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Key Points:
- Extreme warming observed in western U.S. mountains is largely artificial
- Systematic observation bias has overstated higher-elevation warming
- Widely used gridded climate products propagate artificial warming signal

Supporting Information:
- Texts S1–S4 and Figures S1–S7
- Table S1

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Artificial amplification of warming trends across the mountains of the western United States

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Abstract Observations from the main mountain climate station network in the western United States (U.S.) suggest that higher elevations are warming faster than lower elevations. This has led to the assumption that elevation-dependent warming is prevalent throughout the region with impacts to water resources and ecosystem services. Here we critically evaluate this network's temperature observations and show that extreme warming observed at higher elevations is the result of systematic artifacts and not climatic conditions. With artifacts removed, the network's 1991–2012 minimum temperature trend decreases from +1.16°C decade^{-1} to +0.106°C decade^{-1} and is statistically indistinguishable from lower elevation trends. Moreover, longer-term widely used gridded climate products propagate the spurious temperature trend, thereby amplifying 1981–2012 western U.S. elevation-dependent warming by +217 to +562%. In the context of a warming climate, this artificial amplification of mountain climate trends has likely compromised our ability to accurately attribute climate change impacts across the mountainous western U.S.

1. Introduction

Temperatures across Earth's surface are warming globally [*Intergovernmental Panel on Climate Change*, 2013], but certain land areas are warming faster than others [*Meehl et al.*, 2012]. Empirical studies suggest that temperatures at higher elevations are warming faster than at lower elevations [*Diaz and Bradley*, 1997; *Diaz and Eischeid*, 2007; *Wang et al.*, 2013] leading to concern over impacts to important water resources and ecological services provided by mountainous regions [*Beniston*, 2003]. Multiple mechanisms have been postulated to explain elevation-dependent warming including snow-albedo feedbacks, changes in cloud cover, consistent atmospheric decoupling in valley locations, and greater high-elevation sensitivity to water vapor radiative influences [*Rangwala and Miller*, 2012]. Nonetheless, evidence supporting elevation-dependent warming varies by region and is often scarce or equivocal given limited observations at high elevations and the complexities of mountain climate [*Pepin and Lundquist*, 2008; *Rangwala and Miller*, 2012].

Within the mountainous western U.S., elevation-dependent warming is often assumed to be occurring [*Diaz and Eischeid*, 2007; *Clow*, 2010; *Pederson et al.*, 2011] with associated critical impacts to snowpack [*Clow*, 2010; *Pederson et al.*, 2011], forests [*van Mantgem et al.*, 2009; *Williams et al.*, 2010], and alpine tundra [*Diaz and Eischeid*, 2007]. Observations from the high-elevation Snowpack Telemetry (SNOTEL) network and related gridded climate products [*Thornton et al.*, 1997; *Daly et al.*, 2008] have provided much of the evidence for western U.S. elevation-dependent warming [*Diaz and Eischeid*, 2007; *Clow*, 2010; *Pederson et al.*, 2011]. Despite a critical reliance on SNOTEL data to derive popular gridded climate products [*Thornton et al.*, 1997; *Daly et al.*, 2008], assess elevation-dependent warming [*Diaz and Eischeid*, 2007], and attribute related impacts [*van Mantgem et al.*, 2009; *Williams et al.*, 2010; *Pederson et al.*, 2011, 2013], the SNOTEL network has not been evaluated for inhomogeneities—nonclimatic temperature jumps and trends resulting from changes in observation protocols, instrumentation, or station siting. Because inhomogeneities have the potential to significantly bias climate trend analyses and obscure true climate signals, it is necessary to subject SNOTEL observations to the same critical evaluation as other meteorological networks [*Menne et al.*, 2009]. The objective of this study is to evaluate the homogeneity of the SNOTEL temperature record and assess if an elevation-dependent warming signal is evident.

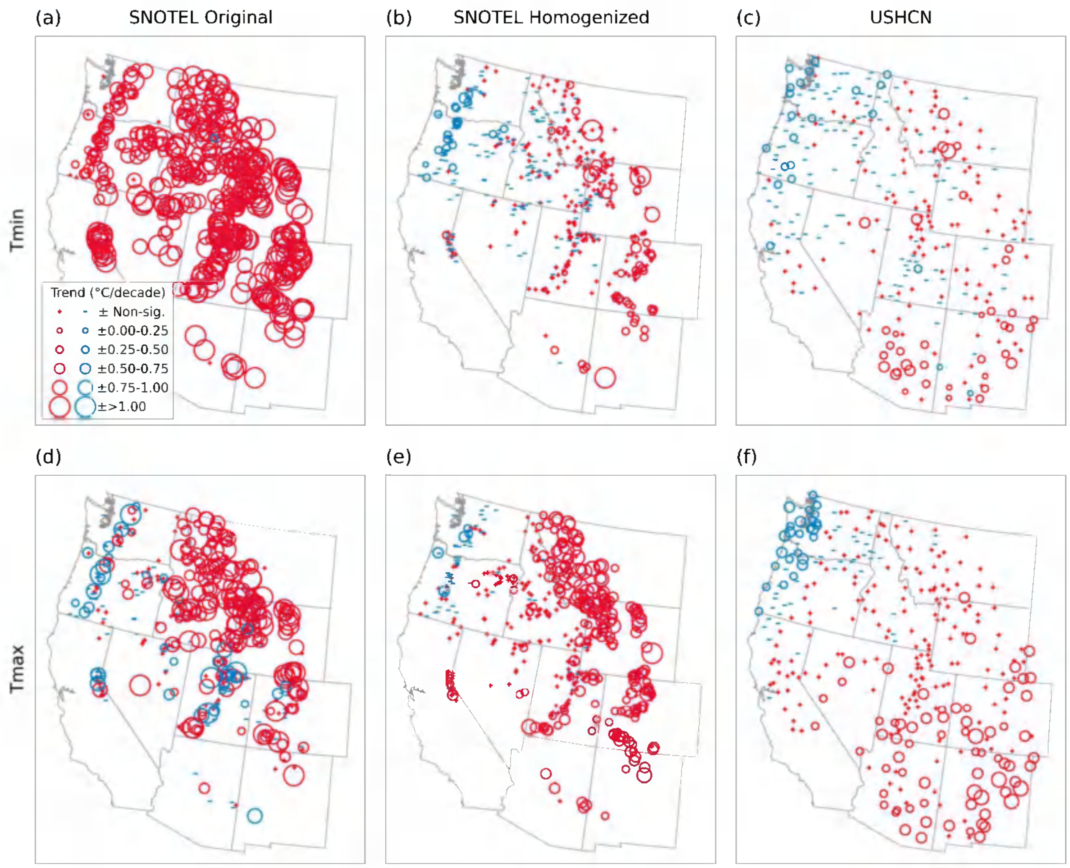

Figure 1. The 1991–2012 annual temperature trends in the western U.S. for higher-elevation SNOTEL ($n = 482$) and lower elevation USHCN ($n = 320$) stations. Minimum temperature (Tmin) trends for (a) SNOTEL, (b) SNOTEL homogenized, and (c) USHCN. Maximum temperature (Tmax) trends for (d) SNOTEL, (e) SNOTEL homogenized, and (f) USHCN. USHCN stations include all USHCN stations within 200 km of a SNOTEL station. Nonsignificant trends have a p value > 0.05 when accounting for temporal autocorrelation.

2. Materials and Methods

We first quantified observed 1991–2012 SNOTEL trends for annual minimum and maximum temperature (Tmin and Tmax) and compared them to those at lower elevation stations from the U.S. Historical Climatology Network (USHCN) [*Menne et al.*, 2009]. USHCN is the primary homogenized station database for the conterminous U.S. and is commonly used in temperature trend analyses [*Menne et al.*, 2009]. We restricted the trend analysis to the period 1991–2012 because a majority of SNOTEL stations started to observe temperature in the early 1990s (Figures 1a and S2a in the supporting information). We obtained all daily Tmin and Tmax observations for SNOTEL stations in the western U.S. ($n = 716$) from the United States Department of Agriculture Natural Resources Conservation Service (USDA NRCS). To remove possible observation errors and invalid values, we applied the automated quality assurance procedures of *Durre et al.* [2010] to the SNOTEL data. We aggregated daily SNOTEL values to monthly values with the requirement that a given month have ≤ 9 days of missing data [*Menne et al.*, 2009]. If this criterion was not met, we marked the month as missing. Over the period 1991–2012, we required a SNOTEL station to have no more than 2 years of missing data in each month. This resulted in a total of 482 input stations (Figure 1). Following the methods of *Oyler et al.* [2014], remaining missing monthly values at individual stations were infilled using surrounding neighboring stations and atmospheric reanalysis data (Text S1). Overall, 0.96% of the 1991–2012 SNOTEL monthly observations were infilled. We obtained corresponding 1991–2012 homogenized monthly Tmin and Tmax observations for USHCN stations (version 2.5.0 20140715) [*Menne et al.*, 2009] from the National Climatic Data Center. For a USHCN station to be included in subsequent analyses, we required it to be within 200 km of a SNOTEL station. A total of 320 USHCN stations met this requirement (Figure 1).

To estimate annual temporal temperature trends for SNOTEL and USHCN, we applied ordinary least squares linear regression to time series of annual temperature anomalies. When assessing trend statistical significance, we used the methods of *Santer et al.* [2000] to account for temporal autocorrelation. To determine if SNOTEL and USHCN trends were significantly different, we calculated the statistical significance of the trend of their anomaly differences [*Santer et al.*, 2000]. We estimated composite western U.S. annual temperature anomaly time series for both SNOTEL and USHCN by first averaging station annual anomalies by each western U.S. state and then taking an area-weighted average of the state anomalies. To calculate 1991–2012 time series of annual anomaly differences between a given SNOTEL station and its surrounding USHCN stations, we calculated a composite anomaly time series from the five closest neighboring USHCN stations and then subtracted it from the given SNOTEL station's anomaly time series.

To identify possible inhomogeneities and homogenize monthly SNOTEL observations, we used the USHCN pairwise homogenization algorithm (PHA) [*Menne and Williams*, 2009; *Menne et al.*, 2009]. PHA uses multiple pairwise comparisons of station time series to identify inhomogeneities in a station record relative to stations in its surrounding region [*Menne and Williams*, 2009; *Menne et al.*, 2009]. To remove inhomogeneities, PHA estimates temperature adjustments for all homogenous segments to bring them into consistency with the current state of the station [*Menne and Williams*, 2009; *Menne et al.*, 2009]. For instance, if PHA identifies a sudden artificial rise in temperature at a station, it will apply a positive adjustment to all temperatures before the artificial rise occurred. Following *Oyler et al.* [2014], we ran PHA over a 65 year time period (1948–2012) using all nonmissing aggregated monthly station observations from both the SNOTEL network and the unhomogenized Global Historical Climatology Network Daily (GHCN-D) database [*Menne et al.*, 2012]. When running PHA, we did not use any metadata specifying dates of station changes. To limit the homogenization process from incorrectly imposing a lower elevation regional climate signal on the higher-elevation SNOTEL stations [*Pielke et al.*, 2007], we required a given input GHCN-D station to not only be within 200 km of a SNOTEL station but also at an elevation greater than or equal to the minimum elevation of its neighboring SNOTEL stations. A total of 1395 GHCN-D stations met this requirement and, along with 716 SNOTEL stations, were used as input to PHA.

3. Results and Discussion

3.1. SNOTEL and USHCN Trends

From 1991 to 2012, SNOTEL displayed a positive significant trend of +1.16°C decade^{-1} ($p < 0.01$, ±0.122°C; standard error) in annual Tmin compared to a slightly positive insignificant trend of +0.069°C decade^{-1} ($p = 0.58$, ±0.123°C) for surrounding lower elevation USHCN stations. The larger warming trend in the SNOTEL network was consistent across all states with 87% of SNOTEL stations displaying a statistically significant ($p \leq 0.05$) positive Tmin trend (Figures 1a and S2a in the supporting information). In contrast, 7% of USHCN stations showed a significant positive trend (Figures 1c and S2a). Differences in annual Tmax trends between the two networks were also noticeable but not as regionally consistent (Figures 1d, 1f, and S2b). Relative to surrounding lower elevation USHCN stations, some states displayed significantly warmer regional SNOTEL Tmax trends (e.g., Idaho, Montana, and Wyoming), while other states had cooler (e.g., Colorado and Utah) or statistically indistinguishable (e.g., Oregon) trends (Figure S2b). In contrast to Tmin, the overall western U.S. annual 1991–2012 Tmax trends for SNOTEL (+0.252°C decade^{-1}; $p = 0.15$, ±0.166°C) and USHCN (+0.261°C decade^{-1}; $p = 0.15$, ±0.173°C) were similar.

The apparent differences in temperature trends between the SNOTEL and lower elevation USHCN data corroborate those found in previous analyses [*Diaz and Eischeid*, 2007; *Clow*, 2010; *Pederson et al.*, 2011], but the more coherent elevation-dependent warming signal in SNOTEL Tmin is not consistent with our understanding of potential driving mechanisms. Analyses of SNOTEL observations in the Southern and Northern Rockies have attributed elevation-dependent warming to a possible snow-albedo feedback [*Clow*, 2010; *Pederson et al.*, 2011]. Although a decrease in snow cover could have an impact on Tmin from changes in longwave emissions at night, a snow-albedo feedback would likely have a larger warming influence on Tmax due to increased absorption of solar radiation during daytime hours [*Rangwala and Miller*, 2012]. Furthermore, the largest elevation-dependent warming for SNOTEL Tmax is during the coldest winter months (Figure S3b in the supporting information), not during the spring and summer when a snow-albedo feedback should be most influential [*Rangwala and Miller*, 2012]. Simultaneous warmer SNOTEL Tmin trends and cooler Tmax trends

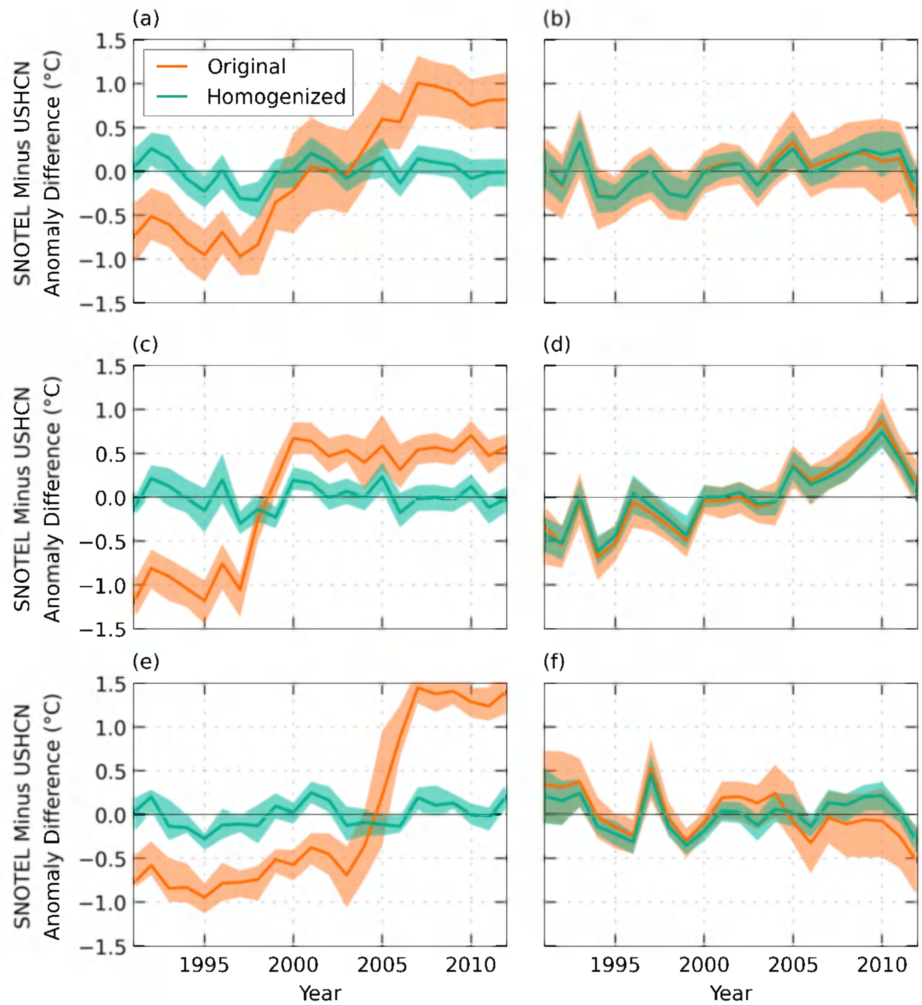

Figure 2. Average differences in 1991–2012 annual temperature anomalies between SNOTEL and USHCN stations for original and homogenized SNOTEL observations. Shaded areas are the interquartile range. (a) Minimum (Tmin) and (b) maximum (Tmax) temperature average annual anomaly differences for entire western U.S. (c) Tmin and (d) Tmax average annual anomaly differences for Northern Rockies of Montana. (e) Tmin and (f) Tmax average annual anomaly differences for Southern Rockies of Colorado.

relative to lower elevation USHCN stations within Utah and the Southern Rockies of Colorado are also inconsistent with a snow-albedo feedback (Figure S2).

Instead of a snow-albedo feedback, a critical assessment of the data indicates that the apparent discrepancy between the SNOTEL and USHCN temperature trends is most likely due to a systematic bias in the SNOTEL observation record. Differences between SNOTEL and USHCN annual Tmin anomalies (SNOTEL minus USHCN) reveal that the trend difference between the two networks is largely driven by sudden steplike increases in SNOTEL Tmin observations that occurred across the western U.S. from the mid-1990s to mid-2000s (Figure 2a). The timing of the step change varies by region (Figures 2c and 2e), but such a distinctive steplike pattern is indicative of network-wide changes that occurred at single points in time. In fact, from the mid-1990s to mid-2000s, a field campaign was conducted to move the SNOTEL temperature sensors to a new standardized meteorological data collection tower at each site with changes in sensor instrumentation, location, and height (Text S2). Such changes often produce nonclimatic variations in observed temperature [*Pielke et al.*, 2007; *Menne et al.*, 2009]. Out of the numerous station changes that occurred, the transition to a new sensor type that was warm biased at colder temperatures was likely the main driver of the systematic increase in SNOTEL Tmin (Text S3 and Figure S4 in the supporting information).

Application of the homogenization algorithm [*Menne and Williams*, 2009] confirmed the direct correspondence between steplike increases in SNOTEL Tmin and changes in sensor type. SNOTEL digital metadata containing exact dates of station changes is generally not available, and only the general period over which changes occurred is known. For a limited set of Utah stations with complete metadata ($n = 52$, Table S1 in the supporting information), homogenization detected an artificial Tmin change point within 1 year of a sensor-type switch at 85% of the stations with an average added bias of +1.53°C. The step increases in Montana and Colorado SNOTEL Tmin (Figures 2c and 2e) and detected change points (Figures S5c and S5e in the supporting information) also coincided with the different timing of the field campaigns in the two states. Overall, 81% of detected change points in the SNOTEL Tmin observations occurred between 1995 and 2006 with an average bias of +1.42°C (Figure S5a).

The direct relationship between network-wide sensor changes and the steplike increases and detected change points in Tmin suggests that recent amplified warming in the SNOTEL network is largely due to network inhomogeneities leading to positive Tmin biases. Once the SNOTEL Tmin observations are homogenized, the step changes in the SNOTEL and USHCN annual anomaly differences are dramatically reduced (Figures 2a, 2c, and 2e). As a result, the regional SNOTEL annual Tmin trend decreases from the +1.16°C decade^{-1} ($p < 0.01$, ±0.122°C) significant trend in the raw data to an insignificant trend of +0.106°C decade^{-1} ($p = 0.39$, ±0.122°C) in the homogenized data, a trend that is statistically indistinguishable ($p = 0.43$) from the lower elevation +0.069°C decade^{-1} ($p = 0.58$, ±0.123°C) USHCN trend (Figures 1 and S2a). We note that the lack of significant positive trends in USHCN and homogenized SNOTEL Tmin likely represents the short time period of the record comparison and does not imply that the western U.S. is not warming [see *Cayan et al.*, 2001; *Bonfils et al.*, 2008; *Das et al.*, 2009; *Meehl et al.*, 2012; *Abatzoglou et al.*, 2014].

In contrast to Tmin, the effect of station changes on annual SNOTEL Tmax trends was less distinctive (Figures 2b, 2d, and 2f). Compared to Tmin, differences between monthly SNOTEL and USHCN Tmax trends had a more defined seasonal pattern with SNOTEL Tmax trends generally warmer than USHCN in winter and cooler than USHCN in summer (Figure S3b). This seasonal pattern and the smaller Tmax changes at the annual temporal scale are likely a result of the temperature-dependent bias of the new sensor. A preliminary analysis of the sensor bias suggests that the bias reverses from positive at colder temperatures to negative at warmer temperatures (Text S3 and Figure S4a). As a result, unlike Tmin which produces a significant positive bias throughout the year, the sensor bias for Tmax switches from positive in winter to negative in summer (Figure S4b). With these counteracting seasonal biases, the overall effect of the sensor change on annual Tmax trends is not as large as it is for Tmin (Figure 2). This seasonal swing in bias also appears to have limited the effectiveness of the homogenization algorithm. The PHA algorithm assumes that an inhomogeneity will result in a bias that is seasonally consistent [*Menne and Williams*, 2009]. The seasonal reversal of the Tmax bias (Figure S3b) makes it difficult for the homogenization algorithm to detect artificial change points and estimate appropriate adjustments [*Trewin*, 2013]. This reasoning is supported by the relatively smaller differences in original versus homogenized SNOTEL Tmax trends in Montana (Figure 2d) where there was a large seasonal swing in the SNOTEL Tmax bias (Figure S3d). Conversely, homogenization had a greater impact on Colorado Tmax SNOTEL trends where the trend differences between SNOTEL and USHCN still displayed a seasonal pattern but were more consistently negative across most months (Figures 2f, S2b, and S3f). With homogenization, Colorado's SNOTEL annual Tmax trend moved from significantly cooler than USHCN to nonsignificantly different (Figures 2f and S2b).

3.2. Implications

Large trend biases in the SNOTEL record, especially for Tmin, have significant implications for our understanding of climate-dependent processes in mountainous environments of the western U.S. Minimum temperatures not only influence snowpack dynamics [*Pederson et al.*, 2011] and various ecological processes [*Dobrowski et al.*, 2013], but two widely used gridded climate products, PRISM [*Daly et al.*, 2008] and Daymet [*Thornton et al.*, 1997], directly ingest raw SNOTEL Tmin observations. While the unhomogenized PRISM and Daymet climate products were never intended for climate trend analyses [*PRISM Climate Group*, 2013], they have been used to analyze elevation-dependent warming [*Diaz and Eischeid*, 2007] in addition to climate-driven impacts to mountain snowpack [*Pederson et al.*, 2013], tree growth [*Williams et al.*, 2010] and mortality [*van Mantgem et al.*, 2009], species distributions [*Crimmins et al.*, 2011], ecosystem productivity [*Turner et al.*, 2011], and a multitude of other climate-driven ecological and hydrological processes across the western U.S.

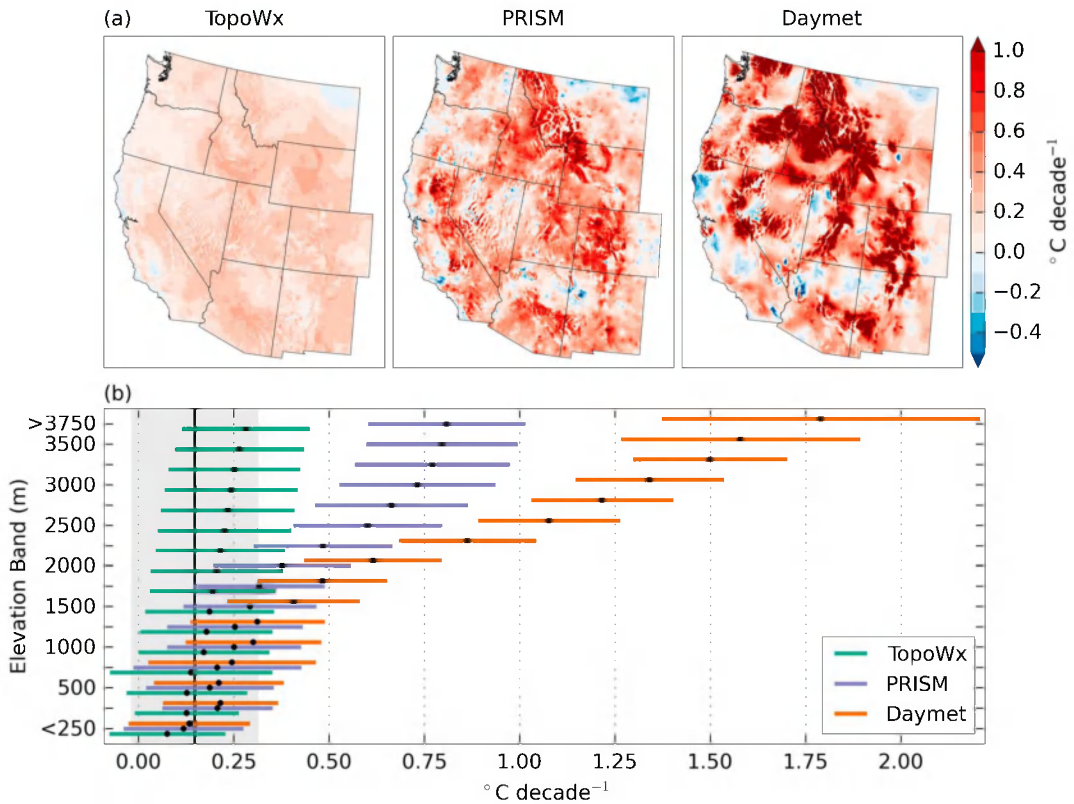

Figure 3. The 1981–2012 annual minimum temperature (Tmin) trends for the western U.S. for the homogenized TopoWx data set, PRISM, and Daymet. (a) Trend maps. (b) Composite trends by elevation band. Dots and colored bars in Figure 3b represent the composite trends and associated 95% confidence intervals for different elevation bands. Vertical black line and light gray envelope in Figure 3b represent the composite lower elevation USHCN western U.S. Tmin trend (+0.148°C decade^{-1}; $p = 0.08$, ±0.081°C) and associated 95% confidence interval.

In both the PRISM and Daymet products, a direct correspondence with the spurious SNOTEL anomalies with increasing elevation confirms that SNOTEL observations have biased the gridded data (Figure S6 in the supporting information). Moreover, the effect of the SNOTEL Tmin inhomogeneity is not simply limited to the short time period of this analysis. Over a longer 32 year period (1981–2012), PRISM and Daymet display a strong pattern of elevation-dependent warming (Figure 3). If we consider the 1981–2012 composite USHCN Tmin trend for the western U.S. (+0.148°C decade^{-1}; $p = 0.08$, ±0.081°C) as the lower elevation trend, PRISM higher-elevation trends are statistically different (t test $p \leq 0.05$) from USHCN starting at 2250 m (Figure 3b). Daymet trends are statically different from USHCN starting at an even lower elevation of 1500 m (Figure 3b). In contrast, TopoWx, a recently developed gridded climate product that uses PHA to homogenize input SNOTEL observations [*Oyler et al.*, 2014], displays coherent Tmin warming across the western U.S. over the 1981–2012 time period (+0.173°C decade^{-1}; $p = 0.04$, ±0.079°C), but has a more muted elevation-dependent warming pattern (Figure 3). TopoWx trends increase with elevation but are never statistically different from the lower elevation USHCN trend (t test $p \leq 0.25$ at all elevations), suggesting that the Tmin elevation-dependent warming signal has not yet fully emerged from the interannual variability associated with the background regional trend (Figure 3b). For every 500 m increase in elevation, the increase in the average 1981–2012 annual Tmin trend for TopoWx is only +0.027°C decade^{-1}, whereas the PRISM increase is 217% greater at +0.084°C decade^{-1} and the Daymet increase is 562% greater at +0.176°C decade^{-1}. For elevations > 3000 m, this results in a composite 1981–2012 annual Tmin trend of +0.760°C decade^{-1} ($p < 0.01$, ±0.10°C) and +1.438°C decade^{-1} ($p < 0.01$, ±0.10°C) for PRISM and Daymet, respectively, compared to +0.250°C decade^{-1} ($p = 0.01$, ±0.084°C) for TopoWx with homogenized input data. Differences between PRISM and Daymet are a result of how their contrasting spatiotemporal interpolation methods extrapolate and enhance the spurious SNOTEL warming (Text S4).

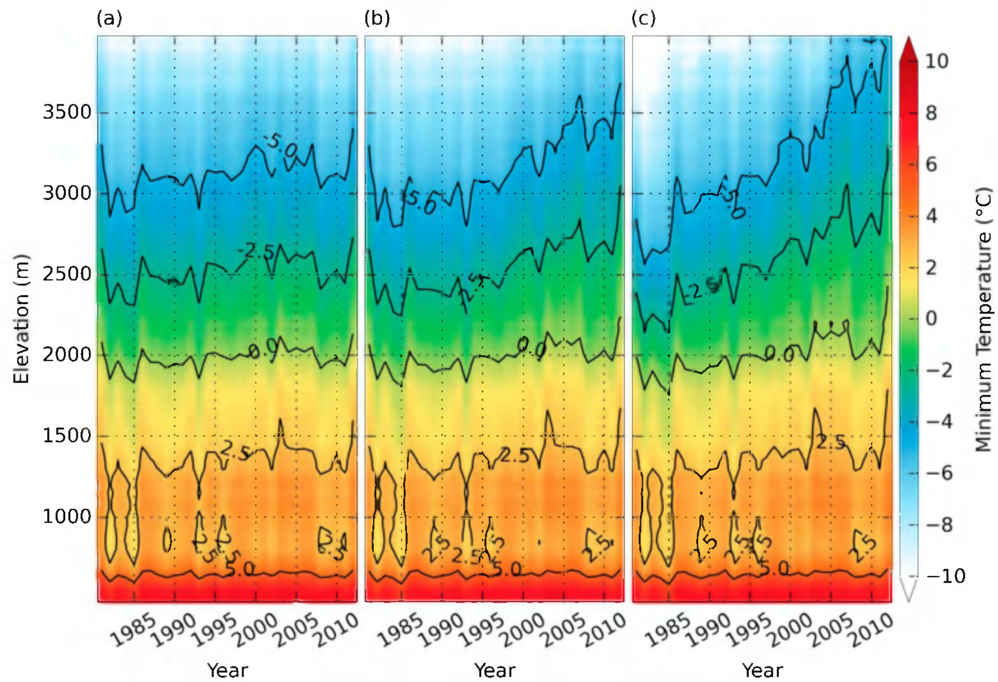

Figure 4. The 1981–2012 annual average minimum temperature with respect to time and elevation for the western U.S. From top to bottom, contour lines are the −5.0°C, −2.5°C, 0.0°C, 2.5°C, and 5.0°C isotherms. (a) Homogenized TopoWx data set. (b) PRISM. (c) Daymet.

The anomalous high-elevation warming trends in PRISM and Daymet overestimate the change in elevation of important isotherms that influence snowpack dynamics (Figure 4) [*Pederson et al.*, 2011, 2013]. This compromises our ability to understand regional and landscape-scale climate impacts across the western U.S. From 1981–2012, during the typical snowmelt season in late spring (April, May, and June), PRISM displays a +274 m ($p = 0.06$, ±140 m) elevation increase in the 0°C Tmin isotherm, an important hydrological and ecological temperature threshold [*Pederson et al.*, 2010]. Daymet displays an even greater increase of +487 m ($p < 0.01$, ±143 m), whereas the homogenized TopoWx data set displays a positive but statistically insignificant change (+66 m; $p = 0.64$, ±140 m). Annually, in contrast to colder, high-elevation isotherms, lower elevation isotherms exhibit more similar changes among all three data sets (Figure 4) further emphasizing the influence of the amplified elevation-dependent warming pattern (Figure 3b).

Robust estimates of historical climate are critical for accurately assessing historical climate change exposure of species and various biotic and physical processes and their associated climate sensitivities [*Glick et al.*, 2011]. Because of the amplified trends, previous analyses may overestimate climate change exposure at high elevations [*Dobrowski et al.*, 2013]. Studies that have related amplified temperature trends to observed biotic processes, such as tree growth [*Williams et al.*, 2010] and mortality [*van Mantgem et al.*, 2009], may overestimate or underestimate the actual process' climatic sensitivity depending on whether a causal linkage between observed biotic impacts and changes in temperature can definitely be made. In other words, biased trend estimates compromise our ability to attribute the causes of biotic and hydrologic climate change impacts.

4. Conclusions

It is clear that widespread systematic inhomogeneities in the SNOTEL network have artificially amplified an elevation-dependent warming signal across much of the western U.S. (Figures 1 and 2) and have significantly biased trend estimates from widely used climate data products (Figure 3). The application of a standard homogenization procedure in this analysis and in the TopoWx gridded data is a first step in addressing the SNOTEL inhomogeneity issues but is not completely effective due to the limited ability of the homogenization to account for strong seasonal dependencies within the SNOTEL sensor bias (Figures S3 and S4). SNOTEL inhomogeneities, while subdued by homogenization, likely remain in the TopoWx product (Text S4 and Figure S7 in the supporting information), especially on a seasonal basis (Figure S3). Additionally, although homogenized data sets like TopoWx produce more spatially coherent trends (Figure 3a) [*Menne et al.*, 2009], it is important to note that homogenization has the potential to over smooth the trend field and remove unique climate phenomena from individual sites [*Pielke et al.*, 2007]. Due to these limitations and the degree of concurrency with which SNOTEL station changes were made in specific regions, a direct empirical model of the sensor bias or a homogenization approach that accounts for seasonally varying biases [*Trewin*, 2013] will be needed to adequately homogenize SNOTEL observations and fully disentangle any elevation-dependent warming from both the inhomogeneities and the background regional climate signal. It is also important to note that although SNOTEL inhomogeneities are severe, this does not negate the possibility that elevation-dependent warming has occurred in the western U.S. [*Diaz and Eischeid*, 2007] or that it will be important in the future [*Bradley et al.*, 2004; *Rangwala et al.*, 2013]. Modeling studies suggest that a snow-albedo feedback can strongly enhance surface warming in mountainous terrain [*Giorgi et al.*, 1997; *Rauscher et al.*, 2008]. Nonetheless, owing to the inhomogeneity issues we have illustrated here and the sparse number of quality high-elevation observations, our ability to fully identify key mechanisms and patterns of elevation-dependent warming continues to remain limited [*Rangwala and Miller*, 2012].

While the mountainous landscapes of the western U.S. have clearly warmed over the latter half of the twentieth century [*Cayan et al.*, 2001; *Bonfils et al.*, 2008; *Das et al.*, 2009; *Meehl et al.*, 2012; *Abatzoglou et al.*, 2014] and there is evidence of climate-driven hydrological and ecological impacts [*Cayan et al.*, 2001; *Barnett et al.*, 2008], we have shown that uncritical use of climate data can lead to erroneous conclusions about trends in mountain climate. As climate change vulnerability and impact assessments are increasingly used to inform natural resource policy and conservation [*Glick et al.*, 2011], climate change impact studies should be assessed to ensure that climate change exposure has been accurately quantified. We call for more dialogue between managers of observational networks, climate data product developers, and data end users to increase community awareness of data product caveats and limitations, and to advance our understanding of the unique drivers and impacts of climate change in mountain environments.

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