Daily to Seasonal Moisture Signals Present in Sub-Annual Tree-Ring Data

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Daily to Seasonal Moisture Signals Present in Sub-Annual Tree-Ring Data

A dissertation submitted in partial fulfillment
of the requirements for the degree of
Doctor of Philosophy in Geosciences

by

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ABSTRACT

In recent decades, there has been an increase in the development of sub-annual earlywood (EW) and latewood (LW) width tree-ring chronologies that have been used to make long-term inferences about discrete seasonal moisture variability for different regions of North America. This doctoral research developed a new network of EW, LW, and adjusted latewood (LWa) tree-ring chronologies from the western Great Plains. These chronologies were used to reconstruct 300+ years of spring and summer moisture variability over the northern and southern Plains. The reconstructions document new information about the long-term seasonal climate history of the Great Plains, including the unusual nature of persistent spring dryness in the mid- to late-19th century and the unprecedented summer drought conditions of the 1930s. Using daily precipitation data, a regional average of four LWa chronologies from eastern Colorado was shown to be significantly correlated (r = 0.83) with the heaviest rainfall day that occurs during the wettest two weeks of year in late-July. A 238-year reconstruction estimates there has been an increase in the frequency of the most extreme events (>90th percentile), especially during the late-20th century. The heaviest midsummer rainfall days are associated with strong Canadian cold fronts and a negatively tilted upper-level ridge over the western United States that increases atmospheric moisture advection from the Gulf of Mexico and Pacific. Sixty-nine Douglas-fir and ponderosa pine LWa chronologies from the Southwest were compared with gridded daily rainfall data to objectively define the length and timing of the precipitation intervals best correlated with independent LW growth during the warm season. The majority of LWa chronologies are significantly correlated with early-season monsoon precipitation and are related to the start of the local monsoon season. Ponderosa pine is largely correlated with sub-monthly precipitation totals during the early-monsoon season across the Southwest. Douglas-fir tends to be correlated with
precipitation summed over longer intervals, and there is substantial geographic variability in terms of the timing of the best warm season precipitation signals. The intensity of the monsoon during the month of July appears to be the primary factor in determining geographically which sites contain a monsoon precipitation signal.
ACKNOWLEDGEMENTS

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DEDICATION

This dissertation is dedicated to my first born son, Finlee. You have completely changed mine and your mother’s world for the better, and I cannot imagine a life without either of you.
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Chapter 2:

CHAPTER 1

Introduction
1. INTRODUCTION

Tree-ring chronologies have been invaluable towards understanding past climate change and variability at local to global scales (Bradley 1985). Cross-dating, the hallmark of dendrochronology and tree-ring research, is based on the scientific principle that climate is the most limiting variable to tree growth within a given stand or forest. This principle has resulted in largescale development of tree-ring chronologies around the globe that have been used to produce highly-resolved interannual reconstructions of past precipitation and temperature variability that in some cases extend across millennia (Fritts 2001).

Since the early-20th century, tree-ring chronologies of total-ring width (TRW) have been the primary tree-ring proxy used in paleoclimate reconstructions. Climate-sensitive TRW chronologies reflect a growth response to local environmental conditions averaged over the months and seasons prior to, and during the entirety of the growing season. As a result, tree-ring reconstructions that incorporate TRW chronologies are often calibrated with instrumental climate variables that are also averaged across several months and seasons. These variables include the Palmer Drought Severity Index (PDSI; Cook et al. 1999; 2007); a metric that combines the effects of precipitation and temperature on drought with a statistical month-to-month autocorrelation coefficient built into its calculation (Palmer 1965), as well as annual precipitation totals (Cleaveland and Duvick 1992; Sauchyn and Skinner 2001) and average annual temperatures (Graumlich 1986; Liu 2009). Particularly for gridded regional or continental-scale reconstructions, such as the North American Drought Atlas (NADA; Cook et al. 1999, 2007) which reconstructs PDSI, integrative variables are chosen to account for the differences in regional climatology and the spatially varying seasonal climate signals inherent in the tree-ring data (St. George et al. 2014). The strong month-to-month persistence of PDSI largely smooths
out these differences, allowing for spatiotemporal comparisons of past climate variability. However, extracting these seasonal climate signals from the tree-ring data can sometimes improve the temporal resolution of tree-ring reconstructions and provide long-term estimates of sub-annual climate variability and dynamics.

The annual growth rings of some conifer and hardwood species exhibit distinctive sub-annual anatomical growth characteristics, referred to as earlywood (EW) and latewood (LW), that have been shown to contain separate seasonal climate information. A few studies from the 20th century had identified distinct seasonal moisture signals present in EW and LW width data (Douglas 1919; Schulman 1942; Cleaveland 1986), but in the last two decades there has been a rapid increase in the development of sub-annual tree-ring width chronologies over North America. Currently, the University of Arkansas Tree-ring Laboratory has obtained the data for 388 paired EW and LW chronologies for North America (Fig. 1), 22 of which were developed throughout this dissertation research (blue circles in Fig. 1). These sites are largely located in the southwestern United States and northern Mexico; a region with distinct cool and summer wet seasons that are equally important to the precipitation climatology. Earlywood chronologies of Douglass-fir (*Pseudotsuga menziesii*) and Ponderosa pine (*Pinus ponderosa*) from this region have been shown to be significantly correlated with cool season precipitation totals (Villaneuva-Diaz et al. 2007; Stahle et al. 2009), while LW chronologies are most responsive to summer precipitation associated with the North American monsoon (Meko and Baisan 2001; Therrell et al. 2002; Stahle et al. 2009; Griffin et al. 2013; Woodhouse et al. 2013). Adjusted latewood chronologies (LWa) are relatively new to the field of dendroclimatology, and represent the independent LW growth not correlated with antecedent conditions (Meko and Baisan 2001). These specialized tree growth chronologies are also correlated with summer rainfall (Stahle et al.
2009; Griffin et al. 2011), and in some cases the correlations with summer moisture can be improved compared to the unadjusted LW chronology (Stahle et al. 2009).

Climate responses analyses and separate seasonal reconstructions that utilize sub-annual tree-ring chronologies have provided long-term paleoclimate perspectives of the separate cool and warm season moisture variability over Mexico and the Southwest. These studies have also provided a useful framework for conducting similar analyses for other regions where EW and LW chronology development is possible. The seasonal moisture signals encoded in EW, LW, and LWa growth chronologies has been examined in the Rocky Mountains of southern Canada, Idaho, and Montana (Watson and Sauchyn 2001; Crawford et al. 2015), the Pacific Northwest (Dannenberg and Wise 2016), the southeastern United States (Stahle et al. 2012), and more recently the south-central United States (Torbenson and Stahle 2018). Unlike the Southwest, most of these regions lack well-defined bi-modal precipitation regimes. However, climate-responses analyses using monthly and seasonal climate data indicate at least some separation in climate signals between EW chronologies that have the highest correlation with regional early growing season moisture, and the LW and LWa climate responses which have a slightly later moisture signal in the late-spring and summer months. Substantially more work can be done to develop comprehensive networks of EW and LW chronologies that can be used in the reconstruction of local to continental-scale seasonal climate variability.

Tree-ring chronologies are typically calibrated with instrumental climate data calculated on monthly, seasonal, or annual timescales. Daily data have not been widely applied in dendroclimatic research, partly due to a lack of high-quality daily records being available in remote regions where tree-ring collections are commonly made. Additionally, given tree-ring chronologies (both annual and sub-annual) tend to be best correlated with climate variables that
reflect climate conditions averaged over several months or seasons, there has not been much utility in using daily datasets for climate-response analyses or continuous (i.e. an estimated value for each year) paleoclimate reconstructions. Daily weather records have been used to identify extreme weather events that have been shown to cause anatomical damage to growth rings. Weather extremes such as growing season freeze events (LaMarche and Hirshboeck 1984; Stahle 1990; Carolina-Barbosa 2010), springtime floods (Ballesteros-Cánovas et al. 2015; Munoz et al. 2018), and mid-growing season weather reversals (Villalba and Veblen 1996; Fritts 2001; Edmonson et al. 2010) may induce distinctive anatomical evidence in the xylem cells of living trees that can then be compared with daily weather records during the instrumental period to infer the history of these episodic events in pre-history. But these types of events and the resulting anatomical damage tend to be highly episodic and the derived paleoclimate records are discontinuous. To date, the interannual relationship between extreme weather events and tree growth has not been examined extensively.

This doctoral dissertation sought to develop a network of EW, LW, and LWa chronologies from semi-arid conifer sites located in the western Great Plains that could be compared with meteorological and seasonal moisture variables and used to make inferences about pre-instrumental climate variability at sub-annual timescales. The analyses and results using the newly developed network are presented in chapters two and three, and chapter four is a meta-analysis of the daily precipitation signals present in Douglas-fir and ponderosa pine LWa chronologies from the southwestern United States.

Chapter two, titled “Separate tree-ring reconstructions of spring and summer moisture in the northern and southern Great Plains,” describes the development of a sub-annual tree-ring network used to reconstruct the spring and summer moisture balances (i.e. Palmer’s Z-index;
Palmer 1965) for two regions of the western Great Plains. Fifteen sites that had been sampled by previous researchers, or developed by myself and Dr. Stahle, were measured for EW and LW width. Chronologies of EW, LW, and LWa width were then compared with instrumental Z-index data to determine the potential spring (March-May) and summer (June-August) moisture signals present in these chronologies. Using EW sites that were correlated with spring and those LW and LWa with summer, two 300+ year reconstructions of separate spring and summer moisture variability were produced for the northern and southern Plains. These reconstructions were compared with instrumental climate data and shown to reproduce similar large-scale ocean-atmospheric modes of variability. The most significant findings from the reconstructions are the unprecedented spring drought conditions of the mid- to late-19th century, and the severity and extensiveness of summer drought in the 1930s. This research was recently published in the journal “Climate Dynamics” on October 11th, 2018 (Howard et al. 2019).

Chapter three, titled “Tree-ring reconstructions of single day precipitation totals over eastern Colorado,” demonstrates the ability to reconstruct meteorological rainfall extremes using specialized LWa chronologies. Exploratory analysis using gridded daily precipitation data and four LWa chronologies from eastern Colorado revealed robust sub-monthly moisture signals. In fact, a regional average of these LWa chronologies is correlated with 14-day total precipitation from July 19th – August 1st at \( r = 0.78 \) over a 50-year period from 1948-1997. Identifying the single wettest day each year within this July 19th and August 1st interval, and then correlating this annual time series with the regional LWa chronology improves the correlation to \( r = 0.83 \). This suggests that it is the heavier rainfall days in midsummer that are primarily responsible for replenishing the soil moisture column and impacting late-season tree growth of ponderosa pine from the Front Range and High Plains of eastern Colorado.
A 238-year reconstruction of midsummer rainfall extremes based on the regional LWa chronology suggests there has been an increase in the frequency of the most extreme (90th percentile) one-day rainfall totals since the late-18th century. The most extreme events are associated with major synoptic-scale weather systems, featuring strong summer cold fronts and a negatively titled upper-level ridge that act to enhance southerly moisture advection from the Gulf of Mexico and Pacific. These synoptic characteristics identified with the reconstructed data are similar to the most intense single day flash floods to impact eastern Colorado, including the July 31st, 1976 Big Thompson Canyon Flood, and the July 27-28th, 1997 flooding of Spring Creek in Fort Collins. The ability to reconstruct continuous extreme precipitation events using specialized tree-ring chronologies may represent a modest paradigm shift within the field of dendroclimatology and could be applicable to other areas of North America where heavy rainfall days are the most important sources of moisture during the growing season.

The methods derived from chapter three, particularly the use of daily observations in a precipitation response interval analysis, led to a larger regional meta-analysis using this method with LWa chronologies from the southwestern United States. In chapter four, titled “The independent latewood growth response to warm season moisture over the southwestern United States,” a total of 69 Douglas-fir and ponderosa LWa chronologies were correlated with total precipitation accumulated over intervals ranging from 10-90 days for each day of the year calculated from the grid points nearest to the tree-ring sites. The results reveal interesting spatial and species-dependent differences in terms of the length and timing of the precipitation signals best correlated with each LWa chronology. Adjusted latewood chronologies of ponderosa pine, especially over the Front and Range and High Plains of Colorado and New Mexico, tend to have the highest correlations with precipitation accumulated over sub-monthly timescales during the
local early-monsoon season in mid- to late-July, while Douglas-fir has a more integrated precipitation response. Douglas-fir also exhibits much greater geographic variability in the timing of these precipitation responses, likely reflecting the geographic influence of the North American monsoon system on the summer climate of the Southwest. The connection between the start of the local monsoon season and the timing of the best precipitation responses for many of these LWa chronologies underscores the importance of using daily precipitation data. Chapter five, “Conclusions” highlights the major findings of chapters 2-4 and summarizes future research priorities.

REFERENCES


Fritts HC (2001) Tree Rings and Climate. Sci Am, 226, 92–100


APPENDIX

Fig. 1. The location of all 388 paired EW and LW chronologies currently available at the University of Arkansas Tree-Ring Laboratory. The 22 blue circles over the western Great Plains are those developed during the course of this dissertation.
CHAPTER 2

Separate tree-ring reconstructions of spring and summer moisture in the northern and southern Great Plains
ABSTRACT

The two most severe droughts to impact the Great Plains in the 20th century, the 1930s Dust Bowl and 1950s Drought, were the result of multiyear moisture deficits during the spring and especially the summer season. Tree-ring reconstructions of the Palmer Drought Severity Index indicate similar droughts in magnitude have occurred in previous centuries, but these reconstructions do not capture the potential distinct seasonal drought characteristics like those of the 1930s and 1950s. Separate tree-ring reconstructions of the spring and summer Z-index based on earlywood, latewood, and adjusted latewood width chronologies have been developed for two regions in the northern and southern Great Plains of the United States. The reconstructions extend from 1651-1990 and 1698-1990, respectively, with instrumental data added from 1991-2017. The four reconstructions explain from 39 to 56% of the variance during the 1945-1990 calibration interval and are significantly correlated with independent moisture balance observations during the 1900-1944 validation period. The reconstructions reproduce similar seasonal sea-surface temperature and 500mb geopotential height spatial correlation patterns detected with the instrumental data. The 1930s is estimated to have been the most extreme decadal summer drought to impact the two regions concurrently in the last few centuries. On average, spring moisture deficits were more severe during the multidecadal droughts of the mid-to late-19th century, but the timing of drought onset and termination differed between the study regions. In the recent two decades the spring moisture balances for the two study regions have largely been opposite, and this has been one of the most extreme periods of anti-phasing in the last few centuries. Seasonal moisture reversals are not randomly distributed in time based on the reconstructed estimates and are related to sea-surface temperature anomalies in the tropical
Pacific and to mid-tropospheric circulation changes over the North Pacific-North American sector during May and June.

1. INTRODUCTION

Precipitation during the spring (March-May) and summer (June-August) months is a vital water resource for the North American Great Plains. The spring and summer seasons account for 70% of the total annual precipitation, with 30% occurring in spring, and 40% in summer (Mock 1996; Wang and Chen 2009). Springtime precipitation can result from several different atmospheric circulation features, most prominently mid-latitude storm systems, frontal boundaries, and leeside cyclogenesis in the central and eastern Rocky Mountains (Mock 1996), During summer, deep convection, and less commonly, synoptic-scale disturbances produce a significant portion of summer rainfall (Dai 2001). The Great Plains low-level jet (GPLLJ) is an important component to both spring and summer moisture (Higgins et al. 1997). Major synoptic weather systems in spring can increase the advection of low-level moisture from the Gulf of Mexico, creating atmospheric environments that promote widespread precipitation over the Great Plains (Hirschboek 1991). Though less frequent than spring, shortwave disturbances and major frontal systems passing over the Great Plains can deliver significant amounts of moisture in summer. But deep convection is more common during the summer months when the GPLJJ reaches its maximum strength and southerly low-level moisture advection is highest over the Great Plains (Weaver and Nigam 2008). Strong southerly moisture advection can generate deep convection at local to regional scales even in the absence of major synoptic forcing (Weaver and Nigam 2011). Convective activity that forms along areas of low-level moisture convergence sometimes evolves into major mesoscale convective systems that move eastward across the
plains during the evening and nighttime hours, providing one of the most important sources of summertime precipitation to the region (Fritsch et al. 1986; Higgins et al. 1997).

The atmospheric and sea-surface temperature (SSTs) teleconnection patterns vary both regionally and seasonally across the Great Plains. The El Niño Southern Oscillation (ENSO) significantly influences precipitation and temperature over the southern and central Great Plains in the winter and spring seasons by causing perturbations in mean atmospheric circulation (Ropelewski and Halpert 1987). The tropical Pacific SST teleconnection weakens into summer, but studies have shown that positive phases of spring-summer ENSO can lead to above normal summer precipitation in the northern Great Plains (Bunkers et al. 1995). Other major ocean-atmospheric modes of variability that influence both spring and summer moisture include the Arctic Oscillation (Hu and Feng 2010); the North Atlantic Oscillation (NAO; Ruiz-Barradas and Nigam 2005), and the Pacific-North American pattern (Leathers et al. 1991). These sources of seasonal forcing primarily alter the latitudinal position and strength of the upper-level westerly jet stream (as with the PNA in spring; Leathers et al. 1991), and the advection of low-level moisture from the Gulf of Mexico (as with the AO and NAO in summer; Weaver and Nigam 2008; Hu and Feng 2010). At longer timescales [e.g. decadal to multidecadal], spring and summer moisture variability are influenced by low frequency SST fluctuations in both the Atlantic and Pacific, manifested in the Atlantic Multidecadal Oscillation (AMO; Enfield et al. 2001) and the Pacific Decadal Oscillation (PDO; Mantua and Hare 2002). The seasonal impacts of these slowly varying modes of SST variability are realized in their decadal-to-multidecadal effects on other teleconnections [e.g. ENSO] and large-scale atmospheric circulation.

The Great Plains has been impacted by a number of sustained multyear drought episodes in both the instrumental and historical period (Woodhouse and Overpeck 1998). Drought in the
Great Plains, and the associated agricultural, ecological, and socioeconomic impacts are primarily the result of severe moisture deficits that accrue during the spring and summer since these are the seasons when the bulk of the precipitation occurs (Karl et al. 1987). The two most severe and sustained droughts to impact the Great Plains in the instrumental period, the 1930s Dust Bowl and 1950s Drought, were the result of multiyear moisture deficits during the spring, but especially the summer season over the regions most impacted by drought conditions. During the 1930s over the central and Northern Great Plains, and in the 1950s over the Southern Great Plains, only modest precipitation deficits were present during the spring season (Figs. 1a and 2a). But during summer, the precipitation anomalies were much more widespread and severe (Fig. 1b and 2b). Temperature anomalies were also more extreme in summer, and these conditions acted to exacerbate drought conditions during the extreme summer droughts of 1934 and 1936 (Figs. 1c, 1d, 2c, 2d; Donat et al. 2015). The intensification of drought from spring to summer during the 1930s and 1950s is also well represented by the instrumental Palmer’s Z-index (Figs. 1e, 1f, 2e, 2f; Palmer 1965); the atmospheric moisture balance calculated from precipitation and temperature measurements but without the strong monthly persistence prescribed for the soil moisture formulation of the full Palmer Drought Severity Index (PDSI). A number of the most intense single-year summer droughts outside of the 1930s and 1950s [e.g. 1988 and 2011] also exhibited a similar intensification of dry conditions from spring to summer, potentially arising through land-surface interactions that cause persistence in atmospheric circulation anomalies across seasons (Hoerling et al. 2013)

However, spring climate conditions are typically not a reliable predictor of summer precipitation totals over the Great Plains, and there have been numerous years of opposing seasonal moisture anomalies in the instrumental record. These seasonal “moisture reversals,”
defined as below-normal spring precipitation preceding a wet summer and vice versa, are largely unpredictable on seasonal timescales (Hoerling et al. 2014) and pose numerous challenges to agricultural and water resource planning. Examples of seasonal moisture reversals include the “flash droughts” of 1980 and 2012, when near-to-above normal spring moisture and average temperatures rapidly transitioned into severe drought conditions in the late-spring and early-summer over the central and southern United States (Karl 1981; Mo and Lettenmaier 2016). Developing proxy records that can separately model spring and summer moisture variability could allow for the examination of the multi-century history of the two most important seasons to the precipitation climatology of the Great Plains, and possibly provide a longer baseline for the frequency of drought intensification events and major moisture reversals. In the Great Plains, it may be possible to reconstruct seasonal climate variability using sub-annual tree-ring proxies from the region.

Investigation of climate variability at interannual, decadal, and multidecadal timescales can be augmented with historical and paleoclimate data (Muhs and Holliday 1995; Meko and Baisan 2001; Mock 1991; Stahle et al 2009; Burnette and Stahle 2013; Griffin et al 2013). Moisture sensitive tree-ring data have provided high-resolution estimates for past droughts and pluvials at local to continental scales, including in the Great Plains. Tree-ring chronologies from the Great Plains have been used to reconstruct variables that tend to integrate climate conditions across several seasons, such as annual precipitation totals (e.g. Cleaveland and Duvick 1992; Sauchyn and Skinner 2001) or more commonly the PDSI (e.g. Stockton and Meko 1983; Stahle and Cleaveland 1988; Cook et al. 1999; Woodhouse and Brown 2001; Cook et al. 2007; St. George et al. 2009). The annual precipitation and soil moisture signals are strong in many tree-ring chronologies of total-ring width (TRW) from the Great Plains. But the integrative nature of
these variables potentially masks important seasonal climate information in the pre-instrumental record.

It is possible to separately estimate seasonal climate conditions from tree rings by using intra-annual earlywood (EW) and latewood (LW) width chronologies. Many North American tree species exhibit a distinct transition between EW (or springwood) and LW (or summerwood) portions of the annual ring, and the year-to-year growth variability in these intra-annual growth characteristics may contain useful proxy information on climate at seasonal timescales (Schulman 1942). For instance, cool season precipitation totals in Mexico and the southwestern United States have been reconstructed from EW width chronologies of Douglas-fir and ponderosa pine (Cleaveland et al. 2003; Villanueva-Diaz et al. 2007, Stahle et al. 2009). Latewood and the so-called adjusted latewood (LWa) chronologies from these regions tend to be correlated with summer moisture and have been used to reconstruct precipitation totals associated with the North American Monsoon (Meko and Baisan 2001; Therrell et al. 2002; Stahle et al. 2009; Faulstich et al. 2012; Griffin et al. 2013; Woodhouse et al. 2013). Torbenson and Stahle (2018) recently produced two separate reconstructions of May self-calibrating PDSI and the summer Z-index for the south-central United States from TRW, EW, and LW chronologies to examine the interannual to multidecadal relationship between cool season soil moisture and the summer moisture balance. Earlywood and LW chronologies have also been used to estimate more integrative moisture variables like PDSI in east-central Mexico (Burns et al. 2014) and annual precipitation totals in the southern Canadian Cordillera (Watson and Luckman 2004). Several studies have also investigated the seasonal climate response of EW and LW chronologies from Douglas-fir and ponderosa pine sites in the Pacific Northwest and interior Rocky Mountains (Watson and Luckman 2002; Crawford et al. 2015; Dannenberg and Wise...
2016). Earlywood and LW width chronologies have yet to be developed for the Great Plains, therefore the potential for seasonal moisture reconstructions has yet to be examined.

In this paper, we develop separate spring and summer reconstructions of Palmer’s Z-index based on a new network of EW, LW, and LWa chronologies for two sub-regions located in the western Great Plains of the United States. The Northern Plains study area includes western North and South Dakota, eastern Montana, and eastern Wyoming, along with EW and LW width chronologies from the vicinity. The Southern Plains study region encompasses areas of southeastern Colorado, western Kansas, northeastern New Mexico, and the panhandles of Oklahoma and Texas, with EW and LW width chronologies developed from escarpment woodlands in this region. The Z-index reconstructions are used to examine the history of both spring and summer moisture variability in the Northern and Southern Plains study areas for the past 300+ years, to investigate the persistence and reversals of moisture conditions from spring to summer, and to explore the possible influence of large-scale ocean-atmospheric variability on changes in spring to summer moisture.

2. DATA AND METHODS

2.1 Instrumental climate data

We used Palmer’s (1965) original formulation of the Z-Index to represent the ‘discrete’ non-overlapping spring and summer moisture balance. The PDSI is first calculated by computing monthly soil moisture departures based on the supply and demand of water at the surface along with local climate conditions (Palmer 1965; Karl 1986; Feng et al., 2017). These monthly values, called the Z-index (or the monthly moisture anomaly index), represent short-term moisture fluctuations sensitive to deficiencies and excesses on monthly timescales. Unlike the PDSI, the
Z-index does not have the statistical autocorrelation coefficient built into its calculation. The Z-index (Palmer 1965) was chosen to represent spring and summer climate variability because drought is often a combination of precipitation and temperature departures, both of which can impact tree growth (Fritts 1965). The Z-index was calculated from gridded precipitation and temperature data obtained from the 4km resolution Parameter-elevation Regression on Independent Slopes Model (PRISM) dataset for the period 1895-2015 (Daly et al. 1994) and then re-gridded to 0.5° resolution.

Monthly precipitation, temperature, and Z-index data were averaged into the spring (March-May) and summer (June-August) seasons. These seasons were analyzed given their importance to the region’s annual rainfall climatology and their high interannual variability (Mock 1996; Seager et al. 2005). We also hypothesized that EW tree growth in this region is best correlated with spring moisture, and LW growth is most responsive to summer rainfall. Gridded monthly SST (Kaplan et al. 1997) and 500mb geopotential height anomaly data from the 20th Century Reanalysis Project V2 provided by the NOAA/OAR/ESRL (Compo et al. 2011) were used to identify the large-scale SST and atmospheric circulation influences on the observed and reconstructed seasonal moisture balances for the two regions. We calculated gridded SST and 500mb height anomalies relative to 1951-1980 climatology.

2.2. EW and LW chronology development

Earlywood and LW width tree-ring chronologies were developed using samples of ponderosa pine (*Pinus ponderosa*) and Douglas-fir (*Pseudotsuga menziesii*) collected from sites located in the western Great Plains. The eastern range of these species in the United States extends east of the Rocky Mountains into the western Great Plains, with extensive stands present
in the Black Hills of South Dakota and eastern Wyoming, and more isolated populations in eastern Montana and western North Dakota (Wells 1965; Little 1971). Southward in eastern Colorado and New Mexico, ponderosa pine stands typically occupy higher elevation sites on isolated bluffs, escarpments, and mesas (Woodhouse and Brown 2001). Populations of Douglas-fir are rare in the western Great Plains, but a stand located on an isolated bluff in the Black Forest region of east-central Colorado was identified and sampled by Woodhouse and Brown (2001). Previous investigators have sampled many of these western Great Plains’ sites and produced chronologies of TRW (e.g. Stockton and Meko 1983; Sieg et al. 1996; Woodhouse and Brown 2001). We obtained the samples from 13 of these sites from the University of Arizona’s Laboratory of Tree-Ring Research archives. We made additional collections at one new site (Sierra Grande) and resampled at another (Kenton) in northeastern New Mexico in the spring of 2015. The 15 sites are clustered into two regions of the western Great Plains, a northern network in the Dakota states and Wyoming, and a southern network in eastern Colorado and New Mexico (Fig. 3; Table 1). Each collection is composed of 15-85 increment core specimens and/or cross-sections from living or dead trees, and the annual rings were dated using dendrochronological methods (Stokes and Smiley 1996).

We implemented the techniques outlined by Stahle et al. (2009) to re-measure each sample for EW and LW width. Chronologies were computed using the signal free method of ring-width standardization (Melvin and Briffa 2008; Cook et al. 2014). Signal free detrending preserves high-to-medium frequency variance by iteratively dividing the long-term growth curve into the original measurement data until the common signals inherent in the individual series are maximized. The data were power transformed and detrended with an age-dependent cubic smoothing spline. Ring-width indices were computed as residuals from the fitted curve, and then
averaged into the mean index chronology using the biweight robust mean (Cook 1985; Hoaglin et al. 2000). Adjusted LW chronologies were also computed using the procedures described by Meko and Baisan (2001). LW growth tends to be highly correlated with the antecedent EW due in large part to biological persistence. Meko and Baisan (2001) used a bivariate regression model to remove this seasonal growth dependency, where the LW predictor chronology was regressed on the EW predictand chronology. The residuals from this regression are the annual values of the adjusted LW chronology (i.e., LWa) that can be subsequently used as a potential predictor of summer moisture independent of the EW signal (Griffin et al. 2011).

2.3. Study regions and seasonal climate response

The 15 available EW chronologies were correlated with the spring Z-index at every grid point in the United States from 1895-1979, and the 15 LW chronologies were correlated with summer (supplemental Figs. 1-4). The 1895-1979 interval represents the common overlap between instrumental data and the tree-ring site with the earliest end date of 1979 (at Teapot Dome, WY; Table 1a). Based on the average spatial correlation patterns, we determined the sites from the northern network are best correlated with a region of the Northern Plains defined by the coordinates 43°-47°N, 107°-101°W, and the shared region of highest correlation for sites in the southern network is defined by the coordinates 35.5°-38.5°N, 105°-99°W. We calculated regional averages of the seasonal Z-indices, and all 45 EW, LW, and LWa chronologies were then individually screened for correlation with the respective regional spring and summer moisture balances. Only EW chronologies were considered as potential predictors for the spring Z-index, and LW and LWa chronologies for the summer Z-index, in order to produce seasonal
estimates that exhibited a similar correlation between spring and summer seen in the
instrumental data for each region.

Based on the correlation analyses with the regionally averaged spring Z-index, we
selected three EW chronologies from the Northern Plains and nine from the Southern Plains as
potential predictors (bolded EW chronologies in Tables 2a,2b). These chronologies were selected
because they contained a significant correlation with the spring Z-index, and the correlation
coefficient was higher compared to summer. The same selection criterion was used for the LW
and LWa chronologies for the summer Z-index (bolded LW and LWa chronologies in Tables 2a
and 2b). The LW and some of the LWa chronologies tend to be best correlated with summer
moisture. However, if the correlation coefficient of the LWa chronology was not significantly
different than the LW at a given site (based on Fisher’s r to z transformation test), the LWa was
selected as the potential predictor due to its independent variability from the EW.

2.4 Regression modeling

Initially, all the potential predictors of the respective seasonal climate variable were
submitted to a principal component regression (PCR) scheme [Cook et al. 1999; Burnette and
Stahle 2010 (as recently modified by D. Burnette, personal communication, 2017)]. Principal
component analysis (PCA; Jolliffe 2002) was computed on the selected predictor chronologies to
identify modes that account for the most variance in the array. Bivariate or multivariate
regression models were then used to calibrate the eigenvector amplitude time series with the
instrumental Z-indices in each region over the common interval of 1945-1990. Following the
initial PCR for each of the four reconstructions, we experimented with removing chronologies
from the predictor pool that had the lowest correlation with the climate variable and recomputed
the regression models. This was done multiple times until the most robust models were produced based on the minimum Akaike information criterion (Aikake 1974).

The reconstructed estimates were verified on independent instrumental data for the period 1900-1944. Standard regression and statistical tests were used to assess the agreement between the reconstructed estimates and instrumental data at interannual time scales, including the explained variance (adjusted $R^2$) in the calibration period, and the Pearson’s correlation coefficient, the sign hit/miss test, the reduction of error (RE), and coefficient of efficiency (CE) in the validation period (Cook and Kairiukstis 1990; Cook et al. 1999; Fritts 2001). Spectral coherence analysis (Percival and Constantine 2006) was used to estimate how well the reconstructed estimates agree with the instrumental data at frequencies ranging from interannual to multidecadal. Early instrumental data from the 19th century were also used as independent validation of the spring and summer moisture balance estimates. Mock (1991) compiled 19th century weather data and computed seasonal precipitation percentiles for eight regions of the Great Plains, two of which were similar to the reconstructed regions in this study. The annual values from the seasonal time series plots from Mock’s (1991) analysis were determined visually (Appendix A), and then tested for correlation with the respective seasonal reconstruction during the common overlap periods. The instrumental variance lost in regression was restored to each reconstruction so that the instrumental data could be used to extend the records to 2017. However, restoration of instrumental variance can be considered a trade-off, given the error in tree-ring reconstructions tend to be amplified.
2.5 Analyses of the reconstructed data

In order to assess the ability of the seasonal tree-ring reconstructions to reproduce the largescale ocean-atmospheric teleconnections seen in the instrumental data, the instrumental and reconstructed Z-indices were correlated with the gridded SST and 500mb height fields during the calibration interval 1945-1990. Because SSTs tend to persist across months and seasons, the spring reconstructions were correlated with the December-May (DJFMAM) SSTs, and the summer reconstructions were correlated with March-August (MAMJJA). The spring (MAM) 500mb geopotential height data were correlated with the spring reconstructions, and the same was done for summer. The significance levels of the correlations at each grid point were calculated after accounting for the potential reduced degrees of freedom due to autocorrelation in either the gridded variable or the Z-index data (Ebisuzaki 1997).

We normalized the four reconstructions (mean = 0.0 and standard deviation = 1.0) to compare across seasons and regions. Time series differences between the spring and summer reconstructions were calculated by subtracting the summer Z-index value from spring. A positive value indicates the Z-index value was higher for spring and vice versa. The regional difference series were then correlated with DJFMAM SST data to identify the potential teleconnections related to differences between spring and summer moisture for the full period 1856-1990. The 135 year period was chosen to provide the longest possible assessment of SST influence related to seasonal moisture differences.

Seasonal moisture reversals were defined by identifying years when the spring and summer Z-index values contained the opposite sign. The largest sign reversals were defined as years when the spring and summer Z-index values were >±0.5 standard deviations. This was done with both instrumental and reconstructed data from 1900-1990 to assess how well the
estimates model the largest seasonal moisture reversals. After normalization, 0.5 standard deviations equate to approximately incipient wet or dry conditions based on Palmer’s (1965) scale. The time intervals between years of the same seasonal moisture reversal type [e.g. dry spring to wet summer or wet spring to dry summer] were calculated, along with their frequency distribution. The potential non-randomness of the time interval frequency distribution was then tested using the Lilliesfors test (Conover 1980; Cleaveland and Stahle 1989). Composite analysis of 500mb height anomalies for sign changes in spring and summer moisture that exceeded $\pm 0.25$ standard deviations in instrumental and reconstructed data were analyzed for the period 1900-1990. The lower 0.25 standard deviation threshold was used simply to increase the sample size of seasonal changes in the instrumental period. The 500mb circulation anomalies for May represented spring, and June for summer because these months often had the most dramatic and significant changes in 500mb heights related to seasonal moisture reversals.

3. RESULTS AND DISCUSSION

3.1 Calibration and validation statistics of the regression models

The regression models used to reconstruct the spring and summer moisture balances are presented in Table 3, and the instrumental and reconstructed time series during the calibration and validation periods along with squared coherence plots are shown in Figure 4. The tree-ring data calibrate 39 to 56% of the instrumental Z-index variance and perform well against the instrumental data during the independent 1900-1944 validation period (Table 3). Spectral coherence analysis indicates that all four reconstructions share significant common variability at p < 0.10 with the respective instrumental data from interannual to multidecadal frequencies (Figs. 4b,d,f,h). The results from the sign hit/miss tests indicate all reconstructions significantly
(p < 0.05) reproduce the correct sign of the seasonal value in the validation period (Table 3). The correlations between the instrumental spring and summer moisture balances from 1900 to 1990 are \( r = 0.47 \) for the Northern Plains and \( r = 0.33 \) for the Southern Plains, similar to the reconstructed data (\( r = 0.52 \) and \( r = 0.32 \)). For the full reconstruction periods, the correlation between the seasonal estimates are 0.53 from 1651-2017 and 0.33 from 1698-2017 for the Northern and Southern Plains, respectively.

The correlations with the independent 19th century weather data from Mock’s (1991) seasonal precipitation data (Appendix A) also provide some additional validation of the seasonal moisture balance estimates (Table 4). The correlations with the respective seasonal moisture variable are all positive, with an \( r \)-value as high as 0.77 between spring moisture variables from the Southern Plains. The overall higher correlations for the Southern Plains may reflect the larger number of 19th century weather stations in this region (Mock 1991), and the stronger calibration and validation statistics achieved with the Southern Plains’ reconstructions.

3.2. Ocean-atmospheric forcing of spring and summer climate in the Northern and Southern Plains

Correlation analyses with gridded SST data illustrate the regional and seasonal differences in large-scale SST teleconnection patterns over the Northern and Southern Plains (Fig. 5). The spring \( Z \)-index for the Northern Plains does not have a strong SST teleconnection signal (Figs. 5a and 5b), but the summer season is positively correlated with an ENSO-like pattern in the tropical Pacific and with SSTs in the Gulf of Alaska (Figs. 5c,5d). The positive correlations between summer moisture in the Northern Plains and spring-summer ENSO is a finding previously reached by Bunkers et al. (1996). Mechanistically, this relationship results
from ENSO’s effect on the strength and positioning of the subtropical Bermuda high in the
Atlantic and the summertime GPLLJ. Positive phases of ENSO during summer tend to enhance
low-level moisture advection over the Great Plains due to SST and sea-level pressure gradients
that develop between the tropical Pacific and subtropical Atlantic (Krishnamurthy et al. 2015).
Both seasons also appear to be negatively correlated with North Atlantic SSTs, with the strongest
signal present in summer (Figs. 5a-d).

Winter-spring ENSO is highly correlated with the spring Z-index for the Southern Plains
(Figs. 5e,5f), with correlation coefficients reaching as high as $r = 0.65$ at some grid points in the
tropical Pacific based on the instrumental data. It is interesting to note that the spring moisture
balances for both regions are also modestly and oppositely correlated with SSTs in the Gulf of
Mexico, likely reflecting the dipole relationship with the GPLLJ. The positive (negative)
correlations over the Northern Plains (Southern Plains) suggest warmer-than-normal SSTs results
in a stronger GPLLJ and further northward transport of low-level moisture, generating higher
precipitation totals over the Northern Plains but deficits to the south (Weaver and Nigam 2008).
The summer Z-index for the Southern Plains is positively correlated with SSTs over much of the
central and eastern Pacific basin, with the strongest signal present near the west coast of North
America (Figs. 5g,h). Other than the Gulf of Mexico, the Atlantic teleconnections associated
with spring and summer moisture over the Southern Plains are relatively weak (Figs. 5e-h).

The spatial patterns of correlation for the Northern Plains’ spring Z-index and the gridded
500mb geopotential height data resembles the negative phase of the PNA (Leathers et al. 1991),
with modest negative correlations that extend across the western United States (Figs. 6a and 6b).
The typical configuration of upper-level atmospheric circulation over North America during
negative phases of the PNA includes a large-scale trough centered over the northern United
States and southern Canada, leading to zonal flow and a more active storm track over the Northern Plains (Leathers et al. 1991). Negative phases of the PNA have also been shown to enhance the GPLLLJ during the warm season months, and these combined upper-level and low-level circulation features have produced some of the wettest precipitation events on record over the north-central United States (Harding and Snyder 2015). A similar pattern of negative correlations extending from the central Pacific into the western U.S. is evident based on the correlations with the summer Z-index for the Northern Plains, along with a region of positive correlations over the Gulf of Alaska and Southern Great Plains (Fig. 6c,d). The summertime pattern once again indicates zonal flow over the northern U.S. and southern Canada directs more shortwave disturbances over the region, leading wetter summers over the Northern Plains.

The spatial patterns of correlation based on the spring Z-index for the Southern Plains are consistent with the expected upper-level atmospheric circulation anomalies associated with ENSO. During positive phases of ENSO, a recurrent upper-level trough and active subtropical jet stream resides over the southern United States and northern Mexico (areas of negative correlation in Figs. 6e and 6f), increasing the frequency of Pacific storms and advection of subtropical moisture over the Southern Plains (Horel and Wallace 1982). The 500mb height patterns for the summer Z-index resemble an amplified ridge-trough pattern over the United States, which would result in northwest flow aloft over the study region and a greater frequency of storm systems moving southward out of Canada over the Southern Plains during the summer months (Figs. 6g, h). These SST and 500mb correlation patterns detected in the reconstructions are remarkably similar to the instrumental data, and indicate these seasonal estimates can reproduce the major ocean-atmospheric modes of variability associated with independent spring and summer moisture over the Northern and Southern Plains.
3.3. The reconstructed spring and summer moisture balances

The four reconstructions presented here offer new insight into the seasonal and spatial characteristics of major pre-instrumental era droughts, and provide a long-term seasonal context for dry conditions in the 1930s and 1950s. The normalized reconstructed spring and summer moisture balances for the two regions are plotted in Figure 7, and the instrumental Z-indices smoothed with a 10-year cubic spline are also plotted from 1895-2017 to illustrate that the seasonal estimates largely track the decadal variability of the instrumental data. The reconstructions indicate that the 1930s Dust Bowl Drought represents one of the few periods when sustained summer dryness impacted both study regions. The estimated values based on the decadal splines indicate that the 1930s Dust Bowl was the worst decade of summer drought to impact the two regions concurrently in the last 300 years (Figs. 7b,7d). From 1931-1940, the summer Z-index values are estimated to have been below normal in the Northern and Southern Plains seven and eight out of the ten years, respectively. The 1930s decade also contains the highest frequency of summer drought years shared between the two regions for any 10-year period over the common 1698-2017 interval. However, spring drought during the 1930s was not as exceptional compared to summer, particularly over the Northern Plains (Fig. 7a,7c). Burnette and Stahle (2013) also noted the unprecedented nature of the summer Dust Bowl Drought based on a 159-year record of July-August precipitation totals from weather stations in eastern Kansas and Missouri. Yet, when precipitation is averaged across the April-August growing season, the decadal moisture anomalies of the 1930s are not substantially more severe than other identified droughts in the 19th century. Similar results are evident based on the reconstructions for the Northern and Southern Plains and highlight the distinct seasonal nature of growing season moisture conditions during the worst drought to impact the United States in the modern era.
Periods when both regions in either season were impacted by sustained droughts do not occur frequently, particularly during summer. Table 5 lists the ten driest non-overlapping decades for each season and the two regions based on a simple ten-year average. Five of the decades listed as the driest based on the spring estimates for the Northern Plains are also listed for the Southern Plains (1810s, 1860s, 1870s, 1930s, 1950s), but only three decades are shared between the summer reconstructions (1780s, 1860s, and 1930s). Worth noting, seven of the driest decades for spring are also listed for summer in both regions, suggesting a seasonal persistence in moisture balance conditions during these decadal moisture excursions.

At interannual timescales, the reconstructed spring and summer Z-indices for the Northern Plains are not correlated with the Southern Plains ($r = 0.08$ and $r = 0.11$ for spring and summer, respectively). The lack of coherence between study regions is not necessarily surprising given the different seasonal teleconnection patterns illustrated in Figures 5 and 6. In fact, there have been several periods when seasonal conditions between the two regions exhibit opposite behavior, including in the recent 21st century during the spring months. Since 2003, there have been six springs with normalized values greater than 1.0 standard deviation in the Northern Plains that have co-occurred with drought conditions ($< -1.0$ standard deviation) to the south. The recent spring of 2011 ranks as the wettest year in the Northern Plains over the full 1651-2017 period (Fig. 7a), and this is also one of the driest springs on record for the Southern Plains (Fig. 7c). The diverging moisture balance anomalies of the early-21st century between the study regions can possibly be attributed to the changing characteristics of the low-level jet and synoptic circulation over North America. Barandiaran et al. (2013) noted that a significant trend in the strength of the low-level jet has led to precipitation changes across the Great Plains in recent decades, with increases in the Northern Great Plains but decreases over the Southern Great
Plains especially over Oklahoma and Texas. These changes have also coincided with a northward shift of the average springtime position of the upper-level jet stream (Wang et al. 2013), which has also contributed to these diverging spatial patterns of moisture. A 20-year running correlation between the spring Z-index data from 1698-2017 suggests significant periods of anti-phasing during the spring season have occurred and been greater in magnitude, but the last 20 years has been one of the most extreme (r = -0.36 from 1998-2017; not shown).

The longest sustained periods of dual-season drought to occur in either region were during the mid- and late-19th century (Fig. 7). Each pair of reconstructions are plotted consecutively from 1840-1900 and 1930-1960, so that estimates of the spring Z-index are followed by the estimated summer Z-index for the same year to provide the most detailed comparison of spring and summer drought conditions during the 19th and 20th centuries (Fig. 8). The duration and persistence of dry conditions in the mid- and late-19th century do not have clear analogs in the instrumental record, especially as it relates to the spring season. Spring and summer drought estimated for the Northern Plains persisted across the two seasons beginning in the spring of 1859, and no positive Z-index value for either season is estimated until the spring of 1878 (Fig. 8a). Conditions improved in the 1880s, but spring and primarily summer drought returned in the 1890s and persisted until the beginning of the 20th century. Drought onset occurred much earlier over the Southern Plains, approximately beginning in 1841 with few years of alleviation until 1865 (Fig. 8b).

While the 1930s Dust Bowl and 1950s drought had numerous years when above-normal or moderately dry springs preceded severe summer drought (Figs. 8c,8d), drought years in the 19th century were often more severe during the spring season. The average seasonal Z-index values for the major 19th century drought intervals are substantially lower in spring for both
regions (Fig. 8a,8b), compared to the lower values in summer during the 1930s and 1950s (Fig. 8c,8d). Sixteen out of the 19 estimated dual-season drought years from 1859-1877 were drier during the spring season over the Northern Plains (Fig. 8a), and of the 16 dual-season drought years from 1841-1865 estimated for the Southern Plains, ten were more severe in spring (Fig. 8b).

The few data sources from the Global Historical Climatology Network (GHCN) also seem to indicate that moisture deficits in the central United States were more severe in the spring season during the droughts of the mid- to late- 19th century (Herweijer et al. 2006). Previous studies have documented that La Niña conditions persisted for multiple consecutive years between the 1840s and 1860s (Cole et al. 2002), providing one explanation for the more frequent and intense spring drought years over the Southern Plains from 1841-1865. However, the weak winter-spring ENSO signal in the Northern Plains (Figs. 5a,5b) suggests drought from 1859-1878 was likely a separate event caused by other mechanisms, perhaps related to random atmospheric variability (Hoerling et al. 2009). Multidecadal oscillations in Atlantic and Pacific SSTs may also have influenced spring and summer drought conditions in the 19th century by affecting the positioning and strength of the GPLJJ (Weaver and Nigam 2008). These characteristics of seasonal drought evolution may reflect in part differences in ocean-atmospheric forcing and the influence of internal atmospheric variability, but perhaps the added anthropogenic land degradation component amplified summer drought conditions in the 20th century (Seager et al. 2005; Cook et al. 2009).
3.4. *Spring to summer moisture changes*

The reconstructions provide an extended proxy record of the frequency and temporal distribution of spring to summer moisture changes, and the potential ocean-atmospheric forcing of these seasonal differences. Of the nine largest (±0.5 standard deviations) seasonal reversals identified in the instrumental data for the Northern Plains, the reconstructions produce the correct signs of the seasonal Z-index values in five of the years. Eleven large reversal years were identified in the instrumental data for the Southern Plains, and the reconstructions reproduce the correct sign for nine of these events. Sign changes in Z-index values between spring and summer, each exceeding at least 0.5 standard deviations from the mean, are estimated to have occurred 26 times in the 367 years of the data for the Northern Plains (7%), and 28 out of 320 years (9%) in the Southern Plains (larger blue and green dots in Figs. 9a and 10a).

The differenced series shown in Figures 9a and 10a illustrate the decadal to multidecadal variability in the relationship between the two seasons and in the occurrence of the largest moisture reversals. Decadal estimates of the reconstructed differenced series are significantly correlated with the instrumental data (r = 0.57 and 0.65 for the Northern and Southern Plains, respectively). For much of the 19th century, summer conditions are estimated to have been more favorable in summer over the Northern Plains, exemplified by the large differences in moisture balance values during the 1820s and from the late-1840s to early-1880s. In the recent decades spring values on average have been higher compared to summer based on the decadal spline, but there has been substantial interannual variability between the two seasons. The differenced estimates for the Southern Plains indicate that for much of the early-18th century conditions were wetter in summer with a greater probability for drier springs to precede wet summers. Conditions appear to have reversed in the late-18th and early-19th centuries, when springs are estimated to
have been wetter. Likewise, in the late-20th century wetter springs punctuated an overall favorable spring and summer growing season, but in the recent two decades springs have become drier over the Southern Plains.

The time intervals between seasonal moisture reversal years seem to exhibit nonrandom behavior. The distributions of return intervals fail to reject the null hypothesis of randomness based on the Lilliefors test for three of the reconstructions (p < 0.05; Figs. 10b, 11b and 11c), the exception being spring drought alleviation events estimated for the Northern Plains (Fig. 10c). Seasonal moisture reversals most often occur one to five years following a reversal of the same sign. However, there are estimated to have been multidecade periods between moisture reversals, the longest being 37 years from 1857 to 1893 for dry springs to wet summers in the Northern Plains. The non-randomness identified in the return intervals of reconstructed moisture reversals is also evident in the instrumental data (not shown).

The spring to summer moisture reversals may be related in part to large-scale ocean-atmospheric teleconnection patterns. Correlation analyses between the differenced series for the Northern Plains and SST data from 1856-1990 indicate a modest relationship with La Niña conditions in the Pacific. This weak La Niña signal can be explained in part by the differences in the ENSO teleconnection from spring to summer (Figs. 5a,b,c,d). Winter-spring La Niña conditions would increase the likelihood of a drier summer over the Northern Plains but would not have a significant impact on the spring climate, thus leading to a higher probability for a larger spring Z-index value. The spatial correlation patterns associated with seasonal differences in the Southern Plains indicate a pattern representative of the positive phase of ENSO in the Pacific (Fig. 10d). El Niño conditions during the winter and spring seasons would favor a wet spring over the Southern Plains, and as a result, an increased likelihood that the spring Z-index
value would be higher compared to the summer. These SST correlation patterns appear to be stable through time based on non-overlapping 46 year correlation periods, though the strongest signals are present in the most recent 46 years (not shown).

Major seasonal moisture reversals sometimes reflect significant changes in mid-tropospheric flow over the Pacific-North American sector from spring to summer, especially for the last month of spring and the first month of summer (i.e., May vs. June). Shown in Figure 11 are composites of the 500mb height anomalies for seasonal moisture reversals exceeding 0.25 standard deviations identified in the instrumental and reconstructed data for the Northern Plains (the anomalies for the Southern Plains are weak and not shown). The largest differences in atmospheric circulation from May to June are observed in both instrumental and reconstructed data over the western United States for wet springs that precede dry summers (Figs. 11g-l). Positive height anomalies over the Gulf of Alaska and a downstream trough in May (Figs. 11g,11j) are replaced with nearly the opposite circulation patterns in June (Figs. 11h,k,l). Wet summers that follow drier springs appear to be connected to slightly-above or near-normal 500mb heights over the western United States in May (Figs. 11a,11d), but the anomaly patterns for June indicate a shift to zonal flow (Figs. 11b,11e). Despite the low 0.25 standard deviation threshold, these moisture reversals are relatively rare events, and their occurrence and distribution over time does not appear to be random. This suggests that atmospheric mechanisms related to major moisture reversals may be influenced by other low-frequency modes of variability that create large-scale environments conducive for significant and sometimes sudden seasonal changes in upper-level circulation over North America.

Analyses of major reversals over the Southern Plains do not indicate any large changes in atmospheric circulation from May to June, or spring to summer over the United States. However,
one of the largest spring to summer moisture reversals over the Southern Plains in both the instrumental and reconstructed series is 1980 (Fig. 10a). This unusual year was associated with a rapid change in atmospheric circulation as discussed by Namias (1982). A deep trough over the southwestern United States during May was replaced within a week in late-May and early-June by a persistent omega-type blocking ridge over the Southern Plains. These observations indicate that some of the strongest spring to summer moisture reversals in the Southern Plains do reflect substantial changes in seasonal atmospheric circulation anomalies, and previous events of this magnitude in the reconstructed record may reflect similar conditions.

4. CONCLUSIONS

Separate seasonal moisture signals are encoded in EW, LW, and LWa width chronologies from the Northern and Southern Plains. Reconstructions of the seasonal atmospheric moisture balance (Palmer’s Z-index) have been developed for both regions, although the reconstruction for the spring season in the Northern Plains study area only represents 39% of the variance in the instrumental record. Estimates of the spring Z-index for the Northern Plains may be improved with additional sampling of sites and species where the potential EW growth is in response primarily to spring moisture. Field sampling strategies tailored towards producing LW and LWa chronologies (e.g. Griffin et al. 2011) could also improve estimates of summer moisture. Nonetheless, select historical observations of weather and climate from the 19th century add independent support for some of the seasonal reconstructions, as do the similar SST and 500mb geopotential height anomaly patterns linked with instrumental spring and summer moisture variability during the 20th century.
The derived reconstructions document the seasonal climate history of the Northern and Southern Plains and highlight the intra-annual to multidecadal variability of regional spring and summer moisture. The 1930s Dust Bowl drought may have been the most extreme summer drought to impact the Northern and Southern Plains in the last few centuries, but drought conditions in spring were not substantially more severe compared to other periods. Our results suggest that the 1930s may have been the only decadal episode of summer drought to simultaneously impact both the study areas since the late-17th century. Comparatively, sustained dual-season drought characterized the mid- and late-19th century, and it appears the driest conditions most often occurred during the spring months. The differences in seasonality associated with these major instrumental and historical-era droughts, and the large-scale ocean-atmospheric dynamics responsible for these estimated seasonal differences, is an important research topic requiring further investigation. The seasonal reconstructions also suggest that the spring and summer climates of the Northern and Southern Plains are largely independent, and in recent decades there has been significant divergence in spring moisture trends possibly attributable to changes in low-level moisture advection from the GPLLJ and position of the upper-level westerly jet stream.

The seasonal reconstructions provide an extended proxy record of the infrequent spring to summer moisture reversals. These seasonal moisture changes in the Northern Plains appear to be related in part to SST anomalies in the Pacific and to changes in mid-tropospheric circulation from spring to summer over the North Pacific and western North America. Separate reconstructions of spring and summer moisture conditions from tree-ring chronologies of EW, LW, and LWa width more broadly across the United States will improve our understanding of
the history of seasonal climate variability and provide insight into the large-scale ocean-atmospheric mechanisms responsible for these seasonal regimes.

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APPENDIX

APPENDIX A

Precipitation percentiles from Mock’s (1991) analysis of 19th century weather for the two regions that are closest to the Northern Plains (NP) and Southern Plains (SP) study area.

<table>
<thead>
<tr>
<th>Year</th>
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<th>NP summer percentiles</th>
<th>SP spring percentiles</th>
<th>SP summer percentiles</th>
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<td>100</td>
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<td>1888</td>
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<td>110</td>
</tr>
<tr>
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<td>70</td>
<td>65</td>
<td>100</td>
<td>70</td>
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<tr>
<td>1890</td>
<td>125</td>
<td>45</td>
<td>100</td>
<td>80</td>
</tr>
</tbody>
</table>
Table 1 Sites of the EW and LW chronologies that were potential predictors for reconstruction of the regional spring and summer moisture balances. The sites were located in two sub-regions of the western Great Plains: (a) the northern network of the western Dakotas and eastern Wyoming, and (b) the southern network of eastern Colorado and northeastern New Mexico. The primary species sampled was ponderosa pine (PIPO), but there is also one site of Douglas-fir (PSME). Included is the year when the expressed population signal (EPS; Cook and Kairiukstis 1990) reaches 0.85, which is generally considered the threshold for a reliable chronology.

**A. Northern network**

<table>
<thead>
<tr>
<th>Rank</th>
<th>Site Description</th>
<th>State</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Elevation(m)</th>
<th>Species</th>
<th>Record length</th>
<th>LW EPS &gt; 0.85</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.</td>
<td>Burning Coal Vein (BCV)</td>
<td>North Dakota</td>
<td>46.43</td>
<td>-102.58</td>
<td>792</td>
<td>PIPO</td>
<td>1592-1990</td>
<td>1647</td>
</tr>
<tr>
<td>2.</td>
<td>Eagle Nest Canyon (ENC)</td>
<td>South Dakota</td>
<td>45.21</td>
<td>-103.07</td>
<td>1090</td>
<td>PIPO</td>
<td>1651-1991</td>
<td>1662</td>
</tr>
<tr>
<td>3.</td>
<td>Reno Gulch (REN)</td>
<td>South Dakota</td>
<td>43.54</td>
<td>-103.36</td>
<td>1740</td>
<td>PIPO</td>
<td>1370-1991</td>
<td>1444</td>
</tr>
<tr>
<td>4.</td>
<td>Buckhorn Mountain (BHM)</td>
<td>South Dakota</td>
<td>43.49</td>
<td>-103.31</td>
<td>1768</td>
<td>PIPO</td>
<td>1600-1991</td>
<td>1681</td>
</tr>
<tr>
<td>5.</td>
<td>Cedar Butte (CED)</td>
<td>South Dakota</td>
<td>43.36</td>
<td>-101.07</td>
<td>785</td>
<td>PIPO</td>
<td>1646-1991</td>
<td>1680</td>
</tr>
<tr>
<td>6.</td>
<td>Teapot Dome (TEA)</td>
<td>Wyoming</td>
<td>43.23</td>
<td>-106.31</td>
<td>1847</td>
<td>PIPO</td>
<td>1483-1979</td>
<td>1716</td>
</tr>
</tbody>
</table>

**B. Southern network**

<table>
<thead>
<tr>
<th>Rank</th>
<th>Site Description</th>
<th>State</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Elevation(m)</th>
<th>Species</th>
<th>Record length</th>
<th>LW EPS &gt; 0.85</th>
</tr>
</thead>
<tbody>
<tr>
<td>8.</td>
<td>Valley View Ranch (VVF)</td>
<td>Colorado</td>
<td>39.07</td>
<td>-104.43</td>
<td>2094</td>
<td>PIPO</td>
<td>1649-1998</td>
<td>1854</td>
</tr>
<tr>
<td>10.</td>
<td>Turkey Creek (TCU)</td>
<td>Colorado</td>
<td>38.21</td>
<td>-104.29</td>
<td>1407</td>
<td>PIPO</td>
<td>1634-2003</td>
<td>1689</td>
</tr>
<tr>
<td>12.</td>
<td>Mesa De Maya (MDM)</td>
<td>Colorado</td>
<td>37.10</td>
<td>-103.62</td>
<td>2060</td>
<td>PIPO</td>
<td>1631-1997</td>
<td>1681</td>
</tr>
<tr>
<td>14.</td>
<td>Kenton (KEN)</td>
<td>New Mexico</td>
<td>36.49</td>
<td>-103.01</td>
<td>1493</td>
<td>PIPO</td>
<td>1635-2015</td>
<td>1684</td>
</tr>
<tr>
<td>15.</td>
<td>Sierra Grande (SIE)</td>
<td>New Mexico</td>
<td>36.43</td>
<td>-103.51</td>
<td>2377</td>
<td>PIPO</td>
<td>1633-2014</td>
<td>1657</td>
</tr>
</tbody>
</table>
Table 2 The EW, LW, and LWa chronologies from the two networks were correlated with the respective regional average spring and summer Z-indices from 1895-1979, and the Pearson’s product moment correlation coefficients are listed. Significant correlations (p < 0.05) are marked by *. The bolded values represent those chronologies used as the initial potential predictors of the respective seasonal climate variable. Note that if the correlation coefficient of the LWa chronology was not significantly different from the LW at a given site, the LWa was preferentially selected as a predictor.

**A. Northern Plains [43°- 47°N, 107°- 101°W]**

<table>
<thead>
<tr>
<th>Sites</th>
<th>EW_r</th>
<th>LW_r</th>
<th>EW_r</th>
<th>LW_r</th>
<th>LWa_r</th>
</tr>
</thead>
<tbody>
<tr>
<td>BCV</td>
<td>*0.49</td>
<td>*0.23</td>
<td>*0.36</td>
<td>*0.54</td>
<td>*0.28</td>
</tr>
<tr>
<td>ENC</td>
<td>*0.48</td>
<td>*0.51</td>
<td>*0.33</td>
<td>*0.60</td>
<td>*0.51</td>
</tr>
<tr>
<td>REN</td>
<td>*0.27</td>
<td>*0.29</td>
<td>*0.28</td>
<td>*0.44</td>
<td>0.08</td>
</tr>
<tr>
<td>BHM</td>
<td>*0.52</td>
<td>*0.22</td>
<td>*0.38</td>
<td>*0.44</td>
<td>0.12</td>
</tr>
<tr>
<td>CED</td>
<td>*0.33</td>
<td>*0.23</td>
<td>*0.43</td>
<td>*0.55</td>
<td>*0.32</td>
</tr>
<tr>
<td>TEA</td>
<td>*0.41</td>
<td>*0.24</td>
<td>*0.54</td>
<td>*0.30</td>
<td>0.17</td>
</tr>
</tbody>
</table>

**B. Southern Plains [35.5°- 38.5°N, 105°- 99°W]**

<table>
<thead>
<tr>
<th>Sites</th>
<th>EW_r</th>
<th>LW_r</th>
<th>EW_r</th>
<th>LW_r</th>
<th>LWa_r</th>
</tr>
</thead>
<tbody>
<tr>
<td>BFE</td>
<td>*0.54</td>
<td>*0.31</td>
<td>*0.42</td>
<td>*0.51</td>
<td>*0.27</td>
</tr>
<tr>
<td>VVF</td>
<td>*0.48</td>
<td>*0.40</td>
<td>*0.44</td>
<td>*0.49</td>
<td>*0.33</td>
</tr>
<tr>
<td>VVR</td>
<td>*0.55</td>
<td>*0.38</td>
<td>*0.45</td>
<td>*0.43</td>
<td>*0.28</td>
</tr>
<tr>
<td>TCU</td>
<td>*0.51</td>
<td>*0.33</td>
<td>*0.42</td>
<td>*0.51</td>
<td>*0.27</td>
</tr>
<tr>
<td>KIM</td>
<td>*0.59</td>
<td>*0.60</td>
<td>*0.34</td>
<td>*0.54</td>
<td>*0.44</td>
</tr>
<tr>
<td>MDM</td>
<td>*0.41</td>
<td>*0.32</td>
<td>*0.24</td>
<td>*0.28</td>
<td>0.15</td>
</tr>
<tr>
<td>COR</td>
<td>*0.62</td>
<td>*0.35</td>
<td>*0.41</td>
<td>*0.53</td>
<td>*0.23</td>
</tr>
<tr>
<td>KEN</td>
<td>*0.66</td>
<td>*0.33</td>
<td>*0.27</td>
<td>*0.67</td>
<td>*0.56</td>
</tr>
<tr>
<td>SIE</td>
<td>*0.52</td>
<td>*0.34</td>
<td>*0.33</td>
<td>*0.40</td>
<td>*0.21</td>
</tr>
</tbody>
</table>
Table 3: The transfer function models’ calibration and validation statistics are listed for the spring (MAM) and summer (JJA) reconstructions for the (1,2) Northern Plains and (3,4) Southern Plains. (a) The transfer function used for the reconstruction, where $\hat{y}_t$ is the estimated Z-index value for year $t$ and $PC_t$ is the value for the principal component time series. (b) The final predictor chronologies used in the PCR scheme (3-letter site codes defined in Table 1), * denotes an EW chronology, + LW, and ^ LWa. (c) $R^2$ adjusted downward for loss of degrees of freedom (Draper and Smith 1981). (d) $r$ = The Pearson product moment correlation coefficient between instrumental and reconstructed data in the validation period. (e) RE = reduction of error statistic (Fritts 2001); CE = coefficient of efficiency (Cook and Kairiukstis 1990). (f) The number of occurrences in the validation period when the reconstructed data contained the same (hit) or different (miss) sign as the instrumental Z-index data.

<table>
<thead>
<tr>
<th>Season</th>
<th>Model&lt;sup&gt;a&lt;/sup&gt;</th>
<th>Chronologies used in final PCR&lt;sup&gt;b&lt;/sup&gt;</th>
<th>$R^2$ adj.&lt;sup&gt;c&lt;/sup&gt;</th>
<th>$r$ &lt;sup&gt;d&lt;/sup&gt;</th>
<th>RE/CE&lt;sup&gt;e&lt;/sup&gt;</th>
<th>Sign hit/miss&lt;sup&gt;f&lt;/sup&gt;</th>
</tr>
</thead>
<tbody>
<tr>
<td>1. MAM</td>
<td>$\hat{y}_t = -0.368 + (1.02 * PC1_t)$</td>
<td>BCV*, BHM*, ENC*</td>
<td>0.39</td>
<td>0.65</td>
<td>0.39/0.39</td>
<td>32/13</td>
</tr>
<tr>
<td>2. JJA</td>
<td>$\hat{y}_t = -0.074 + (-1.23 * PC1_t)$</td>
<td>BCV^+, CED^+, ENC^</td>
<td>0.43</td>
<td>0.80</td>
<td>0.64/0.63</td>
<td>31/14</td>
</tr>
<tr>
<td>3. MAM</td>
<td>$\hat{y}_t = -1.0645 * PC1_t$</td>
<td>BFE*, COR*, KEN*, KIM*, VVR*</td>
<td>0.56</td>
<td>0.78</td>
<td>0.58/0.54</td>
<td>34/11</td>
</tr>
<tr>
<td>4. JJA</td>
<td>$\hat{y}_t = 0.034 + (-1.14* PC1_t)$</td>
<td>COR^+, KEN^, TCU^</td>
<td>0.46</td>
<td>0.73</td>
<td>0.54/0.54</td>
<td>32/13</td>
</tr>
</tbody>
</table>
Table 4 Spring and summer precipitation percentile data for two regions similar to the reconstruction regions were obtained through visual analysis of Figs. 5 and 6 from Mock’s (1991) analysis (see Appendix A). Correlations analyses with the respective regional and seasonal moisture balance estimates over the common period (1877-1890 for the Northern Plains and 1867-1890 for the Southern Plains) was performed.

<table>
<thead>
<tr>
<th></th>
<th>Pearson’s r</th>
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<tbody>
<tr>
<td>NP spring Z-index (1877-1890)</td>
<td>0.45</td>
</tr>
<tr>
<td>NP summer Z-index (1877-1890)</td>
<td>0.51</td>
</tr>
<tr>
<td>SP spring Z-index (1867-1890)</td>
<td>0.77</td>
</tr>
<tr>
<td>SP summer Z-index (1867-1890)</td>
<td>0.53</td>
</tr>
</tbody>
</table>
Table 5 Simple 10-year moving averages were calculated for each normalized reconstruction. The 10-year average for each decade [e.g. 1651-1660, 1661-1670], and the ten driest decades for each seasonal reconstruction were then identified. The 10-year average Z-index value is listed next to the decades.

<table>
<thead>
<tr>
<th>Northern Plains spring</th>
<th>Southern Plains spring</th>
<th>Northern Plains summer</th>
<th>Southern Plains summer</th>
</tr>
</thead>
<tbody>
<tr>
<td>1. 1860s (-1.47)</td>
<td>1860s (-1.05)</td>
<td>1930s (-0.92)</td>
<td>1840s (-0.98)</td>
</tr>
<tr>
<td>2. 1870s (-0.95)</td>
<td>1850s (-0.87)</td>
<td>1890s (-0.70)</td>
<td>1930s (-0.89)</td>
</tr>
<tr>
<td>3. 1710s (-0.81)</td>
<td>1930s (-0.61)</td>
<td>1860s (-0.69)</td>
<td>1810s (-0.69)</td>
</tr>
<tr>
<td>4. 1750s (-0.72)</td>
<td>1810s (-0.55)</td>
<td>1790s (-0.59)</td>
<td>1850s (-0.56)</td>
</tr>
<tr>
<td>5. 1950s (-0.58)</td>
<td>1840s (-0.51)</td>
<td>1710s (-0.46)</td>
<td>1950s (-0.46)</td>
</tr>
<tr>
<td>6. 1700s (-0.56)</td>
<td>1870s (-0.41)</td>
<td>1750s (-0.38)</td>
<td>1970s (-0.40)</td>
</tr>
<tr>
<td>7. 1810s (-0.52)</td>
<td>1950s (-0.33)</td>
<td>1780s (-0.35)</td>
<td>1820s (-0.27)</td>
</tr>
<tr>
<td>8. 1930s (-0.45)</td>
<td>1960s (-0.32)</td>
<td>1870s (-0.27)</td>
<td>1800s (-0.24)</td>
</tr>
<tr>
<td>9. 1890s (-0.29)</td>
<td>1730s (-0.27)</td>
<td>1980s (-0.26)</td>
<td>1860s (-0.16)</td>
</tr>
<tr>
<td>10. 1780s (-0.19)</td>
<td>1970s (-0.21)</td>
<td>1660s (-0.22)</td>
<td>1780s (-0.15)</td>
</tr>
</tbody>
</table>
Fig. 1 (a,b) Normalized precipitation for the (a) spring (MAM) and (b) summer (JJA) seasons are plotted for the Dust Bowl period using the gridded PRISM dataset. (c,d) Same as (a,b) but with temperature data. (e,f) Same as (a,b) but with Palmer’s Z-index. Note that in all cases, conditions intensified and expanded from spring to summer.
Fig. 2 Same as Fig. 1, but for the 1950s Drought from 1951-1956. Similar to the Dust Bowl Drought, the largest anomalies took place during the summer season, but drought conditions were more centered over the Southern Plains and Southeast.
Fig. 3 Locations of the Douglas-fir and ponderosa pine study sites from the western Great Plains used in the analysis. Red circles indicate the northern network (Table 1a) and black circles are part of the southern network (Table 1b). The numbering of each circle corresponds to the sites listed in Table 1.
Fig. 4 The instrumental (dashed lines) and reconstructed (solid black lines) Z-indices for spring (March-May) and summer (June-August) are normalized and plotted together for the (a,c) Northern Plains and (e,g) Southern Plains from 1900-1990. The calibration (1945-1990) and validation (1900-1944) intervals are separated by the vertical dashed line. The statistics of each regression model are presented in Table 3. (b,d,f,h) Shown are squared coherence plots between instrumental and reconstructed Z-indices for the respective calibration and validation periods (solid black line). Dashed lines represent the 95% and 99% confidence thresholds for significant coherence.
Fig. 5 The (a) instrumental and (b) reconstructed spring Z-indices for the Northern Plains reconstructed region were correlated with gridded SST data from 1945-1990 (Kaplan et al. 1997), after the SSTs were averaged to the winter-spring (December-May) season. The (c) Instrumental and (d) reconstructed summer Z-indices for the Northern Plains are correlated with the spring-summer (March-August) averaged SSTs. Black box represents the Northern Plains study region [43°- 47°N, 107°- 101°W]. (e,f) Same as (a,b) for the Southern Plains study region. (g,h) Same as (c,d) for the Southern Plains. Only grid points with significant (p < 0.05) correlation coefficients shown. Significance levels account for the reduced degrees of freedom due to autocorrelation in the SST and Z-index data (Ebisuzaki 1997). Black box represents the Southern Plains study region [35.5°- 38.5°N, 105°- 99°W].
Fig. 6 The (a) instrumental and (b) reconstructed spring Z-indices for the Northern Plains reconstructed region (black box) were correlated with gridded 500mb geopotential height data from 1945-1990 (Compo et al. 2011), after the data were averaged to the spring season. The (c) Instrumental and (d) reconstructed summer Z-index for the Northern Plains were correlated with the summer averaged 500mb geopotential height data. (e,f) Same as (a,b) for the Southern Plains study region. (g,h) Same as (c,d) for the Southern Plains. Only grid points with significant (p < 0.05) correlation coefficients shown. Significance levels account for the reduced degrees of freedom due to autocorrelation in the atmospheric and Z-index data.
Fig. 7 The normalized (a) spring and (b) summer moisture balance reconstructions for the Northern Plains are plotted from 1651-1990. Reconstructions have been fit with a 10-year cubic-smoothing spline designed to emphasize decadal variability (black line; Cook and Peters 1981). (c) The spring and (d) summer moisture balance reconstructions for the Southern Plains are plotted from 1698-1990. Instrumental Z-index values from 1991-2017 are also plotted (dashed gray lines). The 10-year smoothed instrumental spring and summer Z-indices are also plotted from 1895-2017 (red series) with each reconstruction.
Fig. 8 (a) Reconstructed spring (green circles) and summer (blue circles) are plotted consecutively from 1840-1900 for the Northern Plains to highlight the distinct seasonal drought conditions of the 19th century. (b) Same as (a) but for the Southern Plains. (c,d) The estimates are plotted consecutively from 1930-1960 to illustrate the seasonality of the 1930s and 1950s Droughts for the (c) Northern Plains and (d) Southern Plains. The dashed vertical bars represent the time interval of regional drought conditions. The mean values of the two seasons for each of these intervals are also included. Note the more intense spring drought conditions in the 19th century, but summer drought was more severe during the 1930s and 1950s.
Fig. 9 (a) A time series calculated by differencing the normalized reconstructed spring Z-index from summer is plotted interannually for the Northern Plains. Values above zero are years when spring was wetter than summer, and vice versa. The time series has been fit with a 10-year spline to emphasize decadal variability (black line), and a decadal spline fit to a differenced series of the instrumental data has also been included (red line). The blue and green points represent a change in sign of the moisture balance from spring to summer, with blue points indicating a wet spring followed by a dry summer, and green points are the opposite pattern. Large blue and green dots represent seasonal moisture reversals when both the spring and summer Z-index values were 0.5 standard deviations above or below the mean. (b) The histogram of return intervals for dry springs followed by wet summers (p-values test the distribution of return intervals for non-randomness; p < 0.05 = non-random distribution). (c) Same as (b) for wet springs followed by dry summers. (d) The reconstructed differenced series was correlated with December-May (DJFMAM) SST data from 1856-1990. Only grid points with significant (p < 0.05) correlation coefficients are plotted.
Fig. 10 Same as Fig. 10 for the Southern Plains.
Fig. 11 (a,d) Dry springs to (b,e) wet summers and (g,j) wet springs to (h,k) dry summers that exceeded >±0.25 standard deviations were identified in instrumental and reconstructed data from 1900-1990 for the Northern Plains. The 500mb height anomalies for May vs. June were composited for the identified reversal years in instrumental and reconstructed data. (c,f,i,l) Differences in 500mb height from May to June (May – June) are plotted. Shaded regions in the composite maps of May and June indicate significant (p < 0.05) anomalies relative to the monthly climatology calculated from 1981-2010.
CHAPTER 3

Tree-ring reconstructions of single day precipitation totals over eastern Colorado
ABSTRACT

Precipitation is highest during midsummer over eastern Colorado and some of the most damaging Front Range flash floods have occurred during daily rainfall extremes in late-July or early-August. Tree-ring chronologies based only on the last formed latwood cells in ponderosa pine from eastern Colorado are highly correlated with the highest one-day rainfall totals occurring during this midsummer precipitation maximum. A regional average of these “adjusted latewood width” chronologies was used to reconstruct the single wettest day observed during the last two weeks of July. The regional chronology was calibrated with the CPC .25x.25 Daily U.S. Unified Gauge-Based Analysis of Precipitation dataset and explains 68% of the variance in the highest one-day midsummer precipitation totals in the instrumental data from 1948-1997. The reconstruction extends from 1779 to 2017 and indicates that the frequency of one-day rainfall extremes in midsummer has increased since the late-18th century. The instrumental and reconstructed one-day precipitation extremes in midsummer are often associated with unseasonable atmospheric circulation, including a negatively tilted upper-level ridge, slow-moving surface front, and an increase in atmospheric moisture advection from the Gulf of Mexico and Pacific. These synoptic conditions were observed during the catastrophic floods on the Big Thompson River on July 31, 1976, and on Spring Creek near Fort Collins during July 27-28, 1997. Chronologies of adjusted latewood width in semiarid Colorado constitute a proxy of weather timescale precipitation extremes useful for investigations of long term variability and for framing natural and potential anthropogenic forcing of midsummer precipitation extrema in a long historical perspective.

Key words: midsummer rainfall maximum; single day rainfall extremes, flash flooding, Colorado, adjusted latewood width, ponderosa pine, dendrometeorology
1. INTRODUCTION

Extreme midsummer rainstorms have caused some of the most damaging flash floods to impact the Colorado Front Range Urban Corridor, including the Big Thompson Canyon Flood of July 31, 1976 (Maddox et al. 1978), and the flooding of Spring Creek near Fort Collins on July 27-28, 1997 (Doesken and Mckee 1998). The Front Range is vulnerable to severe thunderstorms and flash flooding in midsummer when high atmospheric moisture levels, weak upper level winds, and slow-moving thunderstorms converge over the steep topographic gradient along the eastern Rocky Mountains (Maddox et al. 1979; Mckee and Doesken 1997; Cotton et al. 2003). The frequency and intensity of extreme rainfall events appear to have increased over the United States since 1901 (Wuebbles et al. 2017) and the positive trend has been most pronounced in summer (Karl and Knight 1998). Major trend in precipitation extremes has not been detected for Colorado (Eden et al. 2016), but climate model simulations suggest that the frequency and magnitude of daily rainfall extremes may increase with unabated anthropogenic global warming (Kunkel et al. 2013; Wuebbles et al. 2017). Historical documentary evidence, early instrumental observations, and potentially exactly-dated wood anatomical or sub-annual ring width data might provide a longer historical perspective on midsummer rainfall extremes prior to the onset of heavy anthropogenic weather and climate forcing.

Climate sensitive tree-ring chronologies have been widely used to reconstruct growing season precipitation and the Palmer Drought Severity Index (PDSI; Palmer 1965; Stahle and Cleaveland 1992; Fritts 2001; Cook et al. 2007). Douglass (1920) described tree growth as a response to integrated climate conditions “distributed throughout the year.” The correlation between seasonal precipitation totals and annual ring-width chronologies can in fact be so high that they have been referred to as “integrating pluviometers” (Blasing and Fritts 1976). Because
tree growth tends to utilize soil moisture accumulated during or even preceding the growing season, it has not been possible to develop estimates of daily timescale weather phenomena on a continuous year-by-year basis extending back into prehistory using total ring-width (RW) chronologies. Weather extremes associated with severe growing season freeze events (LaMarche and Hirshboeck 1984; Stahle 1990; Carolina-Barbosa 2010) and mid-growing season weather reversals (Villalba and Veblen 1996; Fritts 2001; Edmondson et al. 2010) may induce distinctive anatomical evidence in the xylem cells of living trees. The meteorological significance of these so-called frost and false ring chronologies can be demonstrated during the instrumental period and then used to infer the history of these episodic events during the pre-instrumental era. But weather sufficiently extreme to cause anatomical damage to tree rings is infrequent, so the derived event chronologies tend to be highly discontinuous.

In this article we describe a strong correlation between chronologies based on the width of the last formed latewood cells in ponderosa pine (*Pinus ponderosa*) and the single wettest day that occurs during a two-week window in midsummer over eastern Colorado. These so-called adjusted latewood width chronologies (LWa) are correlated with total July precipitation, but higher correlations are actually observed when compared with an instrumental time series based on the wettest single rainfall day during the last two weeks of July (July 19 to August 1). Individual heavy rainfall events dominate the climatology of midsummer precipitation, and we use the strong single-day signal in a regional LWa chronology for eastern Colorado to reconstruct these totals each year from 1779 to 1997. These “dendrometeorological” moisture proxies are then used along with instrumental observations to describe the synoptic meteorology and interannual to multidecadal changes in midsummer rainfall extremes, placing them in the context of natural weather and climate variability since the late-18th century.
Warm season (April-September) precipitation contributes over 70% of the annual total in the semiarid Front Range and adjacent High Plains of eastern Colorado [37.75°N-39.75°N, 105°W-103W°; Mahoney et al. 2015], but a large fraction of the warm season totals occurs during two relatively short periods when daily precipitation rates tend to be highest (Fig. 1). These spring (May and early-June) and midsummer (late-July) rainfall maxima are separated by a drier interval that reaches a minimum in late-June and early-July.

The evolution of daily rainfall rates during the warm season can be explained in part by the changing atmospheric circulation over North America from spring to summer. During May and early-June, the upper-level jet stream and the associated Pacific storm track are often still far enough south to influence eastern Colorado. Midlatitude wave cyclones from the Pacific can increase low-level moisture advection from the Gulf of Mexico into Colorado, elevating atmospheric instability and upslope flow along the eastern Rocky Mountains (Hirschboeck 1991). Heavy precipitation, including spring snowstorms, can develop when cold fronts penetrate into these moist unstable environments. The strength and spatial extent of the Great Plains low-level jet increases from late-spring and peaks in early-summer (Fig. 2a; Krishnamurthy et al. 2015), but the northward retreat of the upper-level jet stream results in fewer storm systems penetrating as far southward as eastern Colorado, and thus daily rainfall rates tend to decline from mid-June thru early-July (Fig. 1; Whittaker and Horn 1981).

By mid-to-late-July, upper-level winds over the southern United States and northern Mexico have usually shifted from westerly in the winter and spring months to southeasterly as the subtropical North Atlantic high expands northwestward (Fig. 2a; Adams and Comrie 1997). The North American monsoon circulation also advances northward from northern Mexico into the southwestern United States in July, resulting in a ‘monsoon’ upper-level ridge that resides
over the region during the midsummer months. The influence of the monsoon circulation can extend as far north as Colorado and southern Wyoming from late-July through early-August, funneling moist mid-to upper-level subtropical air from the Pacific into the interior Rocky Mountains and western High Plains (Fig. 2b; Hales 1974; Tang and Reiter 1984). The moisture sources from the Gulf of Mexico and Pacific result in an annual atmospheric humidity maximum over eastern Colorado in late-July and early-August (Hansen et al. 1978). The higher humidity content in the atmosphere, coupled with mechanisms that promote atmospheric instability [e.g. frontal boundaries, short wave disturbances, and diurnal heating of the land surface], provides the necessary thermodynamic environment for precipitation rates that are often greater in magnitude and duration than the daily totals in May and early-June (Fig. 1).

2. DATA AND METHODS

2.1 Daily rainfall data

The gridded daily precipitation data used in this study were acquired from the National Oceanic and Atmospheric Administration (NOAA) Climate Prediction Center (CPC) Daily U.S. Unified Gauge-Based Analysis of Precipitation dataset (Chen et al. 2008). The gridded data were calculated using the optimal interpolation (OI) method from a dense network of observing sites and contain daily precipitation values on a 0.25° latitude by 0.25° longitude grid extending from 1948 to present. Daily totals are calculated for the 24-hour period ending at 12z. Rarely do grids points within the eastern Colorado study region contain zero values due to the nature of the interpolated data. For the purposes of this study, we treated daily precipitation values of less than 1mm as zero at each grid point.
Daily precipitation totals from 40 selected observation stations in eastern Colorado in the Global Historical Climatology Network (GHCN) from the National Climatic Data Center (NCDC) were also used for analyses with the gridded daily data and tree-ring chronologies (Fig. 3). These 40 stations were selected based on their location within or near the eastern Colorado study region (black box in Fig. 3), and the availability of at least 25 years of observations. Daily atmospheric data derived from the 20\textsuperscript{th} Century Reanalysis project (Compo et al. 2011) were used to illustrate the changing spatial patterns of moisture influx from spring to summer (Fig. 2) and to identify synoptic-scale circulation features associated with the largest observed and reconstructed one-day rainfall totals over eastern Colorado. Because most of the continuous station records begin after 1940, we restrict our instrumental analyses to the period 1940-2017. Anomalies in the reanalysis data were calculated relative to the 1981-2010 climatology.

\textit{2.2 Adjusted latewood chronology development}

The range of ponderosa pine extends sparsely eastward from the Rocky Mountain Front Range into the High Plains of eastern Colorado. Native stands with old trees can occasionally be found on fire-protected escarpments and on certain higher elevation microenvironments (Wells 1965). Several annual ring width chronologies of ponderosa pine have been previously developed in eastern Colorado and used to reconstruct the Palmer Drought Severity Index (PDSI; Woodhouse and Brown 2001), streamflow (Woodhouse and Lukas 2006). We remeasured many of these collections for earlywood (EW), latewood (LW), RW, and computed LW\textsubscript{a} chronologies for four sites in eastern Colorado [Black Forest East (BFE), Jefferson County (JFU), Ridge Road (RIR), and Turkey Creek (TCU); Fig. 3]. The procedures outlined by Stahle et al. (2009) were followed to carefully measure EW and LW width with a stage micrometer to 0.001 millimeter
(mm) precision. Chronologies of EW and LW width were computed for each site using the signal free method of standardization (Melvin and Briffa 2008; Cook et al. 2014). Power transformed ring-width indices were calculated as residuals from the fitted growth curve, and then averaged into the mean index chronology using the biweight robust mean (Cook 1985; Hoaglin et al. 2000).

Meko and Baisan (2001) developed statistical methods to derive a discrete monsoon season precipitation signal from Douglas-fir (*Pseudotsuga menziesii*) tree rings in southern Arizona where separate chronologies of EW and LW width can be identified for each year due to the clear anatomical demarcation between these sub-annual features of the growth rings. Earlywood and LW width chronologies have potential to retain separate seasonal climate information [e.g. Watson and Luckman 2002; Stahle et al. 2009; Griffin et al. 2011; Griffin et al. 2013; Woodhouse et al. 2013; Crawford et al. 2015; Dannenberg and Wise 2016; Watson and Luckman 2016; Kerhoulas et al. 2017; Arzac et al. 2018; Torbenson and Stahle 2018; Howard et al. 2019], but the correlation between EW and LW can be so high as to obscure any separate summertime signal that might be recorded in LW width (Torbenson et al. 2016). Meko and Baisan (2001) regressed the LW on EW chronologies to derive the LW regression residuals that are uncorrelated with the EW chronology (i.e., LWa). This manipulation works reasonably well in isolating the independent variability of the last formed LW cells that theoretically should reflect summer moisture independent of climate and growth conditions during the preceding spring and early summer. In fact, robust reconstructions of summer monsoon precipitation have subsequently been derived in the southwestern United States using both LW (Griffin et al. 2013) and LWa chronologies (Stahle et al. 2009, 2015). However, our research indicates that in
In semiarid environments, the variance of the LWa chronologies may be dominated by rainfall variability considerably shorter than the summer season, in some cases even at daily timescales.

Two methods were compared to compute the LWa width chronologies using the final mean index chronologies of EW and LW width at each sampling location. The LW chronologies were regressed on the EW width and the regression residuals were used as the LWa chronology as described by Meko and Baisan (2001). The Kalman Filter (Welch and Bishop 2006) was also used to estimate the LW variability unrelated to EW width, allowing for the potential time dependency of the regression relationship. Because correlation analyses among the LWa chronologies calculated with both methods and with regional precipitation data for eastern Colorado indicate the Kalman filter produced chronologies that are slightly more homogenous among the four sites and better correlated with precipitation, we used the Kalman Filter method to derive the LWa chronologies.

Due to large changes in sample size resulting from fewer old trees in the early years of each chronology, the beginning of all four LWa chronologies exhibit elevated variance. This elevated variance in the early portion of the chronology was reduced using a variance stabilization technique (Meko 1981). A smoothing spline with a 50% frequency response equal to 100-years (Cook and Peters 1980) was fit to the absolute values of the annual LWa indices minus their mean, and the ratios of the fitted spline to the absolute values were computed. The sign was then restored and the mean was added back to each annual value to produce the variance-stabilized LWa chronologies (Meko 1981; Cook and Krusic 2005). The annual values for the four chronologies were then squared to increase skewness and better represent the statistical distribution of the daily precipitation.
2.3 Daily precipitation response of latewood growth

A regional average of the four LWa chronologies was initially correlated with monthly precipitation totals to examine the potential moisture signals present in the tree-ring data. We then implemented a daily precipitation analysis to determine if stronger moisture signals are present in the regional LWa chronology at weather timescales (i.e., daily to weekly). To derive a time series of the single highest daily rainfall total at any location within the 72-grid point study area (black box in Fig. 3), the daily precipitation totals at each grid point were extracted and subset into interval lengths ranging from 1 to 31 days for all 365 days of the year, resulting in 11,315 intervals. The wettest daily total at any of the 72 grid points for each interval was then identified and that single grid point value was used to represent year \( t \) in the wettest single day time series for all intervals 1 to 31. The single wettest day rainfall total time series (\( n = 11,315 \)) produced from these computations were then correlated with the regional LWa chronology to identify the tree-ring response to these single wet days during each interval. These results were compared with the correlations between LWa and monthly and seasonal totals to determine the best timescale response to precipitation.

Identification of the single grid point within the study region was used rather than simple area averaging because the variance of daily rainfall totals at one grid point better replicates daily totals based on individual station data and this method also captures the spatial variability of summertime thunderstorms at localized scales. Precipitation totals at each grid point were also summed for interval lengths ranging from 1 to 365 days for all days of the year (\( n = 133,225 \)), and a regional average of the precipitation totals was calculated. These time series were then correlated with regional averages of EW, and LW, LWa, and RW chronologies to compare the different moisture signals present among the four types of tree-ring data.
2.4. Tree ring reconstruction of one-day precipitation extremes in midsummer

The time series of the wettest single day during intervals of one to eight weeks were computed from the gridded data for the eastern Colorado study area. Because these one-day extremes in midsummer are very highly correlated with the regional LWa chronology, the optimal interval was chosen for the tree-ring reconstruction of single day precipitation extremes in late-July from 1779-2017. Split calibration and validation experiments for the reconstruction of single day rainfall extremes were performed for two sub-periods from 1948-1997 in common to both the regional tree-ring chronology and the gridded instrumental observations. The regional LWa chronology was initially calibrated on the 1973-1997 period using regression, and data from 1948-1972 were withheld for independent verification. The LWa chronology was also calibrated on the earlier period, and the estimates were then independently verified on the later period. The final reconstruction was based on the full common interval from 1948-1997.

3. RESULTS

The regional average LWa chronology from eastern Colorado is significantly correlated with July precipitation totals ($r = 0.68$, $p < 0.0001$), but the correlation with the time series of the highest one-day rainfall amount during the last two weeks of July is actually much stronger ($r = 0.83$, $p < 0.0001$; Fig. 4). The correlation coefficients computed between the regional LWa chronology and the single highest daily precipitation total for all possible 14-day intervals during the year are plotted in Figure 4a (the correlation for Julian day 1 is with the highest daily total between December 19 and January 1). These correlations peak on Julian day 213 when the Pearson correlation with the wettest day in the two week interval reaches $r = 0.83$ (Spearman correlation is $r = 0.74$; not shown). Note the sharp increase in correlations at Julian day 200,
which indicates that the width of the last formed LW cells begins to be influenced by the heaviest one-day rainfall totals beginning in the second week of July, even though the peak response is during the interval from Julian day 200-213, or July 19 to August 1 (Fig. 4a). When this method is applied using the single highest daily rainfall total in late-July observed at the 40 individual weather stations (Fig. 3), the highest correlation is also with the wettest daily total identified at any of the 40 stations during the two week interval in late-July ($r = 0.82$; not shown).

For comparison with the response to one-day totals in midsummer, the LWa chronology was also correlated with precipitation totaled for all possible 7-day to 56-day intervals during the year (Fig. 4a). The highest correlation with these various weekly to monthly precipitation totals reaches 0.78 for the 14-day period also ending on Julian day 213, so that roughly 60% of the LWa variance might be explained by the two-week rainfall total in late-July compared with 68% of the variance associated with the highest one-day total during the same period. Correlation with the LWa chronology then declines as precipitation is summed over longer periods (Fig. 4a).

The high correlation with the single day rainfall totals in late-July coincides with the annual two-week precipitation maximum over eastern Colorado (Fig. 4b). On average, nearly 70% of the total precipitation recorded from July 19 to August 1 is contributed by the heaviest single rainfall day in this area (i.e., 68.25%). In fact, when analyzed at the grid point level for semiarid eastern Colorado, all of the 14-day total can be attributed to a single rain event in some years.

The precipitation response of the EW, LW, and RW chronologies illustrates the major differences in moisture signal compared to the LWa data from eastern Colorado (Fig. 5). The EW, LW, and RW chronologies have an integrated, nearly annual moisture signal and are most
highly correlated with precipitation accumulated over several months prior to and during the growing season. These highest correlations with precipitation over all possible continuous intervals are accumulated over 305, 292, and 308 days during and preceding the growing season for the regional EW, LW, and RW regional chronologies from eastern Colorado, respectively. Using the precipitation response profile for 300-day total precipitation as an example, the highest correlation with both EW and RW is with precipitation accumulated from the previous mid-September to current mid-July (September 21 to July 19, or Julian day 265 of the prior year to day 200 of the current year). The highest 300-day moisture signal for un-adjusted LW is only slightly later (i.e., October 4 to August 2).

The strong relationship between the one-day rainfall totals in midsummer and the regional LWa chronology is present despite the low correlations among the four LWa chronologies. The average correlation among these four chronologies is only $r = 0.34$ during the instrumental period (1947-1997) and $r = 0.28$ for the full common interval (1779-1997). However, the individual LWa chronologies are each well correlated with the gridded time series of the wettest single day in midsummer (range of $r = 0.45$ to 0.65 for the four chronologies). When these four LWa chronologies are averaged the correlation with the wettest single day time series derived from the gridded data reaches $r = 0.83$.

The wettest single day in late-July is also poorly correlated among most instrumental station observations [e.g., the four stations closest to the four tree-ring sites (Denver NWS, Byers, Limon, and Pueblo) are correlated on average at only $r = 0.25$ for the wettest day at each location from July 19 to August 1, 1948-1997, ranging from $r = 0.07$ to 0.65]. However, each of these four time series is well correlated with the wettest single day time series derived from the gridded data (range of $r = 0.38$ to 0.71 for the four stations). When the wettest day in
midsummer is averaged among the four instrumental stations, regardless of date of occurrence within the two week window (similar to what may often happen at the four tree-ring collection sites), the correlation with the wettest single day series based on the gridded data for eastern Colorado improves to $r = 0.80$).

These comparisons reflect the spatially discontinuous nature of midsummer precipitation events over eastern Colorado, but the common signal among individual weather stations or tree-ring sites can be greatly enhanced with regional averaging. The individual weather stations and the LWa chronologies are all highly correlated with the regional midsummer rainfall extremes for eastern Colorado, but the extremes do not always occur on the same day in late-July at each collection site or weather station. However, wet years with heavy single day totals, and dry years with very light single day totals in the gridded regional time series (e.g., Fig. 6), tend to drive the correlations of midsummer one-day precipitation extremes with the tree-ring sites and stations observations.

The regional LWa chronology not only correlates well with the magnitude of the wettest single day totals, but also the spatial extent of these events. The five highest one-day totals correspond with widespread precipitation over most of eastern Colorado and the central Great Plains (Fig. 6a-f). The zonal rainfall pattern in the composite of the five wettest days resembles organized precipitation along a front or by a mesoscale convective system (Maddox 1980) that develops in Colorado and propagates eastward (Fig. 6f). However, the five lowest totals suggest more localized thunderstorm activity (Figs. 6g-l).

The reconstruction of late-July one-day rainfall totals was developed using regression between the LWa chronology (predictor) and the single wettest day at any grid point in the eastern Colorado study area (predictand). The instrumental and the reconstructed values are
plotted in Figure 7a. The reconstructed series explains 68% of the interannual variance in one-day rainfall totals during the full 1948-1997 calibration period (Fig. 7a). Split calibration and validation experiments on two 25-year sub-periods indicate the relationship between the two series is reasonably stable (Table 1a). The distribution of instrumental and reconstructed rainfall values tends to be positively skewed by the most extreme wet years (Fig. 7b,c). Using the wettest single days identified among the four closest instrumental stations produces similar calibration and validation statistics (Table 1b).

The 239-year reconstruction from 1779-1997 documents the interannual to decadal variability of late-July single day rainfall extremes over eastern Colorado (Fig. 8; the variance lost in the regression was restored to the reconstructions from 1779-1997 and the instrumental values were then appended to the time series from 1998-2017). The reconstruction suggests that the frequency of the wettest >90th percentile events more than doubled from the 19th to 20th centuries, while the driest <10th percentile events appear to have decreased since the late-18th century (Fig. 8). These 90th percentile totals tend to be events that impact eastern Colorado and the western United States at the synoptic scale (Figs. 6a-f). But in spite of the apparent increase in the frequency of wet extremes, stochastic volatility analysis (Kastner 2016) does not indicate a significant increase in the overall variance of the reconstructed or observed wettest one-day totals (not shown).

The late-20th century (1960-1999) is estimated to have experienced a high frequency of midsummer one-day rainfall extremes (Fig. 8), but this high frequency of extremes was terminated by persistent below-average conditions during the early-21st century drought. No 90th percentile events have occurred within the study region since 2000, and no single day totals exceeded 55 mm from 2000 to 2017. These characteristics of the late-20th and early-21st
centuries seen in the gridded and reconstructed data are also evident in the precipitation data
recorded at the 40 individual weather stations in eastern Colorado (Fig. 9). Many of the largest
events identified in the instrumental data occur between the 1960s and 1990s, but there is a
noticeable decline in heavy rainfall days after 1999. In fact, the five year-average from 2002-
2006 calculated from the single wettest day identified from among the station data is the lowest
for any period from 1940-2017 (Fig. 9).

Major sub-decadal to decadal periods of drought identified using instrumental and tree-
ring reconstructed PDSI are also apparent in the one-day midsummer rainfall totals (e.g. 1930s
Dust Bowl, early-21st century drought; Worster 1978; Seager 2007; mid-19th century drought;
Woodhouse et al. 2002; Herweijer et al. 2006; Cook et al. 2007; Fig. 8). The Dust Bowl Drought
in particular had a negative impact on the highest midsummer rainfall totals in eastern Colorado.
Seven out of the ten years from 1931-1940 are estimated to have been below average, and just
two other comparable periods of sustained deficits in midsummer single day totals occurred
during the 1840s and 1850s, and in the early-21st century. The reconstruction of midsummer
one-day rainfall values is not correlated with reconstructed PDSI at interannual timescales (r =
0.07 from 1779-1997; not shown), but as these major PDSI droughts suggest, prolonged dryness
appears to reduce daily rainfall extremes in midsummer. The decadal estimates of the wettest
single day of midsummer time series is not strongly correlated with decadal estimates of a
regional average of reconstructed summer PDSI (r = 0.37; Cook et al. 2007), but agreement
among these smoothed time series is highest during the most severe and sustained droughts in
the regional PDSI reconstructions (not shown).

Some of the largest instrumental and reconstructed single day extremes during the 1948-
1997 calibration period are associated with severe midsummer storms and flash flooding along
the Rocky Mountain Front Range (noted in Fig. 8). The single day total in 1997 corresponds to
the July 27-28 flash flooding of Spring Creek near Fort Collins. The Fort Collins flood was
estimated as a 500-year event and remains one of the costliest flash floods in Colorado’s history,
with over $200 million in damages (Grigg et al. 1999). Other notable heavy rainfall days include
July 19, 1985, which is the largest reconstructed value and is considered one of the wettest days
in Colorado’s history (Doesken and McKee 1986). Hourly rainfall rates recorded at many
stations in eastern Colorado were greater than 25.4 mm on July 19, 1985, and widespread reports
of flash flooding and other severe weather hazards including hail, damaging wind, and tornadoes
are documented (Doesken and McKee 1986). The reconstructed heavy rainfall day for July 24,
1965, was part of a persistent pattern of precipitation from July 20-25, including heavy rains on
July 23-24 that led to significant flash flooding of Tucker Gulch in Golden, Colorado (UDFCD
1981). The Big Thompson Canyon flood was not identified as an extreme single-day event for
eastern Colorado in either the instrumental or reconstructed data, but that 1976 event occurred to
the northwest of the study area and was likely recorded in the last-formed latewood cells of
ponderosa pine from the vicinity.

The wettest days in midsummer tend to be part of organized 24 to 48-hour rainfall events
that occur at the synoptic scale. The synoptic meteorology of the five highest single day rainfall
extremes is illustrated with composite maps (Figs. 10a,b). Thunderstorm activity that produces
the wettest days over eastern Colorado in late-July are often associated with a major frontal
boundary located in the central Great Plains, and a surface anticyclone behind the frontal
boundary over the northern U.S. and southern Canada (Fig. 10a). The surface front separates the
cool, northeasterly winds flowing into Colorado and Wyoming from the warmer, moist air from
the Gulf of Mexico to the south. These surface conditions tend to increase moisture advection in
the lower levels of the atmosphere and enhance easterly upslope flow along the eastern Rocky Mountains (Doesken and Mckee 1998). At the 500mb level, a slightly negatively tilted upper-level ridge axis extends from the central Great Plains northwestward into southern Canada (the so-called ‘bent-back ridge’; Maddox et. al 1978). The negatively tilted ridge can contribute to deep southeasterly flow against the eastern Rocky Mountains (Fig. 10b; Cotton et al. 2003). These synoptic conditions generally produce significant hourly and single day precipitation totals over eastern Colorado (Figs. 6a-f), much more intense and widespread than the localized daily convective activity typical of midsummer.

4. DISCUSSION AND CONCLUSIONS

These results demonstrate that the precipitation response of ring width data in ponderosa pine can span a large range of time scales, from the annually integrated precipitation signal of total ring width to the daily precipitation extremes recorded by some adjusted latewood width chronologies. The seasonal to annual precipitation signal recorded by both earlywood and total ring width chronologies has been widely applied, and this seasonal to near annual integration of climate signal in ring width data has been the ruling paradigm of dendroclimatology for 100 years (Douglass 1920). However, adjusted latewood width chronologies of ponderosa pine from eastern Colorado are highly correlated with precipitation at the daily timescale and demonstrate the feasibility for the tree-ring reconstruction of weather timescale precipitation totals, or dendrometeorology.

The precipitation signals recorded by the last formed latewood cells in ponderosa pine or other species in North America have not been thoroughly explored. The few studies that have used adjusted latewood chronologies for moisture reconstruction have been calibrated with
monthly or seasonal moisture data (e.g., Stahle et al. 2009, 2015). Adjusted latewood chronologies also represent just one of several types of tree-ring data that partition the annual growth ring into sub-seasonal timescales. Sub-annual tree-ring data derived with x-ray densitometry (Schweingruber et al. 1978), blue-light intensity (Campbell et al. 2007), or stable isotopes (McCarroll et al. 2014) might potentially record precipitation variability at the daily or weekly time scale, especially in semiarid regions where single events dominate the seasonal totals.

Midsummer is the wettest time of the year over eastern Colorado when daily precipitation rates are highest and the probability of severe weather and dangerous flash flooding is greatest (McKee and Doesken 1990). In fact, the single wettest day in a two-week window of late-July dominates the midsummer precipitation totals over eastern Colorado. These one-day totals in midsummer may be the most important source of soil moisture recharge needed to promote the final latewood cell formation in arid site ponderosa pine of eastern Colorado. Oxygen isotope measurements of the nearly full latewood from mature ponderosa pine in northern Arizona indicate a reliance on winter moisture (Kerhoulas et al. 2017), which is similar to the results shown for the unadjusted latewood width data in Figure 5. The response to single day rainfall events in midsummer over eastern Colorado was only revealed after isolating the last few latewood cells formed at the end of the growing season with the calculation of adjusted latewood width.

The most extreme of these single day rainfall events can generate flash flooding and tend to result from unseasonable midsummer atmospheric circulation over western North America. Maddox et al. (1980) defined the synoptic meteorological conditions similar to those described here for midsummer rainfall extremes as “Type I” events that most commonly lead to significant
and widespread rainfall over eastern Colorado in July and August. These conditions include an upper-level ridge over western North America with an axis tilted to the northwest, a slow-moving surface frontal boundary over the central United States, and an increase in atmospheric moisture advection from the Gulf of Mexico and Pacific. These “Type I” synoptic conditions have been responsible for some of the largest hourly and daily precipitation totals in Colorado (Maddox et al. 1980, Cotton et al. 2003), and have been linked with other notable summertime flash flooding events in western North America, including the Rapid City, South Dakota, flood of June 1972, the Las Vegas, Nevada, flood in July of 1975 (Maddox et al. 1978; 1980), and the June 2013 flash floods in Calgary, Alberta (Milrad et al. 2015).

The 239-year reconstruction of the highest single day rainfall totals in midsummer provides an interesting long-term perspective on daily rainfall extremes over eastern Colorado. Extreme single day rainfall events are estimated to have increased in midsummer since the late 18th century and the period from 1960-1999 had the highest frequency of these extremes over the last 239 years. Coupled with rapid urban development along the Front Range, some of these heavy rainfall extremes led to disastrous flash flooding during the late-20th century. Many of these notorious flash floods occurred on the days identified as the single heaviest daily rainfall total in the eastern Colorado study area. Ponderosa pine and other semiarid conifer species are native to the drainage basins above several cities in the American West that are prone to midsummer flash flooding. Development of adjusted latewood width chronologies in these regions could be useful for analyses of change in the frequency of midsummer precipitation extremes.
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REFERENCES


**APPENDIX**

*Table 1.* The calibration and validation statistics computed for the tree-ring reconstruction of one-day precipitation totals in late-July over eastern Colorado are listed (a). The same statistics for an experimental reconstruction using an average of the largest one-day rainfall totals among the four weather stations closest to the tree-ring collection sites are also listed (b; Byers, Denver, Limon, Pueblo; July 19 to August 1). The statistics include the coefficient of determination, $R^2$, adjusted downward for loss of degrees of freedom (Draper and Smith 1981); the Pearson product moment correlation coefficient, $r$ (Draper and Smith 1981), reduction of error, RE (Fritts 2001); coefficient of efficiency, CE (Cook and Kairiukstis 1990).

**a. Regional LWa chronology**

<table>
<thead>
<tr>
<th>Calibration Period</th>
<th>Adj. $R^2$</th>
<th>Validation Period</th>
<th>$r$</th>
<th>RE</th>
<th>CE</th>
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</thead>
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<td>1948-1972</td>
<td>0.75</td>
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<td>0.49</td>
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<td>1973-1997</td>
<td>0.87</td>
<td>0.71</td>
<td>0.70</td>
</tr>
<tr>
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<td>-----------------</td>
<td>0.83</td>
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**b. Four station single day precipitation extremes**

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<th>Adj. $R^2$</th>
<th>Validation Period</th>
<th>$r$</th>
<th>RE</th>
<th>CE</th>
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<tr>
<td>1948-1997</td>
<td>0.64</td>
<td>-----------------</td>
<td>0.79</td>
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Fig. 1. The daily mean precipitation totals for 1948-2017 are plotted for a regional average of eastern Colorado from the gridded daily precipitation data in the CPC .25x.25 Daily U.S. Unified Gauge-Based Analysis of Precipitation dataset [37.75°N-39.75N°, 105°W-103°W; black box on the map inset]. The red line is a 10-day moving average of the daily means. Note the two peaks in the climatology of daily rainfall during late-spring and midsummer.
Fig. 2. (a) The monthly mean 850mb wind speed and direction (wind vectors; m/s) and meridional wind speed (shaded; m/s) are plotted for June, July, and August. Note the strengthening of the low-level meridional winds over the southern Great Plains in July. (b) The month to month differences in percent relative humidity are mapped. Note the advection of moisture across the southwestern United States and into eastern Colorado from June to July. All maps are based on the 20th Century Reanalysis dataset (Compo et al. 2011).
Fig. 3. This map notes the locations of the ponderosa pine stands used to develop the adjusted latewood width chronologies (open circles labeled BFE, JFU, RIR, TCU and defined in text). The black dots are the locations of the 40 instrumental precipitation stations (station metadata available on request). The box outlines the region with gridded instrumental precipitation data for the eastern Colorado study area [37.75°N-39.75°N, 105°W-103°W].
Fig. 4. (a) The correlations between the regional LWa chronology and the *wettest single day precipitation total* over the eastern Colorado study area during non-overlapping 14-day intervals from 1948-1997 are plotted by Julian date of the year (red line; significance thresholds noted). The highest correlation was computed for the wettest daily total during the 14-day interval extending from Julian day 200 to 213 (July 19 to August 1; $r = 0.83$). The vertical dashed lines denote this best 14-day interval (correlations are plotted at the end date of the 14-day interval). The correlation between the regional LWa chronology and precipitation *totals* accumulated for all possible 7 to 56 day periods during the year are also plotted (gray). The highest LWa correlation with 14-day total rainfall is also from Julian day 200 to 213 ($r = 0.78$), but still below the correlation with the single wettest day during this interval ($r = 0.83$). (b) Box and whisker plot of the non-overlapping bi-weekly precipitation climatology for the study area. The mean, upper, and lower quartiles are plotted in each box, along with the highest extreme values (dots; 1948-2017). Note that the highest two-week mean precipitation also corresponds with the two-week interval in (a) when the LWa chronologies are most highly correlated with daily precipitation (i.e., late-July).
Fig. 5. The correlation of the LWa chronology with all possible non-overlapping 14-day intervals from 1948-1997, which peaks in late-July (red line; from Fig. 4a), is compared with the correlations between the regional chronologies of EW, LW, and RW for eastern Colorado and all continuous 300-day precipitation totals throughout the year (gray and black lines). The gridded daily precipitation data were totaled for all possible 300-day intervals beginning in the previous year and ending on the Julian date noted on the x-axis during the current year for 1948-1997 (January 1 = previous Julian day 66 to current day 1). Note the strong near-annual precipitation signals integrated in the EW, LW, and RW chronologies, all of which include significant correlations during winter, spring, and summer. However, the LWa chronology is strongly correlated with single day precipitation totals only in late-July.
Fig. 6. The gridded instrumental precipitation totals are plotted for the five wettest (a-e) and driest (g-k) single days identified with instrumental data in eastern Colorado during late-July from 1948-1997 (the wet and dry composites are also mapped in f and l). The rank of each one-day precipitation total is listed in parentheses.
Fig 7. (a) The instrumental and reconstructed highest single day precipitation totals in late-July are plotted from 1948-2017 with the instrumental mean. The frequency distributions of the instrumental (b) and reconstructed (c) highest single day precipitation totals in late-July are also illustrated, based on the periods 1948-2017 and 1779-1997, respectively.
Fig. 8. The highest one-day precipitation totals in midsummer were reconstructed from 1779-1997 (gray) and the instrumental data were appended from 1998-2017 (dashed). A smoothed version of the reconstruction that highlights decadal variability is plotted in black. The mean, 90th, and 10th percentile thresholds for 1779-2017 are plotted and extremes above or below these thresholds are noted (asterisks). The major flash floods events that occurred in eastern Colorado on July 24, 1965; July 19, 1985; and July 28, 1997 are also indicated (squares).
Fig. 9. The highest single day precipitation totals identified at the 40 instrumental weather stations during midsummer are each plotted from 1940-2017 (i.e., July 19 to August 1; gray). The maximum single day total identified at any of the 40 stations during this 14-day interval is also plotted (dark blue) and is highly correlated with the reconstruction based on the largest one-day rainfall total in the gridded data. Extremes above or below the 90th and 10th percentile thresholds are indicated (asterisks).
Fig. 10. (a) The composite surface temperature anomaly (°K; shaded) and 850mb wind speed and direction (m/s wind vectors) for the five highest reconstructed single day precipitation totals in midsummer (i.e., July 24, 1965; August 1, 1966; July 20, 1973; July 19, 1985; July 24, 1991). Note the contrast in surface air temperatures from the Northern to Southern Plains and the area of low level convergence over the central United States. (b) Mean 500mb geopotential heights (contours) and 700mb relative humidity anomalies (%) are mapped for these five most extreme wet days of midsummer in the reconstruction. Note the northwest-tilted ridge axis over the Northern Rockies and the elevated moisture levels over the central Rockies and High Plains.
CHAPTER 4

The independent latewood growth response to warm season moisture over the southwestern United States
ABSTRACT

The North American Monsoon System (NAMS) has been a major focus of paleoclimate research. Tree-ring chronologies of latewood (LW) or adjusted latewood (LWa) width from semi-arid conifers from the Southwest have shown to have significant correlations with summer precipitation. We expand upon these previous studies by examining the tree growth response to the summer precipitation using daily precipitation data and 69 LWa chronologies of Douglas-fir and ponderosa pine across the Southwest physiographic region. Our results indicate that all LWa chronologies are significantly correlated with precipitation accumulated over some interval during the warm season from April-September, but there are substantial geographic and species-related differences in moisture response. Ponderosa pine across much of the Southwest is tightly coupled to the start of the local monsoon season and is best correlated with precipitation totaled over time periods of less than a month primarily in late-June and July. On average, Douglas-fir is best correlated with precipitation totaled over monthly to bi-monthly intervals, but the timing of the best tree growth responses and to moisture and relationship with monsoon onset varies geographically and reflects the importance of the NAMS to the regional hydroclimate. Our findings suggest that future tree-ring reconstructions of monsoon precipitation across the NAMS may be improved using daily instrumental data to identify the timing and length of the precipitation intervals that have the highest correlation with the tree-ring chronologies.
1. INTRODUCTION

Warm season precipitation associated with the North American Monsoon system (NAMS) is a vital water resource to the montane ecosystems of northern Mexico and the southwestern United States. The contribution of monsoon precipitation is highest over the Sierra Madre Occidental in northern Mexico extending northward into southern Arizona and New Mexico (Douglas et al. 1993), but the influence of the monsoon circulation may encompass much of the western North American plateau (Hales 1972; Tang and Reiter 1984; Reiter and Tang 1984). Since the mid-20th century, significant changes in the seasonality of monsoon rainfall has been identified over much of the Southwest, with progressively later onset dates and slight declines in July precipitation (Grantz et al. 2007). Climate model projections suggest a continued drying in the early-monsoon season over much of Mexico and the far southwestern United States, but slight increases at the northern peripheries of the NAMS. Late-season monsoon rainfall (i.e. September and October) across the NAMS is expected to significantly increase for (Cook and Seager 2013). Shifts in the distribution of precipitation during the monsoon season will likely impact ecological systems of the Southwest that are reliant upon this growing season moisture source (Ray et al. 2007), underscoring the need to understand the importance of monsoon precipitation and seasonality to plant growth and forest productivity.

The NAMS initially forms over the Sierra Madre Occidental and Mexican plateau in response to the land-ocean pressure gradient that develops during late-spring. As the Northern Hemisphere transitions from spring to summer, the NAMS migrates northward, generally impacting the southwestern United States beginning in the latter half of June (Adams and Comrie 1997; Barlow et al. 1998). The southerly winds of the monsoon circulation advect low-level moisture from the Gulf of California and eastern Pacific Ocean, as well as mid-level atmospheric
moisture from the Gulf of Mexico into the Southwest (Higgins et al. 1997; Wright et al. 2001). The higher atmospheric humidity creates an environment conducive for daily afternoon and evening thunderstorm activity that in some years can last into late-September and October (Adams and Comrie 1997; Higgins et al. 1997; Ellis et al. 2004).

Tree-ring chronologies of annual or total-ring width (TRW) have frequently been used to understand the importance of climate to tree growth and make inferences about long-term hydroclimatic change and variability. The climate sensitivity of annual growth rings reflects environmental conditions most limiting to plant growth, including the local to regional hydroclimate, local site conditions, and physiological processes inherent to a given tree species (Fritts 1976). Ring width chronologies tend to have the highest correlations with climate variations aggregated across several months and seasons (Douglass 1920; Blasing and Fritts 1976). In many temperate North American tree species, the annual growth ring can be further divided into sub-annual earlywood (EW) and latewood (LW) growth components, which can isolate or enhance seasonal climate signals that are not recoverable using the more integrative proxy of TRW. In Mexico and the southwestern United States for example, chronologies of EW width from semi-arid conifers have the highest correlation with moisture during the winter and spring seasons (Cleveleand et al. 2003, Villaneuva-Diaz et al. 2007; Stahle et al. 2009, Woodhouse et al. 2013), but LW and the so-called adjusted latewood (LWa) width chronologies are generally most responsive to summer precipitation associated with the NAMS (Meko and Baisan 2001; Therrell et al. 2002; Stahle et al. 2009; Faulstich et al. 2013; Griffin 2013; Griffin et al. 2013). The seasonal climate signals encoded in EW and LW growth chronologies have also been examined for Douglas-fir and ponderosa pine sites in western Canada (Watson and Luckman 2002), the interior Rocky Mountains (Crawford et al. 2015), the western Great Plains
(Howard et al. 2019), as well with oak (*Quercus*) and short-leaf pine (*Pinus echinata*) in the south-central United States (Torbenson and Stahle 2018). The seasonal moisture variable best correlated with the EW and LW width chronologies varies by region, but a general consensus seems to be that EW is positively correlated with precipitation accumulated prior to and during the regional early growing season, and LW has the highest correlation with summer rainfall. Adjusted LW chronologies, which represents LW growth independent of previous growing season conditions, seems to be most responsive to summer precipitation summed over shorter timescales (i.e. less than seasonal) compared to EW, LW, and TRW width chronologies (Stahle et al. 2009; Griffin et al. 2011; Griffin 2013)

In dendroclimatology, climate-response analyses most often utilize instrumental climate variables calculated on monthly, seasonal, or annual timescales. This is somewhat of a limitation, since tree growth response to environmental conditions is not necessarily constricted to human constructed units of time. The use of daily data can assuage this issue to a degree, assuming instrumental stations or gridded products are available near the tree-ring site and the records extend far enough back in time to robustly compare tree growth with the target variable. But only a few studies have used daily instrumental data in climate-response analyses or tree-ring reconstructions (e.g. Schönbein 2011; Beck et al. 2013; Land et al. 2017). The results from these studies indicate that the growing season moisture and temperature intervals encoded in various tree-ring metrics are not constricted to the beginning and end dates of calendar months or seasons. The use of daily data can often improve the correlations with various tree-ring metrics and reveal long-term multidecadal changes in the tree growth response to growing season climate conditions (Beck et al. 2013). A similar type of analysis would be useful in the
southwestern United States where a distinguishable summer wet season substantially impacts the forested ecosystems of the region,

In this study, we examine the tree growth response to precipitation using g69 Douglas-fir and ponderosa pine LWa chronologies from the southwestern United States using gridded daily precipitation data. The objectives of this study were to determine the length (in days) and timing of the precipitation signals present in each LWa chronology and to assess the local to regional-scale differences in moisture response. Additionally, we define the average onset of the monsoon across the Southwest and determine whether the precipitation response of independent LW growth formation is related to the start of the local summer wet season. The results from this study demonstrate the utility of using daily precipitation data in climate-response analyses, and also highlights the varying growth response to warm season moisture present in this subset of tree-ring data from the Southwest.

2. DATA AND METHODS

2.1 Study region

The 69 sites encompass a diverse physiographic region of the Southwest (Fig. 1; Table 1), with the annual precipitation climatology varying both latitudinally and longitudinally over the study domain (Figs. 2 and 3). Summer monsoon moisture in July and August is the most important source of precipitation to the sites located in the intermontane plateau of southern Arizona and New Mexico, contributing over 50% of the annual total (Figs. 2c,d) with a pronounced peak in daily rainfall totals occurring in early-summer (Fig. 3f). Northward over the Colorado Plateau and Four Corners region the precipitation climatology has a well-defined bimodal regime with the greatest contributions coming from winter and spring moisture
associated with Pacific storms (Fig. 2a; Hirschbeck 1991), and a weaker, but identifiable summer monsoon season (Figs. 2c,d, 3d,e). Eastward over the Rocky Mountain Front Range and High Plains, spring and summer precipitation constitutes over 70% of the annual total (Fig. 2a-d; Mahoney et al. 2015) and is divided into two wet periods occurring in late-spring and mid-summer separated by a drier early-summer (Fig. 3a).

2.2 LWa chronology development

Forty-two Douglas-fir and 27 ponderosa sites with paired EW and LW width data from the Southwest were used to calculate the LWa chronologies (Fig. 1; Table 1). The majority of these sites were sampled and measured by previous researchers (Stahle et al. 2009; Griffin et al. 2011), but collections from one Douglas-fir and 12 ponderosa pine sites in eastern Colorado and New Mexico were remeasured for EW and LW width (Howard et al. 2019). The EW and LW data were standardized to produce mean ring-width index chronologies fit with autoregressive models to remove low frequency growth persistence not related to climate (Meko 1981; Cook 1985). Though EW and LW width chronologies tend to retain separate seasonal climate information and have been used to separately reconstruct cool or warm season precipitation (e.g. Cleaveland et al. 2003; Griffin 2013), the correlation between EW and LW at a given site can be so high and distort any separate summertime signal that might be recorded in LW width. The physiologically persistence between EW and LW can be removed using regression techniques as described by Meko and Baisan (2001). In their analysis, Meko and Baisan (2001) regressed the predictand LW on the predictor EW chronologies to derive the LW regression residuals that are uncorrelated with the EW chronology (i.e., LWa). This manipulation works reasonably well in isolating the variability of LW growth that theoretically should contain a ‘discrete’ climate signal
that is independent of previous biological and climate conditions (Griffin et al. 2011). Simple linear regression was performed using the EW and LW chronologies from each site over the common interval of 1850-1991 to produce the LWa width data.

2.3 Instrumental climate data

The gridded daily data used in this study were acquired from the National Oceanic and Atmospheric Administration (NOAA) Climate Prediction Center’s (CPC) unified gauge-based dataset for the continental United States (Chen et al. 2008). The gridded data are calculated based on the optimal interpolation (OI) algorithm using a dense network of observing sites and contain daily precipitation values on a 0.25° latitude by 0.25° longitude grid extending from 1948 to present (Chen et al. 2008). The precipitation daily totals are calculated for the 24-hour period ending at 12:00 GMT.

2.4 Precipitation interval response analysis

A precipitation response correlation analysis was performed with each of the 69 LWa chronologies and the gridded daily precipitation data to determine the length and timing of the precipitation intervals best correlated with the tree-ring data. To derive time series of precipitation data summed for different interval lengths, the eight grid points surrounding each LWa chronology were extracted and subset into intervals ranging from 10 to 90 days for all 365 days of the year. Inclusion of precipitation totals less than 10 days often resulted in discontinuous annual time series and were excluded. It was also not necessary to examine the relationship with precipitation total over intervals greater than 90 days based on the results of previous climate response analyses using LWa chronologies (Stahle et al. 2009; Griffin et al. 2011; Griffin 2013).
The summed precipitation total for each interval and day of the year was then used to represent year \( t \) in an annual time series from 1948-2017. At a given site with eight surrounding grid points, this results in 265,720 annual time series that were then correlated interannually with the respective LWa data over the common overlap period of 1948-1991. From among the eight grid points, the highest r-value from all interval lengths for every day of the year was then identified. The precipitation interval and associated beginning and ending Julian day of the year that had the highest r-value with the respective LWa chronology was then selected from this subset.

### 2.5 Calculation of monsoon onset date and the 50th percentile of monsoon rainfall

Calculation of monsoon onset has been analytically derived using a variety of methods for different regions impacted by a monsoon season. In North America, Higgins et al. (1997) analyzed the onset of the monsoon season using a threshold-crossing procedure with 5-day smoothed cumulative precipitation indices for three different regionally-averaged areas of the Southwest. Higgins et al. (1997) defined the onset of the monsoon season over southern Arizona each year as the ending date of three consecutive days of +0.5 mm/day increases in cumulative daily precipitation rates. This duration and magnitude of precipitation increases had to be modified for each region because of the differences in evolution of daily rainfall rates across the Southwest. Ellis et al. (2004) used a similar method to define the North American monsoon over Arizona, New Mexico, and portions of southern Colorado, but used the threshold-crossing procedure with daily dew point temperature data to better capture the atmospheric moisture changes associated with seasonal shifts in synoptic-scale circulation patterns. Cook and Buckley (2009) used a two-phase linear regression approach, which objectively identifies significant
change points in time series data, to define both onset and demise of the Indian monsoon using cumulative daily precipitation station data.

For this analysis we identified the closest grid point to each tree-ring site and produced time series of the climatological daily precipitation rates (mm/day) based on the 1948-2017 average and used the approach of Cook and Buckley (2009) to calculate the monsoon onset dates for each of the 69 instrumental daily time series. Because some locations have a bimodal warm season precipitation regime where both spring and summer precipitation constitute a substantial percentage to the annual total [e.g. the Front Range and High Plains of Colorado and New Mexico], inclusion of days prior to Julian day 152 (June 1st) resulted in the detection of significant change points prior to the traditional summer monsoon season. Therefore, the two-phase linear regression was only applied to the daily precipitation data between Julian days 152 through 243 (August 31st). We hypothesized that the beginning Julian dates of the precipitation intervals best correlated with the LWa chronologies would be similar to the onset of the local monsoon season. To test this, the average monsoon onset date based on the LWa data ("tree-ring onset date") was defined as the beginning Julian date of the precipitation interval that had the highest correlation with the respective tree-ring chronology. This date was then differenced from the onset date calculated from the instrumental station data.

We additionally determined the average number of days required to reach the 50th percentile of total monsoon rainfall for those sites where the tree-ring onset date was within 14 days of the instrumental date. Average total monsoon rainfall was defined as precipitation accumulated from the local onset date calculated at the nearest grid point to 60 days after. The number of days required to reach the 50th percentile were then compared to the average length of the precipitation intervals best correlated with the LWa chronologies.
3. RESULTS

3.1 The length of the precipitation signals encoded in independent latewood growth

Results from the precipitation response interval analysis reveals that all LWa chronologies from the Southwest are significantly (p < 0.05) correlated with precipitation accumulated over timescales of around one month (Figs. 4a, b). Thirty-nine LWa chronologies have the highest correlation with precipitation summed over 31 days or less, and 23 of these chronologies contain a strong signal with precipitation summed over less than three weeks (Figs. 4a, b). A higher percentage of ponderosa pine LWa chronologies have the highest correlation with total precipitation summed over sub-monthly timescales compared to the Douglas-fir. The average length of the best moisture interval for all ponderosa pine LWa chronologies is approximately 25.80 days, significantly shorter (p < 0.05 based on a Student’s T test) than the 35.40 day average of the Douglas-fir chronologies.

The length of the best moisture signals may in part be related to the amount of independent variability inherent in each LWa chronology calculated from the linear regression between the EW and LW data. Correlation analyses using the EW and unadjusted LW width data was done for each site from 1850-1991, and these r-values were then correlated with the interval lengths determined from the precipitation response interval analysis (Fig. 4c). The average correlation between each pair of EW and LW chronologies is 0.71, indicating that 50% of the LW variability can be explained by the antecedent growth of EW (Table 1). However, the degree of independent variability is not significantly related to interval length (r = -0.14; Fig. 4c), therefore the length of these precipitation intervals do not appear to be a statistical artifact resulting from the linear regression procedure.
Spatially, the ponderosa pine LWa chronologies that share a robust correlation with sub-monthly moisture primarily cluster over the central and eastern Rocky Mountains, extending eastward across the Front Range and western High Plains of Colorado and New Mexico (Fig. 5a,b). Most of the ponderosa pine LWa chronologies in far southeastern Colorado and northeastern New Mexico are highly correlated with precipitation totals summed over timescales less than three weeks (Fig. 5a). The Ponderosa pine LWa chronologies in eastern Arizona and western New Mexico appear to have a slightly more integrated moisture response, and the highest r-values are with precipitation totals summed over intervals greater than 31 days (Fig. 5c). The Douglas-fir sites with sub-monthly moisture signals are largely located in southern Arizona and New Mexico, as well as within the Four Corners region (Fig. 5d and 5e). Douglas-fir sites with a moisture signal longer than a month are prevalent throughout the study region, but most are present in central and western New Mexico (Fig. 5f).

3.2 Average timing of the warm season precipitation signals

The 69 LWa chronologies are all significantly correlated with warm season moisture, primarily during the early to mid-summer months when the influence of the NAMS over the Southwest is greatest. Plotted in Figure 6 are histograms of the beginning Julian dates associated with the best precipitation interval for each site separated by species. The beginning Julian dates for 47 of the 69 LWa chronologies occur after the first day of summer (Julian day 172; Figs. 6a and 6b). All but two ponderosa pine LWa chronologies contain a summer moisture signal (Fig. 6a), but Douglas-fir is equally split between chronologies correlated with precipitation beginning prior to and after the first day of summer (Fig. 6b).
Despite the diverse geographic range of ponderosa pine sites across the Southwest, the beginning dates of the best precipitation intervals are largely synchronized to the early-summer period (Fig. 6c). The Douglas-fir LWa chronologies exhibit much greater geographic variability, particularly along a latitudinal gradient in eastern Arizona and western New Mexico (Fig. 6d). South of approximately 34°N and west of the eastern New Mexico border [-103°W], the beginning Julian dates for 18 out of the 23 Douglas-fir chronologies all occur after the first day of summer. But north of 34°N and west of -105.5°W, the beginning Julian dates for these 15 Douglas-fir sites occur prior to June 21st. The beginning dates also vary with longitude over southern Arizona and New Mexico, with later dates present in the western sites gradually transitioning to earlier dates in central and eastern New Mexico (Fig. 6d).

3.3 Monsoon onset and its relation to tree growth

The start dates of the local monsoon season varies geographically over the Southwest (Fig. 7), and this appears to influence the timing of the best precipitation intervals for many of the LWa chronologies. Monsoon onset over the entire study region has a 28-day range between Julian dates 171 (June 20th) and 198 (July 19th), with the earliest dates present at grid points in southeastern New Mexico and far southwest Texas (Fig. 7). Over the region most impacted by the monsoon in southern Arizona and southwestern New Mexico, onset dates are slightly later ranging from the last week of June to the first week of July. These dates become progressively later northward, beginning as late as July 19th at the most northern sites in Colorado. The monsoon dates calculated in this study match well with previous studies (i.e. Higgins et al. 1998; Ellis et al. 2004), and indicate two-phase linear regression is an appropriate analytical technique for determining the start dates of the summer monsoon over North America.
Identifying the LWa chronologies with a tree-ring onset date within 14 days of the instrumental onset date highlights the geographic differences in tree growth response to the start of monsoon season. There are 45 LWa chronologies that appear to be connected to the start of the local monsoon season (Fig. 8a). Most of these sites are located south of 34°N in southeastern Arizona and southern New Mexico, and northward over the Front Range and High Plains of Colorado and New Mexico. The LWa chronologies not related to the start of the monsoon season primarily occupy the area north of 34°N and west of -105.5°W in the Colorado Plateau region and west-central Arizona (Fig. 8b). It is worth noting that there is only one site with a tree-ring onset date that is 14 days later than the local monsoon onset.

Independent LW growth of ponderosa pine in the Southwest appears to have a stronger connection to the start of the monsoon season than Douglas-fir. Only two ponderosa pine LWa chronologies contain a tree-ring onset date that is not within 14 days of the instrumental onset date; and these sites happen to have the two of the weakest precipitation signals in the entire network (sites BPM and GPU; Table 1; Fig. 9a). The 42 Douglas-fir chronologies are once again equally split between those connected to the start of the local monsoon season in southern Arizona and New Mexico, and those with a spring and early-summer moisture signal largely north of 34°N in the Colorado Plateau region (Fig. 9b).

4. DISCUSSION

4.1 Possible mechanisms influencing the length of the precipitation signals

The length of the precipitation intervals best correlated with the LWa chronologies do not appear to be influenced by the amount of independent LW growth variability extracted from the simple linear regression models (Fig. 4c). Differences may arise from a combination of
physiological processes inherent to the plant and exogenous environmental variables related to local and regional precipitation climatology. The significant differences in average interval length between Douglas-fir and ponderosa pine over the entire study region suggest biological mechanisms may influence the moisture signals (Fig. 4a,b). However, these species-dependent differences could reflect the location of these chronologies and the characteristics of the regional hydroclimate.

As an example, the 45 LWa chronologies related to the onset of the local monsoon are largely correlated with sub-monthly to monthly precipitation accumulated during the early monsoon season. The average duration of the early monsoon season, calculated as the number of days required to reach the 50th percentile of total monsoon precipitation, is plotted for the 45 sites in Figure 10. The geography of early monsoon duration reflects a similar pattern to that shown in Figure 4 related to the length of the precipitation intervals that have the highest correlations with the tree-ring data. Over eastern Colorado and New Mexico the LWa chronologies (which are primarily ponderosa pine) are largely correlated with sub-monthly moisture totals (Fig. 4a-d), corresponding with a lower average number of days needed to reach the 50th percentile of monsoon rainfall (Fig. 10). This relatively short summer rainfall maximum from mid-July to mid-August has been well documented, and results from a narrow period of the summer season when Gulf of Mexico and Pacific moisture advect into the region and interact with the complex topography of the eastern Rocky Mountains and synoptic systems to generate especially heavy precipitation events (Hansen et al. 1978; Weaver and Doesken 1990). The LWa chronologies over west-central New Mexico and far southern Colorado (nearly all Douglas-fir sites) tend to be correlated with precipitation over longer (i.e. greater than monthly) timescales (Fig. 4a-d), which coincides with
an early monsoon season that appears to last longer compared to any other region of the Southwest (Figs. 3b,c; Fig. 10)

4.2 **Possible mechanisms influencing the timing of the precipitation signals**

All of the 69 LWa chronologies are significantly correlated with precipitation accumulated over some interval during the warm season months, and the majority are most responsive to summer moisture associated with the NAMS. Higher contributions of July precipitation seem to be indicator of which sites are potential proxies for monsoon rainfall. The spatial pattern of July precipitation contribution across the Southwest, with higher percentages in southern Arizona and New Mexico that extends northeastward into eastern Colorado (Fig. 2c), is quite similar to the spatial pattern of LWa chronologies with a tree-ring onset date that is within 14 days of the local onset date (Fig. 8a). Over the Colorado Plateau and west-central Arizona the July precipitation contribution is considerably less (Fig. 2c), and summer rains are primarily delivered as infrequent surges of moisture during periods when the northern boundary of the NAMS expands (Higgins et al. 1997; Castro et al. 2001). With the summer monsoon being a seemingly less reliable year-to-year water source, independent LW growth at these sites appear to reflect a moisture response that integrates precipitation over a period referred to as the “arid-fore” during the late-spring and early-summer months. These interpretations are similar to the findings of Fritts et al. (1965), who observed at Mesa Verde, Colorado that Douglas-fir LW growth largely terminated in mid- to late-June prior to the traditional monsoon season. It is unclear whether this geographic gradient is inherent only to Douglas-fir, though the few ponderosa pine sites within or near the Colorado Plateau seem to exhibit a pure monsoon moisture signal (Fig. 9a). However, latitudinal differences in the stable oxygen and carbon
isotopic composition of ponderosa pine LW has been identified over a transect of Arizona and Utah, likely reflecting the northward decrease in the intensity of the monsoon (Sjezner et al. 2016; 2018).

5. CONCLUSIONS

The use of the daily precipitation data allows for the objective determination of the length and timing of the precipitation intervals that have the highest correlations with independent LW growth chronologies from the Southwest. As our results indicate, using daily precipitation data can refine the climate signals encoded in different tree-ring metrics, particularly in a region like the Southwest with well-defined step-like changes in precipitation totals associated with the onset of the summer monsoon. With the exception of sites located in the Colorado Plateau region and west-central Arizona, the onset of the local monsoon, and the duration of the early-monsoon season appear to be the most important mechanisms influencing the length and timing of the moisture signals present in the LWa chronologies regardless of species. Future reconstructions of monsoon precipitation for different regions of the NAMS could potentially be improved by accounting for local onset dates and the moisture intervals most important to plant growth.

It is important to note that these precipitation intervals are likely dynamic and can vary from year-to-year depending upon the timing and length of the monsoon season, and perhaps endogenous biological variables that influence the duration of cambial activity during the warm season (Wimmer 2002). The length and timing of these precipitation intervals are likely subject to longer, multidecadal changes as well (Land et al. 2017), perhaps related to the long-term climate variability (Carrer and Urbinati 2006), stand dynamics (Laskurain et al. 2018), and the age of individual trees that comprise a given chronology (Carrer and Urbinati 2004). The gridded daily
data set used in this study begins in 1948, therefore longer meteorological rainfall records are needed to address possible changes in the independent LW growth response to precipitation during the warm season. Incorporation of other environmental variables that influence cell growth and cambial activity that can be measured or calculated on daily timescales (i.e. temperature and the vapor pressure deficit) may reveal additional valuable information related to the tree growth response to warm season climate conditions.

The pre-dominant early season monsoon moisture signals suggest that tree growth in the Southwest may be sensitive to the changes in monsoon seasonality projected with continued anthropogenic warming (Cook and Seager 2013). The expected changes in early-season monsoon precipitation, combined with higher summer temperatures and increased evaporative demand (Williams et al. 2013) will potentially increase forest stress (McDowell et al. 2013), but the impacts of summer precipitation changes on the forested ecosystems of the Southwest will likely vary due to the geographic differences in warm season growth response. Continued development of EW, LW, and LWa chronologies more broadly over Southwest and Mexico is needed to further define the influence of the NAMS on tree growth and identify regions where the consequences of seasonal shifts in summer precipitation may be most impactful.

**ACKNOWLEDGEMENTS**

Thank you to Connie Woodhouse and Peter Brown for use of their tree-ring collections. I also thank my dissertation committee, David Stahle, Song Feng, Ralph Davis, and Fred Paillet for guidance and assistance. This study was funded by the National Science Foundation (grant #AGS-1266014).
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APPENDIX

Table 1 Tree-ring chronology metadata. (a) Also included are the r-values based on the correlations between the paired EW and LW chronologies for each site. Correlations were computed over the common overlap period of 1850-1991.

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Fig. 1 Site maps illustrating the geographic distribution of the (a) Ponderosa pine and (b) Douglas-fir LWA chronologies used in this analysis. Note there are several locations with multiple chronologies.
Fig. 2 (a). The percent contribution of December-May precipitation to the annual total calculated for each grid point. (b,c,d) Same as (a) but for (b) June, (c) July, (d) August. Note the influence of the NAMS beginning in July over much of Arizona, New Mexico, and extending northeastward into Colorado and the western Great Plains.
Fig. 3 Climatological daily precipitation rates (mm/day) calculated from the nearest grid points to six sites are shown for different regions of the Southwest, including (a) eastern Colorado (b) central New Mexico, (c) southern New Mexico, (d) southwestern Colorado, (e) north-central Arizona, and (f) southern Arizona. These series have been smoothed with a 5-day running mean (red line).
Fig. 4 (a) Plotted is a histogram of the interval lengths (in days) best correlated with each of the ponderosa pine (PIPO) LWa chronologies. (b) Same as (a) but for the Douglas-fir (PSME) chronologies. Note the differences in the average best interval length between the two species. (c) The paired EW and LW data for each site were correlated from 1850-1991, and the r-values from those correlations were then compared with the best precipitation intervals for each of the 69 LWa chronologies. The X,Y coordinate data are color coded by species (black = ponderosa pine, red = Douglas-fir). The amount of independent variability present in the LWa data does not appear to influence the length of the best precipitation intervals based on the insignificant r-value (r = -0.08).
Fig. 5 (a,b,c) Shown are the ponderosa pine LWa chronologies that are best correlated with precipitation over an interval of (a) less than 21 days, (b) between 21 and 31 days and (c) greater than 31 days. (d,e,f) Same as (a,b,c) but for the Douglas-fir sites.
Fig. 6 (a,b) Histograms of the beginning Julian day of the precipitation intervals best correlated with each LWa chronology are plotted for (a) ponderosa pine and (b) Douglas-fir. Note the relative uniform response to summer precipitation for the ponderosa pine LWa chronologies, but the more diverse timing of response to warm season precipitation for Douglas-fir. (c,d) The beginning Julian days are plotted geographically for (c) ponderosa pine and (d) Douglas-fir.
Fig. 7 The climatological monsoon onset date was determined for the grid point closest to each tree-ring site using two-phase linear regression (Cook and Buckley 2009) on daily average precipitation data from June 1st – September 1st. The average monsoon onset date is based on the 1948-2017 period.
Fig. 8 (a) Shown are the 45 LWa chronologies that are best correlated with a precipitation interval that begins within 14 days of the local monsoon onset calculated from the nearest grid point. (b) Plotted are the other 24 LWa chronologies with a tree-ring onset date 14 days outside the instrumental onset date.
Fig. 9 (a) Shown are the differences in monsoon onset date for all Ponderosa pine sites. (b) Same as (a) for Douglas-fir.
Fig. 10 The number of days required to reach the 50th percentile of monsoon precipitation, calculated from the local onset monsoon date to 60 days after, is plotted for the 47 sites containing that is best correlated with a precipitation interval that begins within 14 days of the instrumental onset date. Note the regional differences in the duration of the early-monsoon season.
CHAPTER 5

Conclusions
1. CONCLUSIONS

Sub-annual tree-ring chronologies of EW, LW, and LWa width have been vital for understanding the seasonal climate characteristics of pre-instrumental era droughts and pluvials over North America and the possible largescale ocean-atmospheric dynamics responsible for interannual to multidecadal oscillations in seasonal climate conditions. This dissertation research expanded the network over North America by developing EW, LW, LWa width chronologies across the western Great Plains. Taken together, the findings from this dissertation present the useful applications of sub-annual tree-ring chronologies, and further refine the methods for examining the tree growth responses to precipitation and reconstructing moisture variability at timescales as resolute as meteorological. The findings and primary contributions for each chapter of this dissertation are summarized below:

The development of EW, LW, and LWa width chronologies across the western Great Plains fills in a critical regional gap in sub-annual tree-ring data for one of the most drought prone regions on this planet (Karl 1986). These tree-ring records, many of which were sampled nearly two decades ago, were vital for chapters two and three of this dissertation. Hopefully, these data are used for additional research projects once they are made publicly available. Some of these chronologies have already been incorporated into larger continental-scale gridded reconstructions of cool and warm season precipitation variability (“The North American Seasonal Precipitation Atlas; Stahle et al. in prep).

The primary contribution of chapter two is the production of separate spring and summer Z-index reconstructions for two regions of the western Great Plains. These reconstructions reproduce the largescale patterns of seasonal ocean-atmospheric forcings seen with the instrumental data, indicating it is possible to assess the major mechanisms associated with long-
term variations in moisture during the two seasons when the bulk of the precipitation occurs. The spring and summer reconstructions reveal previously undocumented information about the seasonal characteristics of major pre-instrumental era droughts, and indicate that the 1930s Dust Bowl was the most extensive and intense summer drought episode to impact the two study regions, and possibly North America, in the last few centuries. Conversely, drought conditions during the mid- to late-19th century were more severe during the spring seasons, a finding reached by previous paleoclimate and climate modeling studies for different regions of North America (Herweijer et al. 2006; Torbenson and Stahle 2018). The varying spring and summer drought characteristics in both instrumental and tree-ring reconstructed data prove the value of developing EW and LW width chronologies for assessing seasonal climate variability, but also underscore the need to explore the ocean-atmospheric dynamics associated with these differing seasonal drought regimes perhaps using climate model experiments.

The discovery that ponderosa pine LWa chronologies from eastern Colorado are proxies for heavy one-day rainfall events during the wettest two weeks of the year in midsummer is by far the most important contribution of chapter three. At the very least, the analytical techniques used to derive daily meteorological precipitation signals demonstrates these methods can work in select areas where part of the annual growth ring forms in response to individual weather events rather than to variations in soil moisture over the length of a growing season. It would be pertinent to expand the results from chapter three, both temporally and spatially, across central and eastern Colorado to provide an extended multi-century history of late-July one-day rainfall events that can provide further paleoclimate context for some of the most devastating midsummer flash floods to impact major Colorado Front Range drainage basins and urban areas.
The preliminary results from the 238-year reconstruction indicate that the most extreme one-day midsummer rainfall events have been increasing over eastern Colorado. These findings are in line with previous studies assessing daily rainfall extremes since the early-20th century (Wuebbles et al. 2017). However, in the last 20-years there has been a noticeable decline in the frequency of one-day rainfall extremes, possibly linked with the ongoing early-21st century drought that has impacted much of the southwestern United States (Seager 2007). The apparent decadal to multidecadal oscillations in one-day midsummer rainfall totals is a surprising finding, and further research is needed to determine the possible land-surface and ocean-atmospheric mechanisms influencing low frequency variability in meteorological timescale weather events.

The use of daily data in climate-response analyses has not been applied widely in dendroclimatic research (Land et al. 2017). However, in a region like the Southwest with a distinguishable summer wet season that is known to impact forest productivity and plant growth (Meko and Baisan 2001; Stahle et al. 2009; Griffin 2011; Griffin 2013, Szejner et al. 2016), using daily data can reveal important information about the relationship between tree-ring chronologies and the timing and seasonality of the NAMS. The LWa chronologies located in areas of the Southwest where July precipitation contribution is higher tend to be connected to both the start of the monsoon and the precipitation that accumulates over the first half of the monsoon season. But where the monsoon has less of an influence on the regional precipitation climatology (i.e. the Colorado Plateau), there is a much weaker connection to the NAMS. This naturally has consequences related to the impacts of future changes in warm season precipitation on natural ecosystems of the Southwest. A broader network of LW and LWa chronologies over Mexico and the Southwest from multiple species is needed to expand this analysis and potentially produce regional reconstructions that account for the local onset dates and the
precipitation intervals most related to tree growth. These regionally “time-dependent”
reconstructions, which could extend from Mexico into the Southwest, may allow for an
interesting multi-century paleoclimate perspective on the strength and extent of the monsoon
across North America, and the possible largescale ocean-atmospheric forcings associated with
spatial and temporal variations of the NAMS.

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