimates based on method d using stations like Olallie Meadows, at elevations near the centroid of the snowpack, capture the essential physics of the hydrological model used in (c).
- Using sensitivity estimates derived in this study, we estimate that in the absence of natural, internal climate variability or a regional circulation response to global anthropogenic forcing, global warming would have produced an 8%–16% decrease in the snowpack in the Cascades over the last 30 yr.
- 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

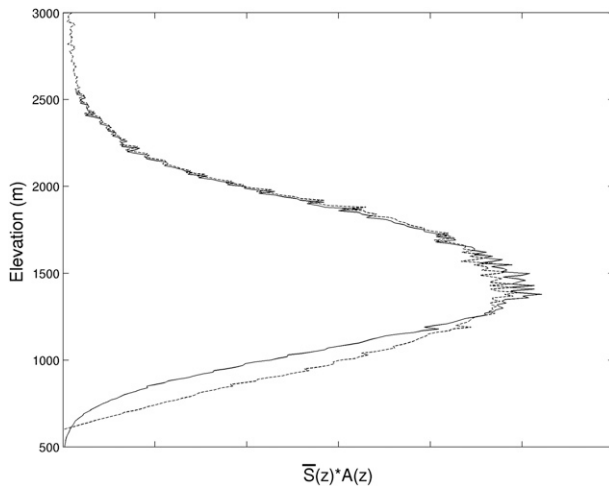


FIG. 9. Comparison of estimated SWE vs elevation using the output of a hydrological model (solid line) and an idealized, linearly increasing $\bar{S}(z)$ for the Cascades (dashed line).

expressed in probabilistic terms rather than as decade-by-decade forecasts of the mean snowpack.

- Simple geometric arguments analogous to those described in section 2a can also be applied to assessing the deterioration of the health of ecosystems that are adapted to the uppermost slopes of mountain ranges

as in the so-called sky islands in the Sonora Desert, where summer temperatures are low enough to permit forest ecosystems to survive. In this case, \bar{S} could be an indicator of the health of the ecosystem, such as the abundance of indicator species, and estimates of the functional form of $\bar{S}(z)$ could be derived from temperature-dependent ecosystem models.

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APPENDIX

Locations and Elevations of Cascade Snow Courses and Nearby HCN Stations

Table A1 lists the 24 snow courses used to estimate basin-integrated 1 April SWE for 1970–2006 as described

TABLE A1. Snow courses used to estimate the basin-integrated SWE for the Cascades, 1970–2006. These snow courses are located in USGS Hydrologic Unit 1711, which drains into the Puget Sound. The snow courses have 1 Apr SWE measurements for at least 33 of the 37 yr in that period.

Snow course	Natural Resources Conservation Service ID	Elevation (m)	Latitude (°)	Longitude (°)
Alpine Meadow	21B48	1067	47.77	121.70
Beaver Creek Trail	21A04	671	48.82	121.20
Beaver Pass	21A01	1122	48.87	121.25
Brown Top Ridge AM	21A28	1829	48.92	121.20
Cayuse Pass	21C06	1615	46.87	121.52
City Cabin	21B03	728	47.32	121.52
Corral Pass	21B13	1829	47.02	121.47
Cougar Mountain	21B42	975	47.30	121.67
Devils Park	20A04	1798	48.75	120.85
Freezeout Creek Trail	20A01	1067	48.95	121.95
Granite Creek	20A06	1067	48.60	120.80
Grass Mountain No. 2	21B27	884	47.22	121.75
Lester Creek	21B29	945	47.17	121.47
Lynn Lake	21B50	1219	47.20	121.78
Marten Lake	21A09	1097	48.77	121.72
Meadow Cabins	20A08	579	48.57	120.93
Mt. Gardner	21B21	1006	47.37	121.57
New Lake Hozomeen	21A30	853	48.95	121.03
Olallie Meadows	21B02	1105	47.37	121.45
Sawmill Ridge	21B31	1433	47.17	121.43
Stevens Pass Pillow	21B01	1241	47.70	121.08
Stevens Pass Sand Shed	21B45	1128	47.75	121.03
Thunder Basin	20A07	732	48.52	120.98
Twin Camp	21B30	1250	47.13	121.78

TABLE A2. Historical Climate Network stations used to estimate wintertime temperature in the Cascades.

HCN station	Cooperative network index	Elevation (m)	Latitude (°)	Longitude (°)
West				
Bellingham	0587	5	48.72	122.52
Blaine	0729	18	49.00	122.75
Buckley	0945	209	47.17	122.00
Cedar Lake	1233	475	47.42	121.73
Clearbrook	1484	20	48.97	122.33
Everett	2675	18	47.98	122.18
Puyallup	6803	15	47.20	122.33
Seattle	7458	6	47.65	122.30
Sedro-Wooley	7507	18	48.50	122.23
Snoqualmie Falls	7773	134	47.55	121.85
East				
Cle Elum	1504	585	47.18	120.95
Conconully	1666	707	48.55	119.75
Stehekin	8059	387	48.35	120.72
Waterville	9012	799	47.65	120.07
Wenatchee	9074	195	47.42	120.32
Winthrop	9376	535	48.47	120.18

in section 2b. Data for these snow courses can be accessed online (ftp://ftp.wcc.nrcs.usda.gov/data/snow/snow_course/wasnow.txt).

Table A2 lists the 16 HCN stations used to estimate wintertime temperatures in the Cascades. Data for these stations also can be accessed online (http://cdiac.ornl.gov/epubs/ndp/ushcn/state_WA_mon.html).

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