SPATIOTEMPORAL MEASURES OF EXPOSURE AND SENSITIVITY TO
CLIMATIC VARIABILITY AND CHANGE: THE CASES OF MODERN SEA LEVEL
RISE AND SOUTHWESTERN U.S. BIOCLIMATE

by

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DEDICATION

I would like to dedicate this dissertation to Evelyn Weiss, who is as sweet as an angel and as tough as nails.
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ABSTRACT

Human activities are the main driver of environmental changes over the past 100-200 years, and threaten the stability of Earth’s environmental systems. One part of Earth’s environment already destabilized due to human activities is its climate. With human-caused changes to Earth’s climate expected to continue, questions arise as to which, where, and when impacts to human and natural systems might occur.

Understanding the vulnerability – the exposure, sensitivity, and resilience – of a system to changes in climate is essential to addressing these questions. This study represents an assessment of system vulnerability to climate change through the cases of modern sea level rise (SLR) and southwestern U.S. (SW) bioclimate. SLR is an important consequence of human-caused climate change, as higher seas have the potential to cause major social, environmental, and economic impacts. With a focus on sensitivity to SLR, we developed a new geospatial dataset that delineates low-lying coastal areas and overlaid this dataset with boundaries of U.S. cities with populations greater than 50,000 to determine areas prone to SLR impacts in this and subsequent centuries. Results demonstrate that potential SLR impacts to 180 U.S. cities will be very local and disproportionate. Recent warm and dry conditions have altered SW bioclimate, and expected further increases in regional temperatures raise concern that anomalous growing conditions will continue to occur and, in cases, worsen in the future. With a focus on exposure of SW vegetation to changing growing conditions, we compared the 1950s and 2000s droughts to take advantage of the opportunity to study mesoscale ecosystem responses to anomalously dry conditions before and during the regional warming. Higher
temperatures and evapotranspirational demand during the more recent drought altered the degree to which climate limited foliar growth. These climatic conditions reduced effects of suboptimal temperatures on foliar growth at lower elevations in winter and higher elevations in summer. They also increased constraints of evapotranspirational demand on foliar growth at lower and middle elevations from spring through summer. Free-tropospheric air temperatures, a strong influence on climate in mountainous areas, support statistical downscaling of projected SW temperatures to assess if and when similar or more anomalous conditions will occur in upcoming decades. Assessing the vulnerability of a system to changes in Earth’s climate like those taking place and projected to happen is a way for environmental sciences to help inform policy decisions that consequently stem from past or potential impacts of climatic hazards.
1. INTRODUCTION

Human activities have been the main driver of environmental changes over the past 100-200 years [Steffen et al., 2007]. Through actions such as expanding utilization of fossil fuels, deforestation, and industrialized agriculture, the human imprint on Earth’s natural environment has become both pervasive and profound. How human influence on the global environment will evolve in upcoming decades is uncertain, but there is growing recognition that its effect has threatened or may threaten the stability of various environmental conditions under which modern societies have developed and prospered [Rockström et al., 2009].

One part of Earth’s environment that has destabilized due to human activities is its climate [Solomon et al., 2007; Rockström et al., 2009]. Fossil fuel combustion and land use change largely have driven atmospheric concentrations of greenhouse gases (GHGs) well above their pre-industrial values. For instance, the atmospheric concentration of carbon dioxide is approaching 394 ppm [National Oceanic and Atmospheric Administration (NOAA); http://www.esrl.noaa.gov/gmd/ccgg/trends], well above its pre-industrial value of 280 ppm. Higher concentrations of GHGs increase radiative forcing, the balance of incoming and outgoing energy at the top of Earth’s atmosphere, resulting in changes to the global climate system such as warmer air and ocean temperatures, greater atmospheric water vapor content, and higher mean sea level [Le Treut et al., 2007; Solomon et al., 2007]. Several positive feedbacks in the climate system such as from increases in atmospheric water vapor and decreases in surface albedo from snow and sea ice loss can amplify warming and further these changes [Le
Warming temperatures, increasing atmospheric water vapor, higher seas, and other associated impacts such as greater numbers of heat waves, extreme precipitation events, and droughts already are occurring globally [Trenberth et al., 2007], due in part to increasing GHG concentrations [Hegerl et al., 2007]. Projections indicate that continued emissions of GHGs at current levels over the 21st century will further magnify these changes and impacts [Solomon et al., 2007].

With human-caused changes to Earth’s climate expected to continue [Solomon et al., 2007; Karl et al., 2009], questions arise as to which, where, and when impacts to human and natural systems might occur. For example, consider rising sea level and urban areas. How much higher will sea levels be by 2100? Which coastal cities might be impacted? Are potential impacts the same in every coastal city? Understanding the vulnerability of a system to changes in climate – that is, the climatic conditions that put people and places at risk of being faced with impacts, and the system qualities that increase or decrease their ability to withstand or respond to such conditions – is essential to addressing these questions [Cutter, 2003; Turner et al., 2003; Dawson et al., 2011].

Vulnerability of a system to human-caused climate change has three components: exposure; sensitivity; and adaptive capacity, or resilience [Turner et al., 2003; Dawson et al., 2011]. Exposure denotes the amount of change of climatic features such as temperature, precipitation, or sea level a system is experiencing presently or is projected to experience. It is a function of change attributes like magnitude, rate, and temporal variability. The sensitivity to that exposure is the degree to which a system is dependent upon the climatic conditions that existed before the change in climate took place.
Different attributes of a system may have different sensitivities to a given climatic change. The ability of a system to experience a change in climatic conditions to which it is sensitive and to regain a natural or desirable state is its adaptive capacity.

Identification and understanding of the three components of a system’s vulnerability – vulnerability science – in the context of climate change provides a basis for explaining interactions between changing climatic conditions and impacted systems [Cutter, 2003]. It integrates present or potential climatic hazards with a system’s susceptibility and capacity to recover. Importantly, vulnerability to global-scale changes in the environment such as those to Earth’s climate can vary strongly in space, even at local scales [Turner et al., 2003]. Thus, vulnerability to climate change can manifest “…geographically in the form of hazardous places…” and, in order to compare places within or between systems, vulnerability assessments must be approached in part through spatial analysis [Cutter, 2003]. Vulnerability science additionally helps describe the range of possible impacts to a system that climate change may have, and anticipates how future vulnerability of a system may be altered under further changes in climate.

One important impact of human-caused climate change is sea level rise (SLR) [Overpeck and Weiss, 2009]. Higher seas have the potential to cause major social, environmental, and economic impacts through increased coastal erosion, saltwater intrusion of freshwater aquifers, temporary flooding, and permanent inundation. Global mean sea level has risen about 210 mm since 1880 with an acceleration of 0.009 ± 0.003 mm yr$^{-2}$ [Church and White, 2011]. Volume of the world’s oceans has grown due to thermal expansion [Bindoff et al., 2007; Purkey and Johnson, 2010], as well as increased
contributions of meltwater from glaciers and ice caps [Jacob et al., 2012] and of both meltwater and land ice discharge from the great polar ice sheets in Greenland and Antarctica [Joughin and Alley, 2011; Rignot et al., 2008; Rignot et al., 2011].

A growing body of recently published work suggests that due to increasing losses, the great polar ice sheets in Greenland and Antarctica will become the dominant contributor to global SLR during the 21st century (e.g., Rignot et al. [2011]; Vermeer and Rahmstorf [2009]). Importantly, these studies and others [Pfeffer et al., 2009] highlight that global SLR influenced by such ice sheet behavior could approach or exceed 1 m by 2100. Furthermore, global GHG emissions over the 21st century will not only influence global SLR in the relative near term of the next ~90 years, but also will continue to drive global SLR over subsequent centuries. This is due to the fact that it will take centuries for the polar ice sheets to fully adjust to the comparatively rapid and largely irreversible global warming that occurs during this century [Solomon et al., 2009]. Temperatures by 2100 may be warm enough to commit Earth to at least 4-6 m of global SLR over following centuries [Meehl et al., 2007; Overpeck et al., 2006; Overpeck and Weiss, 2009; Schneider et al., 2007].

It is important to note that changes in relative sea level – and thus assessment of exposure to global SLR – at a coastal location are not only a function of changes in global mean sea level, but also are determined by a combination of other regional sea level height and local vertical land elevation changes. Other such determinants include changes in ocean climate on decadal time scales [Bromirski et al., 2011] and ocean circulation [Yin et al., 2010; Yin et al., 2009], as well as vertical land movement due to
glacial isostatic adjustment [Lambeck et al., 2010; Mitrovica et al., 2010; Kopp et al., 2010; Mitrovica et al., 2009] and local subsidence [Tornqvist et al., 2008].

In the contiguous U.S., approximately 3.7 million people live within one vertical meter of high tide [Strauss et al., 2012]. Valuable energy and transportation infrastructure, military facilities, and commercial ports are also in low-lying coastal zones [NOAA’s State of the Coast; stateofthecoast.noaa.gov], as are important wetland and conservation areas that are governed by federal law of the Clean Water Act [Titus et al., 2009]. A fundamental way to determine the sensitivity of coastal urban zones, infrastructure, and valued natural systems to different amounts of SLR is to use geospatial analysis of elevation data to delineate low-lying coastal areas with which to intersect socioeconomic and ecological data [Gesch et al., 2009]. Part of the body of work represented in this dissertation addresses this component of vulnerability to SLR by utilizing coastal topography to describe how the susceptibility of potentially impacted areas changes across a plausible range of SLR amounts in this and subsequent centuries.

Another important effect of human-caused climate change is changes in the occurrence and severity of droughts and otherwise dry conditions [Solomon et al., 2007; Karl et al., 2009]. Over the past few decades, droughts have become more common, most notably in tropical and subtropical regions [Trenberth et al., 2007]. This is due largely to variability in sea surface temperatures (SSTs) that perturb atmospheric circulation and hence alter storm tracks, as well as higher atmospheric demand for evapotranspiration driven by warmer temperatures. Additional increased dryness in recent decades has resulted from declining snow packs and reduced soil moisture, also influenced in part by

Precipitation anomalies in future climate scenarios are sometimes subject to notable uncertainty, as it is difficult to project changes in SST-driven perturbations of storm tracks \citep[e.g., El Niño Southern Oscillation \citep{Ropelewski1987, Trenberth1988}] and how they might influence drought frequency and magnitude in coming decades \citep{Meehl2007}. In addition to potential changes in natural modes of climate variability, human-caused alterations to the mean position of the jet stream may influence the occurrence and severity of future droughts. Coherent dynamical changes from stratospheric ozone depletion and expansion of the Hadley circulation have acted to move the mid-latitude storm tracks poleward \citep{Arblaster2006, Seidel2008}, which can lead to decreased precipitation in subtropical regions \citep{McAfee2008}. Future drying of subtropical regions is possible as projections indicate further expansion of the Hadley circulation \citep{Yin2005, Meehl2007}. How other changes in the global climate system influence the variability of large-scale atmospheric circulation, such as rapid warming and sea ice loss in the Arctic \citep{Francis2012}, remains uncertain.

In contrast, warmer temperatures – and the potential for greater evapotranspiration and further declines in snow pack and soil moisture – are virtually assured if radiative forcing of climate continues to increase \citep{Meehl2007, Karl2009}. Higher temperatures also will continue promoting dry conditions as precipitation occurs more often in heavier events with longer and warmer interludes.
[Meehl et al., 2007; Trenberth et al., 2007], and as greater evapotranspirational demand increases the area and intensity of droughts [Karl et al., 2009; Weiss et al., 2009; Weiss et al., 2012].

The southwestern U.S. (SW) is a region where exposure to droughts and otherwise dry conditions is a normal feature of climate [Sheppard et al., 2002], and vegetation sensitivity to shortfalls in moisture manifests as impacts such as reduced growth and increased mortality [Allen and Breshears, 1998; Swetnam and Betancourt, 1998; Weiss et al., 2004; Williams et al., 2010; Woodhouse et al., 2010]. However, warming over the past few decades has profoundly altered the region’s typical conditions of drought and dryness, affecting hydroclimatology through higher freezing levels [Abatzoglou, 2011], lower ratios of snowfall to total winter precipitation [Knowles et al., 2006], increases of rain-on-snow events [McCabe et al., 2007], and shifts to earlier snowmelt [Stewart et al., 2005] and attendant losses in snowpack [Mote et al., 2005; McCabe and Wolock, 2009]. Higher temperatures have also overlaid drought conditions in recent years [Weiss et al., 2009], leading to substantial ecological impacts that range from relatively extreme foliar growth limits [Weiss et al., in review] to widespread mortality [Breshears et al., 2005].

Expectations are that SW temperatures will continue to warm in the 21st century [Christensen et al., 2007; Karl et al., 2009], and that additional spring and early summer dryness may result from poleward shifts of the jet stream [Yin, 2005; McAfee and Russell, 2008]. Such conditions, as well as those during possible future droughts, may expose regional vegetation to intensified constraints on foliar growth through higher
evapotranspirational demand [Weiss et al., in review]. The remainder of the body of work presented in this dissertation addresses this aspect of vulnerability to recent and potential future warming in the SW with particular attention to spatial variability along environmental gradients of this physiographically complex region.

Assessing the vulnerability of a system to environmental changes like those taking place and projected to happen in Earth’s climate is a way for environmental sciences to help inform policy decisions that consequently stem from past or potential impacts of climatic hazards [Cutter, 2003]. Overall, the body of work in this dissertation represents individual parts of vulnerability assessment in the contexts of SLR (sensitivity) and SW bioclimate (exposure). In the following section, I summarize the research appearing in each appendix that includes these parts.
2. PRESENT STUDY

I present the methods, results, and conclusions of this study as a combination of four articles appended to this dissertation. These articles are formatted for publication in professional scientific journals. Two articles are published (Appendixes A and B), one article is in peer review (Appendix C), and one article is being prepared for submission to a peer-reviewed journal (Appendix D). The first three articles are multi-authored articles with myself as the lead author for each. I am currently the sole author of the final article. The following is a summary of the most important findings in this document.

The first article examines implications of recent sea level rise (SLR) science in the context of potential impacts to low-elevation coastal cities in the contiguous U.S. (Appendix A). In this part of the present study, we develop a new geospatial dataset that delineates low-lying coastal areas that are at or below one-to-six meters in elevation and have connectivity to the sea. We overlay this dataset with boundaries of U.S. municipalities that had populations greater than 50,000 as of the 2000 census to determine areas prone to SLR impacts in this and subsequent centuries. In total, we find that 180 municipalities with populations more than 50,000 have land area with elevations at or below six meters and connectivity to the sea. Of these, 20 are municipalities with populations greater than 300,000. An approximate average of 9% of the area within the 180 coastal municipalities lies at or below one meter in elevation, a figure that rises to 36% when considering area at or below six meters. Higher percentages of area at or below one to six meters in elevation typically occur along the Gulf of Mexico and southern Atlantic coasts. Based on this analysis, we discuss several important technical
considerations for developing sound geospatial datasets that delineate low-lying coastal elevations such as vertical error in underlying digital elevation models that we plan to incorporate in future versions of our dataset. We conclude that results in this article demonstrate that potential SLR impacts to U.S. municipalities will be very local and disproportionate, in contrast to the national and international dimensions of and associated efforts to curb greenhouse gas emissions.

Moving from modern global SLR to bioclimate in the Southwestern U.S. (SW), the second article begins our investigation of regional ecological impacts from warming temperatures (Appendix B). Here, we draw particular attention to drought conditions during recent years and the ability of higher temperatures to lead to greater atmospheric demand for evapotranspiration (ET), as measured by vapor pressure deficit. Our approach to examine the warm and dry conditions of the 2000s is to compare them to conditions during the previous major SW drought that occurred in the 1950s. During the 2000s drought, significantly warmer temperatures largely drove greater ET demand during the foresummer, the hot and dry period from late spring through early summer. Significantly warmer temperatures in the 2000s drought continued to primarily drive higher ET demand during the rest of summer and early autumn for parts of the region outside of the core North American monsoon area. Greater ET demand at this time of year during the 2000s drought could have been more spatially extensive had atmospheric humidity been as low as during the 1950s drought in the core North American monsoon area. The foresummer and monsoon are seasons when climate has important ecological implications in the SW that include intensity of the wildfire season and ranching
operations [Ray et al., 2007]. As projections of future climate indicate continued regional warming in upcoming decades [Christensen et al., 2007; Meehl et al., 2007; Karl et al., 2009], results from this article illustrate the potential of higher ET demand to impact ecosystems during these seasons in future droughts.

The next article continues our investigation of the ecological implications of the 2000s drought by examining the degree to which higher temperatures and greater ET demand affected foliar growth along elevational gradients and at tree mortality locations (Appendix C). We once again compare the 1950s and 2000s droughts and take advantage of the opportunity to study mesoscale ecosystem responses to anomalously dry conditions before and during the regional warming that started in the late 1970s. Warmer minimum temperatures during the 2000s drought reduced effects of suboptimal temperatures at lower elevations in winter and higher elevations in summer. Higher ET demand largely driven by warmer temperatures in the more recent drought was more limiting from spring through summer at lower and middle elevations. At many locations of tree mortality, constraints of suboptimal temperatures and ET demand on foliar growth were relatively extreme from early spring through late autumn during the 2000s drought. Our analysis shows that in physiographically complex regions like the SW, seasonality and elevational gradients are important for understanding vegetative responses to warming. It also suggests that continued warming will increase the degree to which ET demand limits foliar growth during future droughts, and expand its reach to higher elevations and other seasons.
The last article shifts our attention from ecological implications of climatic conditions during the 2000s drought to those that might occur in upcoming decades (Appendix D). As warming in the SW has recently combined with a prolonged dry spell to generate extraordinary growing conditions, expected further increases in regional temperatures raise concern that the anomalous growing conditions of recent years may be matched or exceeded, especially during future droughts. To address this, projections of foliar growth limits are needed at spatial scales appropriate to biomes (~5 km).

Unfortunately, present-day climate projections based on global- (~100 to 150 km) and regional-scale (~25 to 50 km) models (GCMs and RCMs) are too coarse for this type of application. Free-tropospheric air temperatures strongly influence the altitudinal gradient of temperature in mountainous areas, and may support a statistical downscaling approach to generate future temperature changes at scales finer than those represented in GCMs and RCMs and in complex terrain. However, before this is done, relationships between free-air and near-surface air temperatures need to be known. This article compares free-air and near-surface air temperatures in the SW along environmental gradients using gridded observational (PRISM) and reanalysis (CFSR) datasets over the latter part of the 20th century. Monthly free-air temperature climatologies at land surface altitudes appear to capture the seasonal variability and spatial features driven by regional climate and elevation that are present in the near-surface air temperature data, and are generally +1° to +3°C warmer throughout the year. Monthly anomalies of the two temperature datasets show strong, widespread co-variability throughout the year, with the majority of correlations above +0.8. Consistency between the two datasets does not appear to
systematically break down along latitude, longitude, or elevation gradients. Results suggest that further development of free-air temperatures for application to fine-scale projections of temperature, and hence foliar growth limits, is justified.
3. REFERENCES


4. APPENDICES
APPENDIX A: IMPLICATIONS OF RECENT SEA LEVEL RISE SCIENCE FOR LOW-ELEVATION AREAS IN COASTAL CITIES OF THE U.S.A.

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IMPLICATIONS OF RECENT SEA LEVEL RISE SCIENCE FOR LOW-ELEVATION AREAS IN COASTAL CITIES OF THE U.S.A.

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Abstract Recently published work estimates that global sea level rise (SLR) approaching or exceeding 1 m by 2100 is plausible, thus significantly updating projections by the Fourth Assessment of the Intergovernmental Panel on Climate Change. Furthermore, global greenhouse gas (GHG) emissions over the 21st century will not only influence SLR in the next ~90 years, but will also commit Earth to several meters of additional SLR over subsequent centuries. In this context of worsening prospects for substantial SLR, we apply a new geospatial dataset to calculate low-elevation areas in coastal cities of the conterminous U.S.A. potentially impacted by SLR in this and following centuries. In total, 20 municipalities with populations greater than 300,000 and 160 municipalities with populations between 50,000 and 300,000 have land area with elevations at or below 6 m and connectivity to the sea, as based on the 1 arc-second National Elevation Dataset. On average, approximately 9% of the area in these coastal municipalities lies at or below 1 m. This figure rises to 36% when considering area at or below 6 m. Areal percentages of municipalities with elevations at or below 1-6 m are greater than the national average along the Gulf and southern Atlantic coasts. In contrast to the national and international dimensions of and associated efforts to curb GHG emissions, our comparison of low-
elevation areas in coastal cities of the conterminous U.S.A. clearly shows that SLR will potentially have very local, and disproportionate, impacts.

1 Introduction

The science surrounding projections of global sea level rise (SLR) as a result of human-caused climate change has grown rapidly in recent years, updating and exceeding the most recent projection by the Intergovernmental Panel on Climate Change (IPCC) of 0.26-0.59 m of global SLR by 2100 for the A1FI emissions scenario (Meehl et al. 2007). Independent estimates of future sea level now indicate that global SLR could approach or possibly exceed 1 m by 2100 (Pfeffer et al. 2008; Vermeer and Rahmstorf 2009), even with moderate reductions of greenhouse gas (GHG) emissions over the rest of this century (Vermeer and Rahmstorf 2009). Importantly, global GHG emissions over the 21st century will not only influence SLR in the relative near term of the next ~90 years, but will also continue to drive SLR over subsequent centuries. Temperatures by 2100 may be warm enough to commit Earth to at least 4-6 m of global SLR over following centuries as the polar ice sheets adjust to the comparatively rapid and largely irreversible global warming that occurs this century (Meehl et al. 2007; Overpeck et al. 2006; Overpeck and Weiss 2009; Schneider et al. 2007; Solomon et al. 2009). Particular to coasts of the conterminous U.S.A., state-of-the-art SLR science also demonstrates that weakening of the Atlantic meridional overturning circulation during this century could result in 1 m of regional SLR occurring along the northern Atlantic coast earlier than the global mean arrival time (Yin et al. 2010; Yin et al. 2009).
These recent studies constitute a worsening prospect of substantial SLR over this and subsequent centuries. As a result, assessing and anticipating potential SLR impacts to coastal areas is now of increased importance. This is especially true where accentuated regional SLR may occur, such as along the northern Atlantic coast of the conterminous U.S.A. In this letter, we report a new geospatial dataset of low-lying coastal elevations for the entire conterminous U.S.A. that helps delineate areas prone to impacts from the amounts of SLR now plausible from continued high global GHG emissions over the next ~90 years. Analysis of elevation data in the context of projected SLR serves a fundamental role in identifying coastal areas that may face issues such as accretion and erosion, temporary flooding, and permanent inundation. Authors of numerous previous studies have analyzed elevation data to map different coastal areas in the conterminous U.S.A. that may be affected by SLR impacts (see review by Gesch et al. 2009). The work in this letter is unique in that it utilizes delineated low-lying coastal areas to make a nationwide assessment of municipalities with elevations at or below 6 m, a possible amount of SLR we commit to this century. Our purpose here is to raise awareness about implications of the latest SLR science by applying our dataset to the calculation of low-elevation areas in coastal cities of the conterminous U.S.A. potentially impacted by SLR in this and following centuries.

2 Methods

We used the National Elevation Dataset (NED; Gesch et al. 2002) with 1 arc-second (~30 m) horizontal resolution to delineate low-lying coastal areas of the conterminous U.S.A.
that may confront SLR issues based on their elevation and 8-way connectivity (Poulter and Halpin 2008) to the sea. Though not explicitly addressed in this letter, identified areas may correspond to potential SLR impacts such as accretion, erosion, flooding, or inundation. The NED is a digital elevation model (DEM) produced by the U.S. Geological Survey (USGS) with 1/9 (~3 m), 1/3 (~10 m), and 1 arc-second versions. At present, only the 1 arc-second NED (NED\text{\textsubscript{larc}}) has complete coverage of the conterminous U.S.A. The NED\text{\textsubscript{larc}} incorporates source data from digital photogrammetry and light detection and ranging (lidar) topographic surveys where available, in addition to 1/3 arc-second and 1 arc-second DEMs with cartographic contour source data. The USGS resamples DEMs of finer resolution for production of the NED\text{\textsubscript{larc}}. The NED\text{\textsubscript{larc}} has an absolute vertical linear error of ±4.75 m (95% confidence interval) for the entire conterminous U.S.A. (Gesch 2007). However, the absolute vertical accuracy of the NED\text{\textsubscript{larc}} along coasts is typically better than the nationwide average (Gesch 2009; Gesch et al. 2009). NED\text{\textsubscript{larc}} values reference the North American Vertical Datum of 1988 (NAVD 88) - the mean sea level at Rimouski, Quebec, Canada - and not a local sea level datum. The average differences between local mean high water (MHW) and the NAVD 88 zero value (i.e., local MHW - NAVD 88) at NOAA tide gauge stations are 0.4 m (range: -0.2-2.7 m) for the Atlantic coast, 0.2 m (range: -0.5-0.7 m) for the Gulf coast, and 1.9 m (range: 1.3-2.9 m) for the Pacific coast. Both the inherent vertical error and NAVD 88 vertical reference of the NED\text{\textsubscript{larc}} place uncertainty on the delineation of low-lying coastal elevations.
To specify the land-sea boundary in the NED\textsubscript{1arc} by where local tides periodically submerge areas, we designated MHW as the shoreline in our analysis using vector shoreline data from the National Oceanic and Atmospheric Administration Coastal Services Center (www.noaa.csc.gov). We created a geoprocessing algorithm that performs iterative cell-by-cell analysis of DEMs to select all cells with elevation values less than or equal to a particular value, and with locations adjacent or connected by cells of equal or lesser value to the sea. We applied this algorithm to the NED\textsubscript{1arc} for each integer value from 1-6 m to delineate areas potentially subject to SLR. We note that our analysis is based on modeled present-day elevations and does not predict future shorelines, nor include processes that may affect local elevation such as glacial isostatic adjustment, tectonics, subsidence, or erosion and accretion (e.g., Englehart et al. 2009; Tornqvist et al. 2008; Zhang et al. 2004).

We utilized municipal boundaries used for the 2000 census by the U.S. Census Bureau (www.census.gov) to calculate land area within coastal cities of elevations at or below 1-6 m landward of the present-day MHW line. The municipal boundaries are geographic information system (GIS) shapefiles that are part of the Census Bureau’s Topologically Integrated Geographic Encoding and Referencing (TIGER) database, and delineate the extent of legally defined boundaries of governmental units used in censuses and survey programs. As some municipal boundaries include area designated as sea in the previously computed elevation datasets, we separated out the portions of municipalities that only occur over areas designated as land. We then calculated the areal overlap
between the land portion of each municipality and our datasets delineating areas with elevations at or below 1-6 m.

Tidal wetlands do occur above the MHW mark as a result of extreme tidal events such as spring high water (e.g., Titus and Wang 2008), and thus may be present within our designated municipal boundaries and lead to an overestimate of land area at or below elevations from 1-6 m. To determine possible effects that high marsh might have on areal overlap calculations of the elevation and municipality datasets, we acquired wetland GIS shapefiles from the U.S. Fish and Wildlife Service (2010). We removed parts within municipal land areas where the estuarine and marine wetland type appeared, and recalculated the areal overlap between the land portion of each municipality and our datasets delineating areas with elevations at or below 1-6 m. We performed all geospatial data analyses with ESRI ArcGIS Desktop™ software. All map figures are in an Albers equal area projection. For this letter, we focus on results for coastal municipalities with populations greater than 50,000 as of the 2000 census, with a particular emphasis on those with populations greater than 300,000 (hereafter major municipalities).

3 Results
Coastal areas of the conterminous U.S.A. with elevations at or below 1-6 m and connectivity to the sea are most apparent along the Gulf and southern Atlantic coasts, especially in southern Louisiana and southern Florida (Fig. A-1). The northern Atlantic and Pacific coasts have relatively less low-lying area visible, although within these regions there are discernable areas with elevations at or below the 6-m mark (Fig. A-S1,
Such places include the Chesapeake Bay near Washington, DC and the San Francisco Bay and San Joaquin-Sacramento river delta in northern California. State averages for the percent of land area within municipalities of elevations at or below 1-6 m reflect these regional-scale patterns (Fig. A-S3, Table A-S1). Average land area percentages of elevations at or below the 1-m contour for municipalities with populations greater than 50,000 surpass the nationwide average of 9.7% in Florida, Georgia, Louisiana, Mississippi, South Carolina, Texas, and Virginia. These states, along with Alabama, Delaware, and New Jersey, have average percentages of municipal land area with elevations at or below the 6-m line greater than the national average of 35.8%.

Along all coasts of the conterminous U.S.A., 20 major municipalities with populations greater than 300,000 (Fig. A-2) and 160 municipalities with populations between 50,000 and 300,000 (Table A-S1) have land area with elevations at or below 6 m and connectivity to the sea. For the major municipalities of Miami, FL, New Orleans, LA, Tampa, FL, and Virginia Beach, VA, land area percentages at or below 1 m of elevation exceed the 180-municipality national average of 9.7% (Fig. A-2, Fig. A-3). These major municipalities, along with Jacksonville, FL (Fig. A-2, Fig. A-S4), also have land area percentages higher than the 180-municipality national average of 35.8% for elevations at or below 6 m. Most striking is that more than 90% of Miami, FL, New Orleans, LA, and Virginia Beach, VA are at or below the 6-m elevation mark (Fig. A-2, Fig. A-3). For New York, NY - the most populous major municipality - land area percentages at or below the 1-6 m elevation contours closely follow those of the 180-municipality national average. Land area percentages at or below the 1-6 m elevation contours for Washington,
DC - the national capitol - are less than those of the 180-municipality average for the nation. Overall, land area percentages along all coasts of the conterminous U.S.A. for both the 20 major municipalities (Fig. A-2, Fig. A-3, Fig. A-S4, Fig. A-S5, Fig. A-S6) and the 160 municipalities with populations between 50,000 and 300,000 (Table A-S1) demonstrate a wide range of land area at or below elevations from 1-6 m.

Tidal wetlands above MHW lead to overestimates of municipal land area percentages for elevations at or below 1-6 m in approximately two-thirds of coastal states in the conterminous U.S.A. Accounting for tidal wetlands above MHW reduces the 180-municipality national mean of land area percentages at or below the 1-6 m elevation contours by an average of 0.5% (Table A-S1). For statewide averages, reductions in municipal land area percentages at or below 1-6 m are less than 2.3% for Delaware, Florida, Louisiana, Maine, Massachusetts, New Jersey, New York, North Carolina, Rhode Island, Texas, Virginia, and Washington. For Georgia, Mississippi, and South Carolina, average decreases in municipal land area percentages at or below 1-6 m range between 2.5% and 9.6%. Tidal wetlands above MHW do not affect statewide averages in Alabama, California, Connecticut, the District of Columbia, Maryland, Oregon, and Pennsylvania. For the 20 major municipalities, tidal wetlands above MHW lessen land area percentages at or below 1-6 m by less than 1.1% for Boston, MA, Houston, TX, Miami, FL, New Orleans, LA, and Seattle, WA. These reductions range between 1.4% and 2.0% for New York, NY and Tampa, FL, and between 0.4% and 5.3% for Jacksonville, FL and Virginia Beach, VA. Tidal wetlands above MHW do not occur in the other 11 major municipalities.
4 Discussion
Several technical considerations are important in order to produce scientifically sound
geospatial datasets of low-lying coastal elevations that help delineate areas prone to SLR
impacts (Gesch et al. 2009; Titus and Wang 2008). For this current version of our dataset,
we incorporated hydrological connectivity and the presence of tidal wetlands landward of
our specified shoreline. Hydrological connectivity models the paths that water from rising
sea levels could traverse on the surface and accounts for features recognizable at the
spatial scale of the DEM that may either facilitate or impede advancing seas such as
channels or levees. In applications of low-lying coastal elevation datasets that quantify
potential socioeconomic and ecological impacts of SLR, omission of hydrological
connectivity may result in overestimations of possible impacts. Exclusion of tidal
wetlands landward of specified shorelines also may generate impact estimates that are too
high. Incorporating tidal wetlands above MHW improved our estimates of land area
percentages for elevations at or below 1-6 m within coastal municipalities of the U.S.A.,
mostly for a small number of coastal cities that include the major municipalities of
Jacksonville, FL and Virginia Beach, VA.

The most important technical consideration that our current dataset does not
incorporate is elevation data uncertainty that stems from the inherent vertical error and
vertical datum of the NED\textsubscript{larc}. These sources of uncertainty affect the elevation zone
delineations that we present and the accuracy of land area percentages that we report. At
the national level, the inherent vertical error of the NED\textsubscript{larc} does not adequately support
delineation of low-lying coastal areas at 1-m increments (Gesch 2007; Gesch et al. 2009).
To be completely suitable for such an analysis, a DEM would minimally need to have a vertical error of ±0.5 m (95% confidence interval), a range of uncertainty equal to the increment of analysis (Gesch et al. 2009). However, vertical error of the NED$_{1\text{arc}}$ along coasts is typically better than the nationwide average (Gesch 2009; Gesch et al. 2009), suggesting that delineation of low-lying coastal areas at increments between 2 and 6 m is fully appropriate in many areas. Furthermore, NED$_{1\text{arc}}$ from source data of highly accurate digital photogrammetry and lidar topographic surveys (e.g., ±29.4 cm 95% confidence interval for lidar; Gesch et al. 2009) in limited coastal areas properly supports analysis at 1-m increments. As a consequence of NED$_{1\text{arc}}$ vertical error, we most confidently interpret our results at the 6-m elevation contour.

At present, the optimal way to address inherent vertical error of modeled elevations is to use DEMs with high vertical accuracy that are based on data from digital photogrammetry and lidar topographic surveys. As a comparison (not shown), areal coverage of our dataset at or below the 1-m elevation contour in the San Francisco Bay region largely underestimates that which is based on lidar data in a similar study by Knowles (2010). An overall underestimation of areal coverage at or below the 1-m elevation mark by the NED$_{1\text{arc}}$ in comparison to the lidar-based 1/9 arc-second NED also appears to be the case in North Carolina, though inherent vertical error of the NED$_{1\text{arc}}$ may reverse this conclusion (Gesch 2009). However, as DEMs of such high vertical accuracy are not yet available for all coastal states of the conterminous U.S.A., use of the relatively coarser and less accurate NED$_{1\text{arc}}$ is necessary for national-level assessments.

Comparison of a DEM with an independent elevation reference such as National
Geodetic Survey global positioning system benchmarks is one way to assess DEM vertical accuracy and generate confidence intervals of both delineated elevation zones and potential SLR impacts (Gesch 2009; Gesch et al. 2009).

Confidence intervals for elevation zones and potential SLR impacts can also incorporate differences between the vertical reference of the DEM and a local sea level datum. Alternatively, VDatum (vdatum.noaa.gov; Parker et al. 2003) or other vertical transformation methods (e.g., Titus and Wang 2008) can systematically adjust DEM values to a particular local datum for analysis of elevation data in the context of projected SLR. For our current dataset, error introduced by the NAVD 88 vertical reference of the \textit{NED}$_{\text{arc}}$ is of greatest regional concern along the Pacific coast, where the average difference between local MHW and the NAVD 88 zero value is 1.9 m. Nonetheless, as relatively little area along the Pacific coast appears for elevations at or below 2-6 m, potential SLR impacts may still be less extensive there than in other regions of the conterminous U.S.A. for elevations at or below approximately 4 m. We plan to integrate technical considerations regarding the vertical uncertainty of DEMs in future versions of our dataset.

More detailed future studies that help identify low-lying coastal elevations prone to SLR impacts also need to consider geoscience issues regarding amount, range, and timing of recent global and regional SLR projections. Addressing these aspects of future sea level enhances estimates and interpretations of elevation zone delineations and potential socioeconomic and ecological impacts due to SLR. For example, elevation zone delineations could track projection model averages and ranges of global and regional
SLR amounts for the 21st century that are calculated under different emissions scenarios established by the IPCC (e.g., Vermeer and Rahmstorf 2009; Yin et al. 2010; Yin et al. 2009). Much as elevation data uncertainty forms confidence intervals for elevation zones and potential SLR impacts, SLR projection averages between and ranges within emissions scenarios establish upper and lower bounds for elevation zones and potential SLR impacts at future points in time. Given that lower, mean, and upper estimates of SLR now approach or exceed 1 m by 2100, analysis of elevation data in the context of SLR projections over the next ~90 years requires DEMs with sub-meter vertical accuracy (Gesch 2009).

Elevation data analysis is only one of the components needed to fully identify coastal areas that may confront potential SLR impacts such as accretion, erosion, flooding, or inundation (Gesch et al. 2009). This is particularly true for natural environments where, in addition to SLR, complex and dynamic processes driven by storms, biota, and sediment flux can change coastlines. The best application of our results is to coastal areas where inundation will be the primary SLR impact. Inundation is the most likely impact in developed areas, such as the municipalities we examined in this letter, as well as in natural areas that have a lack of biota, low sediment influx, gently sloping land, and a small tidal range (Gesch et al. 2009; Kirwan et al. 2010). However, amounts of global SLR as considered in this letter may outpace the ability of most coastal wetlands to adapt to rising sea levels and increase the amount of natural environments potentially impacted by inundation (Kirwan et al. 2010).
Our new geospatial dataset of low-lying coastal elevations for the entire conterminous U.S.A. raises awareness about implications of the latest SLR science and provides a meaningful comparison of coastal regions and cities potentially affected by SLR in this and subsequent centuries. Despite technical limitations stemming from both the NED$_{\text{arc}}$ and our methodology, results clearly highlight that continued high global GHG emissions and the ensuing rise in sea level have the potential to impact numerous coastal cities of the conterminous U.S.A. Potential SLR impacts will be very local and disproportionate, in contrast to the national and international dimensions of and associated efforts to curb GHG emissions.

**Acknowledgements**

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**References**


Fig. A-1. Parts of the conterminous U.S.A. along the Gulf and southern Atlantic coasts with elevations at or below 1-6 m and connectivity to the sea. Labels denote coastal state abbreviations and municipalities with populations greater than 300,000 as of the 2000 census (i.e., major municipalities).
Fig. A-2. Land area percentages of major municipalities (2000 census population greater than 300,000) with elevations at or below 1-6 m and connectivity to the sea. The average land area percentage for all 180 municipalities (2000 census population greater than 50,000) with elevations at or below 1-6 m and connectivity to the sea is on the right.
Fig. A-3. Areas of selected major municipalities (land boundaries in white) with elevations at or below 1-6 m and connectivity to the sea. Map figures for the other 14 major municipalities are in Fig. A-S4, Fig. A-S5, and Fig. A-S6.
Table A-S1. Land area percentages of all 180 municipalities (2000 census population greater than 50,000) with elevations at or below 1-6 m and connectivity to the sea. For each meter level, percentages without parentheses do not have tidal wetland areas landward of the mean high water line subtracted, whereas percentages within parentheses do. As more fully described in the Methods and Discussion, the inherent vertical error and vertical datum of the NED1arc are sources of uncertainty for land area percentages.

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Fig. A-S1. Parts of the conterminous U.S.A. along the northern Atlantic coast with elevations at or below 1-6 m and connectivity to the sea. Labels denote coastal state abbreviations and municipalities with populations greater than 300,000 as of the 2000 census (i.e., major municipalities).
Fig. A-S2. As in Fig. A-S1, but for parts of the conterminous U.S.A. along the Pacific coast.
Fig. A-S3. State average land area percentages of all municipalities (2000 census population greater than 50,000) with elevations at or below 1-6 m and connectivity to the sea. Numbers above state bars indicate the number of such municipalities for a given state. The average land area percentage for all 180 such municipalities is on the right.
Fig. A-S4. Areas of selected major municipalities (land boundaries in white) with elevations at or below 1-6 m and connectivity to the sea.
Fig. A-S5. As in Fig. A-S4.
Fig. A-S6. As in Fig. A-S4.
APPENDIX B: DISTINGUISHING PRONOUNCED DROUGHTS IN THE SOUTHWESTERN UNITED STATES: SEASONALITY AND EFFECTS OF WARMER TEMPERATURES

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DISTINGUISHING PRONOUNCED DROUGHTS IN THE SOUTHWESTERN UNITED STATES: SEASONALITY AND EFFECTS OF WARMER TEMPERATURES

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¹ Department of Geosciences, The University of Arizona, Tucson, Arizona
² Department of Atmospheric Sciences, The University of Arizona, Tucson, Arizona
³ Institute for Environment and Society, The University of Arizona, Tucson, Arizona

Abstract

Higher temperatures increase the moisture-holding capacity of the atmosphere and can lead to greater atmospheric demand for evapotranspiration, especially during warmer seasons of the year. Increases in precipitation or atmospheric humidity ameliorate this enhanced demand, whereas decreases exacerbate it. In the southwestern United States (Southwest), this means the greatest changes in evapotranspirational demand resulting from higher temperatures could occur during the hot-dry foresummer and hot-wet monsoon. Here we examine seasonal differences in surface climate observations to determine how temperature and moisture conditions affected evapotranspirational demand during the pronounced Southwest droughts of the 1950s and 2000s, the latter influenced by warmer temperatures now attributed mostly to the buildup of greenhouse gases. In the hot-dry foresummer during the 2000s drought, much of the Southwest experienced significantly warmer temperatures that largely drove greater evapotranspirational demand. Lower atmospheric humidity at this time of year over parts of the region also allowed evapotranspirational demand to increase. Significantly warmer temperatures in the hot-wet monsoon during the more recent drought also primarily drove
greater evapotranspirational demand, but only for parts of the region outside of the core North American monsoon area. Had atmospheric humidity during the more recent drought been as low as during the 1950s drought in the core North American monsoon area at this time of year, greater evapotranspirational demand during the 2000s drought could have been more spatially extensive. With projections of future climate indicating continued warming in the region, evapotranspirational demand during the hot-dry and hot-wet seasons possibly will be more severe in future droughts and result in more extreme conditions in the Southwest, a disproportionate amount negatively impacting society.

1. Introduction

a. Climate in the southwestern United States

Climate in the southwestern United States (hereafter Southwest) is highly seasonal. There are two distinct times during the year when precipitation typically is expected. From winter through early spring, westerly frontal systems bring both rain and snow, whereas from mid-summer through early autumn, the North American monsoon (hereafter monsoon) and, to a lesser degree, dissipating tropical cyclones from the eastern Pacific Ocean supply rain (see review by Sheppard et al. 2002). Prominence of either of these peaks in the annual precipitation cycle is spatially variable, with a regional gradient from a winter to early spring dominated regime in the northwest to a regime dominated by the monsoon in the southeast. The period from late spring through early summer, often referred to as the foresummer, separates these two relatively wet seasons and is
climatologically the hottest and driest part of the year. Drought in the Southwest thus depends on what modulates precipitation in two different seasons separated by the hot-dry foreshower.

b. Precipitation variability and drought in the Southwest

Drought is a normal part of climate in the Southwest. Over the past millennium, paleoclimate reconstructions indicate recurring periods of drought in the region (e.g., D’Arrigo and Jacoby 1991, Meko et al. 1995, review by Woodhouse and Overpeck 1998). These periods include severe multidecadal droughts, or “megadroughts”, such as the 1100s drought during medieval times that is considered the worst drought in the recent paleoclimate record (Cook et al. 2004, Meko et al. 2007). They also include the late 16th century drought, perceived to be the most severe drought of the past 500 years (Meko et al. 1995, Stahle et al. 2000). Southwest droughts also are conspicuous over the instrumental record of the past approximately 100 years. The most notable ones are the drought of the late 1890s and early 1900s, the drought of the 1950s, and the most recent, and possibly still on-going, drought of the early 2000s (Swetnam and Betancourt 1998, Fye et al. 2003, Seager 2007, Quiring and Goodrich 2008). Several droughts analogous to the one during the 1950s, regarded as the worst of the 20th century, likely occurred during the past 500 years, suggesting a recurrence interval of roughly 50 years for such conditions in the region (Fye et al. 2003, see also Hidalgo 2004). Societal impacts of droughts in the Southwest have been substantial, and are reviewed in other publications (e.g., Schroeder 1968, Woodhouse and Overpeck 1998, Fye et al. 2003, Cook et al. 2007).
Spatiotemporal variability of sea surface temperatures (SSTs) and the consequent shifts in large-scale atmospheric circulation patterns considerably modulate natural drought variability in the Southwest. El Niño-Southern Oscillation (ENSO) in the tropical Pacific Ocean appears to dominate interannual variability of winter and early spring precipitation in the region, and precipitation tends to be below average during La Niña events (Ropelewski and Halpert 1987, Redmond and Koch 1991, Cole et al. 2002, Brown and Comrie 2004). Cool SST anomalies along the equator in the eastern and central Pacific Ocean (i.e., La Niña conditions) change location of convection and atmospheric heating in this region. This change initiates teleconnection patterns, or Rossby wave trains, into the extratropics, weakens and moves poleward the subtropical jet stream, and creates an area of subsidence over the Southwest (Trenberth and Branstator 1992, Hoerling and Kumar 2003, Seager et al. 2003, Seager et al. 2004).

Both the Pacific and Atlantic oceans appear to influence decadal to multidecadal (D2M) variability of winter and early spring precipitation in the region. Cool SST anomalies along the equator in the eastern and central Pacific Ocean and the coast of western North America, in conjunction with warm SST anomalies in the north central Pacific Ocean, tend to promote below average precipitation (Hidalgo 2004, McCabe et al. 2004). The “cool phase” of the Pacific decadal oscillation (PDO; Mantua et al. 1997) or “La Niña-like” conditions (see Seager 2007) typically refer to these SST anomalies. The “cool phase” of the PDO promotes drought in the Southwest similar to La Niña events as previously described, with additional atmospheric perturbations occurring over the northern Pacific Ocean (Harman 1991). Below average precipitation also tends to occur
in concert with warm North Atlantic SST anomalies (Enfield et al. 2001, Hidalgo 2004, McCabe et al. 2004), conditions often referred to as the “warm phase” of the Atlantic multidecadal oscillation (AMO; Enfield et al. 2001). Warm SST anomalies in the northern Atlantic Ocean appear to influence North Pacific climate variability through atmospheric perturbations via the tropics or middle and high latitude atmosphere, in turn weakening and moving poleward the subtropical jet stream (e.g., Dima and Lohmann 2007, Zhang and Delworth 2007).

Spatiotemporal variability of Pacific and Atlantic SSTs also may influence precipitation variability from mid-summer through early autumn in the Southwest. Pacific SST anomalies associated with ENSO and the PDO tend to influence rainfall during this time of year in contrasting circumstances to that of winter and early spring precipitation. Onset of the monsoon tends to be later, and early seasonal precipitation tends to be lower in the region during El Niño events or the “warm phase” of the PDO (Harrington et al. 1992, Higgins et al. 1998, Higgins et al. 1999, Higgins and Shi 2000, Castro et al. 2001, 2007). Warm SST anomalies along the equator in the eastern and central Pacific Ocean and the coast of western North America, in combination with cool SST anomalies in the north central Pacific Ocean, produce teleconnection patterns that favor westerly upper-level winds that inhibit moisture transport into the Southwest and a weaker and more southwardly displaced monsoon ridge (Castro et al. 2001, 2007). Relationships between Pacific SSTs and summer precipitation tend to weaken during the middle and latter parts of the monsoon as the Pacific jet weakens, the monsoon ridge strengthens, and the Bermuda high extends westward. Pacific SSTs also may influence
late summer and early autumn precipitation associated with dissipating tropical cyclones from the eastern Pacific Ocean, as a less favorable environment for cyclone development tends to happen during El Niño events or the “warm phase” of the PDO (Smith 1986, Reyes and Mejía-Trejo 1991, Englehart and Douglas 2001). North Atlantic SST anomalies associated with the AMO may affect precipitation variability from mid-summer through early autumn in similar circumstances to that of winter and early spring precipitation. During the “warm phase” of the AMO, the western part of the Bermuda high tends to be weaker, and moisture transport across northern Mexico and into the Southwest tends to decrease (Sutton and Hodson 2005, Curtis 2007).

Destructive and constructive interference between interannual and D2M SST variability in the Pacific Ocean, and between D2M variability in the Pacific and Atlantic oceans can occur. For example, the tendency of La Niña events to result in below average winter and early spring precipitation in the Southwest is less consistent during a “warm phase” of the PDO than it is during a “cool phase” (Gershunov and Barnett 1998, Gutzler et al. 2002). Pacific SSTs most coherently influence summer precipitation deficits in the region when El Niño events occur during the “warm phase” of the PDO (Castro et al. 2001, 2007). Also, the “cool phase” of the PDO and the “warm phase” of the AMO can enhance regional drought conditions (McCabe et al. 2004). The various effects of Pacific and Atlantic SST anomalies on Southwestern precipitation, together with the unknown causes of “megadroughts” and the roles of other influences on precipitation such as land surface conditions (e.g., Trenberth and Guillemot 1996, Seager et al. 2005), make understanding and predicting droughts in the Southwest difficult.
c. The 1950s and 2000s droughts in the Southwest

Several aspects of SST variability and large-scale atmospheric circulation patterns were similar during the 1950s and 2000s droughts in the Southwest. La Niña events and the “cool phase” of the PDO likely influenced both droughts, as did the “warm phase” of the AMO (Fig. B-1). Geopotential height anomalies from winter through early spring were positive during both periods over the region, indicating an area of subsidence (Fig. B-2; see also Seager 2007, Quiring and Goodrich 2008). In June during both droughts, positive geopotential height anomalies also reflected La Niña events and the “cool phase” of the PDO, indicating less influence from the westerlies and a more northwardly displaced monsoon ridge.

Other aspects of SST variability and large-scale atmospheric circulation patterns notably contrasted during the 1950s and 2000s droughts. Warm SST anomalies in the Indian Ocean occurred during the 2000s drought instead of the cool SST anomalies that typically have coincided with “La Niña-like” conditions (Hoerling and Kumar 2003, Seager 2007). The anomalously warm SSTs in the Indian Ocean as well as the warmest SSTs in the western Pacific Ocean of the 20th century were both part of a warming trend since the mid-20th century and likely contributed to persistent mid-latitude drying throughout the northern hemisphere, including the Southwest. A weak El Niño event that occurred in 2002-2003 did not result in above average precipitation during winter and early spring, contrary to what has tended to happen in the Southwest when warm SST anomalies occur along the equator in the eastern and central Pacific Ocean (Ropelewski and Halpert 1987, Redmond and Koch 1991, Seager 2007). Influences other than ENSO,
such as a northward shift in storm tracks at this time of year as indicated by high index values of the Northern Annual Mode (Thompson and Wallace 1998, McAfee and Russell 2008), may have modulated precipitation variability in the region during these years of the 2000s drought.

The 1950s and 2000s droughts in the Southwest are conspicuous in surface climate records, with combinations of warm temperatures, low precipitation, and strongly negative values of the Palmer Drought Severity Index (PDSI; Palmer 1965), a soil moisture proxy (Fig. B-1). The continuously most negative PDSI values for the region took place during the four-year periods 1953-1956 for the 1950s drought and 2000-2003 for the 2000s drought. Within the region during these years, low precipitation and strongly negative PDSI values varied spatiotemporally. The 1950s drought was focused mainly over Colorado and New Mexico, whereas the 2000s drought was centered more over the Southwest (see also Swetnam and Betancourt 1998, Fye et al. 2003, Hoerling and Eischeid 2007, Quiring and Goodrich 2008). Percent of normal seasonal precipitation in the region from winter through early spring was 74% during the 1950s drought and 86% during the 2000s drought (not shown). From mid-summer through early autumn, 82% of normal seasonal precipitation fell during both droughts. Compared to the 1950s drought, relatively warmer annual temperature anomalies appear to have occurred during the 2000s drought, as did a relatively higher percent of normal precipitation and more strongly negative annual average PDSI values. Warmer temperatures during the 2000s drought, coupled with low precipitation, apparently resulted in greater soil moisture
deficits than during the relatively cooler and drier 1950s drought (see also Easterling et al. 2007).

Both the 1950s and 2000s droughts induced significant vegetation mortality in biomes from deserts to forests (Neilson 1986, Allen and Breshears 1998, Swetnam and Betancourt 1998, Breshears et al. 2005, Gitlin et al. 2005, Mueller et al. 2005, Shaw et al. 2005, Miriti et al. 2007), suggesting that Southwest droughts also have the potential to affect surface climate via impacts on albedo, roughness, and latent and sensible heat fluxes. For example, the 2000s drought resulted in widespread death of piñon pine over 12,000 km² in the Southwest (Breshears et al. 2005). Despite more total precipitation during the years 2000-2003 than 1953-1956, observations indicated that piñon pine die-off during the 2000s drought was significantly greater in magnitude and extent than during the 1950s drought. Warmer temperatures during the 2000s drought, coupled with low precipitation, seemingly drove higher vegetation water stress, increased susceptibility to insect infestations, and more plant mortality than during the relatively cooler and drier 1950s drought.

d. Seasonality and effects of warmer temperatures during drought in the Southwest

Although it is not clear if anthropogenic climate change has affected precipitation variability in the Southwest, it likely influenced the 2000s drought in part through warmer temperatures. Global- to subcontinental-scale climate change detection and attribution studies demonstrate that trends in temperature means and extremes, such as decreasing frost days and increasing heat wave intensity, broke away from expected

The seasonality of temperature and precipitation differences between the 1950s and 2000s droughts in the Southwest is crucial in distinguishing these two dry periods. Higher temperatures increase the moisture-holding capacity of the atmosphere and can lead to greater atmospheric demand for evapotranspiration, especially during warmer seasons of the year. Increases in precipitation or atmospheric humidity ameliorate this enhanced demand, whereas decreases exacerbate it. In the Southwest, this means the greatest increases in evapotranspirational demand resulting from higher temperatures could occur during the hot-dry foresummer and hot-wet monsoon. Moisture deficits in winter and early spring can carry over into the hot-dry foresummer and exacerbate this enhanced demand. From mid-summer through early autumn, moisture deficits can exacerbate this enhanced demand during the hot-wet monsoon. Our goal in this paper is to use surface climate observations during the pronounced Southwest droughts of the 1950s and 2000s to calculate seasonal differences in temperature and moisture conditions, and determine how they affected evapotranspirational demand.
2. Methodology

We examined seasonal differences of surface climate observations during the 1950s and 2000s droughts in the Southwest using gridded observational data compiled by the PRISM Group at Oregon State University. PRISM data are meteorological station data interpolated to 4-km grid cells using a human-expert and statistical knowledge-based system (Daly et al. 2002, http://www.prism.oregonstate.edu). We used PRISM monthly means of maximum temperature (Tmax, °C), minimum temperature (Tmin, °C), and dew point temperature (Tdmean, °C). We calculated three-month standardized precipitation index values (SPI-3, unitless) using PRISM monthly precipitation amounts from December 1949 through January 2007 (McKee et al. 1993). We also calculated vapor pressure deficit (VPD, kPa), the difference between saturation and actual vapor pressure, using PRISM data and the formula:

\[ a \exp\left( \frac{bT_{\text{mean}}}{T_{\text{mean}} + c} \right) - a \exp\left( \frac{bT_{d\text{mean}}}{T_{d\text{mean}} + c} \right), \]

where \( a = 0.611 \text{ kPa}, \ b = 17.502, \ c = 240.97 \text{ °C}, \ T_{\text{mean}} = \text{monthly mean temperature in °C}, \) and \( T_{d\text{mean}} = \text{monthly mean dew point temperature in °C} \) (Campbell and Norman 1998). Both Tdmean and VPD are measures of moisture condition in the atmosphere, with the latter being an estimate of the atmospheric demand for evapotranspiration. The spatial domain for our study was the Southwest region between 27° N to 43° N and 117° W to 100° W (Fig. B-3). PRISM data only were available for domain areas within the United States, and thus our analysis only included the Southwest region of the United States, rather than extending into Mexico.
We defined the analysis periods for the 1950s and 2000s droughts in this study by the four-year periods 1953-1956 and 2000-2003, respectively, after Breshears et al. (2005) (see also Fig. B-1). We used monthly means of the variables Tmax, Tmin, Tdmean, and VPD from December 1952 – January 1957 and December 1999 – January 2004 to calculate three-month seasonal values centered on each calendar month for both drought periods. For example, the seasonal January (i.e., DJF) 1953 Tmax value was the average of Tmax monthly means from December 1952, January 1953, and February 1953. It was unnecessary to calculate seasonal values for SPI-3, as by definition it incorporates precipitation amounts over a 3-month period (McKee et al. 1993). Thus, for statistical tests described below, we used a seasonal value for each variable and month from January 1953 through December 1956 for the 1950s drought, and from January 2000 through December 2003 for the 2000s drought.

We performed difference of means tests to determine local statistically significant differences (i.e., at an individual grid cell) between seasonal mean values of the 1950s and 2000s droughts for each variable and season. We regarded values of a particular variable and season in a drought period (e.g., Tmax values for DJF from 1953-1956) as temporally independent for sample size calculation. We also considered whether or not results from one season were independent from results in other seasons for a particular variable. In general, the e-folding time through the domain indicated that results from seasons with non-overlapping months could be viewed as independent. To address possible spatial autocorrelation in fields of joint difference of means tests, we carried out non-parametric field significance tests with a permutation randomization approach of 500
iterations (Livezey and Chen 1983, Wilks 2006). All tests were conducted at the 95% level.

3. Results

In this section, we present the results from seasonal difference of means and field significance tests for the five surface climate variables being analyzed during the 1950s and 2000s Southwest droughts: three-month standardized precipitation index (SPI-3); maximum temperature (Tmax); minimum temperature (Tmin); dew point temperature (Tdmean); and vapor pressure deficit (VPD). Although we show figures for results of all seasons for each variable, we only describe locally significant results of seasons that are field significant. We interpret the significant results in the following sections.

a. SPI-3

Seasonal SPI-3 during the 1950s and 2000s droughts in the Southwest shows significant differences only from mid-autumn through early winter, and for early summer (Fig. B-4). Significant positive differences from SON through NDJ mostly range from +1.0 to +2.0, indicating that the 2000s drought was wetter than the 1950s drought in these seasons, largely over Arizona and western New Mexico in SON and OND, and over most of New Mexico in NDJ. Significant negative differences are very few during SON and OND, but increase across northern Arizona, southeastern Utah, and western Colorado in NDJ with values from -1.0 to -2.0, indicating that the 2000s drought was drier than the 1950s drought in these areas at this time of year. Significant negative differences with
values from -1.0 to -2.0 also occur in MJJ, mainly over northern Arizona and adjacent areas of neighboring states. Significant positive differences are very few during this warm season.

\textit{b. Tmax}

Seasonal Tmax for the 1950s and 2000s Southwest droughts shows significant differences only from early through late summer, and demonstrates that maximum temperatures were warmer in the more recent drought at this time of year (Fig. B-5). In MJJ and JJA, significant positive differences cover a large area of the region with most values from +2°C to +4°C. Areas displaying significant positive differences decrease in JAS, with values generally from +1°C to +3°C. Significant negative differences are very few during these warm seasons. Areas of significant differences largely are widespread west of a longitudinal margin through central Colorado and central New Mexico, most notably in MJJ and JJA.

\textit{c. Tmin}

During the 1950s and 2000s droughts in the Southwest, significant differences of seasonal Tmin occur from mid-spring through early autumn (Fig. B-6). Values of significant positive differences cover a large area of the region throughout these seasons and mostly range from +1°C to +4°C, showing that minimum temperatures were warmer during the more recent drought. Significant negative differences are very few during
these seasons. As with Tmax, areas of significant differences generally are widespread west of a longitudinal margin through central Colorado and central New Mexico.

d. Tdmean

Seasonal Tdmean during the 1950s and 2000s Southwest droughts shows significant differences only from late spring through late autumn (Fig. B-7). Lower dew point temperatures predominate in the hot-dry foresummer whereas higher dew point temperatures predominate during and after the hot-wet monsoon in the more recent drought. From AMJ through JJA, significant negative differences with values from -1°C to -3°C are present around the four corners area and extreme southern New Mexico and western Texas. In contrast, significant positive differences with values +1°C to +4°C occur in central New Mexico at this time of year. Significant positive differences become more widespread from JAS through OND with values largely from +2°C to +6°C, occurring principally in Arizona and New Mexico. Significant negative differences are very few during these seasons. Spatial patterns of significant differences for seasonal Tdmean broadly resemble those of precipitation as measured by SPI-3.

e. VPD

As the difference between saturation and actual vapor pressure, VPD estimates atmospheric demand for evapotranspiration. Seasonal VPD of the 1950s and 2000s droughts in the Southwest shows significant differences only from mid-spring through late summer, in mid-autumn, and in early winter (Fig. B-8). Significant positive
differences cover large areas from MAM through JAS with most values from +0.15 kPa to +0.75 kPa, indicating that greater evapotranspirational demand occurred in the more recent drought during the hot-dry fo夏季 and hot-wet monsoon. These areas generally are widespread west of a longitudinal margin through central Colorado and central New Mexico, similar to spatial patterns of significantly warmer maximum and minimum temperatures. Significant negative differences of VPD are very few during these seasons. In SON and NDJ, areas with significant negative differences are present, with values from -0.15 kPa to -0.30 kPa, indicating less evapotranspirational demand during the 2000s drought than the 1950s drought at this time of year, most notably in southeastern Arizona and central New Mexico. Significant positive differences are very few during these seasons.

4. Discussion

Several significant differences in seasonal temperature and moisture conditions distinguish the 2000s drought from the 1950s drought. Precipitation during the 2000s drought was greater than the 1950s drought largely over Arizona and western New Mexico from mid- to late autumn and over most of New Mexico in early winter. It was less in early winter across northern Arizona, southeastern Utah, and western Colorado, as well as in early summer mostly over northern Arizona and adjacent areas of neighboring states. Maximum temperatures during the 2000s drought were warmer than the 1950s drought throughout summer in much of the region. Minimum temperatures associated with the 2000s drought were warmer than the 1950s drought from mid-spring through
early autumn throughout much of the region, as well. From late spring through mid-
summer during the 2000s drought, lower dew point temperatures than during the 1950s
drought primarily over northern Arizona and southern Utah contrasted with areas of
higher dew point temperatures in central New Mexico. Higher dew point temperatures
during the more recent drought were widespread from late summer through late autumn,
mainly in Arizona and New Mexico. Evapotranspirational demand during the 2000s
drought from mid-spring through late summer was greater in many areas of the
Southwest than during the 1950s drought. In mid-autumn and early winter during the
2000s drought, evapotranspirational demand was lower than during the 1950s drought,
most notably in southern Arizona and central New Mexico. Apart from these significant
differences, seasonal temperature and moisture conditions during the 1950s and 2000s
droughts were similar.

The significant differences in seasonal precipitation illustrate how spatial patterns
of snowfall and rainfall can vary within the region between pronounced Southwestern
droughts. Our results agree with Breshears et al. (2005) and Quiring and Goodrich (2008)
in that some aspects of the 1950s drought were significantly drier than the 2000s drought,
specifically seasonal precipitation amounts in mid- and late autumn over Arizona and
western New Mexico, and in early winter over most of New Mexico. Decreased influence
of the monsoon in early September, dissipating tropical cyclones from the eastern Pacific
Ocean in September and October, or westerly frontal systems from October through
January may have caused the lower precipitation during the 1950s drought in these
seasons (Smith 1986, Adams and Comrie 1997, Sheppard et al. 2002). Other aspects of
precipitation in our results contrastingly show that the 2000s drought was significantly
drier than the 1950s drought, in particular seasonal precipitation amounts in early winter
across northern Arizona, southeastern Utah, and western Colorado, and in early summer
mostly over northern Arizona and adjacent areas of neighboring states. Decreased
influence of westerly frontal systems from November through January as well as in May,
or perhaps a weaker monsoon in July may have promoted less precipitation during the
2000s drought in these seasons (Higgins and Shi 2000, Sheppard et al. 2002, Grantz et al.
2007).

Our results also are consistent with the Breshears et al. (2005) conclusions that the
2000s drought in the Southwest was significantly warmer than the 1950s drought, and
that the warmer temperatures probably drove greater soil moisture deficits and higher
plant mortality. In particular, maximum temperatures were warmer throughout summer,
minimum temperatures were warmer from mid-spring through early autumn, and
evapotranspirational demand was greater from mid-spring through late summer in much
of the region during the more recent drought. Greater positive anomalies of 500mb
heights in June during the 2000s drought (Figs. B-2D, B-2F, and B-2H) partly reflect
these warmer conditions. Warmer temperatures increase the saturation vapor pressure of
the atmosphere and, without a corresponding increase in actual vapor pressure from
higher atmospheric moisture (i.e., dew point temperatures), drive higher vapor pressure
deficits. Lower dew point temperatures during the more recent drought from late spring
through mid-summer around the four corners area and extreme southern New Mexico and
western Texas increased evapotranspirational demand, in addition to that resulting from
warmer temperatures. Higher dew point temperatures during the 2000s drought in the core area of the monsoon in the United States (Comrie and Glenn 1998), specifically in central New Mexico during summer and southeastern Arizona in late summer, compensated the warmer temperatures enough to maintain evapotranspirational demand similar to what it was during the relatively cooler and drier 1950s drought.

Through our analysis we determined how seasonal differences in temperature and moisture conditions affected evapotranspirational demand during the pronounced Southwestern droughts of the 1950s and 2000s. In the hot-dry foresummer during the 2000s drought, much of the Southwest experienced significantly warmer temperatures that largely drove greater evapotranspirational demand. Lower atmospheric moisture at this time of year over parts of the region also allowed evapotranspirational demand to increase during the more recent drought. Significantly warmer temperatures in the hot-wet monsoon during the 2000s drought primarily drove greater evapotranspirational demand, as well, but only for parts of the region outside of the core monsoon area. Had atmospheric moisture during the more recent drought been as low as during the 1950s drought in the core monsoon area at this time of year, greater evapotranspirational demand during the 2000s drought could have been more spatially extensive.

5. Implications and Conclusions

The hot-dry foresummer and hot-wet monsoon are times during the year when climate has important societal implications in the Southwest. Climatologically hot and dry conditions from late spring through early summer influence the annual peak in urban
water demand, air quality with respect to ozone and particulate matter levels, incidence of valley fever, and intensity of the wildfire season (see review by Ray et al. 2007). Climatologically hot and wet conditions from mid-summer through early autumn influence populations of mosquito species known to be vectors for dengue fever and West Nile virus, incidence of valley fever, decline of the wildfire season, and ranching operations. Drought can negatively affect many of these seasonal circumstances. For example, below average precipitation in winter and early spring can worsen air quality during the hot-dry foresummer. From mid-summer through early autumn, drought can allow the wildfire season to continue further into the normally hot-wet monsoon. With projections of future climate indicating continued warming in the region (Christensen et al. 2007, Meehl et al. 2007b), evapotranspirational demand during the hot-dry and hot-wet seasons possibly will be more severe in future droughts and result in more extreme conditions in the Southwest, a disproportionate amount negatively impacting society.

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**References**


FIG. B-1. Time series of monthly Niño 3.4 (A), PDO (B), and AMO (C) indices, along with annual temperature anomalies (D), percent of normal annual precipitation (E), and annual average PDSI (F) for Arizona, Colorado, New Mexico, Utah, and the Southwest from 1945 through 2007. Dashed rectangles highlight the 1950s (1953-1956) and 2000s (2000-2003) Southwest droughts. Annual temperature anomalies and percent of normal annual precipitation are relative to 1945-2007. NOAA CDC and JISAO provided SST index data. NOAA NCDC provided state and regional data.
FIG. B-3. Map of study domain. The PRISM Group at Oregon State University provided elevation data. PRISM data only are available for domain areas within the United States.
FIG. B-4. Difference of means and field significance test results for seasonal SPI-3. Color gradation quantifies unitless differences between seasonal mean values of the 2000s and 1950s droughts (e.g., DJF SPI-3 mean_{2000-2003} – DJF SPI-3 mean_{1953-1956}). Cross-hatched areas are locally significant at the 95% level. Seasonal maps with a red ‘X’ are not field significant at the 95% level. Positive (negative) values indicate wetter (drier) SPI-3 during the 2000s drought than the 1950s drought.
FIG. B-5. As in Fig. B-4, but for Tmax. Units are °C. Positive (negative) values indicate warmer (cooler) Tmax during the 2000s drought than the 1950s drought.
FIG. B-6. As in Fig. B-4, but for Tmin. Units are °C. Positive (negative) values indicate warmer (cooler) Tmin during the 2000s drought than the 1950s drought.
FIG. B-7. As in Fig. B-4, but for Tdmean. Units are °C. Positive (negative) values indicate higher (lower) Tdmean during the 2000s drought than the 1950s drought.
FIG. B-8. As in Fig. B-4, but for VPD. Units are kPa. Positive (negative) values indicate higher (lower) VPD during the 2000s drought than the 1950s drought.
APPENDIX C: CLIMATIC LIMITS ON FOLIAR GROWTH DURING MAJOR DROUGHTS IN THE SOUTHWESTERN U.S.A.

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CLIMATIC LIMITS ON FOLIAR GROWTH DURING MAJOR DROUGHTS IN THE SOUTHWESTERN U.S.A.

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Abstract

[1] Pronounced droughts during the 1950s and 2000s in the Southwestern U.S.A. (SW) provide an opportunity to compare mesoscale ecosystem responses to anomalously dry conditions before and during the regional warming that started in the late 1970s. This year-round warming has produced fewer cool season freezes, losses in regional snowpack, an 8-10 day advance in spring onset, and hotter summers, all of which should affect vegetation differently across seasons and elevations. Here, we examine indices that represent climatic limits on foliar growth for both drought periods, and evaluate these indices for areas that experienced tree mortality during the 2000s drought. Relative to the 1950s drought, warmer conditions during the 2000s drought reduced effects of suboptimal temperatures on foliar growth at lower elevations in winter and higher elevations in summer. Higher evapotranspirational (ET) demand largely driven by warmer temperatures in the more recent drought was more limiting to foliar growth from spring through summer at lower and middle elevations. At many locations of tree
mortality, constraints of suboptimal temperatures and ET demand on foliar growth were relatively extreme from early spring through late autumn during the 2000s drought. Our analysis shows that in physiographically complex regions like the SW, seasonality and elevational gradients are important for understanding vegetative responses to warming. It also suggests that continued warming will increase the degree to which ET demand limits foliar growth during future droughts, and expand its reach to higher elevations and other seasons.

1. Introduction

[2] An increase in temperatures since the late 1970s, partly attributed to human-caused climate change and positively reinforced by Pacific climate variability [Hegerl et al., 2007; Meehl et al., 2007; Barnett et al., 2008; Bonfils et al., 2008], has profoundly impacted hydroclimatology and ecosystem dynamics in western North America. Hydroclimatic impacts include higher freezing levels [Abatzoglou, 2011], lower ratios of snowfall to total winter precipitation [Knowles et al., 2006], increases of rain-on-snow events [McCabe et al., 2007], and shifts to earlier snowmelt [Stewart et al., 2005] and attendant losses in snowpack [Mote et al., 2005; McCabe and Wolock, 2009]. Well-documented ecological impacts consist of fewer cool season freezes [Weiss and Overpeck, 2005], earlier flowering and leafout [Cayan et al., 2001; Bowers, 2007], earlier emergence for insects [Forister and Shapiro, 2003], advanced breeding for birds [Brown et al., 1999], and increases in large fires [Westerling et al., 2006]. Another potential
impact of the ongoing regional warming is alteration of vegetative responses to recent and future droughts, specifically through changes to climatic limits on foliar growth.

[3] Drought can substantially impact vegetation in the southwestern U.S.A. (SW), a physiographically complex region that spans warm deserts in the lowlands and cool, wet boreal forests in the highlands. For example, major droughts in the 1950s and 2000s induced considerable plant mortality in deserts, grasslands, woodlands, and forests that resulted in ecotone shifts and altered community composition and structure [Neilson, 1986; Allen and Breshears, 1998; Swetnam and Betancourt, 1998; Breshears et al., 2005; Negrón et al., 2009; McAuliffe and Hamerlynck, 2010]. Studies link mortality during the 2000s drought in particular to the ability of warmer temperatures to drive higher water stress [Breshears et al., 2005; Adams et al., 2009] and intensified insect outbreaks [Logan et al., 2003; Raffa et al., 2008]. Indeed, temperatures during the 2000s drought were above average and warmer than temperatures during the 1950s drought (Figure C-1), and drove higher atmospheric demand for evapotranspiration (ET) throughout much of the region [Weiss et al., 2009]. Comparison of the 1950s and 2000s droughts thus offers a unique opportunity to investigate mesoscale ecosystem responses such as mortality and foliar growth to two remarkable climatic events.

[4] In addition to drought variability in the SW, recent warming overlays a highly seasonal climate and complex topography. Temperature follows the typical seasonal cycle from a winter minimum to a summer maximum, whereas most of the precipitation occurs at two distinct times of the year [Sheppard et al., 2002]. From winter through early spring, westerly frontal systems deliver both rain and snow, whereas from mid-
summer through early autumn the North American monsoon (hereafter monsoon) and, to a lesser degree, dissipating tropical cyclones from the eastern Pacific Ocean supply rain. Prominence of either of these peaks in the annual precipitation cycle is spatially variable, with a regional gradient from a winter to early spring dominated regime in the northwest to a regime strongly influenced by the monsoon in the southeast. The period from late spring through early summer, often referred to as the foresummer, separates these two relatively wet seasons and is climatologically the hot and dry part of the year. Terrain ranges from elevations below sea level to above treeline (Figure C-2), and promotes relatively lower temperatures and greater precipitation at higher elevations due to adiabatic cooling and orographic effects, respectively. The seasonality and environmental gradients in this subtropical region lead to ecological phenomena such as short growing seasons for high-elevation forest and tundra during summer [Inouye, 2008], and bimodal growing seasons at lower elevations and latitudes that receive both cool and warm season precipitation [Weiss et al., 2004; Crimmins et al., 2008; Crimmins et al., 2011].

[5] Differences in temperature and moisture between the 1950s and 2000s droughts in the SW may have approached or crossed thresholds of climatic limits on foliar growth, i.e., climatic controls on photosynthesis and transpiration, both seasonally and along elevational gradients. As detailed by Weiss et al. [2009], the main distinction between these events is that much of the region experienced warmer temperatures that increased ET demand during the foresummer in the 2000s drought. Warmer temperatures and greater ET demand continued into mid-summer and early autumn during the 2000s drought, but mainly for parts of the region where monsoon rainfall does not dominate the
annual precipitation cycle. Such changes in climatic conditions can act on vegetation if the changes approach or cross thresholds at which foliar growth becomes inactive or unlimited. For example, conditions during the 2003 summer heat wave in the Swiss Alps approached or crossed both suboptimal temperature and ET demand thresholds, with the particular threshold approached or crossed dependent on a location’s elevation [Jolly et al., 2005a]. Vegetative growth at higher elevations increased due to a longer growing season in a temperature-limited environment, whereas it decreased at lower elevations as a result of enhanced water stress in a moisture-limited environment.

[6] In our SW case, we anticipated that warmer conditions during the 2000s drought relative to the 1950s drought led to similarly divergent changes in how climate limited foliar growth of regional vegetation. We also hypothesized that due to the SW’s highly seasonal climate and complex topography, spatiotemporal variability associated with these two droughts additionally changed climatic constraints on foliar growth. In particular, we expected: (1) that greater ET demand in the foresummer during the 2000s drought [Weiss et al., 2009] was more limiting to foliar growth at lower elevations than during the 1950s drought; (2) that these more limiting conditions of the 2000s drought continued outside of the southeast part of the region from mid-summer through early autumn; (3) that warmer summers during the more recent drought reduced effects of suboptimal temperatures on foliar growth at higher elevations; and (4) that warmer springs and autumns reduced effects of suboptimal temperatures on foliar growth at lower elevations.
[7] To examine these anticipated effects of drought and a warming climate on vegetation, we compared seasonal values of indices that represent climatic limits on foliar growth based on suboptimal temperatures and ET demand between the 1950s and 2000s droughts, and inspected these differences along elevational gradients for the entire region. We also assessed the extremeness of climatic limits on foliar growth with these indices at locations where tree mortality occurred during the 2000s drought. Our intent was not to compare mortality during these major droughts, but rather to use differences in climatic limits on foliar growth between the two events to show how warmer temperatures may already be impacting vegetation across a physiographically complex region.

2. Data and Methods

2.1. Indices of Climatic Limits on Foliar Growth

[8] We computed indices that represent climatic limits on foliar growth during the 1950s and 2000s droughts using gridded observational data compiled by the PRISM Group at Oregon State University. PRISM data are meteorological station data interpolated to 4-km grid cells using a human-expert and statistical knowledge-based system [Daly et al., 2002; http://www.prism.oregonstate.edu]. We used PRISM monthly means of maximum temperature (°C), minimum temperature (°C), and dew point temperature (°C). We also calculated monthly means of vapor pressure deficit (kPa), the difference between saturation and actual vapor pressure, using PRISM data and the formula:
\[ a \exp\left( \frac{bT_{\text{mean}}}{T_{\text{mean}}+c} \right) - a \exp\left( \frac{bT_{\text{dmean}}}{T_{\text{dmean}}+c} \right), \quad (1) \]

where \( a = 0.611 \text{ kPa} \), \( b = 17.502 \), \( c = 240.97^\circ \text{C} \), \( T_{\text{mean}} \) = monthly mean temperature (°C), and \( T_{\text{dmean}} \) = monthly mean dew point temperature (°C) [Campbell and Norman, 1998]. Both dew point temperature and vapor pressure deficit are measures of atmospheric moisture, with the latter being an estimate of the atmospheric demand for ET. The spatial domain for our study is the region between 27°N to 43°N and 117°W to 100°W (Figure C-2). PRISM data were only available for domain areas within the U.S.A., and thus our analysis only included the southwest region of the United States, rather than extending into Mexico.

[9] We used monthly means of minimum temperature and vapor pressure deficit to calculate monthly index values of limits on foliar growth based on suboptimal temperatures and ET demand, respectively [Jolly et al., 2005b]. We also computed a monthly foliar-growth-limit index based on photoperiod (i.e., daylength), which is a function of solar declination and latitude [Campbell and Norman, 1998], for the PRISM grid mesh. Known minimum temperature, vapor pressure deficit, and photoperiod limits on foliar growth define corresponding index values that vary linearly between inactive (i.e., foliar growth is not occurring) and unconstrained (i.e., foliar growth is unlimited) thresholds (see Figure 1 of Jolly et al. [2005b]). The inactive threshold for minimum temperature is -2°C, and the unconstrained is +5°C. For vapor pressure deficit, the inactive threshold is +4.1 kPa and the unconstrained is +0.9 kPa. Photoperiod has inactive and unconstrained thresholds of 10 and 11 hours, respectively. Beyond the
inactive threshold, index values equal zero, whereas index values equal one beyond the unconstrained threshold.

[10] As these thresholds represent measurements of climatic limits on foliar growth from a global array of species, we felt they were appropriate for our generalized analysis of the spatially variable vegetation in the SW. Inactive and unconstrained thresholds of minimum temperature and photoperiod primarily reflect the phenological states of dormancy and budburst, respectively (S. Running, personal communication). In contrast, vapor pressure deficits above the unconstrained threshold value indicate conditions that limit foliar growth through physiological stress in leaves. We considered vapor pressure deficit an appropriate surrogate for precipitation, as increased atmospheric humidity coincides with precipitation events and seasons in this semi-arid region (see discussion in Jolly et al. [2005b]).

[11] We also calculated monthly values of a growing season index (iGS), defined as the product of the monthly values of foliar-growth-limit indices based on suboptimal temperatures (iTmin), ET demand (iVPD), and photoperiod [Jolly et al., 2005b]. Values of the iGS represent a generalized model of climatic constraints on foliar growth that integrates the individual effects of these three limits during the year. Jolly et al. [2005b] showed that iGS is a robust proxy of foliar growth variability in space and time, and is appropriate for historical analyses. By definition, iGS ranges from zero (inactive) to one (unconstrained).
2.2. Comparison of the 1950s and 2000s Droughts

[12] We defined the analysis periods for the 1950s and 2000s droughts as the four-year periods 1953-1956 and 2000-2003, respectively, after Weiss et al. [2009] (Figure C-1). We used monthly values of iTmin, iVPD, and iGS from December 1952 to January 1957 and from December 1999 to January 2004 to calculate three-month mean seasonal values centered on each calendar month for both drought periods. For example, the seasonal January (i.e., DJF) 1953 iTmin value is the average of iTmin monthly values from December 1952, January 1953, and February 1953. Thus, for statistical tests described below, we used a seasonal value for each foliar-growth-limit index and month from January 1953 through December 1956 for the 1950s drought, and from January 2000 through December 2003 for the 2000s drought.

[13] We performed non-parametric Wilcoxon rank sum tests to determine local statistically significant differences (i.e., at individual grid cells) between seasonal location parameters (analogous to mean values) of the 1950s and 2000s droughts for iTmin, iVPD, and iGS. We regarded values of a particular foliar-growth-limit index and season in a drought period (e.g., iTmin values for DJF from 1953-1956) as temporally independent for sample size calculation. We also evaluated whether or not results from one season were independent from results in other seasons for each index. In general, results from seasons with non-overlapping months can be viewed as independent. To address possible spatial autocorrelation in fields of joint Wilcoxon rank sum tests, we carried out non-parametric field significance tests with a permutation randomization
approach of 500 iterations [Livezey and Chen, 1983; Wilks, 2006]. We conducted all tests at the 95% level.

2.3. Extremeness of Climatic Limits at Mortality Locations

[14] To characterize the extremeness of climatic limits on foliar growth during the 2000s drought, we calculated percentile values of seasonal iTmin, iVPD, and iGS on a cell-by-cell basis for all four-year seasonal medians from 1895 through 2003. From the percentile data, we selected grid cells that intersect with locations where tree mortality occurred in SW woodlands and forests during the 2000s drought. Similar mortality data of the 1950s drought were not available, so our evaluation was limited to sites that experienced mortality during the more recent drought. Geographic information system shapefiles produced by DIREnet [www4.nau.edu/direnet/index.html] delineate these locations and represent a suitable regional sample of tree mortality during the 2000s drought. DIREnet creates regional mortality shapefiles by tree species through use of U.S. Forest Service (Intermountain, Rocky Mountain, and Southwest Regions) aerial survey data of tree mortality. These aerial surveys annually map trees undergoing foliage discoloration and defoliation from biotic and abiotic agents while disregarding those that appear to have died in previous years. We utilized annual mortality shapefiles from 2000 through 2003 for Abies concolor (white fir), A. lasiocarpa (subalpine fir), Picea sp. (spruce), Pinus edulis and P. monophylla (pinyon pine), P. flexilis (limberpine), P. ponderosa (ponderosa pine), and Pseudotsuga menziesii (douglas fir). We accessed these mortality data in August 2010. For the purpose of this study, we merged mortality
shapefiles of individual years and species to create one mortality shapefile for the four-year period 2000-2003 and for all of the above species. We performed all geospatial data analyses with ESRI ArcGIS Desktop™ software.

3. Results

[15] We first present the results from seasonal Wilcoxon rank sum and field significance tests in map form for foliar-growth-limit indices based on suboptimal temperatures (iTmin), ET demand (iVPD), and integrated growing season limits (iGS). We display maps for results of all seasons in order to show how differences in climatic limits on foliar growth between the 1950s and 2000s droughts vary across the region and change at different times of the year. However, we only describe locally significant results for seasons that are field significant. We also present these results as scatterplots of significant seasonal differences versus elevation. We then show the extremeness of seasonal climatic limits on foliar growth at locations where tree mortality occurred in SW woodlands and forests during the 2000s drought in bar graph form.

3.1. Wilcoxon Rank Sum and Field Significance Tests

3.1.1. iTmin

[16] Seasonal iTmin during the 1950s and 2000s droughts in the SW shows differences that are field significant only in mid-winter and from mid-spring through mid-autumn (Figure C-3). Locally significant positive differences occur in various areas throughout the region in DJF and from MAM through SON. Values mostly range from
+0.10 to +0.40, indicating that suboptimal temperatures in these seasons were less limiting to foliar growth during the 2000s drought than the 1950s drought. Locally significant negative differences are few during these seasons. The predominantly positive locally significant differences in the region move from lower elevations in mid-winter to higher elevations in mid-summer and back again (Figure C-4).

3.1.2. \textit{iVPD}

[17] Seasonal iVPD for the 1950s and 2000s droughts shows field significant differences only from mid-spring through late summer, and indicates that ET demand was more limiting to foliar growth during the more recent drought at these times of the year (Figure C-5). In MAM and AMJ, locally significant negative differences are widespread in southern New Mexico, Arizona, southern Utah, and the western periphery of the region. Locally significant and widespread negative differences decline in southern New Mexico and southeastern Arizona, expand into northern Utah, and appear in western Colorado from MJJ through JAS. Values generally range from -0.05 to -0.30. Locally significant positive differences are few during these seasons. In the mid-spring through late summer period, the mostly negative locally significant differences occur mainly at lower elevations, and extend into middle elevations during summer (Figure C-4).

3.1.3. \textit{iGS}

[18] Field significant differences of seasonal iGS between the 1950s and 2000s droughts appear only in mid-winter and from mid-spring through late autumn (Figure C-
6). In DJF and from MAM through OND, locally significant positive differences occur in various areas throughout the region and largely correspond to iTmin results (Figure C-3). Values mostly range from +0.10 to +0.40 and indicate that integrated climatic limits in these areas and seasons were less constraining to foliar growth during the more recent drought (Figure C-6). In contrast, locally significant negative differences from MAM through JAS that in general correspond to iVPD results (Figure C-5) mainly occur in the southern and western parts of the region, and are especially widespread in Arizona from AMJ through JAS and in Utah during JJA and JAS (Figure C-6). Values mostly range from -0.05 to -0.30 and show that integrated climatic limits in these areas and seasons were more constraining to foliar growth during the more recent drought. Locally significant negative differences are few in DJF and from ASO through OND. To a great extent, the elevational profiles of locally significant seasonal iGS differences follow a combination of those of iTmin and iVPD (Figure C-4). Locally significant positive differences largely move from lower elevations in mid-winter to higher elevations in mid-summer and back again, whereas locally significant negative differences mainly occur at lower and middle elevations from mid-spring through late summer. A transition from locally significant negative to locally significant positive differences is evident at middle elevations from late spring through late summer.

3.2. Extremeness of Climatic Limits at Mortality Locations

[19] We present the extremeness of seasonal climatic limits on foliar growth at locations where tree mortality occurred in SW woodlands and forests during the 2000s
drought in bar graph form (Figure C-7). Each seasonal panel contains information about foliar-growth-limit indices for the 2000s drought organized along percent and elevation axes. Successive 500-m ranges comprise the elevation axis. For each 500-m elevational range, we show two bars. One depicts the percent of cells whose foliar-growth-limit index values are above the 95\textsuperscript{th} percentile within a given elevational range. The other represents the percent of cells whose values are below the 5\textsuperscript{th} percentile within a given elevational range. Of the 12685 grid cells that intersect with mortality locations, almost all occur between elevations of 1500 and 3500 m (Table C-1).

For example, 100\% of the cells between 500 and 1000 m in elevation have iGS values that are below the 5\textsuperscript{th} percentile during JJA of the 2000s drought (Figure C-7). Values for iGS are also below the 5\textsuperscript{th} percentile for the majority of cells within the 1000-1500, 1500-2000, and 2000-2500 m elevational ranges. Very few to no cells within these elevational ranges have values above the 95\textsuperscript{th} percentile. Between 2500 and 3000 m, however, 14\% of cells have values below the 5\textsuperscript{th} percentile, whereas 20\% of cells have values above the 95\textsuperscript{th} percentile. In further contrast, the majority of cells in the 3000-3500 and 3500-4000 m elevational ranges have values above the 95\textsuperscript{th} percentile. Less than 3\% of the cells in each of these upper elevational ranges have values below the 5\textsuperscript{th} percentile. Overall, these percentages indicate that integrated climatic constraints were extremely limiting for foliar growth during mid-summer of the 2000s drought below 2500 m in elevation at most mortality locations. Above 3000 m, these constraints were extremely unlimiting for a majority of these locations. Again, bar graphs do not depict mortality locations of the 1950s drought.
3.2.1. iTmin

[21] Many seasonal iTmin values of the 2000s drought at locations of tree mortality are above the 95th percentile, indicating that suboptimal temperature constraints were extremely unlimiting for foliar growth during this time period in several of these areas (Figure C-8). These relatively extreme high values mainly occur from early spring through late autumn and vary in elevation depending on the season, similar to the seasonal variability in elevation of locally significant iTmin differences between the two droughts (Figure C-4). No more than 19% of cells have iTmin values above the 95th percentile in elevational ranges below 3000 m from NDJ through JFM of the 2000s drought (Figure C-8). Above 3000 m, no cells have such values. Between 18% and 53% of cells with values above the 95th percentile are in elevational ranges from 1000 to 3000 m during FMA. The percent of cells with values greater than the 95th percentile increases for upper elevational ranges in MAM and AMJ, with these relatively extreme high values appearing at a majority of cells in almost all ranges above 1500 m. From MJJ through JAS, the percent of cells with values above the 95th percentile declines at lower elevational ranges while continuing to account for a majority of cells in ranges above 3000 m. The number of cells with values above the 95th percentile increases during ASO, SON, and OND at most lower elevational ranges, comprising between 20% and 50% of the total in many cases. The percent of cells with these relatively extreme high values declines at elevational ranges above 3000 m during these seasons. Seasonal iTmin values below the 5th percentile occur much less frequently throughout the year, and total no more than 4% of cells within any elevational range and season.
3.2.2. iVPD

[22] In contrast to iTmin, many seasonal iVPD values at mortality locations during the 2000s drought are below the 5\textsuperscript{th} percentile, suggesting that constraints of ET demand were extremely limiting for foliar growth during this time interval in many of these areas (Figure C-9). These relatively extreme low values appear mostly from early spring through late autumn and vary in elevation depending on the season, similar to the seasonal variability in elevation of locally significant iVPD differences between the two droughts (Figure C-4). Less than 1\% of cells in any elevational range have iVPD values below the 5\textsuperscript{th} percentile from NDJ through JFM during the 2000s drought (Figure C-9). In FMA, cells with values below the 5\textsuperscript{th} percentile are between 1\% and 49\% of the total in elevational ranges from 1000 to 3000 m. Relatively extreme low values account for a higher percent of cells in each of these elevational ranges during MAM, and appear in all elevational ranges during AMJ. Values below the 5\textsuperscript{th} percentile occur at a large majority of cells in lower elevational ranges during these two seasons. From MJJ through JAS, all elevational ranges have a large majority of cells with values below the 5\textsuperscript{th} percentile. Cells with these relatively extreme low values are not as numerous as in these seasons for any elevational range in ASO, most notably at upper ranges. Nonetheless, between 50\% and 60\% of cells have relatively extreme low values in elevational ranges from 500 to 2500 m at this time of year. The number of cells with values below the 5\textsuperscript{th} percentile decrease further in SON and OND, comprising between 0\% and 43\% of the total in elevational ranges. Seasonal iVPD values above the 95\textsuperscript{th} percentile comprise less than 1\% of cells within any elevational range and season.
3.2.3. iGS

[23] Seasonal iGS values above the 95th percentile or below the 5th percentile occur at many locations of tree mortality during the 2000s drought, showing that integrated climatic constraints were extremely unlimiting or limiting, respectively, for foliar growth in several of these areas from 2000 through 2003 (Figure C-10). The relatively extreme values mainly appear from early spring through late autumn and vary seasonally in elevation, similar to the seasonal variability in elevation of locally significant iGS differences between the two droughts (Figure C-4). Less than 19% of cells have iGS values above the 95th percentile in elevational ranges below 3000 m from NDJ through JFM of the 2000s drought (Figure C-10). No cells have such values above 3000 m. Also during these seasons, less than 1% of cells in any elevational range have values below the 5th percentile. In FMA, cells with values above the 95th percentile are between 18% and 50% of the total in elevational ranges from 1000 to 3000 m, whereas cells with values below the 5th percentile comprise less than 1% in any of these elevational ranges. Relatively extreme high values account for a greater percent of cells for upper elevational ranges in MAM and AMJ. In contrast, lower elevational ranges show increasing percentages of relatively extreme low values during these seasons. From MJJ through JAS, the amount of cells with values above the 95th percentile remains large in the 3000-3500 and 3500-4000 m elevational ranges, with percentages between 49% and 90%. The number of cells with values below the 5th percentile during these seasons is considerable at elevational ranges below 2000 m, where percentages vary from 48% to 100%. In elevational ranges above 2000 m, these relatively extreme low values appear in
1% to 55% of cells. The number of cells with values above the 95th percentile comprises 7% to 67% of elevational ranges above 1000 m in ASO, SON, and OND. These percentages decrease at upper elevational ranges as this time of year advances, whereas they increase at most lower elevational ranges. Between 1% and 53% of cells with relatively extreme low values occur in ASO, with larger percentages at lower elevational ranges. Seasonal iGS values below the 5th percentile appear infrequently during SON and OND, and total no more than 6% of cells within any elevational range.

4. Discussion

[24] Warmer temperatures in the SW appear to be shifting climatic limits on foliar growth in a complicated, yet understandable, way. In our comparison of the 1950s and 2000s droughts, examination of these limits both seasonally and along elevational gradients is important for this understanding. Minimum temperatures near the range from -2°C to +5°C strongly conform in space to the intricate regional terrain, and move from low elevations in winter to the highest elevations in summer and back again (Figure C-11). As a result, warmer minimum temperatures that are widespread throughout the region in many seasons during the more recent drought [Weiss et al., 2009] are spatially and temporally restricted in changing constraints of suboptimal temperatures on foliar growth. Despite this restriction, warmer minimum temperatures during the 2000s drought still lessen constraints of suboptimal temperatures on foliar growth from low-elevation subtropical desert in mid-winter [Weiss and Overpeck, 2005] to high-elevation forest and tundra in the summer [Inouye, 2008].
[25] Warmer minimum and maximum temperatures that drive higher vapor pressure deficits during the 2000s drought for much of the SW from mid-spring through late summer [Weiss et al., 2009] also have a spatially and temporally restricted effect on the constraints that ET demand places on foliar growth. Vapor pressure deficits near the range from +0.9 kPa to +4.1 kPa distinctly follow the varied terrain of the region as well, but occur at only the lowest elevations in winter and cover all but the highest elevations in summer (Figure C-11). Consequently, constraints of ET demand on foliar growth are higher from spring through summer of the 2000s drought, most notably for scrublands, grasslands, and relatively xeric woodlands and forests [Brown, 1994]. For differences between the two droughts, the transition between climatic limits of suboptimal temperatures and ET demand is most noticeable during summer at elevations from 2000 to 3000 m (Figure C-4). Integration of both suboptimal temperatures and ET demand throughout the year shows that warmer temperatures affected foliar growth constraints along the entire regional elevation gradient during the more recent drought.

[26] At locations of tree mortality in SW woodlands and forests during the 2000s drought, warmer temperatures also appear to push climatic limits on foliar growth towards relatively extreme levels from early spring through late autumn. As with seasonal differences between the two droughts, relatively extreme values of foliar-growth-limit indices reflect the strong seasonal and elevational connections described above. They also show a pronounced transition during summer from more limiting ET demand at lower elevations to less limiting suboptimal temperatures at higher elevations. Interestingly, vapor pressure deficits during summer of the 2000s drought also limit foliar
growth at relatively extreme levels for mortality locations at higher elevations (Figure C-9). However, the influence of ET demand on constraining foliar growth in these settings is not as large as that of suboptimal temperatures, which continues to dominate the integrated climatic limits on foliar growth (Figure C-10). Nonetheless, as temperature-driven water stress and mortality have likely increased throughout elevational ranges of forests in western North America [van Mantgem et al., 2009], physically based process modeling (e.g., Tague and Band [2004]) is warranted to determine the influence of relatively extreme ET demand at all mortality locations.

[27] The importance of elevation in determining seasonal climatic limits on foliar growth highlights the role that spatial variability of climate plays in impacting vegetation during SW droughts. During the 1950s drought, the epicenter of anomalously dry conditions was in the southeastern part of the region, over the U.S.A.-Mexico borderlands and southern High Plains, whereas the epicenter of the 2000s drought was more regionally centered over the Colorado Plateau. Physiography and vegetation shift dramatically from extensive lowlands dominated by desertscrub and grassland in the south to expansive highlands clad in woodlands and forests to the north, across a gain of ~1500 m in elevation (Figure C-2). The different spatial patterns of climatic conditions during the 1950s and 2000s drought likely interacted with regional landscapes to produce dissimilar patterns in ecosystem responses, affecting which species and populations were impacted. For instance, tree mortality in the 1950s drought was most severe in the south [Swetnam and Betancourt, 1998], whereas during the 2000s drought, it was more extensive to the north [Breshears et al., 2005]. In our study, more limiting ET demand
from spring through summer during the 2000s drought (Figure C-5) impacted middle and high elevations to the north, which are commonly occupied by stands of *Pinus edulis* (pinyon pine) and *P. ponderosa* (ponderosa pine). Future studies comparing mesoscale ecosystem responses to the major droughts of the 1950s and 2000s should consider both recent, region-wide warming and intra-regional moisture differences, and their interaction with the complex SW physiography.

[28] Our analysis suggests that further warming [Christensen et al., 2007; Karl et al., 2009] and episodic drought in the SW during this century will continue to produce climatic conditions that approach or cross thresholds at which foliar growth varies between inactive and unconstrained throughout the elevational gradient. And, at least in the case of tree mortality locations during the 2000s drought, future climatic constraints on foliar growth under such circumstances will be relatively extreme. Thresholds based on suboptimal temperatures and ET demand will shift to progressively higher elevations for a given season, or similarly occur earlier (later) at the start (end) of the growing season for a given location. Consequently, suboptimal temperature constraints will become less affecting on foliar growth in space and time, whereas the area over, length of time during, and level to which ET demand limits foliar growth will increase. This last suggestion is of particular concern as regional warming already is a likely contributor to water stress and tree mortality [Breshears et al., 2005; van Mantgem et al., 2009], and is projected to become increasingly so [Adams et al., 2009; Williams et al., 2010]. Under current rates of global greenhouse gas emissions, higher temperatures in the SW that influence future droughts by potentially raising ET demand are virtually assured.
Amelioration of higher ET demand by monsoon moisture for parts of the region is uncertain, as simulations of monsoon variability are inconsistent [Liang et al., 2008]. Increases in ET demand during future regional droughts will not only intensify its constraint on foliar growth where and when it already occurs, but expand its reach to higher elevations and additional seasons.

[29] Acknowledgments. The NOAA-funded Climate Assessment for the Southwest (CLIMAS) project (NOAA grant #NA16GP2578) provided support to JLW and JTO for this study. We thank DIREnet and Kirsten Ironside for assistance with shapefiles of tree mortality, and Christopher Castro for comments that helped improve the manuscript.

References


Table C-1. Number of grid cells that intersect with locations where tree mortality occurred in SW woodlands and forests during the 2000s drought by 500-m elevational ranges (grid cell sum = 12685).

<table>
<thead>
<tr>
<th>Elevational Range</th>
<th>Grid Cell Count</th>
</tr>
</thead>
<tbody>
<tr>
<td>$3500 &lt; x \leq 4000m$</td>
<td>141</td>
</tr>
<tr>
<td>$3000 &lt; x \leq 3500m$</td>
<td>1230</td>
</tr>
<tr>
<td>$2500 &lt; x \leq 3000m$</td>
<td>3543</td>
</tr>
<tr>
<td>$2000 &lt; x \leq 2500m$</td>
<td>5041</td>
</tr>
<tr>
<td>$1500 &lt; x \leq 2000m$</td>
<td>2472</td>
</tr>
<tr>
<td>$1000 &lt; x \leq 1500m$</td>
<td>241</td>
</tr>
<tr>
<td>$500 &lt; x \leq 1000m$</td>
<td>17</td>
</tr>
</tbody>
</table>
Figure C-1. Time series of annual average temperature anomalies (a), percent of normal annual precipitation (b), and annual average Palmer Drought Severity Index (PDSI; c), a soil moisture proxy [Palmer, 1965; Heim, 2002], for Arizona, Colorado, New Mexico, Utah, and the SW from 1945 through 2007. Dashed rectangles highlight the 1950s (1953-1956) and 2000s (2000-2003) SW droughts. Annual temperature anomalies and percent of normal annual precipitation are relative to 1945-2007. NOAA NCDC provided state and regional data.
Figure C-2. Map of study domain. The PRISM Group at Oregon State University provided elevation data. PRISM data were only available for domain areas within the U.S.A.
Figure C-3. Wilcoxon rank sum and field significance test results for seasonal iTmin. Color gradation quantifies unitless differences between seasonal median values of the 2000s and 1950s droughts (e.g., DJF iTmin median$_{2000-2003}$ – DJF iTmin median$_{1953-1956}$). White areas in U.S.A. denote locations with index values of zero (inactive) or one (unconstrained) for both drought periods. Cross-hatched areas are locally significant at the 95% level. Seasonal maps with a red ‘X’ are not field significant at the 95% level. Positive (negative) values indicate that suboptimal temperatures were less (more) limiting to foliar growth during the 2000s drought than the 1950s drought.
Figure C-4. Scatterplots of locally significant differences between seasonal median values of the 1950s and 2000s droughts and corresponding elevations for iTmin (red), iVPD (blue), and iGS (green). Positive (negative) values indicate that the 2000s drought was less (more) limiting than the 1950s drought to foliar growth for a given index. Scatterplots in gray refer to seasonal maps that are not field significant at the 95% level.
Figure C-5. As in Figure C-3, but for iVPD. Positive (negative) values indicate that ET demand was less (more) limiting to foliar growth during the 2000s drought than the 1950s drought.
Figure C-6. As in Figure C-3, but for iGS. Positive (negative) values indicate that integrated climatic limits were less (more) limiting to foliar growth during the 2000s drought than the 1950s drought.
Figure C-7. Bar graph characterizing the extremeness of iGS values during JJA at locations of woodland and forest tree mortality during the 2000s drought. For each 500-m elevational range (i.e., $500 < x \leq 1000$, $1000 < x \leq 1500$, …), bars depict the percent of cells whose iGS values are above the 95th percentile and below the 5th percentile within that range. Values above the 95th percentile indicate that integrated climatic constraints were extremely unlimiting for foliar growth, whereas values below the 5th percentile suggest that these constraints were extremely limiting. This bar graph does not depict mortality locations of the 1950s drought.
Figure C-8. As in Figure C-7, but for iTmin values in all seasons. Values above the 95th percentile indicate that suboptimal temperature constraints were extremely unlimiting for foliar growth, whereas values below the 5th percentile suggest that these constraints were extremely limiting. Bar graphs do not depict mortality locations of the 1950s drought.
Figure C-9. As in Figure C-8, but for iVPD. Values above the 95th percentile indicate that constraints of ET demand were extremely unlimiting for foliar growth, whereas values below the 5th percentile suggest that these constraints were extremely limiting.
Figure C-10. As in Figure C-8, but for iGS. Values above the 95\textsuperscript{th} percentile indicate that integrated climatic constraints were extremely unlimiting for foliar growth, whereas values below the 5\textsuperscript{th} percentile suggest that these constraints were extremely limiting.
Figure C-11. Conceptualization of how suboptimal temperature and ET demand limits on foliar growth vary along the elevational gradient of the SW for (a) winter, (b) spring, (c) summer, and (d) autumn. Suboptimal temperature has inactive and unconstrained thresholds at minimum temperatures of -2°C and +5°C, respectively. For ET demand, the inactive threshold is +4.1 kPa and the unconstrained +0.9 kPa, as measured by vapor pressure deficit. Both suboptimal temperature and ET demand have a spatially and temporally restricted influence on the constraints that they individually place on foliar growth. Warmer temperatures during the 2000s drought affected foliar growth constraints along the entire elevational gradient with less limiting suboptimal temperatures and more limiting ET demand over large parts of the region.
APPENDIX D: POTENTIAL USE OF GRIDDED FREE-AIR TEMPERATURES IN DETERMINING AND PREDICTING FOLIAR GROWTH LIMITS IN THE SOUTHWESTERN U.S.

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POTENTIAL USE OF GRIDDED FREE-AIR TEMPERATURES IN DETERMINING AND PREDICTING FOLIAR GROWTH LIMITS IN THE SOUTHWESTERN U.S.

Jeremy L. Weiss

Abstract

[1] Unusually warm and dry conditions during the last decade drove foliar growth limits to extreme levels at many woodland and forest locations in the southwestern U.S. (SW). Further warming and drying is expected in coming decades, as indicated by present-day global- (~100 to 150 km) and regional-scale (~25 to 50 km) models (GCMs and RCMs). However, projections of foliar growth limits are needed at spatial scales appropriate to biomes (~5 km). Free-tropospheric air temperatures strongly influence the altitudinal gradient of temperature in mountainous areas, and could support a statistical downscaling approach to generate future temperature changes in complex terrain and at scales finer than those represented in GCMs and RCMs. This study compares free-air and near-surface air temperatures in the SW along environmental gradients using gridded observational (PRISM) and reanalysis (CFSR) datasets over the latter part of the 20th century. Monthly free-air temperature climatologies at land surface altitudes appear to capture the seasonal variability and spatial features driven by regional climate and elevation that are present in the near-surface air temperature data, and are generally +1° to +3°C warmer throughout the year. Monthly anomalies of the two temperature datasets show strong, widespread co-variability throughout the year, with the majority of correlations above +0.8. Consistency between the two datasets does not appear to systematically break down along latitude, longitude, or elevation gradients. Further
development of free-air temperatures for application to fine-scale projections of temperature, and hence foliar growth limits, is justified.

1. Introduction

[2] Warming over the past few decades in the southwestern U.S. (SW) has been some of the most rapid in the country [Karl et al., 2009]. In recent years, this warming coincided with a prolonged drought, exacerbating atmospheric demand for evapotranspiration (ET) [Weiss et al., 2009] and inducing broad-scale tree mortality through higher water stress [Breshears et al., 2005; Adams et al., 2009] and intensified insect outbreaks [Logan et al., 2003; Raffa et al., 2008]. Warming in the region is projected to continue [Solomon et al., 2007; Karl et al., 2009], and its role in raising ET demand and contributing to water stress and tree mortality is expected to increase [Adams et al., 2009; Weiss et al., 2009; Williams et al., 2010].

1.1. Recent Variability of SW Foliar Growth Limits

[3] Another potential vegetative response to ongoing warming in the SW is changes to foliar growth limits based on suboptimal temperatures and ET demand. These limits represent the degree to which climatic conditions constrain foliar growth, and are defined by known thresholds at which foliar growth is inactive or unconstrained [Jolly et al., 2005]. Minimum temperatures of -2°C and +5°C establish the inactive and unconstrained thresholds of suboptimal temperatures, respectively. Vapor pressure deficits of +4.1 kPa and +0.9 kPa set these respective thresholds of ET demand. Changes
in climate such as higher temperatures can act on foliar growth if the changes include minimum temperatures or vapor pressure deficits within the above ranges. Foliar growth limits are highly variable in the SW through space and time (i.e., seasons) due to the region’s complex physiography (Figure D-1) [Weiss et al., in review].

[4] Anomalous warmth during the 2000s drought (2000-2003) in the SW [Weiss et al., 2009] reduced the effects of suboptimal temperatures at lower elevations (< 2000 m) during winter and higher elevations (> 2000 m) in summer relative to another pronounced regional drought during the 1950s (1953-1956) [Weiss et al., in review]. Warmer temperatures also drove more limiting ET demand from spring through summer at lower and middle elevations (< 3000 m). Furthermore, higher temperatures drove both of these limits of foliar growth to relatively extreme levels at tree mortality sites of the 2000s drought throughout most of the growing season. Weiss et al. [in review] suggest that further increases in regional temperatures will intensify the constraints that ET demand places on foliar growth during future droughts, and expand their reach to higher elevations and other seasons.

[5] Not only does the 2000-2003 period show anomalous constraints on foliar growth in comparison to a previous drought, but so do several other recent years relative to the entire observational record for three SW woodland and forest biomes (Figure D-2) in which much of the recent tree mortality occurred (Figure D-3). In the late 1990s and throughout the 2000s, suboptimal temperatures are less limiting during spring, summer, and autumn within the Rocky Mountain subalpine conifer forest, Rocky Mountain montane conifer forest, and Great Basin conifer woodland (Figure D-4) [Brown et al.,
Warmer, less limiting minimum temperatures appear during these years from May through October in subalpine forests, with an unprecedented occurrence of foliar growth being almost completely unconstrained by suboptimal temperatures in July and August. For montane forests, reduced effects of suboptimal temperatures emerge from April through June and in September and October. Less limiting suboptimal temperatures are evident in regional woodlands from March through May and in October.

[6] The most limiting ET demand for all three SW woodland and forest biomes during the observational record appears during summer in recent years, coincident with the warm and dry conditions of the recent drought (Figure D-5). In subalpine forests, limits of ET demand on foliar growth only appear in July of the 2000s. This was the case for montane forests, as well, along with an unusual number of years with relatively more limiting ET demand in June. In woodlands, the most limiting ET demand during the observational record occurred in June and July of the 2000s, with comparatively high ET demand additionally taking place in May of this decade.

[7] Integrating limits of both suboptimal temperatures and ET demand on foliar growth suggests that increases in ET demand during the 2000s had greater impact in the lower elevation woodlands and montane forests than in the higher elevation subalpine forests of the SW. This effect on foliar growth in woodlands and montane forests manifests as a shift in the optimal atmospheric conditions for foliar growth from a peak in mid-summer to peaks in both early and late summer (Figure D-6). In subalpine forests, integrated growing season limits are less constraining from May through October and a relatively high occurrence of nearly unconstrained foliar growth emerges in July and
August, similar to foliar growth constraints influenced solely by minimum temperatures. Comparable variability of integrated growing season limits takes place in montane forests during April, May, September, and October, and in woodlands during March, April, and October. More limiting ET demand in recent years seems to have driven more constraining integrated growing season limits during June and July in both montane forests and woodlands.

1.2. Projecting Variability of SW Foliar Growth Limits

[8] Changes to foliar growth limits can alter plant growth, adjust the uptake of carbon by terrestrial ecosystems, and impact the carbon cycle [Cleland et al., 2007]. With the expectation that increasing temperatures will be a mainstay of future climate in the SW [Solomon et al., 2007; Karl et al., 2009], an important question is whether climatic limits on foliar growth in woodlands and forests of the SW during future dry spells and years will exceed those attained in the late 1990s through the 2000s. Key to addressing this is determining the amount and timing of warming in upcoming decades with projections of future regional climate.

[9] Projections of the region’s climate through either global-scale general circulation models (GCMs) or regional-scale downscaling efforts (e.g., statistical or dynamical, the latter of which utilizes regional climate models, or RCMs) are useful in supplying increases of future temperatures. The current generation of GCMs provides robust estimates of increases in future temperatures [Randall et al., 2007] at relatively coarse horizontal resolutions (~100 to 150 km), although a limited number of advanced
GCMs are now operating at horizontal resolutions typical to RCMs (~25 to 50 km). Regional-scale downscaling generates projections of future climate at finer spatial scales than those of the current generation of GCMs in part for climate change impact assessments [Maurer et al., 2007]. Statistical, or empirical, downscaling uses cross-scale relationships based on observations and integrates them with GCM climate projections [Bader et al., 2008]. Various methods of statistical downscaling include linear and non-linear regression, analogues, and weather generators. Strengths of this approach are that it is computationally inexpensive, can generate information at scales finer than those of RCMs, and can be applied to a wide range of variables. Potential drawbacks of this approach are the assumption that cross-scale relationships between observations hold through time (i.e., are stationary), and that the broader regional climate as generated by the GCM is accurate. Dynamical downscaling utilizes RCMs (i.e., numerical models based on fundamental conservation laws and similar to GCMs) to resolve features that are too fine in scale to be represented well or at all in GCMs. Although this approach represents dynamical processes that are not explicitly defined in statistical downscaling, limitations include that it is computationally expensive, and that additional uncertainties may be introduced through parameterization, as well as through bias and scale mismatches with parent GCMs [Bader et al., 2009; Pielke and Wilby, 2011]. Despite the usefulness of both global- and regional-scale climate projections, a need still exists to get information on possible future temperatures at relatively finer spatial scales across elevational gradients and complex terrain [Pepin and Lundquist, 2008; Dobrowski et al., 2009] that corresponds to spatiotemporal variations of foliar growth limits in the SW.
1.3. **Free-tropospheric Air Temperature**

[10] The free troposphere is a relatively stable and less complex atmospheric layer above the turbulent atmospheric boundary layer [Stull, 2006]. In contrast to the boundary layer, it has minimal interaction with Earth’s surface, and free-tropospheric air temperatures are less spatially variable. Nonetheless, the vertical profile of free-air temperatures (i.e., the decrease of free-air temperatures with height) strongly influences the altitudinal gradient of temperatures in mountainous areas [Barry, 2008; Dobrowski et al., 2009], and can be especially dominating on mountain summits and freely draining slopes than surface radiative and microclimatic effects [Pepin and Seidel, 2005; Pepin and Lundquist, 2008]. Studies at a global scale [NRC, 2000; Pepin and Norris, 2005; Pepin and Seidel, 2005] and for the domain of the western U.S.A. [Pepin et al., 2005] generally report differing trends in near-surface air and free-air temperatures, with near-surface air temperatures often warming more quickly than free-air temperatures over recent decades. However, as relationships between near-surface and free-tropospheric air temperatures vary on local scales [Pepin and Losleben, 2002; Seidel and Free, 2003], and as the relative influences of free-air temperatures and landscape-scale physiography on local temperature are largely unknown [Dobrowski et al., 2009], comparison of free-air and near-surface air temperatures along environmental gradients remains valuable for studies of climate change and its impacts in mountainous regions like the SW.

[11] If anomalies of near-surface and free-tropospheric air temperatures co-vary, it may be possible to get information on possible future temperatures at spatial scales finer than those in GCMs and RCMs. Both GCMs and RCMs have multiple levels in the
vertical direction for individual horizontal grid cells that describe smoothly varying characteristics of Earth’s atmosphere at different altitudes, including temperature (Figure D-7). Horizontal and vertical interpolation of temperature from such model output can be applied to land surface elevations on a finer grid to estimate free-air temperature at the same altitude of the land surface. This would allow for comparison to near-surface air temperatures, and if favorable, approximation of near-surface air temperatures from free-air ones. Here, we examine the suitability of this technique by comparing near-surface air and interpolated free-air temperatures along regional environmental gradients using gridded observational and reanalysis datasets over the latter part of the 20th century. Favorable results from this approach facilitate projecting the amount and timing of warming in upcoming decades, and the corresponding changes in foliar growth limits at the spatial scale of SW woodlands and forests.

2. Methods

2.1. Data

[12] We used PRISM data compiled by the PRISM Group at Oregon State University as the gridded near-surface (2-m) air temperature dataset for comparison with interpolated free-air temperatures. PRISM data are meteorological station data interpolated to 4-km grid cells using a human-expert and statistical knowledge-based system that accounts for the elevation and vertical layer in which a grid cell occurs, temperature observations at several meteorological stations of varying proximity to a grid cell, nearness of grid cell to a coast, and the topographic orientation of a grid cell [Daly et
We calculated monthly mean temperature from 1979-1999 (see below for explanation of analysis period) using PRISM monthly means of maximum temperature (°C) and minimum temperature (°C). The spatial domain for our study is the region between 27°N to 43°N and 117°W to 100°W (Figure D-1). PRISM data were only available for domain areas within the U.S.A., and thus our analysis only included the southwest region of the United States, rather than extending into Mexico.

[13] We interpolated free-air temperatures from Climate Forecast System Reanalysis (CFSR) data [Saha et al., 2010; http://dss.ucar.edu/pub/cfsr.html] developed by NOAA’s National Centers for Environmental Prediction (NCEP). CFSR, the most recent NCEP reanalysis effort, is a global, high-resolution, coupled atmosphere-ocean-land surface-sea ice model that provides estimates for the states of these components of the Earth system. We used CFSR monthly means of geopotential height (m) and temperature (converted from °K to °C) resolved on a 0.5° by 0.5° grid. CFSR data include both geopotential height and temperature at 37 different isobaric surfaces that range from 1000 mb to 1 mb. We used the 16 levels between 1000 mb and 500 mb (1000, 975, 950, 925, 900, 875, 850, 825, 800, 775, 750, 700, 650, 600, 550, and 500 mb), which span from the lowest (in altitude) available isobaric surface to a pressure level that is always above the highest land surface elevations in the SW (Figure D-7). Although CFSR data are available for the years 1979-2010, we only utilized data through 1999, as we will perform a future comparison with Intergovernmental Panel on Climate Change (IPCC)
Fourth Assessment Report (AR4) simulations of the 20\textsuperscript{th} Century Climate in Coupled Models (20C3M) scenario that, for some models, end in 1999.

2.2. **Interpolation Methods for Reanalysis Data**

[14] We first interpolated CFSR geopotential heights (m) and temperatures (°C) of the isobaric surfaces from the 0.5° by 0.5° (~50 km by 50 km) CFSR grid to the approximately 4 km by 4 km PRISM grid using linear, spline, and cubic interpolation methods [Pepin and Norris, 2005; Pepin and Seidel, 2005]. With these interpolated heights and temperatures at pressure levels, we calculated a linear lapse rate between sequential isobaric surfaces for each grid cell. To compute free-air temperature, we applied individual lapse rates (i.e., a linear lapse rate between sequential isobaric surfaces) to surface elevations from the PRISM digital elevation model (DEM) that fell within the height (i.e., altitude) ranges of the corresponding isobaric surfaces (Figure D-7). We used the lapse rate between the 1000 mb and 975 mb isobaric surfaces to elevations that fell below the altitude of the 1000 mb level.

2.3. **Comparison of Free-air and Near-surface Air Temperatures**

[15] To compare the CFSR interpolated free-air temperature data statistically to the PRISM near-surface air temperature data, we performed difference of means tests for paired samples to determine local statistically significant differences (i.e., at an individual grid cell) between monthly mean values from 1979 through 1999. We treated differences between temperature values of a particular month (e.g., 1979-1999 January values) as
temporally independent for sample size calculation. To address possible spatial autocorrelation in fields of joint difference of means tests for paired samples, we carried out non-parametric field significance tests with a permutation randomization approach of 500 iterations [Livezey and Chen, 1983; Wilks, 2006]. All tests were conducted at the 95% level. For additional comparison, we calculated Pearson’s linear correlation coefficients between monthly anomalies of CFSR interpolated free-air temperature data and PRISM near-surface air temperature data. As CFSR assimilates radiosonde and satellite data but no near-surface air temperature observations, we considered CFSR interpolated free-air temperature and PRISM near-surface air temperature data as independent.

3. **Results**

[16] We present results only for the CFSR interpolated free-air temperature data that are based on altitude and temperature of pressure levels that we interpolated using a spline method. We found that results based on pressure level interpolated data using the linear and cubic interpolation methods are similar to those presented below.

3.1. **Climatologies of Free-air and Near-surface Air Temperatures**

[17] Monthly temperature climatologies of the CFSR interpolated free-air temperature data appear to capture the seasonal variability and spatial features driven by regional climate and elevation that are present in corresponding climatologies based on the PRISM near-surface air temperature data over the 1979-1999 period (Figures D-8
through D-11). Application of linear lapse rates between subsequent pressure levels produces reasonable temperature variability along elevational gradients (compare Figure D-1 with Figures D-8 through D-11, CFSR interpolated free-air data). The CFSR free-air data characterize temperature along elevational gradients in finer detail than PRISM near-surface air temperature data. For example, features such as mountain ranges and the Grand Canyon in northern Arizona (Figure D-1) look more resolved in the CFSR dataset than in the PRISM data.

3.2. Differences of Monthly Climatologies

[18] Monthly free-air temperatures based on interpolated CFSR data for the SW from 1979 through 1999 show significant differences from corresponding PRISM near-surface air temperature data throughout the year (Figures D-12 and D-13). Significant positive differences are widespread in all months and mostly range from +1° to +3°C, indicating that, in large part, the free-air temperatures estimated from interpolated CFSR data are warmer than near-surface air temperatures based on PRISM data. Additional significant positive differences from +4° to +6°C exist in isolated areas in south-central Colorado, northeastern Utah, and the northern periphery of the region from December through February. Significant negative differences occur to a lesser extent in all months, and are most common from February through May in the northern part of the region, with values largely between -1° and -3°C. For these areas, the free-air temperatures based on interpolated CFSR data are cooler than the near-surface air temperatures based on PRISM data.
[19] For all months, the spatial variability of positive and negative differences results in weak to medium correlations between significant differences and latitude, longitude, and elevation (Table D-1). Correlations between the differences and latitude are significantly positive for January, from June through September, and December, and are significantly negative from February through May, October, and November. For the differences and longitude, correlation values are significantly positive from November through June and significantly negative from August through October. Significantly positive correlations exist between the differences and elevation from June through August, whereas significantly negative correlations occur for all other months of the year.

3.3. Correlations of Monthly Anomalies

[20] Monthly anomalies of the CFSR interpolated free-air and PRISM near-surface air temperatures show similar variability throughout the year and the SW (Figures D-14 and D-15) for the 1979-1999 period. Significant positive correlations are widespread in all months and mostly are above +0.8, indicating that free-air and near-surface air temperatures strongly co-vary in large part on a monthly time scale. Additional but fewer significant positive correlations above +0.4 also exist in several parts of the region during all months. Few highly localized areas have non-significant correlations, most notably in January, March, April, May, July, August, and December. No significant negative correlations occur between monthly anomalies of the CFSR interpolated free-air and PRISM near-surface air temperatures.
[21] For all months, the spatial variability of significantly positive correlations between anomalies of CFSR-based free-air and PRISM-based near-surface air temperature results in weak to medium associations between the correlation values and latitude, longitude, and elevation (Table D-2). Associations between the correlations and latitude are significantly positive for April, and from June through November, and significantly negative in the remaining months. For the correlations and longitude, linear relationships are significantly positive from November through February, June, August and September, and significantly negative in April, May, and October. Significantly positive associations exist between the correlations and elevation only in October, whereas significantly negative correlations occur for all other months of the year.

4. Discussion

[22] Interpolating free-air temperatures from lapse rates derived from subsequent atmospheric pressure levels and applying them to land surface elevations on a finer grid appears to produce reasonable values, suggesting that further development of this technique for the study of climate change impacts in the SW is justified. As the vertical profile of free-air temperatures strongly influences the altitudinal gradient of temperatures in mountainous areas [Barry, 2008; Dobrowski et al., 2009], it is encouraging that monthly climatologies and anomalies of CFSR-based free-air and PRISM-based near-surface air temperatures display similar spatial and temporal variability. Temperature values at PRISM grid cells are a function of elevation, the vertical layer in which a grid cell occurs, temperature observations at several
meteorological stations of varying proximity, topographic orientation, and proximity to a coast [Daly et al., 2002]. In contrast, our methodology simply uses the elevation of a grid cell and the interpolated lapse rate of subsequent pressure levels in which the grid cell occurs to calculate surface temperature. Nor is it discouraging that the two temperature datasets analyzed here show significant differences. Apart from uncertainties of input observations for both CFSR and PRISM data, free-air and near-surface air temperatures are not measures of the same variable, as boundary layer dynamics and landscape-scale topoclimatic effects such as slope and aspect can strongly influence the latter [Pepin et al., 2005; Pepin and Seidel, 2005; Dobrowski et al., 2009]. Local physiographic effects on temperature may explain some of differences between the datasets in our study [Dobrowski et al., 2009].

[23] It is also encouraging that biases of differences between the monthly climatologies of free-air and near-surface air temperatures along environmental gradients of latitude, longitude, and elevation in the SW appear to be governed by regional physiography. For example, in the case of latitude, positive correlations during December and January (Table D-1) point to the occurrence of relatively high positive differences in south-central Colorado, northeastern Utah, and the northern periphery of the region. Negative correlations from February through May stem from the presence of negative differences in the northern part of the region centered on western Colorado. Negative correlations between the differences and elevation also exist for these months, as several parts of this area are at higher elevations relative to the rest of the region (Figure D-1). Despite the presence of biases along environmental gradients (Tables D-1 and D-2), our
technique for interpolating free-air temperatures to a land surface grid of finer resolution does not look to have systematic errors along latitude, longitude, or elevation. Nonetheless, determining reasons for these biases (e.g., local topoclimatic effects) is another step we will take in another study to further evaluate this approach.

[24] The comparability of our CFSR-based free-air temperatures to PRISM-based near-surface air temperatures also suggests that our technique may be applicable to output from GCMs and RCMs for projecting changes to free-air temperatures, which could lead to projecting foliar growth limits in SW woodlands and forests. As an initial step, we will apply our methodology utilizing pressure level heights and temperatures to interpolate free-air temperatures from 20C3M runs of IPCC AR4 for comparison to the 1979-1999 free-air and near-surface air temperature data presented here. We will then similarly generate projections of interpolated free-air temperatures data from the IPCC Special Report on Emissions Scenarios (SRES) A1b scenario between 2000 and 2050. This scenario has the greatest temperature increase between the present and the middle of the 21st century, our time domain of interest, and thus is most similar to the temperatures already observed over the past ~12 years. We also foresee using RCM climate projection data that dynamically downscales GCM AR4 output in similar fashion, as well as IPCC Fifth Assessment Report (AR5) data, which are now becoming available.

[25] As temperature is an important factor on moisture balance in the region [Stephenson, 1998], further warming in the SW will be a primary driver of future ecological changes, including those potentially driven by the limits of suboptimal temperature and ET demand on foliar growth in woodland and forests. Moisture
variability will also play an important role in the limits of ET demand, and hence overall growing season conditions. Whereas rising temperatures are virtually assured if radiative forcing of climate continues to increase, moisture anomalies in future climate scenarios are sometimes subject to notable uncertainty [Solomon et al., 2007; Karl et al., 2009]. To incorporate moisture variability into our projections of foliar growth limits based on ET demand, we intend to follow an approach similar to Gutzler and Robbins [2010], who combined projections of statistically downscaled near-surface air temperature trends for the 21st century with temperature and moisture variability from the 20th century. Trends for 21st century projected temperatures in our analysis, however, will be based on an integration of interpolated free-air temperature projections and topoclimatic factors [Dobrowski et al., 2009]. Integrating projected changes in temperature with observed temperature and moisture variability will permit us to generate scenarios of foliar growth limits for upcoming decades at relatively fine spatial scales, and compare them to those from the late 1990s through the 2000s.

[26] **Acknowledgements.** The NOAA-funded Climate Assessment for the Southwest (CLIMAS) project (NOAA grant #NA16GP2578) provided support to JLW for this study. The CFSR data for this study are from the Research Data Archive (RDA), maintained by the Computational and Information Systems Laboratory (CISL) at the National Center for Atmospheric Research (NCAR), in dataset number ds093.2. NCAR is sponsored by the National Science Foundation (NSF).
References


Table D-1. Correlation values between local statistically significant differences of paired sample difference of means tests for monthly CFSR-based free-air and PRISM-based near-surface air temperatures and latitude, longitude, and elevation. Pearson’s linear correlation coefficients ($r$) values in bold have $p$-values less than 0.05.

<table>
<thead>
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<th>differences v. elevation</th>
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<td>+ 0.0843</td>
<td>+ 0.1192</td>
<td>- 0.1326</td>
</tr>
<tr>
<td>Feb</td>
<td>- 0.1452</td>
<td>+ 0.2463</td>
<td>- 0.3540</td>
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<tr>
<td>Mar</td>
<td>- 0.4154</td>
<td>+ 0.3242</td>
<td>- 0.5945</td>
</tr>
<tr>
<td>Apr</td>
<td>- 0.4145</td>
<td>+ 0.2184</td>
<td>- 0.5541</td>
</tr>
<tr>
<td>May</td>
<td>- 0.2568</td>
<td>+ 0.1572</td>
<td>- 0.2812</td>
</tr>
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<td>Jun</td>
<td>+ 0.0014</td>
<td>+ 0.0986</td>
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<td>Jul</td>
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<tr>
<td>Oct</td>
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<td>- 0.2576</td>
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<td>- 0.1574</td>
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<td>Dec</td>
<td>+ 0.0289</td>
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<td>- 0.1798</td>
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Table D-2. Correlation values between statistically significant correlations of monthly anomalies CFSR-based free-air and PRISM-based near-surface air temperature anomalies and latitude, longitude, and elevation. Pearson’s linear correlation coefficients (r) values in bold have p-values less than 0.05.

<table>
<thead>
<tr>
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<th>correlation v. latitude</th>
<th>correlation v. longitude</th>
<th>correlation v. elevation</th>
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Figure D-1. Map of the SW and the spatial domain we used in this study. The PRISM Group at Oregon State University provided elevation data. PRISM data were only available for domain areas within the U.S.A.
Figure D-2. Map of the SW with spatial distributions of the woodland, montane, and subalpine biomes [Brown et al., 2007]. These biomes occur successively along much of the middle and higher elevations in the SW. The inset locates the SW and the woodland, montane, and subalpine biomes within western North America, and shows the spatial domain we used for this study (dashed white line).
Figure D-3. As in Figure D-2, but with locations where tree mortality occurred in the SW during the 2000s drought (white areas). Geographic information system shapefiles produced by DIREnet [www4.nau.edu/direnet/index.html] delineate these locations and represent a regional sample of tree mortality. DIREnet creates regional mortality shapefiles by tree species through use of U.S. Forest Service (Intermountain, Rocky Mountain, and Southwest Regions) aerial survey data of tree mortality. Annual mortality shapefiles from 2000 through 2003 include the species *Abies concolor* (white fir), *A. lasiocarpa* (subalpine fir), *Picea sp.* (spruce), *Pinus edulis* and *P. monophylla* (pinyon pine), *P. flexilis* (limberpine), *P. ponderosa* (ponderosa pine), and *Pseudotsuga menziesii* (douglas fir). We accessed these mortality data in August 2010.
Figure D-4. Contour plots of monthly suboptimal temperature limits on foliar growth for the subalpine, montane, and woodland biomes in Arizona, Colorado, New Mexico, and Utah from 1895 through 2010. Values are based on a suboptimal temperature index that linearly varies between zero (foliar growth is inactive) to one (foliar growth is unconstrained) [Jolly et al., 2005]. Index values are biome averages based on 3,841, 5,438, and 12,207 grid cells for the subalpine, montane, and woodland biomes, respectively. See Weiss et al. [in review] for methodology and data used to generate index values.
Figure D-5. As in Figure D-4, but for monthly ET demand limits on foliar growth. Values are based on a vapor pressure deficit index that linearly varies between zero (foliar growth is inactive) to one (foliar growth is unconstrained) [Jolly et al., 2005].
subalpine

montane

woodland

ET demand

completely limited
completely unlimited
Figure D-6. As in Figure D-4, but for monthly integrated growing season limits. Values are defined as the product of the monthly foliar-growth-limit indices based on suboptimal temperatures (Figure D-4), ET demand (Figure D-5), and daylength, and range from zero (foliar growth is inactive) to one (foliar growth is unconstrained) [Jolly et al., 2005].
Figure D-7. Conceptual diagram of multiple levels in the vertical direction for individual horizontal grid cells. In this study, we used 16 pressure levels (1000-500 mb) from the Climate Forecast System Reanalysis (CFSR) dataset to calculate 15 lapse rates that we applied to the PRISM digital elevation model (DEM) in order to estimate free-air temperature at land surface altitudes (see Methods). Monthly mean geopotential height, or altitude, (m) and air temperature (°K) are available at each pressure level.
Figure D-8. Monthly 1979-1999 climatologies of PRISM-based near-surface air and CFSR-based free-air temperatures in the SW for January, February, and March. Free-air temperatures are based on lapse rates calculated from a spline interpolation of monthly mean geopotential heights and temperatures. The grid resolution for both sets of data shown here is approximately 4 km by 4 km.
Figure D-9. As in Figure D-8, but for April, May, and June.
Figure D-10. As in Figure D-8, but for July, August, and September.
Figure D-11. As in Figure D-8, but for October, November, and December.
Figure D-12. Paired sample difference of means test results for monthly mean PRISM-based near-surface air and CFSR-based free-air temperatures from January through June. Color gradation quantifies mean differences between monthly mean values of the two datasets from 1979 through 1999 (e.g., the mean of January CFSR-based free-air temperature – January PRISM-based near-surface air temperature). Cross-hatched areas are locally significant at the 95% level. All monthly maps are field significant at the 95% level. Positive (negative) values indicate that the CFSR-based free-air temperatures are warmer (cooler) than PRISM-based near-surface air temperatures.
Figure D-13. As in Figure D-12, but for the months from July through December. All monthly maps are field significant at the 95% level.
Figure D-14. Pearson’s linear correlation coefficients between monthly anomalies of CFSR interpolated free-air temperature data and PRISM near-surface temperature data. Cross-hatched areas indicate correlation coefficients with a $p$-value < 0.05.
Figure D-15. As in Figure D-14, but for the months from July through December.
APPENDIX E: PERMISSIONS
APPENDIX A: IMPLICATIONS OF RECENT SEA LEVEL RISE SCIENCE FOR LOW-ELEVATION AREAS IN COASTAL CITIES OF THE U.S.A., is reproduced with kind permission from Springer Science and Business Media.

APPENDIX B: DISTINGUISHING PRONOUNCED DROUGHTS IN THE SOUTHWESTERN UNITED STATES: SEASONALITY AND EFFECTS OF WARMER TEMPERATURES is reprinted with permission from the American Meteorological Society.