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Key Points:

- Decadal low-snow intervals, particularly during the 19th century, are part of the natural Oregon Cascade snowpack variability
- Snow "whiplash" events are a common although variable feature of the Oregon Cascades' reconstructed record
- In our 331-year record, only one other event matches the severity of the Oregon Cascades 2014–2015 snow drought

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The Severity of the 2014–2015 Snow Drought in the Oregon Cascades in a Multicentury Context

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Abstract The western United States (US) is a hotspot for snow drought. The Oregon Cascade Range is highly sensitive to warming and as a result has experienced the largest mountain snowpack losses in the western US since the mid-20th century, including a record-breaking snow drought in 2014–2015 that culminated in a state of emergency. While Oregon Cascade snowpacks serve as the state's primary water supply, short instrumental records limit water managers' ability to fully constrain long-term natural snowpack variability prior to the influence of ongoing and projected anthropogenic climate change. Here, we use annually-resolved tree-ring records to develop the first multi-century reconstruction of Oregon Cascade April 1st Snow Water Equivalent (SWE). The model explains 58% of observed snowpack variability and extends back to 1688 AD, nearly quintupling the length of the existing snowpack record. Our reconstruction suggests that only one other multiyear event in the last three centuries was as severe as the 2014–2015 snow drought. The 2015 event alone was more severe than nearly any other year in over three centuries. Extreme low-to-high snowpack "whiplash" transitions are a consistent feature throughout the reconstructed record. Multi-decadal intervals of persistent below-the-mean peak SWE are prominent features of pre-instrumental snowpack variability, but are generally absent from the instrumental period and likely not fully accounted for in modern water management. In the face of projected snow drought intensification and warming, our findings motivate adaptive management strategies that address declining snowpack and increasingly variable precipitation regimes.

1. Introduction

The western United States (US) is a global snow drought hotspot (Huning & AghaKouchak, 2020b), and has experienced significant mountain snowpack declines (~15%–30%) since the mid-1900s (Mote et al., 2018). In particular, the Cascade Range (Cascades) in the US Pacific Northwest (PNW) has undergone the most dramatic declines during the instrumental era (Mote et al., 2005, 2018), with the largest climate sensitivities and reduced snowpack predictability (Livneh & Badger, 2020). Oregon has sustained the greatest reductions (Mote, 2003; Mote et al., 2005) and contains approximately half of "at-risk" Cascades snow (Nolin & Daly, 2006). Analysis of observational data sets dating to the mid-20th century suggest a 16% loss in Cascades snowpack independent of internal variability from 1930 to 2007 (Stoelinga et al., 2010).

Oregon Cascade snowpacks act as natural reservoirs for water supply that are slowly released in the spring and summer months when demand is highest (Barnett et al., 2005; Siirila-Woodburn et al., 2021). Located adjacent to the state's largest human population centers, they supply up to 75% of annual societal, economic, agriculture, and ecosystem water demands (United States Department of Agriculture, Natural Resources Conservation Service, 2022). The potential impacts of snow drought (Harpold et al., 2017) in Oregon were brought into sharp focus during the 2014–2015 snow drought event (spanning the 2014 and 2015 water years) when near-average winter precipitation was accompanied by exceptionally warm temperatures, resulting in precipitation primarily falling as rain rather than snow (a "warm snow drought") (Cooper et al., 2016; Harpold et al., 2017; Mote et al., 2016). Along with much of the western US, 76% of Oregon's long-term snow monitoring sites reached synchronous record-breaking low April 1 Snow Water Equivalent (SWE) measurements in 2015 (Mote et al., 2016; Natural Resources Conservation Service, 2015). This produced severe and costly ecological, societal, and economic consequences throughout the state (Northwest Interagency Coordination Center, 2016; Oregon

Department of Fish and Wildlife, 2015; Oregon State University, 2018; The Washington Times, 2015). This drought culminated in Oregon's Office of the Governor declaring a drought emergency and $2.5 million in agricultural sector drought mitigation costs alone (Oregon Office of the Governor, 2015; Oregon Water Resources Department, 2015; The Oregonian, 2015).

Relative to other western US mountain ranges, Oregon Cascade snowpacks are particularly susceptible to atmospheric warming trends (Mote et al., 2005) due to mild mean winter air temperatures that are close to the freezing level (Livneh & Badger, 2020; Mote et al., 2005). Warming winter and spring temperatures, particularly since the 1970s, are likely the primary driver of observed snowpack declines (Cayan et al., 2016; McCabe & Wolock, 2009), despite increases in precipitation (Mote, 2003; Mote et al., 2005). Resulting changes in the amount, timing, and phase of winter precipitation over the past several decades have led to earlier snowmelt and peak streamflow timing (Cayan et al., 2001; Dettinger & Cayan, 1995; Fritze et al., 2011; Kapnick & Hall, 2012; Regonda et al., 2005; Stewart et al., 2005), decreases in annual streamflow (Luce & Holden, 2009), more severe forest pest and pathogen outbreaks (J. M. Vose et al., 2012; Kolb et al., 2016), and increased frequency and severity of large wildfires (Dennison et al., 2014; Littell et al., 2009; Westerling et al., 2006). The vulnerability of water supplies and ecosystems to Oregon Cascade snow droughts is exacerbated by the steep and fast-draining watersheds that characterize the range, paired with low reservoir storage capacity (as low as ∼25% of annual streamflow runoff) (Graf, 1999; Tague & Grant, 2004). Further, aging reservoir and dam infrastructure and operational guidance are not adequately designed for recent shifts in the snow-to-rain ratio, earlier snowmelt timing, and unpredictable transitions from very dry to very wet years (climate "whiplash" events) (Li et al., 2017; Siirila-Woodburn et al., 2021; Swain et al., 2018; Vicuna & Dracup, 2007). Projected declines in SWE will likely result in reduced hydroelectric generation, which is currently used to offset non-renewable energy sources and supplement renewable sources, such as solar, in the peak demand season (mid-to-late summer) (A. Marshall & Chen, 2022).

Understanding long-term Oregon Cascade snow drought variability is critical to inform mitigation and adaptation strategies aimed at lessening statewide impacts of projected snowpack decline. Observational SWE data sets are commonly leveraged to examine snow drought behavior across the western US since the 1950s (e.g., Mote, 2003, 2006; Mote et al., 2005, 2008, 2018; Fyfe et al., 2017; Harpold et al., 2012; Hatchett & McEvoy, 2018; Kapnick & Hall, 2012; Pierce et al., 2008; Regonda et al., 2005). However, short instrumental data availability makes it impossible to constrain the full range of natural snowpack variability prior to recent anthropogenic climate change (B. I. Cook et al., 2018; Huning & AghaKouchak, 2020a) or to identify possible recent departures from relatively stationary long-term hydroclimate (Milly et al., 2008). To address these challenges, we introduce the first annually-resolved, multicentennial Oregon Cascade snowpack reconstruction using tree-ring records. We leverage this new record to: (a) examine the magnitude and duration of snow droughts and snow "whiplash" events, (b) quantify intervals of persistent below-average snowpack, and (c) assess the severity of the 2-year 2014–2015 snow drought in a multicentennial context.

1.1. Study Region

The Oregon Cascades extend from the Columbia River to the northern border of California, and encompasses three arboreal Level III ecoregions: the Cascades, the Eastern Cascades Slopes and Foothills, and the Klamath Mountains (Environmental Protection Agency, 2020) (Figure 1). The range is located approximately 120 km inland with elevations as low as 750 m asl and some major volcanic peaks exceeding 3,000 m asl including the highest, Mount Hood (3,427 m asl) (Figure 1c). Winter snowpacks (Figure 1b) provide runoff that serves PNW communities' water and energy supply needs throughout the year, especially during summer and autumn. In the Willamette Valley, where over 70% of Oregon's population lives, approximately 80% of the summer streamflow used to meet municipal, agricultural, and ecosystem demands is derived from the >1,200 m asl snow zone (Brooks et al., 2012; Jaeger et al., 2017).

The hydroclimate of the Oregon Cascades (between ∼121°–123° West and ∼42°–45.5° North) is characterized by mild and wet winters with temperatures averaging approximately −2°C, and dry but moderately warm summer and fall temperatures averaging 23°C (University of California Merced Applied Climate Science Lab, 2022). Orographic lifting of moisture-rich frontal air masses derived from the Pacific Ocean and transported by the westerly midlatitude storm track, primarily during the cool-season, yields approximately 760–1,395 cm total mean annual snowfall, and varies with elevation (Stoelinga et al., 2010; Western Regional Climate Center, 2022). The dominant seasonal snowpack regime (Hatchett, 2021) is classified as maritime, with a relatively warm and deep snow cover type, higher maximum snow accumulations with shorter accumulation periods (<220 days), and an earlier onset of the snowmelt season compared to inter-mountain and continental western US snowpack

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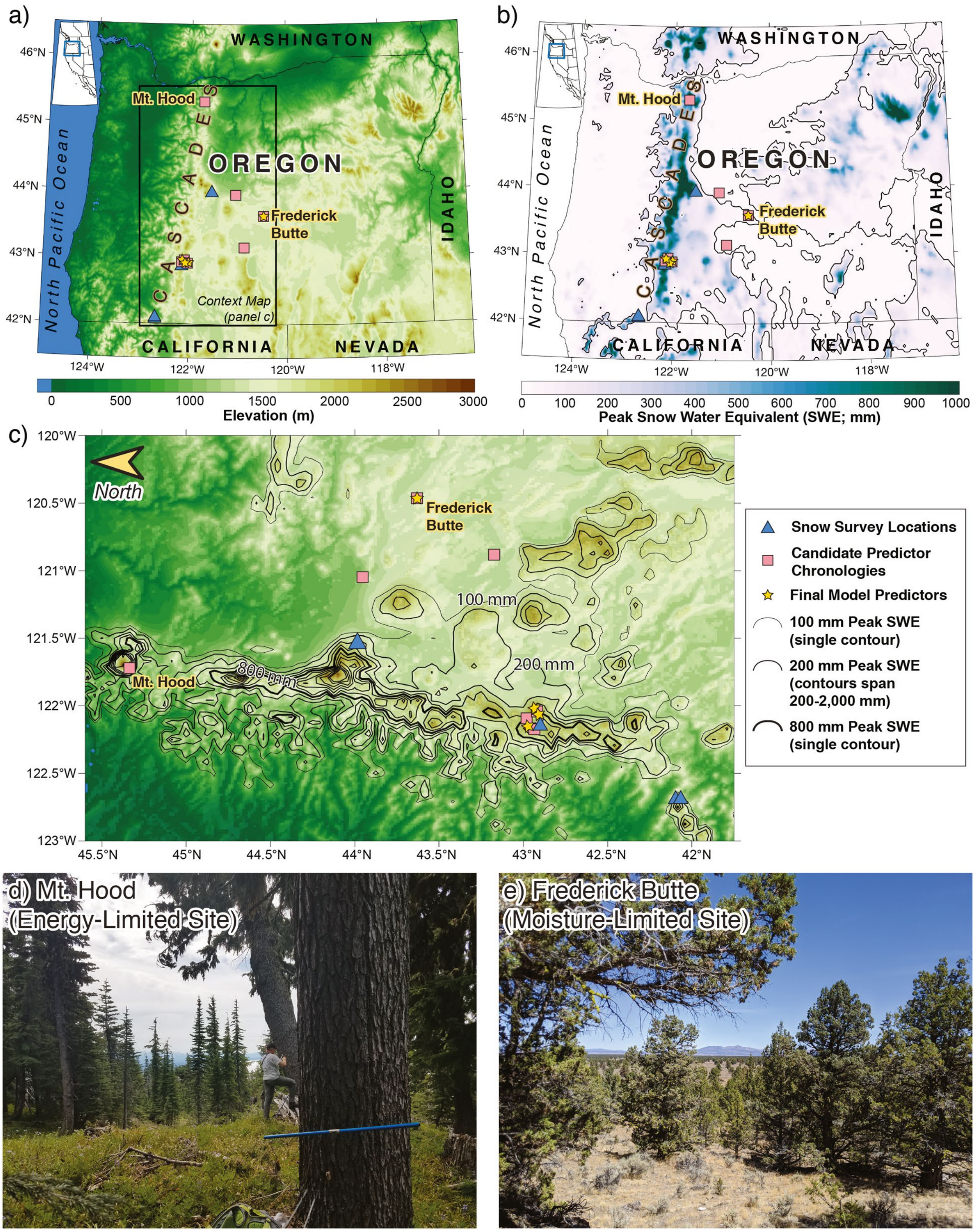

Figure 1.

regimes (Trujillo & Molotch, 2014). The Oregon Cascades were identified as highly sensitive ("at-risk"; (Nolin & Daly, 2006)) in the transition from seasonal to ephemeral snowpacks during low-snow years, as seen in water year 2015 (Hatchett, 2021). Regional April 1 snowpacks may accumulate and persist as low as 750 m asl (E. Sproles et al., 2013). These snowpacks vary semi-independently both from those in Washington (Cayan, 1996; Hatchett, Koshkin, et al., 2022; Mote et al., 2018) and also south of the Pacific dipole transition zone located between 40 and 42°N in northern California (Wise, 2010). Oregon Cascade snowpack dynamics are also influenced by internal large-scale ocean-atmosphere climate modes, as well as forced (external) mechanisms including anthropogenically-driven warming (Barnett et al., 2008; Pierce et al., 2008). Shifts in tropical convection and patterns of both tropical and midlatitude sea surface temperatures (SSTs) associated with low frequency modes of Pacific climate variability (Intergovernmental Panel on Climate Change, 2022), such as the Interdecadal Pacific Oscillation (Power et al., 1999) and the El Niño Southern Oscillation (ENSO), influence large-scale North Pacific atmospheric pressure and temperature gradients (Patricola et al., 2020). The resulting Rossby wave patterns, manifesting as the Pacific-North-America Pattern, affects the position and strength of the Pacific stormtrack and atmospheric river landfall frequencies, factors that ultimately influence the distribution and magnitude of regional cool-season precipitation (Gibson et al., 2020; Hudson et al., 2019). While El Niño (La Niña) typically results in low (high) PNW April 1 SWE and influences western North America by creating a north-south spatial dipole pattern (Cayan, 1996), Pacific decadal variability further amplifies winter ENSO conditions when they are in phase (e.g., negative phase of La Niña) (Gershunov & Barnett, 1998). The spatial pattern of ENSO-related SST anomalies (particularly during strong El Niño events) also influences the ENSO-western US precipitation teleconnection (Patricola et al., 2020). Together, these climate modes account for 61% of year-to-year variation in the western US snowpack regime (McCabe & Dettinger, 2002). Internal climate variability can offset snowpack declines (Siler et al., 2019) and also introduce large uncertainties in both short- and long-term snowpack projections even though the loss of snowpack is expected to be strongly exacerbated by the influence of anthropogenically-induced warming (Mankin et al., 2015, RCP8.5) (Rhoades et al., 2022, SSP585).

Along with much of the western US, the Oregon Cascades have experienced increases in winter temperatures (Stewart et al., 2005), winter rain-to-snow ratios (Knowles et al., 2006; Lynn et al., 2020), rain-on-snow event frequencies (McCabe et al., 2007), flood risk (Hamlet & Lettenmaier, 2007), and freezing elevations (Abatzoglou, 2011), as well as earlier average peak snowpack dates (E. Sproles et al., 2013). Rising winter and spring temperatures are primarily driving these shifts, with the greatest risk to low-elevation, ephemeral snowpack (Hatchett, 2021; Nolin & Daly, 2006). Oregon Cascade April 1 SWE is strongly influenced by winter and spring temperatures, rather than precipitation alone (Cayan, 1996). Furthermore, increasing wildfire size and severity are substantially (>200%) increasing the solar radiation absorbed by snowpack in burned forests by decreasing snow albedo and canopy interception (Gleason et al., 2013, 2019; Hatchett, Rhoades, et al., 2022; Koshkin et al., 2022). Throughout the 20th to early-21st centuries, mean annual temperatures have increased by approximately 0.8°C across the PNW, with a projected mean annual increase of 2.03°C (2.59°C) by the mid-century (2036–2065) under the RCP4.5(RCP8.5) emission scenario (R. Vose et al., 2017). Model-based projections for the Cascades estimate peak snowpack declines of 50%–70% (Siirila-Woodburn et al., 2021, incorporating RCP4.5, RCP8.5, and SRES A2 emission scenarios) with decreases in snow-derived runoff from 78% to 44% by 2100 (Li et al., 2017, RCP8.5). Cooper et al. (2016) suggests that with each 1°C warming, peak SWE declines may reach up to 30% in some Oregon Cascade watersheds, with a 28% shift in snow to rain; and under 2°C warming, snow is virtually eliminated at low- and mid-elevations. Very high-end emissions scenarios provide a signal that can help diagnose the sensitivity of the climate system to perturbations or to establish worse case scenarios; however, at the moment these (e.g., RCP8.5) are unlikely to occur (Hausfather & Peters, 2020).

1.2. Tree-Ring Based Snowpack Reconstruction in Oregon

Tree-ring records have been successfully used to estimate pre-instrumental variability in temperature (Anchukaitis & Smerdon, 2022; Esper et al., 2018), precipitation (Stahle & Cleaveland, 1994; A. Williams

Figure 1. (a) Map of the Oregon Cascade topography and tree-ring chronology network. Snow courses used in the Oregon Cascade Snow Water Equivalent (SWE) record are shown as blue triangles. Chronologies included in the reconstruction model predictor pool are pink squares. Final selected model predictors are gold stars. (b) As in (a) but showing 1982–2017 median peak snowpack based on a gridded 4 km horizontal resolution SWE product (Zeng et al., 2018). The black contour delineates seasonal from non-seasonal snowpacks (Hatchett, 2021). (c) Cascade-specific context map (from panel a) of hillshade and topography, overlain with peak SWE as contours, with 100, 200 and 800 mm threshold values. Photographs of the (d) Mt. Hood site, an energy-limited location, and (e) the Frederick Butte site, a moisture-limited location.

et al., 2021), soil moisture drought (E. R. Cook & Jacoby, 1977; E. R. Cook et al., 2004; Stockton & Meko, 1975), glacier mass balance (Pederson et al., 2004), major climate modes (E. R. Cook et al., 1998; Fritts et al., 1979; LaMarche, 1974; Lough & Fritts, 1985), and streamflow (Stockton & Jacoby, 1976; C. A. Woodhouse & Lukas, 2006). High-resolution, precisely- and annually-dated dendrohydrological reconstructions are used worldwide to supplement and extend instrumental hydroclimate data sets on timescales relevant to both water resource management and climate change mitigation and adaptation (Galelli et al., 2021; Loaiciga et al., 1993; D. M. Meko & Woodhouse, 2011; D. Meko et al., 2012; Sauchyn et al., 2015; C. Woodhouse & Lukas, 2006; C. Woodhouse et al., 2016). Even ~50 years of additional hydroclimate data can significantly alter and better inform water management-relevant statistical analyses, including more precise calculations of frequency and return periods (Huning & AghaKouchak, 2020a), and refining predictions of future hydrologic changes (Coulthard et al., 2016; Hart et al., 2010; Rodenhuis et al., 2009; Whitfield et al., 2010). A recognition of the urgent need for extended snowpack records has resulted in several tree-ring derived reconstructions directly targeting snowpack (Coulthard et al., 2021). These reconstructions exist throughout the western US (Anderson, 2010; Anderson et al., 2012; Barandiaran et al., 2017; Belmecheri et al., 2016; Brice et al., 2021; Harley et al., 2020; Lepley, 2018; Pederson et al., 2011; Timilsena & Piechota, 2008; Thornton, 2018; Touchan et al., 2021; Tunnicliff, 1975; C. A. Woodhouse, 2003). In Oregon, tree-ring derived reconstructions document ongoing hydroclimate shifts in Upper Klamath Basin water year precipitation (Malevich et al., 2013), Columbia River Basin streamflow (Littell et al., 2016), Crater Lake water levels (D. L. Peterson et al., 1999), and Central Oregon soil moisture (Pohl et al., 2002). Snowpack reconstructions for the adjacent North Cascades (Harley et al., 2020) and Sierra Nevada (Belmecheri et al., 2016) both demonstrated the magnitude of the 2014–2015 snow drought was unprecedented relative to the past several centuries. However, no such records exist for Oregon Cascade snowpacks. Thus, a regional perspective on natural and forced interannual snowpack variability is missing.

Two distinct types of tree-ring chronologies sensitive to snowpack fluctuations exist in the Oregon Cascades, allowing the development of snowpack reconstructions (Coulthard et al., 2021). *Energy-limited* chronologies are derived from high-elevation conifer species that experience deep maritime snowpacks and whose ring widths are often negatively correlated with annual maximum SWE as a result of snow truncating the length of the growing (energy) season (Heikkinen, 1985; D. W. Peterson & Peterson, 1994, 2001; D. W. Peterson et al., 2002). These late-melting snowpacks are thought to control soil temperature, the timing of soil moisture availability to tree root systems, and cambium reactivation in a variety of conifer species including Silver fir (*Abies alba*), Subalpine larch (*Larix lyallii*), and Mountain hemlock (*Tsuga mertensiana*) (Coulthard et al., 2021; Emmingham & Waring, 1977; Kirdyanov et al., 2003). *Moisture-limited* chronologies are derived from mid- to low-elevation conifer species growing in semiarid and/or rainshadow environments that primarily depend upon snowmelt-derived soil moisture for growth (Coulthard et al., 2021). In contrast with energy-limited snow proxies, annual radial growth is positively correlated with annual maximum SWE (D. W. Peterson & Peterson, 2001; D. W. Peterson et al., 2002). Responsive species include Western juniper (*Juniperus occidentalis*), Douglas-fir (*Pseudotsuga menziesii*), and Ponderosa pine (*Pinus ponderosa*).

2. Materials and Methods

2.1. Development of an Oregon Cascade Composite SWE Record

Oregon April 1 snow course SWE records (based on manual snow surveys) located west of −121° and above 900 m asl were acquired from the United States Department of Agriculture Natural Resources Conservation Service's (NRCS) National Water and Climate Center (United States Department of Agriculture, Natural Resources Conservation Service, 2022). We elected to use snow course data sets due to their earlier (~1930s) start dates, compared to NRCS' Snow Telemetry (SNOTEL) stations (records typically begin in the early ~1980s). Additionally, snow course data sets are more regionally representative, correlate with one another over larger distances, and have less short-scale variability than SNOTEL sites (Dressler et al., 2006). April 1 typically represents maximum annual snow accumulation in the Cascades (Cayan, 1996; Mote et al., 2005; Serreze et al., 1999) and a 900 m asl elevation threshold was applied in order to account for lower-elevation SWE fluctuations while not being strongly biased toward the ephemeral snow zone. The Oregon Cascades rain-snow transition zone fluctuates around 400–1,200 m asl (Marks et al., 1998; Tague & Grant, 2004; E. Sproles et al., 2013) and low- to mid-elevation stations may not accurately represent April 1 peak SWE accumulation in this region (Roth & Nolin, 2017). Finally, SNOTEL station data sets were not of sufficient duration for robust tree-ring model calibrations.

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Observational snow data presents many challenges. Given snow courses' still relatively short duration, a tendency for substantial missing data, and potential dissimilarity even between neighboring locations, we applied a series of tests to identify a robust and spatiotemporally representative SWE record for the Oregon Cascades. We retained only those snow course records that were at least 50 years in length and extended through at least 2018, that did not have any consecutive missing years, and that had no more than one missing year within any given 35 year period. Missing values in the records that were retained were replaced with the corresponding station's long-term record mean. Any record that failed a Shapiro-Wilk test for normality were omitted from further analysis (R Core Team, 2019; Royston, 1982). The resulting pool of SWE records were truncated to the length of the shortest data set and we calculated the simple arithmetic mean. This mean SWE time series (hereafter, the "Oregon Cascade SWE record") allows straightforward and quantitative interpretations, as opposed to composite SWE records derived from alternative methods (e.g., Principal Components Analysis), which result in dimensionless indices and is better suited for larger data sets.

2.2. Tree-Ring Data

The tree-ring data used in this study were derived from (a) existing and potentially energy- and moisture-limited tree-ring measurements for the state of Oregon archived at the National Oceanic and Atmospheric Administration's International Tree-Ring Data Bank (ITRDB) (National Oceanic and Atmospheric Administration, 2020), and (b) new and updated chronologies with established statistical correlations with SWE (Coulthard et al., 2021) from Mt. Hood, Oregon (*Tsuga mertensiana*, sampled in 2019) and Frederick Butte, Oregon (*Juniperus occidentalis*, sampled in 2020). At these sites, two or three cores were extracted from a minimum of 20 trees at each site using an increment borer. Cores were mounted on slotted boards and surfaced following standard dendrochronological procedures (Stokes & Smiley, 1968). Each tree-ring sequence was visually and graphically dated and then measured (including the early-wood (EW), late-wood (LW), and total ring width (RW)) to a resolution of 0.001 mm using a Velmex measuring stage and the Tellervo measurement software (Brewer, 2014). The boundary between early- and late-wood was defined as the point where >50% of the xylem cells showed a reduction of cell volume and increased lignification surrounding the cell walls (Crawford et al., 2015; Stahle et al., 2009). Crossdating accuracy was confirmed statistically using the program COFECHA (Holmes, 1983). Where raw measurement series already existed at the sample site, the updated series collected for this study were crossdated with existing measurements from the ITRDB (National Oceanic and Atmospheric Administration, 2020).

Each tree-ring measurement series was detrended by fitting a negative exponential growth curve and dividing RW measurements by the value of that curve to remove age-related (non-climatic) growth trends (E. R. Cook, 1985; E. R. Cook & Kairiukstis, 1990; Fritts, 1976; Stokes, 1996) using the Dendrochronology Program Library (dplR) package (Bunn, 2008) in the R programming language (R Core Team, 2019). A negative exponential curve was selected based on sensitivity tests (including considering stiff smoothing splines, Friedman's super smoother, and Signal-free methods), since it best preserved medium- to high-frequency variability and conservatively removed age-related growth trends without introducing any "end-effect" biases (E. R. Cook & Peters, 1997). Tree-ring measurement series for each site were then standardized as dimensionless growth indices and an arithmetic mean was used to estimate the final standard chronologies (E. R. Cook & Kairiukstis, 1990). Retaining the persistence within tree-ring data provides invaluable information to accurately estimate annual SWE conditions, therefore we used standard chronologies, rather than pre-whitened chronologies, which removes autocorrelation. Standard tree-ring series statistics were used to examine chronology quality and common signal, including the expressed population signal (EPS) (Wigley et al., 1984), series intercorrelation, average mean sensitivity, and the mean correlation coefficient (RBAR) (E. R. Cook et al., 1990). The EPS statistic is used to assess the adequacy of the sample size for capturing a hypothetical population growth signal (Wigley et al., 1984). Chronologies were truncated where EPS fell below 0.80 (Wigley et al., 1984). Furthermore, chronologies which did not extend forward in time to at least 2005 and backward in time to at least 1800 were omitted to maintain a minimum >40 years calibration period and longer segment length.

To identify tree-ring chronologies that could be used as proxies for Oregon Cascade SWE, Pearson's product moment cross-correlation tests were calculated to examine the strength of linear relationship between each chronology (including time-lagged relationships, $t - 1$, $t + 1$), and the Oregon Cascade SWE record (E. R. Cook et al., 1999). We tested the lagged relationships since tree-ring chronologies often retain temporally autocorrelated information about hydroclimate variability even after age-related growth trends are removed in the detrending process (E. R. Cook & Kairiukstis, 1990; Coulthard et al., 2020; Lepley et al., 2020). For example, a forward-lagged tree ring ($t + 1$) from 1979 may inform on SWE in year t (1980) based on autocorrelation in the climate system itself. That is, a high snowpack in 1980 may be more likely given a high snowpack in 1979 (E. R. Cook & Kairiukstis, 1990). A backward-lagged tree ring ($t - 1$) from 1981 may inform on SWE in year t (1980) based on autocorrelation in the tree biology and metabolism (e.g., carbohydrate storage), such that snowpack conditions in 1980 pre-condition growth (via carryover soil moisture) in the following year (E. R. Cook & Kairiukstis, 1990). Only chronologies that were significantly correlated ($p < 0.05$) with the Oregon Cascade SWE record were retained in the candidate predictor pool for the reconstruction.

2.3. Reconstruction Model

The candidate predictor chronologies were entered into a forward stepwise multiple linear regression model to calibrate the reconstruction model for the Oregon Cascade SWE record (E. R. Cook & Kairiukstis, 1990). We used an initial calibration period from 1966 to 2010. The best model was selected based on a suite of statistical diagnostics including the model explanatory power (R^2 and adjusted R^2 statistic), assessment of model fit (analysis of the model residuals, Durbin-Watson (DW), and variance inflation factor (VIF) statistics), the standard error (SE) of the model estimate, and an estimate of the statistical significance of the regression equation (F-statistic). We used a leave-one-out cross-validation procedure to verify that the model was statistically robust (Fritts, 1990; D. Meko, 1997; Michaelsen, 1987; Snee, 1977). In this procedure, we calculated the average error of the reconstruction's predictions for the withheld observations (root means square error of the cross-validation (RMSEv)), and a measurement of skill of the model's prediction (the reduction-of-error (RE) statistic) (Gordon, 1982). We also used a split-sample cross-validation procedure to ensure that correlations were stationary over time, and not dominated by a couple of decades. Based on this suite of diagnostics, the final model was then calibrated with the complete Oregon Cascade SWE record and used to reconstruct historical SWE over the available length of the predictor tree-ring data sets (E. R. Cook & Kairiukstis, 1990). We did not use a nesting approach, so the reconstruction is constrained by the temporal extent of the youngest (shortest) tree-ring chronology. Lastly, in order to then compare the reconstruction to the most recent observational SWE values, we used quantile mapping to correct for bias using the overlapping period of observations and reconstruction (Gudmundsson et al., 2012; Robeson, 2015; Robeson et al., 2020).

2.4. Runs Analysis and Snow “Whiplash”

To quantify the magnitudes and duration of Oregon Cascade snow droughts, we conducted a runs analysis (González & Valdés, 2003; Griffin & Anchukaitis, 2014; Salas, 1980). We first extended the bias corrected reconstruction (hereafter, the reconstruction) out to 2018 using the observed Oregon Cascade SWE record for the period 2012 to 2018 (Griffin & Anchukaitis, 2014; A. Williams et al., 2021; A. P. Williams et al., 2022). We then converted the reconstruction to anomalies by removing the long-term mean. We calculated the 10th and 90th percentiles relative to the period of record (1688–2018) to identify extreme snow droughts and pluvial events. We also compared these bottom 10th percentile years over the reconstructed record's instrumental period (1966–2018) to the Oregon Cascade SWE record's bottom 10th percentile years, over their period of common overlap (1966–2018), as a check on the ability of the reconstruction to reliably identify and quantify observed snow drought years. The magnitude and duration of every single- and multi-year run of April 1st SWE values below the 10th percentile was calculated. Although multiyear snow droughts may be punctuated by one or more near/above average snowpack years, we disallowed such interruptions by defining multi-year snow droughts as consecutive back-to-back bottom 10th percentile values. We also identified and recorded the occurrence and frequency of extreme snow “whiplash” events in our reconstruction following the method and terminology used by Swain et al. (2018), identifying these as a ≤20th percentile SWE year followed by a ≥80th percentile SWE year. Lastly, we evaluated our “whiplash” analysis using the more conservative definition used in the runs analysis, setting the thresholds to a ≤10th percentile year followed by a ≥90th percentile year.

3. Results

3.1. SWE and Tree-Ring Records

From the screening process outlined in Section 2.1, a total of 154 snow course stations which fall within the specific spatial and elevation parameters were downloaded from the NRCS (United States Department of

Table 1
Oregon Cascade Snow Course Network (United States Department of Agriculture, Natural Resources Conservation Service, 2022) Selected for the Composite Instrumental Snow Water Equivalent Record

Site ID	Site name	Elev. (m)	Latitude	Longitude	Start year
21F03	Tangent	1,667	43.99N	−121.54W	1952
22G32	Ski Bowl Road	1,850	42.07N	−122.69W	1966
22G31	Mount Ashland Switchback	1,959	42.09N	−122.69W	1966
22G05	Park H.Q. Rev	2,002	42.90N	−122.14W	1943

Agriculture, Natural Resources Conservation Service, 2022). From these, only 15 stations which had lengthy and consecutive records remained, and from those, four were normally distributed. Based on analysis of the available instrumental data from the Oregon Cascades, the Tangent (21F03), Mount Ashland Switchback (22G31), Ski Bowl Road (22G32), and Park H.q. Rev (22G05) snow courses were retained for this study (Figure 1). Each snow course was statistically significant to one another ($p < 0.0001$) and had a Pearson's correlation coefficient $r > 0.80$ with every other record. Their elevations range from 1,667 to 2,002 m asl (Table 1; Figure 1) and they cover a common data period of 1966–2018. It should be noted that the observed period of record does indeed extend beyond 1966 (often as far back as the 1930s). However, we refer to the observed period as the 1966–2018 period derived from the steps outlined in Section 2.1, which still covers >50 years of instrumental data.

Relaxing the quality control criteria outlined in Section 2.1 (to allow stations with a record of at least 35 years, had no more than two consecutive missing years, and no threshold for the number of missing years in any given 35 year period) resulted in eight stations but with a shorter period spanned by all the records (1984–2018). Over their period of overlap, our four station composite and the eight station record are highly correlated ($p < 0.00001$; $r = 0.98$). Because the four station record spans a longer period with few missing data, we elected to use this record for our reconstruction. Furthermore, the 2014–2015 snow drought is at least as severe in the eight-station record. Thus, the inferences about that event relative to the paleoclimate record are not affected by the choice of quality control thresholds. It is also important to recognize the differences in spatial scale. While this study examines snowpack trends, interannual anomalies, and snow droughts over a large (state-wide) spatial scale, the finer-scale locations (i.e., individual watersheds) have characteristics unique to them, and thus, are not entirely represented in this reconstruction.

Six updated tree-ring EW, LW, and RW chronologies were developed using new and existing samples collected at Mount Hood (ITRDB or045) and Frederick Butte (ITRDB or093). Series intercorrelation values for the updated chronologies used in this study ranged from 0.65 to 0.81, average mean sensitivity values from 0.33 to 0.47, and RBAR values from 0.32 to 0.58 (Table 2). Combined with the ITRDB (National Oceanic and Atmospheric Administration, 2020) tree-ring measurement data sets, the lengths of the final EPS-truncated chronologies extend from 530 AD to 2019 AD. Chronology screenings (see Methods) resulted in 12 chronologies that are significantly ($p < 0.05$) correlated with the Oregon Cascade SWE record, and make up the candidate predictor pool membership for the reconstruction model (Figure 1 and Table A1).

Table 2
Chronology and Time Series Information for Updated Chronologies

Chronology	Length (years)	Length truncated EPS (years)	# series	Series intercorrelation	Average mean sensitivity	Total RBAR
MHDU RW	1706–2018	1706–2018	67	0.65	0.33	0.32
MHDU EW	1745–2018	1819–2018	36	0.67	0.39	0.37
MHDU LW	1745–2018	1844–2018	36	0.40	0.36	0.18
FBU RW	870–2019	870–2019	78	0.81	0.47	0.53
FBU EW	1763–2019	1763–2019	32	0.80	0.43	0.58
FBU LW	1763–2019	1895–1969	32	0.21	0.29	0.12

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Table 3
Oregon Cascade April 1 Snow Water Equivalent Reconstruction Model Statistics, Model Predictors, and Reconstruction Period

R^2	R^2a	F	SE	RE	RMSEv	RMSE	Predictors	Calibration period	Reconstruction period
0.58	0.54	13.73	21.75	0.47	23.06	**19.81**	or099 RW	1966–2010	1688–2011
							FBU RW		
							or101 RW t-1		
							or103 RW		

Note. The root means square error statistic of the bias corrected reconstruction is identified in bold.

3.2. Oregon Cascade Snowpack Reconstruction

The Llao Rock RW (or099), Frederick Butte Update RW (FBU), Mount Scott RW (or101, t-1), and Pinnacle RW (or103) entered the stepwise calibration model as predictors for the reconstruction (Figure 1, Table 3). The final model explains 58% of regional Oregon Cascade snowpack variance (Table 3), and the final reconstruction spans the period 1688 to 2011 and captures nearly the full range of variability of the instrumental SWE record (Figure 2).

Our reconstruction model is skillful and has reasonable regression statistics (Table 3). The adjusted R^2 statistic indicates minimal variance inflation from the number of predictors, the reconstruction displays no significant trend or pattern in the residuals, and has acceptable DW, VIF, Portmanteau Q, and F test statistics. Furthermore, the residuals are not autocorrelated (Figure A1), indicating that the final reconstruction does not possess autocorrelation structure that is not also present in the instrumental target. The small SE statistic suggests minimal uncertainty of the model's predictions in the calibration period. The positive RE statistic (possible range from $-\infty$ to $+1$), suggests the reconstruction provides skillful estimates of Oregon Cascade April 1 SWE beyond the mean value. The difference between the SE and RMSEv statistics is small (1.31), reflecting a minimal difference in the magnitude of model validation and calibration prediction error (Table 3). Lastly, the RMSE (calculated from the entire period of overlap with the Oregon Cascade SWE record) of the bias corrected reconstruction is 19.81, which is a minimal increase above the RMSE of the raw reconstruction (19.22). The split-sample test yielded similar results and helped ensure correlations were stationary over time.

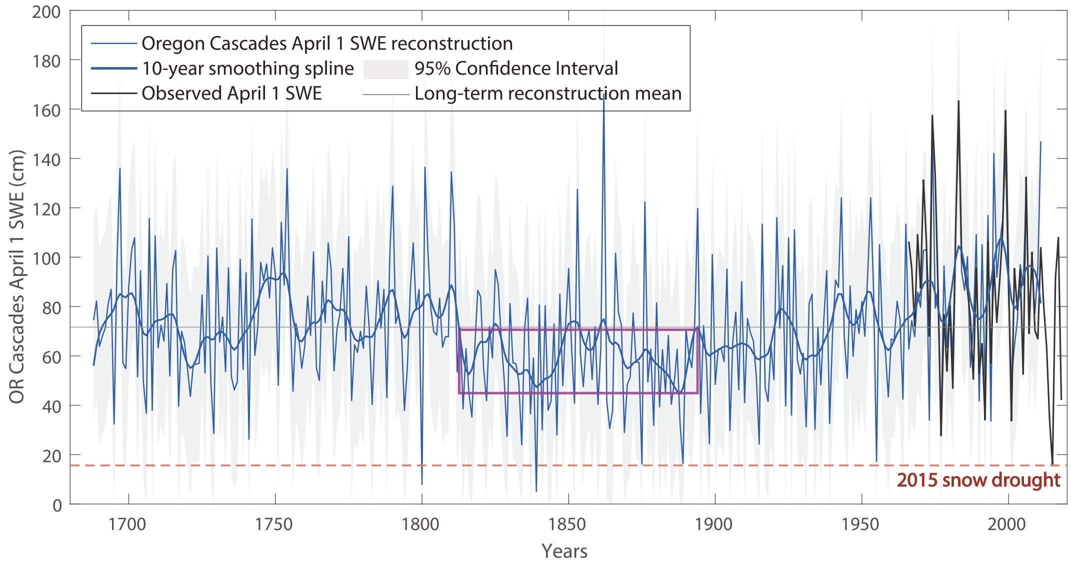

Figure 2. Time series plot of the Oregon Cascade April 1 Snow Water Equivalent (SWE) bias corrected reconstruction (blue line, 1688–2011) and Oregon Cascade April 1 SWE observed record (black line, 1966–2018). The gray shaded envelope represents 95% confidence intervals of the reconstruction, calculated from the root means square error. The reconstructed long-term mean and a 10-year smoothing spline (1688–2011) is shown by the horizontal light gray line and bold blue line, respectively. Persistent 19th-century below-the-mean SWE identified within the purple box.

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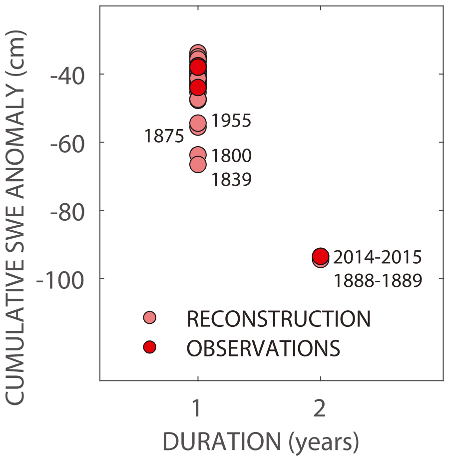

Figure 3. Runs analyses including reconstructed and observed snow drought events, defined as consecutive years falling below the 10th percentile from the long-term reconstructed mean. Snow drought event magnitudes (*y*-axis) are calculated as the cumulative Snow Water Equivalent departure during the duration of the event from the long-term reconstructed mean value, and snow drought durations are shown on the *x*-axis. Event years for the most severe 1- and 2-year events are listed next to their respective data point.

3.3. Runs Analysis and Snow "Whiplash"

The runs analysis identified 31 snow drought events from 1688 to 2018 in the bottom 10th percentile (Figure 3). Single-year snow droughts, from most to least severe, occurred in 1839, 1800, 2015, 1875, and 1889. Only two multi-year snow droughts (consecutive bottom 10th percentile years) occurred, one in 1888–1889 and the other in 2014–2015 (Figure 3). The reconstruction captures three out of six known snow drought years (1977, 2014, 2015) within the instrumental period in the lowest 10th percentile. However, it fails to capture the full severity of the 1992 (18th percentile reconstructed vs. 6th percentile observed), 2001 (23rd percentile reconstructed vs. 4th percentile observed), and 2018 (15th percentile reconstructed vs. 10th percentile observed) snow drought years. We identified 22 low-to-high snow "whiplash" events in our reconstruction (Figure 4), with the frequency of events varying between 0 and 5 per 30 year period. "Whiplash" events were absent in the early/middle 1800s, and most common (5 per 30 year period) in the late 1900s.

4. Discussion

4.1. Snowpack Reconstruction

Our final reconstruction model spans the period 1688–2011 and nearly quintuples the length of existing SWE records for the Oregon Cascades. It explains 58% of Oregon Cascade instrumental snowpack variance, which is comparable to other snow (Harley et al., 2020; Mood et al., 2020) and hydroclimate (Graumlich, 1987; Knapp et al., 2002, 2004; Littell et al., 2016; Lutz et al., 2012; Malevich et al., 2013; Pohl et al., 2002) reconstructions in the region. While the raw (prior to bias correction) reconstruction faithfully captured the direction (wet vs. dry) of observed extreme SWE values and followed closely in magnitude, it still underestimated the full variance in observed SWE values. Indeed, most

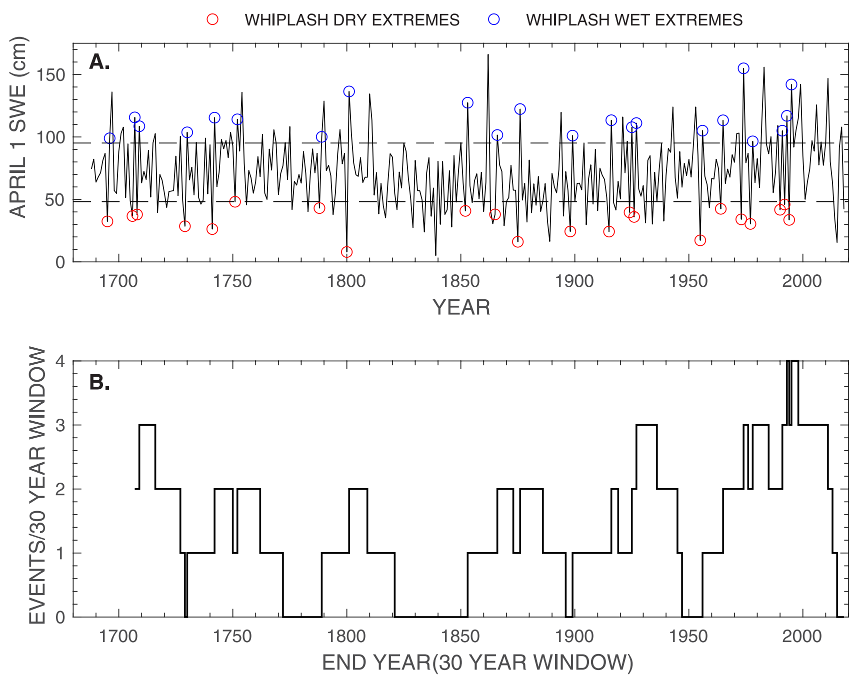

Figure 4. Snow "whiplash" analysis following Swain et al. (2018). (a) The time series of reconstructed and instrumental Snow Water Equivalent with dry-to-wet extreme events shown in red and blue circles. The dashed horizontal lines show the 20th and 80th percentile thresholds used for the analysis. (b) Frequency of "whiplash" events per 30 year window. The *x*-axis indicate the last year of the 30 year window over which the frequency of events is calculated.

tree-ring reconstructions using multiple linear regression will, by their nature, lack the full range of variance or fail to capture extreme values (Esper et al., 2005). We therefore use quantile mapping for bias correction (see Methods) in order to allow us to better compare the reconstruction directly with the observed SWE record and importantly, the 2014–2015 snow drought event.

The nearly three-and-a-half century record is dominated by medium- (decadal) to high- (annual) frequency interannual-scale shifts between high and low SWE years (Figure 2). These signals are characteristic of observed (Cayan, 1996; Losleben & Pepin, 2004; Serreze et al., 1999) and projected (Lute et al., 2015, RCP8.5) Cascade SWE dynamics and that remain most important for modern hazard management in the region. More specifically, the steep terrain and limited reservoir storage capacity in the Oregon Cascades makes this region especially vulnerable to snow droughts, such that reservoir capacity will be rapidly depleted and insufficient to supply peak warm-season demand (Graf, 1999; Raymond et al., 2014).

While tree-ring reconstructions can suffer from asymmetries in their ability to capture extremes (Wise & Dannenberg, 2019), our reconstruction reproduces the direction and general magnitude of both high and low SWE values. This may be in part due to our use of incorporating both energy- and moisture-limited tree-ring proxies in our reconstruction, rather than solely one or the other. For example, moisture-limited chronologies more accurately reflect low extremes but may experience a "saturation effect" wherein they become insensitive to additional moisture inputs in wet years (D. Meko & Graybill, 1995; Wise & Dannenberg, 2019; Yang et al., 2011). On the other hand, energy-limited chronologies may record the timing of a "functional snow level" at which the cambium activates in trees, allowing them to capture both high and low extremes (Coulthard et al., 2021). While moisture-limited chronologies alone could misdiagnose cool-season rain as snow, energy-limited chronologies should more faithfully record both warm and dry snow droughts and help overcome this potential source of error (Coulthard et al., 2021). By utilizing these two proxy types in tandem, future work could begin to disentangle the *type* of snow drought (warm vs. dry) in the pre-instrumental record. Given that the energy- (moisture-) limited trees persist at relatively high- (low-) elevations, we are able to capture spatially-integrated snowpack information across the varying elevations at which they were sampled. This is especially important for the Oregon Cascade maritime snowpack, which is sensitive to temperature fluctuations as a function of elevation (Mote, 2003; Mote et al., 2005; E. Sproles et al., 2013). In combination, our reconstruction suggests that the two proxies increase the likelihood of capturing the full range of SWE extremes (Figure 2), and should be of special consideration in future SWE reconstructions.

4.2. Snow Droughts

The multiyear 2014–2015 snow drought in the Oregon Cascades is indistinguishable within uncertainties from the 1888–1889 snow drought as the most severe event in our 331-year reconstruction, based on their combined magnitude and duration (Figure 3). The reconstructed 1888–1889 SWE value is slightly more negative than the observed 2014–2015 value, but uncertainties in the reconstruction prevent us from ranking them. This suggests that the 2014–2015 snow drought event was at the extreme negative margin of the range of natural variability of Oregon Cascade SWE. The 2015 snow drought conditions in our record were third in severity, with only 1839 and 1800 being more severe (Figures 2 and 3). However, it is likely that temperature anomalies during 2014–2015 were hotter than in the 19th century (Anchukaitis et al., 2017). This inference is strengthened by Graumlich (1987)'s reconstruction of PNW annual precipitation, which identified 1889 as the most severe drought within the 1675–1975 period. The 1888–1889 event in the Oregon Cascades was most likely caused by a deficit of snow, rather than warm temperatures and precipitation falling as rain. Nonetheless, this further emphasizes the widespread effects that warming temperatures will have on a low-to-no snow future (Rhoades et al., 2022; Siirila-Woodburn et al., 2021), compounded by increased evaporative demand (Albano et al., 2022). In the McKenzie River Basin, a major tributary to the Willamette River Basin, the 2014 and 2015 snow drought years serve as an analog for projected snow conditions under +1°C and +2°C warming, respectively (E. A. Sproles et al., 2017).

While the neighboring North Cascades (Harley et al., 2020) and Sierra Nevada (Belmecheri et al., 2016) report 2014–2015 to be the most severe snow drought in over three centuries, this is not the case for our Oregon Cascades reconstruction. Instead, the 2014–2015 Oregon Cascade snow drought is essentially tied with the only other highly unusual compound event that has occurred since 1688 (1888–1889, Figure 3). In fact, the 1888–1889 snow drought is not captured in the North Cascades at all (Harley et al., 2020) nor the Sierra Nevada (Belmecheri et al., 2016) snowpack reconstructions. This may be due in part to the differences in reconstruction length and bias correction. Here, we opted to address limitations in the reconstruction in order to confidently compare to the observed data. The Harley et al. (2020) and Belmecheri et al. (2016) reconstructions (which both extend to only 1980) rely on appending the instrumental period to the reconstruction without additional rescaling. Additionally, compared with other North American snowpack reconstructions, recent (mid-20th century to present) Oregon Cascade snowpack does not yet demonstrate a sustained or multi-decadal departure from the stationarity of pre-industrial mean snowpack variability (Harley et al., 2020; Pederson et al., 2011). Rather, discrete annual and multi-year snow droughts (Figure 2) differentiate modern Oregon snowpack variability from past centuries. Both the magnitude and duration of future snow droughts in the Oregon Cascades are however expected to worsen due to projected atmospheric warming and subsequent snowpack losses (Barnett et al., 2005; Collins et al., 2013; Fyfe et al., 2017; Gergel et al., 2017; Li et al., 2017; Lute et al., 2015; A. M. Marshall et al., 2019; Mote & Salathé, 2010; Siirila-Woodburn et al., 2021).

Other paleohydrological reconstructions indicate the severe 1839 snow drought evident in our record was spatially widespread across the broader PNW region (Graumlich, 1987) and extended into the North Cascades (Harley et al., 2020) and southwestern British Columbia (Mood et al., 2020). Pre-instrumental Oregon Cascade snow drought years also correspond with those documented in the interior PNW in 1695, 1706, 1729, 1829, 1898, 1926, 1931, 1934, 1939, and 1955, as well as the second most severe single-year snow drought in our record: 1800 (Knapp et al., 2002). Some years fall within the range of multi-year snow drought events recorded by tree rings in southwestern British Columbia (1816–1817, 1904–1905, and 1914–1915; (Mood et al., 2020)), periods of cool-season precipitation deficit in the interior PNW (1793–1795 and 1833–1834; Knapp et al. (2004)), as well as deficits in both of those respective regions during the widespread Dustbowl Drought in 1934–1935 and 1934–1937. The fact that our record reflects only a single snow drought year during these multi-year intervals is partially due to our very conservative definition of drought (10th percentile). If we instead apply a 30th percentile threshold, following the regionally-relevant snow drought work of others (e.g., Hatchett, Rhoades, et al., 2022; Huning & AghaKouchak, 2020b; Siirila-Woodburn et al., 2021), then there are 78 snow drought events over the entire period of record (1688–2018) in the Oregon Cascades, including three 3-year events, but the 1888–1889 and 2014–2015 events remain the most severe on record.

4.3. Persistent Low-Snow Intervals

Multi-decadal periods of above-the-mean SWE occurred during the mid-1700s, early 1800s, and late 1900s in the Oregon Cascades (Figure 2), consistent and coincident with high SWE periods in southwestern British Columbia (Mood et al., 2020) and high Columbia River annual streamflow (Littell et al., 2016). Relatively higher snowpacks in the North Cascades (Harley et al., 2020) and Sierra Nevada (Belmecheri et al., 2016), as well as increased precipitation in the upper Klamath basin (Malevich et al., 2013) and across the broader PNW (Graumlich, 1987), also occurred in the early 1800s. In contrast, intervals of below-average SWE persisted in the Oregon Cascades briefly during the early/mid-1700s and throughout much of the 1830s–1880s. This is consistent with Harley et al. (2020), who estimated somewhat below-average SWE during the early-to-mid 1700s in the North Cascades, which suggests regional snowpack synchronicity throughout the north and south Cascades at that time. This is further supported by Mood et al. (2020) in southwestern British Columbia. Belmecheri et al. (2016) notes above-average SWE in the Sierra Nevada during this time, which would be consistent with the position of the Pacific dipole (Hudson et al., 2019; Wise, 2010).

The 1830s–1880s were a particularly low-snow interval in the Oregon Cascades, which experienced the longest-duration of below-average conditions in our reconstruction, encompasses the most severe snow drought event (1888–1889; Figure 3) and two severe single-year snow droughts (Figure 3). This overlaps with widely-documented warming temperatures in the mid- to late-19th century across the Northern Hemisphere following the end of the Little Ice Age (Anchukaitis et al., 2017; Wilson et al., 2016). Perhaps more importantly, it coincides with sustained periods of low annual precipitation in southern and northern Oregon in the first and latter halves of the 1800s, respectively (Graumlich, 1987). Interestingly, snowpack deficits during this time are not reflected in any other major western US snowpack reconstructions (Belmecheri et al., 2016; Harley et al., 2020; Pederson et al., 2011). In fact, Harley et al. (2020) and Pederson et al. (2011) estimate higher than average snowpack during this time in the North Cascades and northern Rockies. Yet when compared with the Steinman et al. (2012) lake sediment records in northern Washington, this period of sustained below-the-mean Oregon Cascade SWE overlaps with periods of below-average winter precipitation across the PNW. Perhaps this could be a result of our tree-ring proxies being imprinted by low-frequency internal modes of climate

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variability. Nonetheless, the incongruity of this feature of sustained low Oregon Cascade snowpack with other tree-ring-derived snow paleorecords warrants further investigation.

4.4. Snow "Whiplash"

Snow "whiplash" events are a consistent feature of historical Oregon Cascade snowpack variability since 1688. However, we find that their frequency is dependent on how they are defined and the time period over which their frequency is evaluated. When using a 20th and 80th percentile threshold three nearly consecutive 1990–1991, 1992–1993, and 1994–1995 low-to-high "whiplash" events occur in the recent observed period, in part driving the highest frequency of events (5 per 30 years) in our record from approximately 1960 to 2000. However, other periods of frequent "whiplash" are also seen in the early-to-mid 1700s, the late 1800s, and the early 20th century. Using a more conservative threshold of 10th and 90th percentiles for defining these events reveals nine occurrences since the 17th century, spread throughout the length of the reconstruction. This natural feature of the Oregon Cascade hydroclimate can create unpredictable runoff timing and water storage scenarios (Vano et al., 2019) that needs to be accounted for by modern water management paradigms and infrastructure. This type of "whiplash" may also present a greater regional challenge to reservoir management than the opposite case (wet-to-dry) by prompting aggressive water storage in response to a period of shortage, and priming reservoirs for overtopping or requiring large releases in the event of extreme precipitation later. For example, in California, severe, multi-year drought conditions preceding a record wet winter—coupled with limited and aging regional dam and reservoir infrastructure (Doss-Gollin et al., 2020)—led to a combination of both severe water supply shortages followed by a near-failure of the emergency spillway and potential dam collapse from overtopping (White et al., 2019) within a single water year during the 2017 Oroville Dam Disaster.

Several other regional paleorecords have been interpreted to suggest a pattern of increased Pacific hydroclimate variability in the latter 20th century (Black et al., 2014, 2018). Yet, in regions across the western US that are sensitive in the transition from rain to snow, interannual peak SWE variability is actually expected to decline (A. M. Marshall et al., 2019, RCP8.5). Natural variability may actually be slowing the decline in western US snowpack since the 1980s, which would otherwise have substantially greater losses (Siler et al., 2019). In the North Cascades, one interpretation of tree-ring records suggest snow variability has intensified over the twentieth and twenty-first centuries relative to the prior several centuries (Harley et al., 2020), while our findings here suggest that apparent increases in frequency during the 20th century may be driven by a few events in the 1990s. Some disparity in "whiplash" interpretations may be due to a difference in the manner in which these events have been defined and calculated. For example, Harley et al. (2020) used a measure of year-to-year volatility while we adopt the "whiplash" approach used by Swain et al. (2018). Nonetheless, our findings and those of Harley et al. (2020) underscores that different modern snowpack dynamics characterize these regions of the Cascades. We emphasize that while "whiplash" events are evident throughout the record, the quantitative analysis of these extreme events is sensitive to how they are defined and tabulated, including the selection and calculation of percentile thresholds. Our reconstruction is significantly correlated with that by Harley et al. (2020) ($r = 0.42$, $p < 0.001$), but the magnitude of this correlation as well as the various distinct features discussed above across a range of temporal scales emphasizes the importance of examining the local and regional patterns when considering the impact of both forced and internal variability on snowpack dynamics (Mankin & Diffenbaugh, 2015).

5. Summary and Conclusions

We developed and analyzed the first multicentennial tree-ring snowpack reconstruction for the Oregon Cascades, which extends the regional snowpack record back in time to 1688. Our record nearly quintuples the length of existing snowpack records. Our reconstruction demonstrates that the cumulative severity of the 2014–2015 snow drought is indistinguishable within uncertainty compared to the only other multi-year snow drought event in the past 331 years: 1888–1889. The 2015 snow drought considered alone is more severe than nearly any other year in the past ~3.5 centuries, with only two years (1839 and 1800) slightly surpassing it in magnitude, and within the range of uncertainty.

We find that snow "whiplash" events are a common feature in the pre-instrumental Oregon Cascade snowpack system, suggesting that modern water resource planning should continue to anticipate these events, especially in a changing climate as occasional wet years remain possible amidst megadrought (Hatchett et al., 2016). These snow "whiplashes" are commonly characterized by the potentially problematic order of events wherein an extreme low snow year is followed by an extreme high snow year. Combined with observed western US trends of earlier snowmelt and peak streamflow timing (Fritze et al., 2011; Stewart et al., 2005), as well as increasing rain-on-snow and rain-instead-of-snow event frequencies (Lynn et al., 2020; McCabe et al., 2007), our results highlight that water management paradigms will benefit from changes aimed at increasing water availability and management flexibility (Siirila-Woodburn et al., 2021). Finally, our reconstruction also demonstrates that persistent below-average snow intervals that are not evident in instrumental records have in fact occurred in the past, including as recently as the late 1800s.

Collectively our results indicate that while the recent 2014–2015 snow drought event was the most severe in the observed period of record, it still falls within the extreme tails of the distribution of SWE reconstructed over the last three centuries. This suggests that existing instrumental snowpack records do not capture the full range and statistical distribution of natural Oregon Cascade snow drought variability. These include the variable frequency of year-to-year snow "whiplash" events and persistent multi-decadal low-snow intervals. However, we emphasize that the results presented here are indicative of large-scale, statewide snowpack trends, and will likely differ for individual watersheds as there are distinct local-scale factors which would need to be considered. Nonetheless, these aspects of local hydroclimate may ultimately challenge existing water supply and storage strategies that are based on the relatively brief instrumental record and assumptions of stationarity to inform long-term planning (Milly et al., 2008; Siirila-Woodburn et al., 2021). Given that both the magnitude and duration of PNW snow droughts are only expected to worsen in the face of a warming climate (A. M. Marshall et al., 2019; Siirila-Woodburn et al., 2021), the new insights from our study can be incorporated into regional water resource management and drought adaptation strategies to both improve water supply reliability and to mitigate flood risk.

Appendix A

See Table A1 and Figure A1.

Table A1
Candidate Predictor Pool for the Oregon Cascade Snowpack Reconstruction Model, and Their Respective Correlation Coefficients With the Oregon Cascade April 1 Snow Water Equivalent Record

Site ID	Site name	Species	r	Length (years); EPS >0.80	Latitude	Longitude
or092	Table Rock-Arrow Gap Update	JUOC	0.38	530–2010	43.17N	−120.88W
or094	Horse Ridge Update	JUOC	0.32	886–2010	43.95N	−121.05W
or098	Castle Rock	TSME	−0.50	1688–2012	42.92N	−122.05W
or099	**Llao Rock**	TSME	−0.65	1638–2012	42.97N	−122.15W
or100	Lightning Springs	TSME	−0.60	1513–2012	42.93N	−122.17W
or101	Mount Scott	TSME	−0.52	1638–2012	42.93N	−122.02W
or101 t−1	**Mount Scott**	TSME	0.39	1637–2011	42.93N	−122.02W
or102	Pumice	TSME	−0.51	1613–2012	42.98N	−122.10W
or103	**Pinnacle**	TSME	−0.61	1688–2012	42.91N	−122.07W
MHDU*	Mount Hood Update	TSME	−0.38	1706–2018	45.33N	−121.72W
FBU*	**Frederick Butte Update**	JUOC	0.48	870–2019	43.63N	−120.47W
FBUew*	Frederick Butte Update; earlywood	JUOC	0.49	1763–2019	43.63N	−120.47W

Note. Predictors used in the final model are in bold. An * indicates chronologies developed by the University of Nevada, Las Vegas Hydrology, Tree-Rings, and Climate Lab for this study.

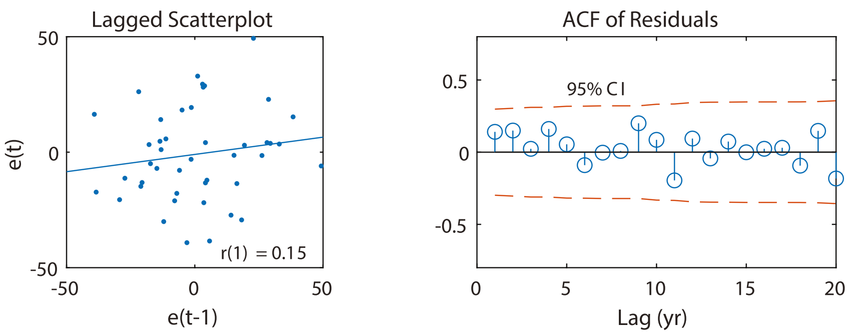

Figure A1. Lagged scatter plot of model residuals (left) and autocorrelation function plot (ACF) (right) to assess any potential trend or autocorrelation in the model residuals.

Data Availability Statement

Previously developed tree-ring data are publicly available from the National Centers for Environmental Information's International Tree-Ring Data Bank (ITRDB) repository (National Oceanic and Atmospheric Administration, 2020) at https://www.ncei.noaa.gov/products/paleoclimatology/tree-ring. The updated Mount Hood and Frederick Butte, Oregon raw ring width measurements (total, earlywood, latewood), correlation statistics, and chronology data used for the Oregon Cascade snowpack reconstruction have been uploaded to the ITRDB, and are available at https://www.ncei.noaa.gov/access/paleo-search/study/36471 for the Mt Hood Update (OR105), and at https://www.ncei.noaa.gov/access/paleo-search/study/36472 for the Frederick Butte Update (OR106). The SWE reconstruction will be archived at the time of publication at the NOAA World Data Service for Paleoclimatology at https://www.ncei.noaa.gov/products/paleoclimatology/climate-reconstruction.

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