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Contributing factors for drought in United States forest ecosystems under projected future climates and their uncertainty

Charles H. Luce
USDA Forest Service, cluce@fs.fed.us

James M. Vose
USDA Forest Service

Neil Pederson
Harvard Forest, neilpederson@fas.harvard.edu

John Campbell
USDA Forest Service

Connie Millar
USDA Forest Service

See next page for additional authors

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Authors
Charles H. Luce, James M. Vose, Neil Pederson, John Campbell, Connie Millar, Patrick Kormos, and Ross Woods

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Contributing factors for drought in United States forest ecosystems under projected future climates and their uncertainty ☆

Charles H. Luce [a,*], James M. Vose [b], Neil Pederson [c], John Campbell [d], Connie Millar [e], Patrick Kormos [f], Ross Woods [g]

[a] USDA Forest Service, Rocky Mountain Research Station, Boise, ID, United States
[b] USDA Forest Service, Southern Research Station, Raleigh, NC, United States
[c] Harvard Forest, Petersham, MA, United States
[d] USDA Forest Service, Northern Research Station, Durham, NH, United States
[e] USDA Forest Service, Southwestern Research Station, Albany, CA, United States
[f] USDA Forest Service, Pacific Southwest Research Station, Albany, CA, United States
[g] Faculty of Engineering, University of Bristol, Bristol, UK

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ABSTRACT

Observations of increasing global forest die-off related to drought are leading to more questions about potential increases in drought occurrence, severity, and ecological consequence in the future. Dry soils and warm temperatures interact to affect trees during drought; so understanding shifting risks requires some understanding of changes in both temperature and precipitation. Unfortunately, strong precipitation uncertainties in climate models yield substantial uncertainty in projections of drought occurrence. We argue that disambiguation of drought effects into temperature and precipitation-mediated processes can alleviate some of the implied uncertainty. In particular, the disambiguation can clarify geographic diversity in forest sensitivity to multifarious drivers of drought and mortality, making more specific use of geographically diverse climate projections. Such a framework may provide forest managers with an easier heuristic in discerning geographically diverse adaptation options. Warming temperatures in the future mean three things with respect to drought in forests: (1) droughts, typically already unusually hot periods, will become hotter, (2) the drying capacity of the air, measured as the vapor pressure deficit (VPD) will become greater, and (3) a smaller fraction of precipitation will fall as snow. More hot-temperature extremes will be more stressful in a direct way to living tissue, and greater VPD will increase pressure gradients within trees, exacerbating the risk of hydraulic failure. Reduced storage in snowpacks reduces summer water availability in some places. Warmer temperatures do not directly cause drier soils, however. In a hydrologic sense, warmer temperatures do little to cause "drought" as defined by water balances. Instead, much of the future additional longwave energy flux is expected to cause warming rather than evaporating water. Precipitation variations, in contrast, affect water balances and moisture availability directly; so uncertainties in future precipitation generate uncertainty in drought occurrence and severity projections. Although specific projections in annual and seasonal precipitation are uncertain, changes in inter-storm spacing and precipitation type (snow vs. rain) have greater certainty and may have utility in improving spatial projections of drought as perceived by vegetation, a value not currently captured by simple temperature-driven evaporation projections. This review ties different types of future climate shifts to expected consequences for drought and potential influences on physiology, and then explains sources of uncertainty for consideration in future mortality projections. One intention is to provide guidance on partitioning of uncertainty in projections of forest stresses.

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1. Introduction

Observations of global die-off in forests has raised concerns about forest responses to drought and the linkages between drought and climate change (Allen et al., 2010, 2015), leading to

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* Corresponding author.
E-mail address: cluce@fs.fed.us (C.H. Luce).

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questions about adapting forest and rangeland management for drought resilience (Vose et al., 2016a). At present, there is substantial disagreement about whether climate change will increase drought occurrence, frequency, or severity (Dai, 2011; Seneviratne et al., 2012; Sheffield et al., 2012; IPCC, 2013; Roderick et al., 2014, 2015; Trenberth et al., 2014; Cook et al., 2015). Despite this uncertainty, there is agreement that forests will be more affected by drought in a warmer environment whether through stronger metabolic demand, reduced opportunity for carbon fixation, thermal mortality, leaf desiccation, or greater potential for cavitation of the fluid transport system within tree stems and branches (e.g. Adams et al., 2009; McDowell et al., 2011; Choat et al., 2012; Anderegg et al., 2013; Allen et al., 2015; Körner, 2015; Mackay et al., 2015).

Significant drought mortality has already occurred in U.S. forests, with the majority of the drought stress and mortality found in western states (Millar and Stephenson, 2015; Clark et al., in press) and a lesser, though still noteworthy, increase in the southeastern U.S. since the late 1990s (Olano and Palmer, 2003; Starkey et al., 2004; Berdanier and Clark, 2016). As an example of the magnitude of effect, the area of forests burned by large fires in the Forest Service's Monitoring Trends in Burn Severity (MTBS) database between 1984 and 2006 in 9 western states (excluding most of CA and NV) was 5.7 Mha (Dillon et al., 2011), and between 1997 and 2010, Bark beetle mortality was estimated at 5.4 Mha (Meddens et al., 2012). Much larger areas have been affected if non-forest lands are considered, if more recent years are added, or for a full accounting of affected regions.

Although "drought" is frequently treated as a technical term quantified with varying metrics, it is used with very broad meaning in public discourse. Inconsistent and variable definitions can make assertions made about shifting drought and drought effects difficult to either question or defend. Simpler concepts, terms like "dry" and "warm" are a useful way to break down meaning about drought that can be more easily tied to typical climate projections for purposes of describing effects on forests at large spatial scales. For example, in the broadest sense, we can examine "dry" and "warm" relative to changing averages. While increasing warmth has high certainty (IPCC, 2013), future precipitation is uncertain in most places, with only general patterns of moistening and drying associated with hemisphere-scale atmospheric circulation being agreed upon features of future climates (Fig. 1). This moisture uncertainty provides slight feedback in temperature uncertainty; for example, some of the drying locations are expected to experience exacerbated warming due to drying. Because increasing temperature is virtually certain, the range of precipitation predictions generates a breadth of potential vegetation outcomes around likely temperature effects.

Although there is uncertainty in precipitation change at annual time scales, some greater certainty exists for shorter time scales, and an improved approach may be to focus on precipitation variability and extremes. Predictions and observations of increasing precipitation variability (Pagano and Garen, 2005; Luce and Holden, 2009; Seager et al., 2012; Hamlet et al., 2013) suggest a future that may be warmer and ***both wetter and drier***, depending on the time scale of examination. We can interpret moisture trends from the perspective of annual averages, seasonal or monthly values, or even shorter time frames such as the hottest days and driest weeks. While an annual scale trajectory might point toward warmer and wetter in a given location, lengthening dry spells between storms would increase the frequency of forest drought stress (Knapp et al., 2008; Heisler-White et al., 2009; Ross et al., 2012) as could drying during the summer season. Focusing on variability shifts our view toward extremes that may shift independently of averages (see, for example, figures in Jentsch et al. (2007), Field et al. (2012), Anderegg et al. (2013)). Although much of the United States is projected to get wetter in general, particularly in forested regions, some specific atmospheric and hydrologic behaviors will likely contribute to increasing dryness for time scales of days to months. These are not typically the time scales associated with mortality of long-lived species, but increased short-term moisture stress on a more regular basis during the growing season creates an important ecological context affecting growth and mortality (e.g. see examples in Knapp et al. (2008), Heisler-White et al. (2009)) contingent on environmental characteristics.

The objective of this synthesis is to identify the physical and hydrologic characteristics of drought that are most relevant for understanding how drought impacts forests from the scale of individual trees up to the forest ecosystem. We also clarify the terminologies used when discussing changing droughts and changing forests and explain the individual roles of precipitation, evapotranspiration, and snowmelt timing in contributing to drought-related stresses.

2. Characterizing specific mechanisms of drought in the context of forest responses

In mechanistic terms, drought relates to the fraction of full soil recharge after each precipitation event (i.e., how much it rains), the frequency of precipitation events (i.e., how often it rains), energy available for evaporation (usually net radiation) and atmospheric demand (i.e. the vapor pressure deficit, or difference between current atmospheric water content and water content at saturation). This balance between soil water supply and tree water demand determines drought severity from the perspective of the forest. Even though drought mortality may arise through external agents like fire (e.g. Littell et al., 2016) or insects and pathogens (Kolb et al., 2016), these are ultimately mediated through plant physiological responses to drought (Phillips et al., 2016).

A large range of physiological processes are implicated in drought mortality and productivity declines, and though the literature highlights substantial uncertainty about process (McDowell and Sevanto, 2010; Sala et al., 2010; Anderegg et al., 2013; Hartmann, 2015; Körner, 2015; McDowell et al., 2015), there is a convergence on two competing alternatives: hydraulic failure, or the formation of air/vapor blockage in xylem (e.g. Sperry, 2000; Sperry et al., 2002), and carbon starvation when stomata allowing gas exchange (and thereby photosynthesis) are kept closed for extended periods (e.g. McDowell et al., 2008). These alternatives reflect the trade-off between strategies that encourage stomatal closure at the cost of reduced carbon fixation versus those that risk hydraulic failure but maximize carbon fixation (e.g. Ambrose et al., 2015). In what may actually represent end-members on a spectrum of diverse strategies, isohydric and anisohydric behaviors are used by trees to regulate risks versus growth in environments of varying aridity (e.g. Franks et al., 2007; Klein, 2014). In short, trees vary in their physiological responses to drought, and geographically and topographically varying differences in climate changes will interact with these physiological responses in potentially unique ways.

Shifts in climate are expected to change hydrology and consequently the nature of soil moisture drought and evaporative demand. Climatic shifts can broadly be characterized as shifts in temperature, which are relatively certain in their magnitude and direct consequence, and shifts in precipitation, which have large uncertainty in magnitude (Table 1). One response to uncertainty is to set aside potential precipitation variability and analyze temperature induced changes conditioned on mild average changes in precipitation. In Table 1, we offer both wetting and drying seasonal trends in precipitation as context, in part because there is spatial variation in seasonal precipitation projections across the

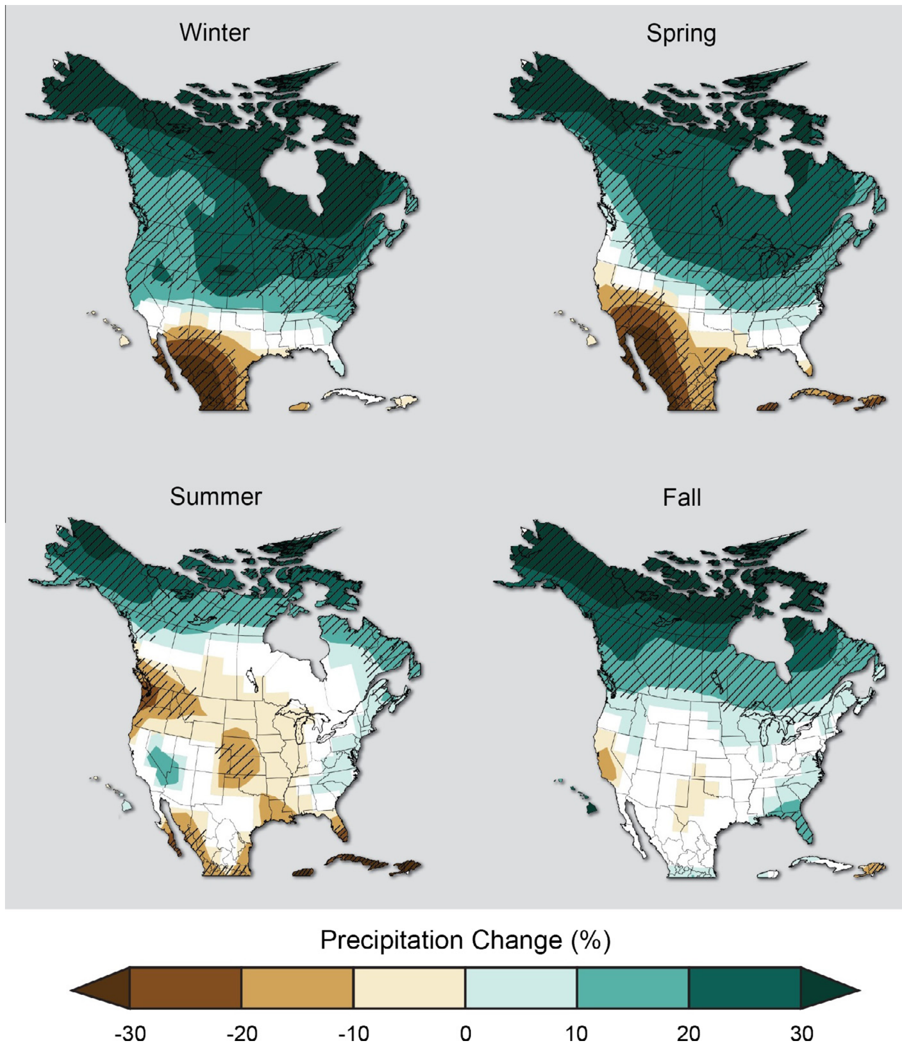

Fig. 1. Seasonal precipitation change for 2071–2099 compared to 1970–1999 under CMIP5 RCP8.5. Hatched areas indicate areas where projected changes are statistically significant and most models agree on the sign of the change. White areas indicate where projected changes are less than might be expected from natural variability. From Fig. 2.15 in Walsh et al. (2014a).

US (Fig. 1), and in part because there is sizable uncertainty in most locations (on the order of ±20%, IPCC (2013)). It is important to contemplate alternative futures, which can yield uncertainty that potentially outweighs temperature induced uncertainty alone (e.g. Wenger et al., 2013).

Many of the pathways described in Table 1, are well represented and described in the literature. Snowpack changes, for instance, are commonly noted for the western U.S. (Barnett et al., 2005; Knowles et al., 2006) where less snow is likely to fall as snow and more as rain (e.g. Pierce et al., 2008; Klos et al., 2014) and snowpacks are melting out earlier (e.g. Cayan et al., 2001; Stewart et al., 2005; Luce et al., 2014) causing forests to endure a longer dry season. Similarly longer dry spells between precipitation events (Fig. 2) (Giorgi et al., 2011) are recognized for altering ecological balances (Knapp et al., 2008; Heisler-White et al., 2009; Ross et al., 2012). Increased runoff consequences of increased precipitation intensity (IPCC, 2013), however, may be overestimated for forests, where high infiltration capacities routinely handle severe events (Hewlett and Hibbert, 1967; Harr, 1977), and the soil drying consequences of more runoff from infiltration capacity mediated runoff may be more appropriate for agricultural systems. Clearly some of the more extreme high intensity events cause some degree of flooding, even from forests now (e.g. Armstrong et al., 2014; Frei et al., 2015), particularly in summer-wet environments where soil saturation mediated runoff is more common. Reduced canopy wetting time may also reduce direct evaporation losses. Warmer temperatures are clearly a risk to forests with respect to direct temperature mortality from yet warmer droughts, but also by increasing respiration, exacerbating carbon starvation effects (Breshears et al., 2005; Adams et al., 2009; Allen et al., 2010, 2015; McDowell, 2011). Effects of warming on plant water demand as mediated by increased vapor pressure deficit (VPD), however require further discussion.

Water vapor diffuses through the stomata at a rate governed by the vapor pressure deficit. This has led to predictions of greater evapotranspiration, and soil drying, with warming temperatures (e.g. Dai, 2013; Cook et al., 2015), but it is important to track context (e.g. water availability and the energy balance) when making such predictions. Noting that the energy balance dictates only mild increases in annual scale evapotranspiration (Roderick et al., 2014, 2015), shorter term rate increases in evapotranspiration would lead to faster draw down of plant and soil reservoirs during a given dry spell, but not necessarily greater annual scale evapotranspiration. In semi-arid forests, the short-term faster response would be balanced in whole by a longer period without water to evaporate, leading to more sustained warmer temperatures, driven (perhaps

Table 1
Expected climate shifts related to hydrologic and biological outcomes in forests.

Climate shift	Drought related outcomes	Relevance to forests	Shift projection confidence
Warming related			
Reduced snowfall fraction	Longer summer dry period in summer-dry (S-D) climates	Increased risk of fire, hydraulic failure (HF), and low non-structural carbon balance (LC)	High
Greater vapor pressure deficit	Increased ET rate when moisture and energy inputs are available Reduced soil moisture in summer-wet (S-W) climates	Increased HF risk Reduced growth and increased LC and HF risk	High
Warmer temperatures	Additional warming over drought induced warming	Increased respiration, heat related mortality, and LC risk	High
Precipitation related			
Longer dry spells between storms	Greater drawdown of soil moisture and plant moisture in S-W climates Little consequence in S-D	Increased HF and LC risk, reduced growth in dry years	Regional; Medium
Increased intensity	Little change in soil moisture in forest soils because of increased throughfall and high infiltration capacity		Low
Annual precipitation			
Wetter	Depends on timing	Uncertain	Regional; Low
Drier	Depends on timing	Uncertain	Regional; Low
Winter precipitation			
Wetter	Could moderate snowfall decline at high elevations	Reduced HF, LC, and fire risks	Regional; Low
Drier	Would exacerbate snowfall impacts	Increased fire, HF, and LC risks	Regional; Low
Spring precipitation			
Wetter	Wetter summer soils for S-D climates	Decreased fire, HF, and LC risks	Regional; Low
Drier	Drier summer soils in S-D climates	Increased fire, HF, and LC risks	Regional; Low
Summer precipitation			
Wetter	Wetter soils and shorter dry spells in S-W climates Wetter soil and more frequent wetting in S-D climates	Reduced fire, HF, and LC risks Reduced fire, HF, and LC risk	Regional; Low
Drier	Increased drought length and severity in S-W climates More infrequent wetting in S-D	Increased LC and HF risks in wet forests Increased LC and fire risks	Regional; Low

ironically) by a lack of evapotranspiration (Yin et al., 2014), and lengthened periods without photosynthesis.

In more mesic environments, the energy balance needs to be considered. Evapotranspiration models based solely on warmer temperatures, which can increase evapotranspiration based on either increased VPD ***or*** increased net radiation (such as the Penman-Monteith equation based potential evaporation), overpredict the evapotranspiration response to warming because the consequences of the evaporative cooling of leaves on the VPD is not modeled. This is a particular issue when the land surface hydrology model used with the original general circulation model (GCM) partitions increased longwave radiation energy one way, but an "aftermarket" hydrology or evapotranspiration model applied to the predictions of the original GCM uses just the temperature or VPD prediction to calculate an increased evapotranspiration rate without consideration of the energy partitioning (Milly, 1992).

A "back of the envelope calculation" offers an impression of what the energy constraint means for predictions of soil drying. Partitioning of net radiation increases associated with increased greenhouse gases will largely (81% of increased radiation) result in warming temperatures versus increased evapotranspiration (19%), most of which is expected over oceans (Roderick et al., 2014). Applying this partitioning for a mesic forest, using additional heating of 34 W/m^2 (=8.5 $\times$ 4, see IPCC AR5 (2013) for explanation) would add capacity for an additional 0.01 mm/h of evapotranspiration without cooling relative to GCM projections, or $\sim$4 mm/mo for 30 12-h days (during photosynthesis); a relatively small amount for most forest water budgets.

Despite large uncertainty, it is important to pay attention to the potential effect of precipitation changes. Variation in annual precipitation is the dominant driver of interannual variations in annual runoff, not potential evapotranspiration (Milly and Dunne, 2002; Vose et al., 2016b), and precipitation has historically been a stronger driver of variability in hydrologic droughts in snow-driven watersheds of the Pacific Northwest (Kormos et al., 2016). The availability of water for streams somewhat parallels the supply available to trees; they both draw from the soil reservoir recharged by precipitation events, as evidenced by a strong correlation between tree growth (reflecting soil water availability) and streamflow (e.g. Cook and Jacoby, 1977; Woodhouse et al., 2010; Lutz et al., 2012). This may help explain why some have noted that droughts severe enough to affect forest ecosystems may be driven as much or more by deficits in precipitation, as by changes in potential evaporation (e.g. Breshears et al., 2005; Holden et al., 2012; Abatzoglou and Kolden, 2013).

3. Projecting changes in climatic conditions and drought

Much of this section draws on data from a number of locations within IPCC AR5 (2013). Other studies are also consulted where noted, to clarify context. This concise review of specific expectations for climate characteristics relevant to drought is meant to be integrated with the hydrological and ecological processes outlined above. It also provides a more informed basis for the ensuing discussion on uncertainty.

3.1. Changes in temperature

Temperatures are expected to increase 4–7 °C by the end of the century across the continental U.S., with stronger increases in the interior than near the coasts. Temperature changes have higher certainty than most climate variables. Summer relative humidity is expected to decrease by 4–8% over land, with weakest declines in the Southwest, where summer humidity is already low. Projected temperature increases in Alaska Range from 4 to 9 °C, with greater increases farther north. Projected temperature increases around Hawaii are in the 3–4 °C range. Both Alaska and Hawaii have nearly no projected change in relative humidity. Projections of warmer temperatures accompanied by constant or decreasing relative humidity predict a greater vapor pressure deficit, however.

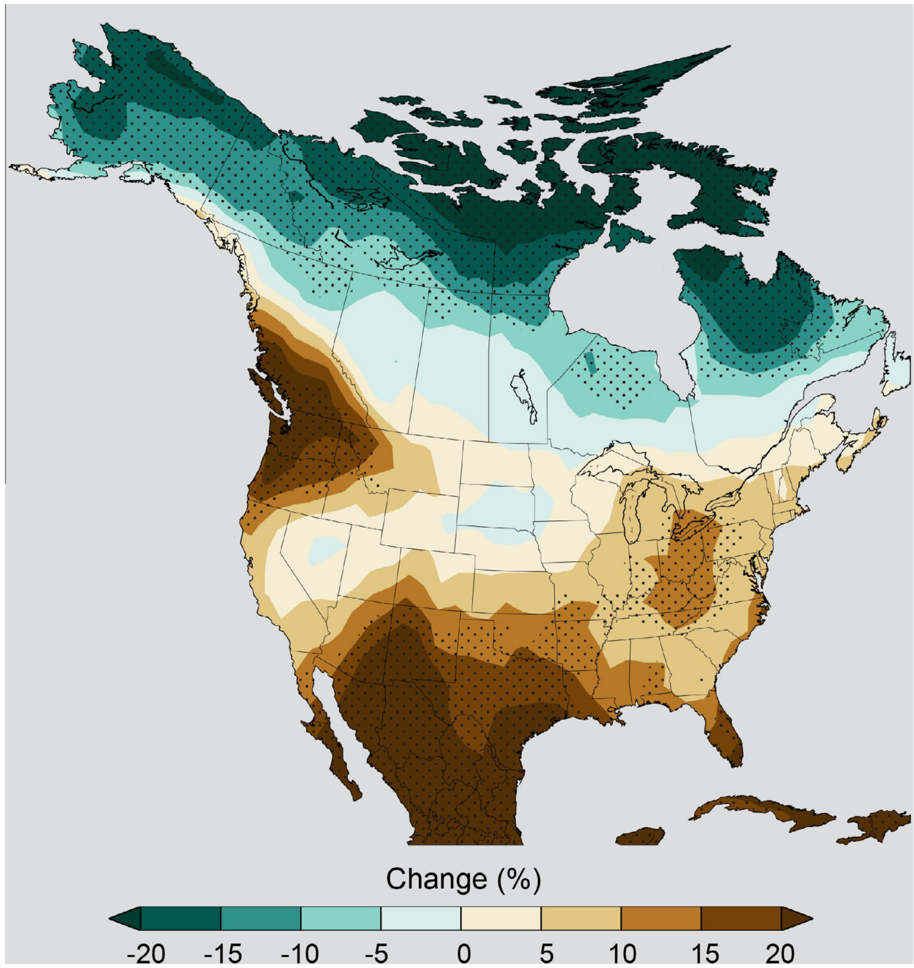

Fig. 2. Change in the number of consecutive dry days (days receiving <0.04 in. (1 mm) of precipitation) at the end of this century (2070–2099) relative to the end of last century (1971–2000) under CMIP5 RCP 8.5, mean of 25 models. Stippling indicates where at least 80% of the models agree on the sign of change. Fig. 33.36 in Walsh et al. (2014b).

3.2. Changes in precipitation

Projections of changes in annual and seasonal precipitation in IPCC (2013) AR5 WG1, show great diversity over the U.S., following the general pattern of dry areas getting drier and wet getting wetter (Held and Soden, 2006) (Fig. 1). Precipitation projections over the continental US are uncertain. Winter precipitation (DJF) increases on the order of 0–10% (with large differences among models) are expected over most of the continental U.S. except for the southwest where declines of 0–10% are projected. Alaska shows increases of 10–50%, increasing with latitude. Hawaii has a minor and uncertain decline. Summer precipitation (JJA) is projected to decline (0–20%) over most of the continental U.S., except for the east and gulf coasts where 0–10% increases are projected. Alaska shows increases of 0–20%, increasing with latitude. Hawaii has a minor and uncertain increase.

Precipitation intensity is also expected to increase, and if total precipitation does not increase with it, there is an expectation of fewer or shorter precipitation events, resulting in longer interstorm periods, increasing the average number of consecutive dry days (e.g. Giorgi et al., 2011). The maximum number of consecutive dry days (precipitation <1 mm) in a year is projected to change in many places in the U.S. with the most notable changes in the southwestern U.S. (AZ, NM, TX), the Pacific Northwest (WA, OR, ID), and the Ozarks and Appalachians (Fig. 2). In the southwestern U.S., increases are expected to occur in summer months in association with changes in the North American Monsoon (IPCC, 2013 AR5 Ch. 14). A change to an increase in the number of dry days during the growing season would have a significant impact in the northeastern U.S. where the number of dry days has decreased significantly since the 1990s (e.g. Bishop and Pederson, 2015), leaving current forest communities more vulnerable to disruptions in growing season moisture supply.

Duration of summer dry spells are expected to increase in western U.S. mountains, where less snowpack accumulation and earlier melt combine to extend the dry summers and their consequences for forests (Barnett et al., 2008). This prediction is tied largely to projected temperature changes, which yield a decreased fraction of precipitation as snow over winter months as well as earlier melt (Knowles et al., 2006; Pierce et al., 2008; Woods, 2009; Luce et al., 2014).

Associated shifts in runoff and soil moisture projections due to changes in precipitation are complex. Generally, patterns of expected runoff change correlate strongly to patterns of expected future precipitation changes, but future soil moisture is expected to decline in most land areas (IPCC, 2013). These differential trends in soil moisture and runoff are explained conceptually by the fact that increased precipitation in future climates is expected through increased precipitation intensity (Wuebbles et al., 2014), meaning that more of the water will become runoff. Again, there is some uncertainty about the validity of this generalization for forest soils.

3.3. Teleconnection mechanisms

Teleconnections from tropical sea surface temperature (SST) patterns are a primary control on drought occurrence in the U.S. (Rajagopalan et al., 2000; Dai, 2011), particularly related to precipitation. Despite substantial inter-model variability in projections of ENSO, it is expected to continue to be a dominant mode of climate variability. The interannual variability driven by ENSO provides some insights into future drought, in so far as it reflects variance in precipitation. Broadly, the sense that wet places get wetter while the dry get drier can also be applied to temporal variations in precipitation as controlled by ENSO (Seager et al., 2012). Because a warmer atmosphere can hold (and release) more water, circulation dynamics leading to greater runoff (P-E) will be enhanced in contrast to those that do not. An increase in interannual variability of P-E of about 10–20% is expected across most of the continental U.S. except the southwestern U.S., where a decline in variance is expected (Seager et al., 2012). Increases in interannual variability on the order of 30–40% are expected in Alaska.

Projections of the key climate phenomena feeding moisture to the continental U.S. in the summer, the North American Monsoon System (NAMS) and North Atlantic tropical cyclones, have uncertain projections (IPCC, 2013 AR5 Ch. 14), though with a tendency toward drier conditions according to the climate models with the strongest historical performance (Sheffield et al., 2013; Maloney et al., 2014). The most consistent projection for NAMS relevant to drought is an increase in the number of consecutive dry days by 15–40% (interquartile range). Multiyear anomalies of NAMS precipitation have been linked to forest mortality (Goeking and Liknes, 2012).

3.4. Changes in wind speed

Although wind speed is acknowledged as a key component of evaporative demand, and trends over the last half-century indicate reduced evaporative demand resulting in part from slower winds globally (e.g. Roderick et al., 2009; McVicar et al., 2012), future expectations of summer surface wind speeds are not well discussed in IPCC (2013). Rather much of the discussion focuses on upper atmosphere wind speeds and jet stream position, which help little in exploring summer evaporative demand over the U.S. GCM projections of surface wind speed for evapotranspiration projections also vary greatly and may be a primary factor in uncertainty of future atmospheric demand (Johnson and Sharma, 2010).

Changes in upper atmosphere wind speeds will likely shift precipitation in mountainous regions relative to expectations expressed by GCMs due to wind interaction with terrain and changes in orographic enhancement of precipitation (Houze, 2012). Actual changes in wind speed perpendicular to mountain fronts will depend on factors such as mountain range shape and existing pressure patterns. In the continental United States, projections reflect stronger warming to the north than the south, reducing meridional (north-south) temperature and pressure gradients, reducing winter westerlies flowing over some mountain barriers and reducing precipitation in those mountains relative to nearby low areas (Luce et al., 2013). Because many western U.S. forests are in mountainous areas, owing to this orographic precipitation enhancement, it will be important to consider regional wind changes in concert with GCM precipitation projections.

4. Uncertainties in using climate projections to predict changes in drought regimes

Most of the runoff, soil moisture, and temperature projections in IPCC AR5 (2013) rely on the land surface hydrology models embedded within the GCMs to calculate the hydrology and energy balances. Because the latent heat of evaporation is a substantial component of the energy balance, it contributes in important ways to predicting future temperature changes. Post processing of GCM outputs has also been applied to examine how changes in precipitation amount, timing, and form (snow vs. rain) interact with energy available for evapotranspiration to estimate fine-scale details of potential drought future conditions (e.g. Sheffield et al., 2004; Wood et al., 2004; Elsner et al., 2010; Vano et al., 2012; Hamlet et al., 2013; Cook et al., 2015). Other approaches directly estimate PDSI values using the projected temperature and precipitation changes (e.g. Dai, 2013).

An important consideration in interpreting output of these post-processing simulations is that they can "double-count" the effects of increased incoming longwave radiation on ET (Milly, 1992), first for heating within the original land surface model embedded in the GCM and later for evapotranspiration in the secondary analysis. This problem is increasingly well known for applications of the original formulation of the PDSI, which uses temperature explicitly through the Thornthwaite (1948) evaporation model, but is generalizable to other potential evapotranspiration schemes where the evapotranspiration formula need not maintain the energy balance of the original GCM Land Surface Model (Milly and Dunne, 2011). The Penman-Monteith potential evaporation formulation (Monteith, 1965), which has been adapted for use in PDSI and as part of more complex water balance models (Sheffield et al., 2012; Cook et al., 2014), implicitly carries a strong temperature dependence in the calculation of the reference evaporation as a function of the vapor pressure deficit term as well, where it can overestimate the consequences of warming on evaporation (Roderick et al., 2015). When atmospheric demand is allowed to substitute for energy in potential evapotranspiration estimation, as it is in some applications of the Penman-Monteith, the modeler must apply care in appropriately tracking the energy balance independently (Milly, 1992). Distinguishing between contributions to, and consequences of, future drought from increased evapotranspiration rates versus precipitation lapses may be helpful in interpreting this kind of work. An awareness that most incoming energy increases are expected to drive warming temperatures rather than ET provides important context (Roderick et al., 2014).

5. Uncertainties in linking future drought regimes to forest responses

In forests, outcomes of drought can take many years to manifest. At the same time, we recognize that individual extreme and extended droughts predispose or contribute to many different mortality pathways for forests, whether through, insects, disease, fire, or starvation (e.g. Breshears et al., 2005; Kolb et al., 2016; Phillips et al., 2016). Decadal variations and century-scale trends expected from climate change may represent increasing drought pressure to existing forest communities, but at individually short time scales (for example, a persistent addition of a few more dry days each summer). A key question is how common sub-annual scale drought pressures will be on forests and the degree to which the forest communities will naturally adjust to such pressure. The question of persistence relates to interannual to interdecadal scale climate variability, and how it will shift with a changing climate (e.g. teleconnections or climate modes such as El Niño Southern Oscillation, the Pacific Decadal-scale Oscillation, or the Atlantic Multi-decadal Oscillation). There is evidence that the inability of GCMs to capture low frequency modes of internal climate variability leads to underestimation of risks of persistent drought (Ault et al., 2014). Multiple years of drought are more stressful to forests than single year droughts, and the relative risks of reduced precipitation for several years to decades is not well represented in GCMs.

A related issue is that trends in means may or may not reflect changes in extremes. Projections specifically of changes in extremes or variance or identification of their trends is much more informative with respect to drought impacts (Seneviratne et al., 2012). Drought is an extreme in moisture availability, and several recent studies show increased variance along with lower annual precipitation in some western U.S. mountains (Pagano and Garen, 2005; Luce and Holden, 2009; Luce et al., 2013). Shifts in extremes may result from shifts in the entire distribution without a change in variability, or they may result from a shift in the variability with no shift in the mean, and a shift in variance or mean could change the probability of exceeding a threshold or proceeding into novel weather (Jentsch et al., 2007; Field et al., 2012; Anderegg et al., 2013). These kinds of changes are important both in the context of a trend acting on existing vegetation, wherein a single crossing into unprecedented weather or drought severity is a potential concern, and in the context of potential future plant communities, which may be shaped more by the extremes in future climate than the means. Extremes, and the events associated with them, will likely be critical determinants of ecological change (Easterling et al., 2000; Parmesan et al., 2000; Dale et al., 2001; Jentsch, 2007; Millar and Stephenson, 2015).

This sense of drought as an extreme is of particular concern with respect to GCM outputs, which are poor at representing interannual variability (e.g. Sperna Weiland et al., 2010). Only a few GCMs accurately recreate the El Niño Southern Oscillation (ENSO) pattern, a driver of interannual scale variability in weather across much of the world (Seager et al., 2012; IPCC, 2013). Most outputs of GCM information are ensemble averages of several realizations from a given model and across models. This allows comparison of climatic averages across models, and maps of these average changes are the common maps of change shown in IPCC reports. Common downscaling procedures draw directly from this kind of information to specify an average difference for a given month or season for each GCM grid cell (Wood et al., 2004). For example, interannual variability in VIC hydrologic projections (e.g. Vano et al., 2012) is a legacy of the historical time series on which the changes in the averages are placed.

GCMs are also more challenged by precipitation estimates than other climate characteristics (e.g. Johnson and Sharma, 2009; Blöschl and Montanari, 2010). GCMs show substantial agreement with metrics like pressure and temperature, but notable discrepancies in precipitation, and the differences among the models are not well understood (IPCC, 2013). Some of the issue is almost certainly that precipitation processes occur at scales much smaller than those of GCM grid cells (e.g. Rasmussen et al., 2011). While GCMs can model general temperature, temperature stratification, and vapor variables that are more or less encouraging of precipitation, they ultimately must rely on sub-grid scale parameterizations to estimate precipitation. That is to say that semi-empirical equations or rules are applied instead of solutions to partial differential equations derived from the basic physics, as is done for temperature and pressure. One consequence of the large grid cell size is also that most GCMs produce what amounts to a persistent drizzle (e.g. Pitman et al., 1990; Gao and Sorooshian, 1994) reflecting the general scale-related issue that it is almost always raining somewhere within a GCM cell, while it is usually only a small proportion of a GCM cell that would actually be experiencing precipitation. In addition, GCMs do not model the control that mountains place on precipitation generation, which has led to efforts to regionally downscale the GCMs to better reflect topographic influences on precipitation in mountainous areas using regional climate models, which represent topography and its effects on climate with finer resolution (Salathé et al., 2010; Rasmussen et al., 2011).

6. Management and research-need implications

Drought forms an important context for forest adaptation to climate change. Climate change adaptation planning is difficult because of complex uncertainties and contingencies (e.g. Millar et al., 2007). The general uncertainties surrounding future physical drought realization and biological response pathways as outlined above form a particularly complex set of problems, and we have identified challenges as priorities for research and management innovation:

- Improve understanding of how species respond to low soil moisture, higher VPD, and higher temperatures as distinct and combined phenomena
- Examine how species in mesic to wet regions respond to extended droughts
- Improve observation and modeling of the climatology and hydrology of mountain regions
- Reduce uncertainty in seasonal precipitation projections from GCMs and RCMs
- Replace temperature based ET projections with combined energy and mass balance based projections
- Design adaptation approaches robust to uncertainties

Although some aspects of future drought, particularly the context of annual precipitation, seem uncertain at this time, there are insights that can be drawn from even general tools such as those discussed by the IPCC (2013). Overall, warming temperatures will stress trees by more rapidly depleting food for respiration, where the strongest effects might be seen in drought adapted species that respond through more conservative stomatal closure. Greater spacing between precipitation events and greater VPD between events, however, may challenge species that favor potential for growth over protection against cavitation. While some recovery from embolisms is possible, there is a carbon cost (e.g. Mackay et al., 2015), trees with more conservative stomatal opening thresholds seem to be more resilient (e.g. Ambrose et al., 2015), and species with more conservative growing season wood anatomy are less sensitive to drought variance (Elliott et al., 2015). There are tradeoffs in attempting to choose for resilience to metabolic challenges versus water tension related challenges; so it is probably important to examine the large range of regional projections of growing season precipitation (Fig. 1) and interstorm spacing (Fig. 2) in contrast to those for atmospheric drying as a basis to hedge options by considering plants with a range of appropriate response (e.g. biodiversity in adaptation). General patterns of wetting and drying are expected to reinforce current moisture regimes, with the dry getting drier and the wet getting wetter, so we may expect to see appropriately adapted species within the mix of available local biodiversity. The combination of future projections and observations of historical and contemporary responses to drought provide a starting point for decisions about suitable species, genotypes, and management practices to increase resilience to future drought and warming.

Because the western U.S. has seen severe extended droughts, there has been opportunity to watch these forests during "natural experiments." While the eastern U.S. has been more fortunate in not seeing such extreme droughts, there is greater uncertainty in how they might respond in the event that the current pluvial period is interrupted with droughts like those seen in the paleoclimatic record (Pederson et al., 2013; Stahle et al., 2013). Although future projections for the eastern U.S. show slightly wetter conditions, the potential for severe extended drought remains, and moistening conditions may only serve to increase biomass and consequent vulnerability to such an event.

With important exceptions noted above, the history of forest-drought relationship research has relied heavily on application of summary indices of "drought" incorporating combinations of precipitation and temperature variations to indicate intensity and duration of "drought." Ironically, although these indices impose a precise mathematical definition to drought in each individual case, there are so many applied in so many different contexts that the term "drought" almost loses meaning. Given that these indices draw on relationships between temperature and moisture that are more correlative than causative (implying that they may not be true in the future due to rising CO_2 concentrations), there may be benefit in extracting individual components of drought, temperature and moisture states and their duration and frequency because these indices may mask how separate climatic variables influence trees, species, and forests. Summarizing precipitation and drying influences into a single index implies a potentially unrealistic equifinality in drought response regardless of the pathway. While temperature, vapor pressure deficit, and related drying processes might show stronger influence on forest response in some places, precipitation variability may be more important in others (Martin-Benito and Pederson, 2015), and capturing that geographic diversity in sensitivity is important to leverage against estimated geographic diversity in expected climatic changes. Better projections of future forest drought and better prescriptions for adaptation in a changing climate, will require a combination of improved projections of moisture and energy fluxes as well as improved understanding of specific mechanisms of tree response to particular moisture and energy state forcings. This forms the basis for a strategy for developing stronger inferences about changing forests under changing climates.

Some of the exceptions noted above already frame this strategy with respect to ecological outcomes, but the physical hydrology and climatology lag in terms of clarifying both what has happened historically and future expectations. In particular, sparse observations of climatology in rough terrain and mountains (Holden et al., 2011; Dettinger, 2014; Henn et al., 2015) pose a relevant and difficult context for interpreting past forest changes, making future projection more uncertain. Emphasizing improvements to climate and hydrologic understanding that target drivers of physiological response to forests would likely benefit adaptation efforts.

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