LETTER

Snowpack signals in North American tree rings

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Abstract

Climate change has contributed to recent declines in mountain snowpack and earlier runoff, which in turn have intensified hydrological droughts in western North America. Climate model projections suggest that continued and severe snowpack reductions are expected over the 21st century, with profound consequences for ecosystems and human welfare. Yet the current understanding of trends and variability in mountain snowpack is limited by the relatively short and strongly temperature forced observational record. Motivated by the urgent need to better understand snowpack dynamics in a long-term, spatially coherent framework, here we examine snow-growth relationships in western North American tree-ring chronologies. We present an extensive network of snow-sensitive proxy data to support high space/time resolution paleosnow reconstruction, quantify and interpret the type and spatial density of snow related signals in tree-ring records, and examine the potential for regional bias in the tree-ring based reconstruction of different snow drought types (dry versus warm). Our results indicate three distinct snow-growth relationships in tree-ring chronologies: moisture-limited snow proxies that include a spring temperature signal, moisture-limited snow proxies lacking a spring temperature signal, and energy-limited snow proxies. Each proxy type is based on distinct physiological tree-growth mechanisms related to topographic and climatic site conditions, and provides unique information on mountain snowpack dynamics that can be capitalized upon within a statistical reconstruction framework. This work provides a platform and foundational background required for the accelerated production of high-quality annually resolved snowpack reconstructions from regional to high (<12 km) spatial scales in western North America and, by extension, will support an improved understanding of the vulnerability of snowmelt-derived water resources to natural variability and future climate warming.

1. Introduction

Anthropogenic climate change has caused widespread mountain snowpack declines and snow droughts across western North America (Pierce et al 2008, Mote et al 2018) that are projected to continue at an increasingly rapid pace over the 21st century (Barnett et al 2005, Mankin and Diffenbaugh 2015). Snowpack stores water during the cool season and releases it during the warm season when human, ecosystem, and evapotranspiration demands are highest (Barnett et al 2005). In the western United States, more than 50% of total regional runoff and two thirds of major reservoir inflows originate from snow (Li et al 2017).
Recent changes to the quantity and timing of continental freshwater runoff (Barnett et al. 2008) are contributing to large-scale water supply shortages and hydrologic drought (AghaKouchak et al. 2014), flood intensification (Cayan et al. 2016), forest mortality (Anderegg et al. 2013), insect outbreak (Mitton and Ferrenberg 2012), and longer and more active fire seasons (Westerling et al. 2006), with profound consequences for ecosystems and human welfare (Vörösmarty et al. 2010).

Studies of snowpack dynamics have previously employed observational and model simulated datasets to investigate the causes and consequences of recent snowpack declines and snow droughts (Pierce et al. 2008, Abatzoglou 2011, Harpold et al. 2012, Pederson et al. 2013, Mao et al. 2015, Shukla et al. 2015, Cooper et al. 2016, Safeeq et al. 2016, Berg and Hall 2017, Hatchett and McEvoy 2018). One drawback of these studies is that they are based on the relatively short (typically <50 years) and strongly temperature forced observational record and are therefore unlikely to capture the full range of natural variability and potential extremes within mountain snowpack systems (Mankin and Diffenbaugh 2015). Tree-ring records offer an exceptional opportunity for reconstructing, quantifying, and understanding long-term mountain snowpack dynamics on timescales relevant to hydroclimate modeling and climate adaptation planning (Bradley 2011, Meko and Woodhouse 2011, Woodhouse et al. 2016).

Dendroclimatology has a rich history of using tree-ring width, wood density, and stable isotope records to study past hydroclimate variations at local to hemispheric scales, with these efforts mainly focused on the development and interpretation of drought- and temperature-sensitive tree-ring chronologies and reconstructions (Gagen et al. 2011, Hughes 2011, Villalba et al. 2011). Other tree-ring records sensitive to cool-season precipitation (rain and/or snow) have been widely used to reconstruct annual runoff mainly in semi-arid environments (Stockton and Jacoby 1976, Woodhouse et al. 2006, Meko et al. 2007), and in a limited number of cases chronologies sensitive to snow water equivalent (SWE) have been used to reconstruct snowmelt-derived streamflow (Hart et al. 2010, Starheim et al. 2013, Coulthard and Smith 2016, Coulthard et al. 2016, Littell et al. 2016, Martin et al. 2019, Welsh et al. 2019) and glacier mass balance (Lewis and Smith 2004, Larocque and Smith 2005b, Wood et al. 2011, Marcinkowski and Peterson 2015) in temperate environments. From a water resources standpoint, existing tree-ring based streamflow, cool-season precipitation, and drought reconstructions do not differentiate between the forms of precipitation (rain, snow) that have substantially different effects on surface hydrology, water supply, storage, and runoff timing (Berghuijs et al. 2014, Barnhart et al. 2016, Li et al. 2017), making targeted snowpack reconstructions critical. But while tree-rings may reflect metrics of snowpack variation in a range of settings, they have rarely been used to reconstruct snowpack directly, and there remains little high space/time resolution paleoclimate knowledge of global mountain snowpack dynamics over the last 2000 years.

Woodhouse (2003) developed the first tree-ring-based snowpack reconstruction for the Gunnison River basin, Colorado, showing twentieth century snowpack dynamics were representative of the past 400 years, a result later supported by Timilsena and Piechota (2008). SWE reconstructions for the Pacific Northwest based on high-elevation conifer chronologies exhibited variability consistent with North Pacific ocean-atmosphere circulation (Larocque and Smith 2005a, Mood et al. 2015) and showed 2015 was likely the lowest snowpack in three or four centuries (Coulthard 2015, Harley et al. 2020). Similarly, Belmecheri et al. (2016) combined chronologies sensitive to winter precipitation and early spring temperature in the Sierra Nevada of California to reconstruct April 1 SWE, indicating snowpacks were likely the lowest in 500 years. Using a spatially distributed multi-species chronology network, Pederson et al. (2011b) generated the first set of broad-scale April 1 SWE reconstructions, revealing snowpack reductions since 1980 across the American Rocky Mountains were unusual in magnitude and north-south synchrony with respect to the last millennium, and likely driven by springtime warming (Pederson et al. 2013). Other notable approaches to reconstructing snowpack include spatially interpolated methods (Barandiaran et al. 2017), use of tree-ring δ13C records (Liu et al. 2011), reconstructing snow cover metrics (Qin et al. 2016), and reconstructing total (Yadav and Bhutiyani 2013) and maximum (Masiokas et al. 2012) winter SWE. Beyond analyses that form the basis for statistical snowpack reconstructions, snow signals contained in total, early, and late-wood ring width, as well as tree-ring isotope, records have also been examined at local- to regional-scales (Heikkinen 1985, Graumlich and Brubaker 1986, Peterson and Peterson 1994, Larocque and Smith 1999, Vaganov et al. 1999, Peterson and Peterson 2001, Peterson et al. 2002, Kirdyanov et al. 2003, Nakawatase and Peterson 2006, Schmidt et al. 2010, Marcinkowski et al. 2015, Watson and Luckman 2016, Carlson et al. 2017, Leonelli et al. 2017). These studies highlight interesting relationships between tree rings and snow that are typically less straightforward than with rainfall and temperature, since snow reflects the combined seasonal influences of precipitation, surface energy balance, topography, and elevation on tree growth through multiple potentially lagged or distal mechanisms.

Leveraging the temporal resolution and long record lengths of tree-ring chronologies combined
with their high spatial replication, spatial field reconstruction is one of the most powerful approaches for examining pre-instrumental drought using dendroclimatology (Cook et al. 1999). ‘Drought Atlases’ that now represent much of the terrestrial Earth (Cook et al. 1999, 2010, 2015, Palmer et al. 2015, Stahle et al. 2016) have been widely used to place recent extremes within a context of long-term variability (Fye et al. 2003, Cook et al. 2004, Woodhouse et al. 2005), examine drought-related environmental processes (Westerling and Swetnam 2003, Hessl et al. 2004, Gray et al. 2006), test the long-term stability of ocean-atmosphere climate forcings (Cole and Cook 1998, Hidalgo 2004, Herweijer et al. 2007, Anchukaitis et al. 2019), and operate as real-world targets for climate model simulations (Seager et al. 2005, Yoshimori et al. 2005, Meehl and Hu 2006, Seager et al. 2009). Given local spatial snow-depth variability in mountainous terrain, snowpack reconstructions with high (<12 km) gridded spatial resolution are particularly desirable, but while the seasonality of drought signals contained in the Atlases is tuned to precipitation-as-snow in some regions (St. George et al. 2010), and there is a growing focus on refining cool- and warm-seasonal precipitation signals in tree-ring based Drought Atlases (Stahle et al. 2020), snowpack has not yet been targeted directly for high resolution gridded reconstruction. Such an advance would enable the questions above to be addressed relative to changing snowpack dynamics for the first time, providing an improved understanding of the variability and vulnerability of snow-dominated water resources to climate change, and how that may ultimately impact society and ecosystems under future warming.

Building on this existing body of work and prompted by the need to understand rapidly changing North American snowpack dynamics, our goal is twofold: (1) to quantify and interpret how and where snow-related signals are captured and preserved in western North American tree-ring chronologies and (2) to develop a network of snow-sensitive paleoenvironmental proxy data suitable for use in high time/space resolution snowpack reconstructions. Specifically, we evaluate the type and density of snow related tree-ring signals for producing high-resolution reconstructions, examine the potential for regional bias in snowpack reconstructions due to the type of snow information contained in local chronologies, and discuss the strengths and limitations of each proxy type for reconstructing dry versus warm snow droughts.

2. Materials and methods
2.1. Tree-ring data and chronology development
We compiled raw total tree-ring width measurements from 954 sites for this study. These sites had all previously been identified as hydroclimatically sensitive and of high quality (including extended age, confirmed crossdating accuracy, and common growth signal strength), and include collections or compilations from the Missouri River Basin (Martin et al. 2019), the Columbia River Basin (Littell et al. 2016), the North American Drought Atlas (Cook et al. 1999), and collections developed by B.L. Coulthard and D J Smith (Coulthard and Smith 2016, Coulthard et al. 2016). All raw measurement series were power transformed to normalize age-related growth trends, variance-stabilized using a 67% spline, detrended using a negative exponential curve or where that failed an age-dependent spline, and standardized using a biweight robust mean to develop new versions of each tree-ring chronology for this study (Cook 1985, Cook and Kairiukstis 1990). The expressed population signal (EPS) (Wigley et al. 1984) was used as an estimate of the adequacy of the sample size for capturing the hypothetical population growth signal, and all chronologies were truncated at the conventional but arbitrary EPS value of 0.85. Standard chronologies were used to preserve the potential influence of snowpack conditions on tree growth prior to the year of ring formation (Cook 1985). Elevation metadata were not available for most chronologies since they have not been previously required in standard tree-ring data formats and repositories, including the International Tree Ring Databank (https://www.ncdc.noaa.gov/data-access/paleoclimatology-data/datasets/tree-ring).

2.2. Influence of snow on radial tree growth
As a first-pass test for whether the tree-ring chronologies are significantly influenced by annual snowpack variability, we calculated site-by-site Pearson’s and Spearman’s rank-order correlations between each chronology and its local (nearest) instrumental April 1 SWE record, a general measure of peak annual snow water content (Bohr and Aguado 2001). Snow-course records based on manual snow surveys, and Snowpack Telemetry (SNOTEL) records based on a network of remote snow pillow sensors, were accessed from the Natural Resources Conservation Service (www.wcc.nrcs.usda.gov/snow). For each chronology, the closest SWE record within 300 km that had a minimum of 30 years of common data and no consecutive missing SWE data years, was selected. Sensitivity testing at 50 km increments demonstrated the 300 km radius optimizes SWE record availability for the largest number of chronology sites while minimizing the influence of heterogeneous mountainous terrain on snowpack over larger areas. We used the full suite of 1596 North American snow-course or SNOTEL records west of 90° longitude (all stations within the general snowpack accumulation zone >900 m asl), and the set of chronologies significantly (p < 0.05) correlated with observed April 1 SWE make up our Western Paleosnow Network (WPN).

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To further refine seasonal snow signal in the WPN, we sought to disentangle interacting influences of monthly and seasonal precipitation and temperature—which are integrated in the annual April 1 SWE measurement—on the annual radial growth increment. We therefore also compared each tree-ring chronology with the closest point in the 0.5° gridded Global Precipitation Climatology Centre precipitation data (GPCC, version 7) (Yoshimori et al 2005, Becker et al 2013) and the 0.5° gridded Climatic Research Unit temperature data (CRU TS4.0) (Mitchell and Jones 2005) over the common data period 1901–2013. Site-by-site correlation (precipitation) and partial correlation (temperature) analysis of each chronology with its local (nearest) gridded climate data were calculated using the seasonal correlation (SEASCORR) procedure developed by Meko et al (2011). We tested individual monthly and seasonal precipitation and temperature records integrating 2, 3, or 5 months, within a 14-month window starting in the August prior to the growth year and ending in the following September. Significance was estimated using exact simulation (Percival and Constantine 2006). Next, we identified groups of snow-sensitivechronologies with similar monthly and seasonal climate responses using k-means cluster analysis on the 112 monthly and seasonal correlation and partial correlation coefficients that were generated for each chronology (Touchan et al 2014). The optimal number of clusters was determined by iterative testing of the gap statistic and silhouette plots (Rousseeuw and Kaufman 1990, Tibshirani et al 2001).

We expected that tree-ring chronologies in the WPN might relate differently to local instrumental April 1 SWE than cool-season precipitation, since the former is an ‘instantaneous’ end-of-season measure of snowpack that integrates both accumulation and melt over the cool season, while the latter is a meteorological measure of total rain and snow during that season (De Jong et al 2005). Knowing how the two are linked at each sample site is important for both interpreting dendroclimatological snow signal and determining where and how that signal can be used to reconstruct different types of snow drought (Harpold et al 2017). We therefore also tested the pointwise correlation of local gridded cool-season (November through April) precipitation at each chronology site with the nearest instrumental April 1 SWE record that was previously selected for that chronology.0

The 0.5° gridded climate datasets were used for the SEASCORR and cool-season precipitation analyses since they extend into Canada. We also performed these analyses using 5 km gridded precipitation and temperature records (Vose et al 2014) for chronologies located within the United States, over the same data period, to ensure that the spatial resolution of the gridded datasets did not significantly influence the results.

3. Results

3.1. Tree ring chronologies

Pearson’s correlations indicated 326 chronologies were significantly ($p < 0.05$) correlated with April 1 SWE and together make up the WPN (figure 1), representing all major mountain ranges in the western United States and Canada. Results were only negligibly different using Spearman’s rank-order correlation tests to account for potential non-linear associations between tree-rings and April 1 SWE (figure S1 (available online at stacks.iop.org/ERL/16/034037/mmedia)), 320 significant chronologies. Chronologies significantly negatively correlated with SWE are concentrated in the Cascade Range, British Columbia Coast Mountains, and southern Canadian and northern American Rocky Mountains (hereafter referred to as the northern Rocky Mountains), and those significantly positively correlated with SWE are mainly located in the southern semi-arid portions of the study domain, particularly the southern Sierra Nevada, Great Basin, Colorado, and Arizona (figure 1). Those weakly correlated with SWE are almost all positively influenced by that variable, and generally follow a pattern of stronger correlation with lower latitude/higher aridity. Some are also located at lower forest border, semi-arid sites in the northern part of the study domain (e.g. in rainshadow zones). Overall, the central latitudes of the study domain bracketing 42° have the fewest and weakest SWE proxies. The WPN includes 19 tree species with the majority of chronology start dates in the 1600s and end dates in the early 1980s though the late 2010s (figure 2(b)). The full network spans the time period 1 to 2013 CE with 13% of the chronologies exceeding >1000 years in length (figure 2(b)). Millennial-length chronologies include lower forest border to high-elevation Juniperus, Pinus, and Pseudotsuga positively correlated with SWE, and high-elevation Larix lyallii (LALY) and Pinus albicaulis (PIAL) negatively correlated with SWE.

3.2. Climate relationships

Three tree-ring chronology clusters were selected based on the absence of below average silhouette scores, and also based on the largest gap change, which was for 3 clusters (table S1, figure S2). The results of this analysis suggest 3 robust chronology clusters exist within the WPN based on dominant and coherent seasonal climate signals across the network (figures 2(a), 3 and S3). Cluster 1 consists of 235 chronologies primarily located in the American Southwest. These chronologies are positively correlated with SWE and cool-season precipitation, and also integrate an independent but weaker negative response to summer (MJJ) temperature (figures 1, 2 and 3(a)). Pseudotsuga menziesii (PSME),
*Pinus ponderosa* (PIPO), and *Pinus edulis* (PIED) account for 82% of the chronologies in this cluster (figure 3(b)). Cluster 2 is comprised of 32 chronologies mostly located in southern California and positively correlated with SWE. Chronologies in cluster 2 also exhibit overall positive correlations with cool-season precipitation, but are distinct from cluster 1 in that they do not integrate summer temperature information, and also lack a correlation with prior fall precipitation (figures 2 and 3(b)). This cluster is dominated by *Quercus douglasii* (QUDG; 59%; figures 1 and 2). Cluster 3 consists of 59 chronologies mainly located in the Cascade Range, British Columbia Coast Mountains, and northern Rocky Mountains, that are negatively correlated with SWE (figures 1 and 2). In aggregate, chronologies in cluster 3 are not significantly correlated with precipitation or temperature alone in any month or season (figure 3(c)). High-elevation PSME, *Tsuga mertensiana* (TSME), *Picea engelmannii* (PCEN), *Pinus flexilis* (PIFL), and LALY represent 76% of this cluster.

To highlight where tree-ring chronologies might differently relate to snow versus rain, local cool-season precipitation correlations with instrumental April 1 SWE at every WPN tree-ring chronology location are mapped in figure 4 by cluster. Overall, SWE is most strongly coupled with cool-season precipitation at semi-arid chronology sites represented by clusters 1 and 2. In comparison, instrumental April 1 SWE tends to be more weakly coupled with cool-season precipitation at the chronology sites represented by cluster 3, with the exception of a small number of sites in Colorado and Utah. These relationships were also apparent when using high-resolution 5 km gridded climate datasets (figure S4).

### 4. Discussion

#### 4.1. Mechanisms governing the snow proxy types

Cluster 1 is made up of chronologies that are positively correlated with SWE (figure 1) and concentrated in the American Southwest (figure 2(a)). The SEASCORR analysis demonstrates these proxies respond positively to cool-season precipitation and negatively to summer temperature (figure 3(a)). In other words, these are traditional ‘moisture-limited’ tree-ring chronologies that benefit from snowmelt-derived soil moisture during the growing season and are hindered by warm conditions that enhance evapotranspiration during the growing season (Fritts 1971). Radial tree growth in the Southwest is more strongly associated with winter precipitation than anywhere else in the northern hemisphere (St. George and Ault 2014) and has been used extensively to reconstruct regional cool-season precipitation and drought (Cook et al 1999, Stahle et al 2007), streamflow (Stockton and Jacoby 1976, Woodhouse et al 2006, Meko et al 2007), and in a smaller number of cases snowpack (Woodhouse 2003, Pederson et al 2011b, Barandiaran et al 2017). That well over half of the hydroclimate-sensitive chronologies initially considered in this study are not actually correlated with April 1 SWE demonstrates that physiological growth processes governing traditional moisture- and SWE-sensitivity are not necessarily one and the same. SWE-correlated chronologies likely capture and integrate important and distinct information related to snow, such as microsite-level snowpack characteristics, cool-season rain to snow proportions, springtime temperatures and snow melt rate/timing, relative humidity, or soil properties.
Within the WPN, cluster 1 chronologies are *moisture-limited snow proxies* that generally have the strongest correlations with SWE, are most ubiquitous on the landscape, have been most widely used by dendroclimatologists in previous soil moisture or precipitation reconstructions, and appear to have excellent potential for snowpack reconstruction based on their correlation with SWE data and overall ubiquity. The chronologies mainly consist of PIED, PIPO, and PSME, mid- to high-elevation semi-arid species that generally rely directly upon snow melt-water availability to support tree growth in spring (figure 5(b)). They are concentrated in Colorado, the Great Basin, and the southwestern Sky Islands of Arizona and New Mexico, with the exception of a small number of low-elevation QUDG, QULO, QUMA chronologies whose site types are closer to those in cluster 2 (figure 2(a)). A caveat of this proxy is that precipitation delivered as rain may be ‘misdiagnosed’ by trees as snowfall; however, this can be alleviated through site-level meteorological analysis and avoidance of settings where cool-season precipitation and April 1 SWE are decoupled, or through more detailed site-level studies of source water use in trees. Indeed, a large number of strongly SWE-correlated chronologies at sites with well-coupled cool season precipitation and April 1 SWE exist across the cluster 1 spatial domain (figure 4, cluster 1).

**Cluster 2** consists of chronologies that are positively correlated with SWE (figure 1), but that are distinguished from cluster 1 by a pure cool-season precipitation signal lacking a summer temperature influence (figure 3(b)). The QUDG chronologies that dominate this cluster are reliant on cool-season rainfall to recharge soil moisture for growth, and are well-known for their very strong sensitivity to rainfall (Stahle et al. 2013, Griffin and Anchukaitis 2014). Distributed around the central valley of California and windward foothills of the Sierra Nevada, these sites are too low in elevation to be directly affected by snowpack, but rather are synoptically influenced by the same weather systems that deliver local precipitation as rain and at higher elevation as snow (figure 5(a)). This growth-climate relationship has been exploited by Belmcheri et al. (2016) to reconstruct the leading mode of Sierra Nevada snowpack in combination with a separate early spring temperature reconstruction, and by Griffin and Anchukaitis (2014) to reconstruct temperature-independent drought. We consider the cluster 2 chronologies with their positive relationship to distal snowpack and a lack of additional temperature sensitivity as a special case of *moisture-limited snow proxy*. Similar to cluster 1, a caveat of this proxy relates to potential decoupling between low-elevation rainfall and high-elevation snowfall, which can be avoided through site-level comparisons of meteorological precipitation and SWE data. In fact, figure 4 suggests these precipitation forms are exceptionally strongly correlated at QUDG sites. In the field, cluster 2 chronologies can be found at the lower forest border between high desert grasslands and mid-elevation mixed conifer forests where they are often prone to drought and wildfire (Stahle et al. 2013).

**Cluster 3** consists of chronologies negatively correlated with SWE (figure 1) and generally and in the aggregate uncorrelated with either precipitation or temperature (figure 3(c)). Year-to-year radial growth variation of these *energy-limited snow proxies* is primarily limited by deep and late-lying snowpacks that truncate the length of the growing season, resulting in a smaller annual growth increment in years with high SWE (Heikkinen 1985, Peterson and Peterson 2001). Distributed mainly at mid- to high-
elevations in the Cascade Range, British Columbia Coast Mountains, and northern Rocky Mountains (figures 2(a) and 5(c)), energy-limited snow proxies are less common on the landscape and less commonly used in dendroclimatology since they lack traditional statistical relationships to temperature and precipitation. They have similar lifespans to moisture-limited snow proxies in the WPN (figure 3(b)).

The specific physiological processes by which late-lying snowpacks inhibit spring radial tree growth are not well-known and have not been experimentally tested. Most studies refer to the early hypothesis of Kozlowski (1964) that, at sites where deep snowpacks insulate soils from freezing, those same snowpacks also maintain cool soil moisture temperatures and high soil moisture viscosity. Trees cannot uptake water through their roots and initiate hydraulic and metabolic processes until snowpacks wane to a threshold at which soil moisture temperatures increase and viscosity decreases (Kozlowski
This hypothesis is consistent with the morphology of some energy-limited snow proxy species: Mountain hemlock (TSME), for instance, are shallow-rooted and require unfrozen, snow-insolated soils more so than other species like larch (LALY), which can endure frozen soils and have deep taproots (Krajina and Peterson 1970). Peterson and Peterson (2001) also suggest increased cloudiness associated with cool springs and late-lying snowpacks could reduce solar radiation and increase the frequency of photoinhibition following cold nights. These earlier observations and the results of our analysis here both indicate that the physiological processes of deep snowpack limitation varies among species and likely site characteristics, and motivates new physiological studies to support process-based climate reconstructions.

In contrast to moisture-limited snow proxies that record a partial or full cool-season climate signature, energy-limited snow proxies record the timing of some ‘functional snow level’, or a threshold in declining seasonal snow quantity at which the
cambium activates in trees (Kirdyanov et al. 2003). This timing serves as an indirect measure of minimum cool-season SWE, with an earlier initiation of growth reflecting a lower SWE. A strength of this measure is that it accounts for both snow gains and losses over the cool season, including losses from rain on snow events and sublimation, as well as spring and summer temperature dynamics. Together, the gains and losses in snowpack are expressed in an ‘instantaneous’ snow measurement at the inception of the conifer growing season. The capacity of these proxies to effectively integrate both precipitation and temperature information is evident in the SEASCORR analysis in figures 3(c) and 3c where, though significantly correlated with instrumental April 1 SWE, they otherwise show no significant correlation with precipitation or temperature alone in any season. A closer look at the link between April 1 SWE and cool-season precipitation at each of the energy-limited snow proxy sites in figure 4 (cluster 3) suggests why: April 1 SWE and cool-season precipitation are only weakly to moderately related at these sites. In the northern Rocky Mountains this may be due to the particularly strong influence of temperature on annual snowpack dynamics (Pederson et al. 2011a), and in the Pacific Ranges due to cool-season temperatures oscillating around 0 °C, rain-on-snow events (these increase the cool season precipitation measurement but decrease the April 1 SWE measurement), and sublimation as an active process throughout the cool season (sublimation has no influence on the cool season precipitation measurement but decreases the April 1 SWE measurement) (DeWalle and Rango 2008).

Pederson et al. (2011a) found strong and linear relationships between annual SWE amount and melt timing over the northern Rocky Mountains. This link between the amount of snow in a given year and the amount of time it takes to melt suggests functional snow level, which is itself a function of melt timing, can reasonably operate as a surrogate for total cool-season SWE despite within-year and year-to-year differences in snow density and energetics (Kormos et al. 2014), canopy interception and shading (Marks et al. 1998, Storck et al. 2002), wind redistribution and loading (Pomeroy et al. 2009), forest disturbance (Harr 1986, Pomeroy et al. 2012), and the timing and magnitude of spring warming (Westerling et al. 2006) that may differently affect the rate and magnitude of snowmelt prior to the inception of tree growth. The threshold low snow depth and/or density at which trees start growing will also likely vary depending on the tree species, sample site characteristics, and other factors (Vaganov et al. 1999, Kirdyanov et al. 2003). Partly as a result of these complexities, the signal in energy-limited snow proxies is reduced relative to clusters 1 and 2, and high sample replication is important for maximizing chronology signal-to-noise ratios (Cook and Briffa 1990). Recent evidence suggests enhanced and earlier spring warming in mountain environments in recent decades may cause non-stationary SWE signals in these spring temperature sensitive proxies (Kirdyanov et al. 2003, Coulthard 2015, Kemp-Jennings 2017), making rigorous signal stability tests imperative when using them. In the field, these forest stands are typically located on north-facing slopes and/or in protected topographic features such as basins where snow can accumulate and persist. Microsite-level snow conditions often vary dramatically among neighboring trees, and individuals can be carefully selected for sampling based on vegetation assemblages and soil depressions that suggest the presence of late-lying snow over multiple years, emphasizing the importance of careful fieldwork when reconstructing snowpack dynamics using these proxies.

4.2. A data-driven approach to snow reconstruction using tree rings
An enhanced understanding of the distinct snow signals recorded by the different snow proxy types is crucial to expand and improve the reconstructions and interpretation of past western North American snowpack dynamics using tree rings. Our results suggest moisture- versus energy-limited snow proxies have different yet complimentary strengths for reconstructing dry snow droughts that are caused by a deficit of precipitation, compared with warm snow droughts that are caused by warm conditions converting the form of precipitation from snow to rain (Harpold et al. 2017). Both high- and low-elevation moisture-limited snow proxies are exceptional recorders of dry snow drought, though without careful site selection based on meteorological analysis these proxies might classify a year with substantial cool-season rainfall and very little snowpack accumulation (warm snow drought) as a high snowpack year. Meanwhile, energy-limited SWE proxies are powerful for recording both dry and warm snow droughts, and though their statistical correlations with SWE are relatively weaker, their growing season-mediated physical link between SWE and tree growth is distinct from that of trees directly reliant on snow meltwaters. Thus, used in tandem energy-limited SWE proxies can potentially serve as an additional independent check on the presence of rainfall misdiagnosis in moisture-limited snow proxies. The incorporation of temperature-sensitive tree-ring proxies in snowpack reconstructions along with the proxy types described here also has the potential to refine snowpack reconstructions even further by incorporating information related to spring and/or summer snow melt rate and timing. This approach has been harnessed in the most statistically robust snowpack reconstructions developed to date (Pederson et al. 2011b, Belmecheri et al. 2016). Overall, it is important to consider site-level timing and form of cool-season precipitation when selecting proxy sample sites, conducting correlation analyses between climate
and radial-growth, interpreting snow proxies, and developing and interpreting snowpack reconstructions. We provide a spatial roadmap of general patterns in figure 4.

Energy-limited snow proxies blend cool-season precipitation and spring/summer temperature information to such an extent that for their median composite only April 1 SWE is statistically linked to radial tree growth. For this reason, instrumental or modeled SWE data should always be used to identify energy-limited snow proxies rather than raw temperature or precipitation data alone. This blended signal is in many ways advantageous, since SWE has an important and distinct role within the changing global water cycle (Huntington 2006), hydrological modeling (McNamara 2017), and water supply management (Mankin et al 2015). Although the influence of precipitation versus temperature on radial tree growth might fluctuate from year to year, disentangling these climate drivers will matter for some research questions more than others.

Dendrochronologists have long recognized that tree-ring widths record dry extremes better than wet, an issue of optimality whereby beyond a certain quantity water is no longer the limiting factor to growth (Fritts 1976, Meko and Woodhouse 2011). Consistent with the findings of Wise and Dannenberg (2019), moisture-limited snow proxies in western North America are likely to best capture the magnitudes of low SWE years, and despite the fact they are not limited by water to grow, energy-limited chronologies seem to do the same (Goulthard and Smith 2016, Coulthard et al 2016). Our understanding of the role of snow in controlling early-season cambial activity suggests sub-annual resolution (earlywood/late-wood) and/or quantitative wood anatomy chronologies that target the early growing season when either snow meltwaters or functional snow level timing most strongly influence cambial activity (Kirdyanov et al 2003, Shamir et al 2019) have strong potential for further enhancing extreme capture in both directions.

5. Conclusions

The quality and spatial extent of snow-sensitive chronologies in the Western Paleosnow Network demonstrates there is now ample opportunity for employing both moisture- and energy-limited snow tree-ring width records to reconstruct pre-instrumental snowpack dynamics across western North America, and to address pressing questions about the pre-instrumental variability of mountain snowpacks that have thus far been unfeasible. Quantifying long-term ranges of internal and forced variability, detecting recent extremes and state-shifts that fall outside that range, and employing snowpack reconstructions as out-of-sample tests in climate models, is essential for predicting how snowpack declines will affect future human welfare and ecosystem resilience under climate change and on management-relevant timescales.

Our results indicate three distinct types of tree-ring chronologies record snowpack fluctuations in western North America: moisture-limited snow proxies that include a secondary spring temperature signal, moisture-limited snow proxies without a coherent temperature signal, and energy-limited snow proxies. Each proxy type is based on distinct physiological tree-growth mechanisms and has advantages and disadvantages as an environmental archive. A precise understanding of the radial growth and hydroclimate processes they represent is therefore essential for developing snowpack reconstructions wisely. Experimental studies could further clarify physical snow-water-tree growth relationships (e.g. snow removal experiments paired with cambial phenology and isotopic source water analysis), while sub-annual and quantitative wood anatomy chronologies focused on early-season cambial activity may be avenues for refining Snow Water Equivalent signal in each proxy type. New Snow Water Equivalent-sensitive tree-ring datasets older than 1500 AD and updated to present are needed, especially in the energy-limited zone and in the latitudinal transition from energy to moisture limitation where historical sampling effort has been lower (figure 1). Data coverage in the last thirty years will be especially important for explicitly incorporating that period of widespread snowpack deterioration within model calibrations and assessment. Temperature-sensitive tree-ring chronologies provide an additional source of information that, used in tandem with the Western Paleosnow Network network investigated here, can further enhance reconstructions of past snowpack.

Data availability statement

The data that support the findings of this study are openly available at the following URL/DOI: https://www.ncdc.noaa.gov/data-access/paleo climatology-data/datasets/tree-ring.

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