CHAPTER 2

Characterizing Drought for Forested Landscapes and Streams

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Introduction

The changing nature of drought is a growing global concern (Cook and others 2015, Dai 2011, Seneviratne and others 2010, Sheffield and Wood 2008b, Trenberth and others 2014, Wilhite and others 2014). Drought can be a severe natural disaster with substantial social and economic consequences affecting large areas with extended durations (Wilhite and Buchanan-Smith 2005). Although it is clear that shifts in circulation patterns, energy for evapotranspiration, and air temperatures are changing in ways that enhance the consequences of drought, there is only weak consensus about the effects of climate change on drought occurrence (IPCC 2013, Seneviratne and others 2012, Trenberth and others 2014). Some of that uncertainty stems from the complex nature of quantitatively defining drought, but also because some of the changes in drought characteristics are only partially reflected in traditional drought metrics [Palmer Drought Severity Index (PDSI) and the Standardized Precipitation Index (SPI)]. Furthermore, although these traditional metrics have adequately reflected the consequences of meteorologically derived moisture deficits on agricultural commodities and water supply, there is a poorer (although improving) understanding of how drought interacts with forests and rangelands and their associated aquatic habitats. Understanding the potential impacts of future drought on forests and rangelands requires knowledge of how droughts impact forest, shrub, and rangeland structure (covered in other chapters in this assessment) and how drought projections are characterized in the General Circulation Model (GCM) output.

The purpose of this chapter is to explore drought as a hydrometeorological phenomenon and reflect broadly on the characteristics of drought that influence forests, rangelands, and streams. It is a synthesis of understanding about drought processes, hydrology, paleoclimatology, and historical climate variability, and how this understanding can help predict potential future droughts and their consequences to forests and rangelands. It describes alternative approaches for characterizing drought and highlights additional work that could inform projection and adaptation for future droughts.

What Is Drought?

Drought is a lack of water. This simple wording implies potentially complex stories. The most important implication is that drought is characteristic of time, not of place. A place can be dry or wet, but droughts occur in given locations over time. For example, the Sonoran Desert may have a lack of water for many purposes, but conceptually it has just the right amount of water for the Sonoran Desert, and it occasionally experiences droughts. Wet places, like the coastal rain forests in Washington State, also experience droughts when there is an unusually dry summer. Consistently dry seasons would not merit the designation of "drought" in and of themselves; however, variations in how dry or long the dry season is from year to year are relevant. In places like the Western United States, where there is a coincidence of the dry season with the growing season every year, it does not make sense to frame summer as a "drought" so much as to note the seasonal aridity of the location (Seneviratne and others 2012, Wilhite and Buchanan-Smith 2005).

Although drought is generally defined from a climatic perspective (precipitation levels that are much lower than the annual average), short-term moisture fluctuations also provide important ecological context. A wet period leading to a vegetation flush, followed by a long dry spell, for instance, may not constitute a "drought" relative to long-term averages, but the inability to meet the temporarily increased water demand would nonetheless represent a meaningful drought. From the perspective of a plant, it may look like a temporary enhancement of seasonal aridity contrasts. An example of this is seen in fire risks in rangeland ecosystems where wet springs can produce an overabundance of fine fuels from invasive annual grasses (Littell and others 2009, Swetnam and Betancourt 1998, Westerling and others 2003). Similarly, such fluctuations have been tied to forest mortality, with timing having an important influence on outcomes (Anderegg and others 2013).

This descriptive definition also speaks to a purpose for water. Lack of water is most relevant when water requirements for sustaining terrestrial and aquatic ecosystems or providing for human uses cannot be met. In natural ecosystems, those requirements broadly relate to climatological expectations (in a statistical sense) for water supply. Similarly, in rational agricultural management, climatological expectations should still provide a basic norm for distinguishing drought conditions, although sometimes drought is claimed when slightly dry conditions endanger marginally suited crops. Generally, drought is perceived as a concern during the growing season or warm season, as these are when there is a demand for water. However, droughts can have substantial ecological consequence related to lack of snow cover, particularly for wildlife (McKelvey and others 2011) or plants that are protected from cold extremes by a layer of snow.

Drought Classification

Most of the indices applied to classify drought are tied to common meteorological measurements, like precipitation and temperature, because these are widely available with relatively long records. The long records make it possible to compare particular precipitation amounts or water-balance estimates to the local climatological distribution, in essence gauging risk or probability levels of current meteorological events. The meteorologically based indices are commonly applied for drought forecasting based on weather forecasts, given that weather is the most fundamental driver and also the most fundamental uncertainty in drought estimation. Despite common application, there is some concern about the utility of drought indices in understanding conditions related to changes in land cover, land use, or climate (Alley 1984, Sheffield and others 2012).

The context of utility or need for water frames much of the traditional way of defining and describing drought. Drought is most commonly thought of in terms of the harm it can induce. It is broadly seen as one of the most potentially severe natural disasters in terms of either human lives or capital loss because it covers large areas and can lead to famine for large populations (Wilhite and Buchanan-Smith 2005). Drought has generally been framed in four classes or "types" (Wilhite and Glantz 1985): (1) meteorological drought, (2) hydrologic drought, (3) agricultural drought, and (4) socioeconomic drought (table 2.1). These four classes are not mutually independent, but refer to different ways to measure, identify, or conceptualize drought conditions. Almost all types of drought relate to meteorological conditions leading to a lack of water, such as a lack of precipitation or excess demand from evapotranspiration. Hydrologic, agricultural, and socioeconomic drought are filters placed on meteorological drought to frame how they affect human demands and values for water, ranging from food production to electrical power production, recreation, and wildlife management. Generally the context is economic, although environmental benefits of water are recognized as well. They also frame different time scales of response to meteorological forcing. Most of these drought types frame drought as an event. In this framing, a drought is a type of disturbance, or even disaster, with a distinct occurrence. This framing is particularly relevant to crops of limited lifespan, seasonal cycles of water

Table 2.1—The four drought classes

Drought class	Primary focus	Estimation
Meteorological	Dry weather	Indices built from weather station or weather forecast information targeted toward soil moisture and agricultural or hydrologic drought. Weather changes rapidly, but indices aggregate weather over time.
Hydrologic	Streamflow and reservoir levels	Measured or modeled runoff and reservoir levels. Some meteorological drought indices target this particular outcome.
Agricultural	Productivity and survival of crops	Measured or modeled crop yields. Some meteorological drought indices target this particular outcome.
Socioeconomic	Economic outcomes of drought	Measured and modeled financial consequences of water supply or agricultural production deficits.

demand (municipal supply, hydropower demand, water-oriented recreation), and other activities where short-term interruptions in water supply are economically relevant (e.g., manufacturing).

Meteorological Drought

Meteorological drought definitions refer to atmospheric components of the water balance, precipitation and evapotranspiration, where low precipitation and high evapotranspirative demand lead to a relative lack of water. In practice, estimates of these components or proxies for their estimation (e.g., temperature) are generally used. Despite the name, meteorological drought metrics use modeling or simple logic to reframe short-term meteorological measurements into indices of impacts to terrestrial and aquatic systems. Meteorological drought can be framed strictly in terms of the amount of precipitation over a time period, as with the SPI (McKee and others 1993), or processed with temperature information to estimate a soil water balance, as in the PDSI (Palmer 1965). The Keetch-Byram Index model (Keetch and Byram 1968) has a similar conceptualization (though with feedback between moisture level and loss rate) directed towards vegetation and fuel moisture levels. Details of these indices are discussed later in this chapter.

Hydrologic Drought

Hydrologic drought is focused more directly on water available as streamflow or in surface or near-surface storage, such as reservoirs or shallow aquifers. While hydrologic drought follows from meteorological drought, its measurement or estimation focuses primarily on the various uses of streamflow or reservoirs, such as aquatic habitat, irrigation, hydroelectric power generation, or recreation. There is a temporal lag between unusually dry meteorological conditions and actual stream and lake levels, and factors other than those indexing meteorological drought may influence actual outcomes for streams, shallow groundwater, lakes, or reservoirs. For instance, precipitation intensity or its distribution in time can markedly influence its potential for generation of runoff in both the short and long term. Multiple brief or small rainfall events may increase the amount of precipitation received, but much of that water would likely be returned to the atmosphere through subsequent evaporation after being intercepted in trees or only shallowly penetrating the soil. On the other hand, focusing an equivalent amount of water into a single, brief storm with high intensity may result in increased runoff but little recovery in soil or vegetation moisture levels.

For some purposes, a lack of streamflow represents a drought more than a lack of precipitation. A lack of water can yield profound impacts for stream fishes, as it affects volume of available fish habitat, stream temperature (Isaak and others 2012), and food supply to fishes (Harvey and others 2006). In western U.S. mountains, runoff from accumulated snowpacks feeds reservoirs and irrigation to downstream farms. Although these farms may get occasional rain in the summer months, that precipitation amount may never really be sufficient to support crops. In such a case, the local precipitation or water balance is not as relevant as the potential hydrologic drought of the stream providing irrigation water.

Simple meteorological indices of drought do not consistently predict the streamflow consequences of a drought. While there is a connection between a lack of precipitation and a lack of streamflow, there are subtleties of shift in precipitation characteristics, or even of changes in vegetation or soil characteristics that can change the runoff efficiency of precipitation events that occur within the context of a prolonged drought (Guardiola-Claramonte and others 2011, Potter and Chiew 2011). There can also be mismatches between return intervals of precipitation amounts (or index values) and streamflow volumes (Potter and Chiew 2011), reflective of the complexity of runoff generation processes.

Agricultural Drought

Agricultural drought, also termed soil moisture drought by Seneviratne and others (2012), is tied to productivity and mortality of crops. It is functionally related to the soil moisture reservoir. Once precipitation has stopped, soils dry through evaporation at the surface and via evapotranspiration of water through the crop. The contribution of direct evaporation at the surface to total evapotranspiration declines as the soil dries. As the soil moisture content drops, roots cannot uptake water as rapidly, and plant productivity falls off as stomata remain closed for a greater fraction of the time. Eventually, the water in the soil is bound so strongly by capillary and osmotic potentials that plant roots cannot extract more water. Crops can die before this level is reached simply if productivity falls below the needs for maintenance respiration. Metrics of agricultural drought (PDSI) were developed for annual crops or vegetation managed as an annual crop. As such, they are best suited for shallow-rooted plants and may overestimate drought experienced by deep-rooted perennial forest and shrub species.

Socioeconomic Drought

Socioeconomic drought follows from both agricultural and hydrologic drought. Agricultural losses from drought represent a fairly direct economic impact, but this also cascades into lack of materials for agricultural manufacturing and support industries and services. Similarly, a lack of water for municipal supply, manufacturing, irrigation, hydroelectric production, or recreation can reverberate through the economy. These economic impacts may substantially lag behind the meteorological drought event that triggers socioeconomic consequences. The relationship between drought and its economic consequences also varies with associated historical decisions (chapter 11). This complex interplay of drought, environmental consequences, and impacts to humans, followed by human-mediated responses to drought (e.g., technological advances and altered land values), leads to broadening economic ramifications over time and space.

Drought Influences on Forest and Stream Ecosystems

For forest and stream ecosystems, meteorological and hydrologic drought frameworks are useful for characterizing impacts of a given drought event. For example, meteorological or hydrologic drought may presage or correlate to fire events (chapter 7) or insect outbreaks (chapter 6). Most of the work in later chapters and information on drought changes in this chapter relate to the common metrics above. However, it is worthwhile to also reflect on ways that drought influences forest ecosystems outside of the traditional harm-oriented framework. It serves as context for drought metrics, trends, and projections described later.

Forests offer a unique challenge to the traditional framing of drought because they persist over long time scales. Forests have adapted to droughts both through their resilience to drought effects and resistance to drought occurrence. While fires, insect outbreaks, and other forest mortality events are tied to drought, it is generally only the most severe droughts that produce large-scale or landscape-scale changes. Most droughts reduce productivity or carbon fixation in trees rather than kill them. As the climate changes, drought may become a stronger driver for changes in vegetation species composition and life form (shrub, woodland, forest) (Dale and others 2001, Jentsch and others 2007, Luce and others 2012). Reframing drought as a driver that determines relative fitness among species adds new context for characterizing drought. To address this, we can draw on ecological theory and our understanding of the hydrometeorological process to assess how long-term changes in drought climatology (frequency, severity, or spatiotemporal scaling) might drive large-scale changes in vegetation.

In addition to drought impacts on terrestrial vegetation, there are impacts to forest aquatic ecosystems. Forests are generally the largest source for high-quality water supplies, and they provide extensive habitat for coldwater fishes (Rieman and others 2003). Drought-related changes in forests cause changes in runoff generation, with the potential for immediate negative impacts on stream ecology. However, drought is also a fundamental driver of instream ecological processes, and some highly prized aquatic species are present because they are better adapted to drought-related effects to streams and surrounding forests than are other species. Changes in drought characteristics may consequently be a driver of long-term and large-scale changes to instream ecological processes. Fish species are substantially less long-lived than trees, but presence of a given population in a stream is a comparably persistent aspect of the problem. The interaction of drought with local population and metapopulation dynamics is a more important consideration than losses from a single generation. For example, increased drought-related disturbance may promote migratory life histories over resident life histories (Dunham and others 2003). Another example is the range of drought adaptations of some fishes in the Great Plains (Falke and others 2011, Fausch and others 2002). Local extirpations of some fishes could result if small streams become both more unpredictable in low flows and less productive (reducing the potential for migratory life histories).

Ecological Drought Characterization

The impacts of dry conditions on ecological processes are about tradeoffs. Some biota benefit from drought primarily because other biota are more negatively impacted. This view of drought ecology—where drought stress is viewed as a driver of ecosystem processes—is part of what makes application of traditional drought metrics difficult. Although indices with single values are easier to present and apply in quantitative analysis, ecological drought characterization is more complex and multidimensional. Some of the key dimensions are:

- **Severity**–defined as degree of moisture deficit
- **Frequency**–level of deficit, alternatively probability
- **Temporal patchiness**–autocorrelation in time that incorporates duration and short-term variability
- **Spatial coherence**–or spatial distribution across the landscape
- **Correlation with other factors** (e.g., season or temperature)

Some current metrics conflate one or more of the aforementioned dimensions. Calculation of PDSI, for example, incorporates lack of precipitation, air temperature, and duration of an event. SPI is a simple metric of severity, but the complete range of options for SPI values with varying temporal footprints reveals that it also can offer temporal autocorrelation information. It is likely that drought cannot be measured or characterized along any one of these dimensions absent some others; however, exploring the different dimensions can be informative for explaining component contributions in part for ecological drought.

The most natural and direct measures of drought severity for forest and stream ecosystems are probably reflected in measures such as soil moisture, streamflow, or fuel moisture. These measures have a particular value at a given time, but they reflect spatial and temporal integration of precipitation and evapotranspiration. Consequently, there is some need to explicitly recognize the role of time if only meteorological information is used. A day without precipitation occurs often, but multiple days without precipitation cause drier soils and reduced streamflow. Nonetheless, we can see that with an appropriately informative footprint (discussed briefly below), variations in the amount of precipitation received or precipitation less evapotranspiration are informative quantities.

Frequency of a given level of severity is critical in understanding the ecological role of drought because the relationship of the frequency of mortality-inducing drought to generation or recovery times is a fundamental descriptor of ecosystem dynamics. At one end of the spectrum, species that mature more slowly than the frequency at which mortality-inducing weather events occur are not well fit to the local climate. At the other end, species that take advantage of frequent disturbance may not compete well with species that invest toward longer term gains when disturbance is infrequent. In most cases, disturbances will occur within a species lifetime, and there are different strategies and adaptations related to different frequencies. Relationships between frequency and severity are commonly embedded within those adaptations. It is unusual to have severe fire or insect outbreaks on a frequent basis, simply because it is hard to regrow adequate fuel or food to sustain the next severe event. Although fire, pathogen, and insect outbreaks are general examples in this area, trees adapted to low-productivity arid sites, such as bristlecone pine (*Pinus aristata*), offer another example where the isolation offered by frequent severe drought builds a degree of resilience to other mechanisms of mortality spread by abundance of neighbors.

The concept of *temporal patchiness* is a metric to describe how slowly moisture states vary. Broadly, high year-to-year autocorrelation in moisture reflects long-term dry states, whereas low autocorrelation reflects high variability in moisture states over short timeframes. Both ends of this spectrum can have substantial impacts on ecological processes and disturbance regimes. For example, high contrasts in moisture over relatively short (seasonal to interannual time scales) can increase the severity of drought-related stressors such as fire or insects. Such is the case in shrublands with wet winters and springs that promote heavy growth of annual grasses (e.g., *Bromus* spp.) and lead to more severe and larger fires during the following dry summer (Abatzoglou and Kolden 2011, Littell and others 2009). Substantial growth in a forest in a wet year contrasted to dry conditions the next year can lead to increased moisture stress because of increased leaf area. While duration is well captured by common drought indices, the high contrast risk is not, and the two are related in a given climate and soil. While the former is important to critical water supply levels and agricultural crops, the latter has greater context in less regulated systems.

Meteorological drought has substantial *spatial coherence*, with a footprint of much greater scale than is typical for landscape ecology. Severe meteorological droughts, sometimes termed "megadroughts" may encompass multiple regions of the United States at a given time (e.g., Coats and others 2014, Cook and others 2014a). While this scale has potential importance for distribution of fire suppression equipment or response to insect outbreaks, more-local scales are also relevant because a particular meteorological drought may play out differently within a landscape. North-facing slopes retain moist conditions longer into the growing season than south-facing slopes, which is reflected in productivity and plant species. One consequence is that fire severity can vary substantially across topographic position, such as aspect or riparian proximity (Dillon and others 2011, Dwire and Kauffman 2003). The outcome for vegetation communities is that the spatial scales of drought-related mortality are related to the grain of topographic variation, or hillslope facets, in historical examinations (Hessburg and Agee 2003). The severity of the meteorological drought can, itself, have some influence on this patchiness, and topographic relationships to fire severity are damped in the driest years (Dillon and others 2011).

Correlation of drought to other meteorological characteristics is an increasingly important area of study. If we look through the list of key dimensions, there is a sense of increasing levels of information or organization (sensu Blöschl 1996) of outcomes for drought events. At the basic level, there is the question of how dry it is (*severity*), followed by how dry an event is relative to other events (*probability/frequency*), followed by how quickly or slowly the moisture levels transition (*temporal patchiness*), and how different places within a landscape experience a given meteorological drought (*spatial coherence*). The next level of organization is how drought might relate to other factors of importance in ecology. In the following section, we discuss the linkage between temperature and drought because it has substantial relevance to the broad discussion on climate change and drought (Dai 2011, Sheffield and others 2012, Trenberth and others 2014), particularly in the context of forest ecology (Adams and others 2009, Allen and others 2010, Breshears and others 2005).

The causal relationship between temperature and drought arises from interactions mediated by solar heating. Clearly, warmer air has lower relative humidity (given a fixed specific water content in the air), and warmer air has the capacity to impart more sensible heat to the energy balance. However, the great majority of the heat in the energy balance is incoming solar radiation. Some solar heating goes to evaporating water and some goes into heating soil, vegetation, and air. As a consequence, air temperature and evapotranspiration are correlated simply because they share the same driver (i.e., solar radiation), not because air temperature substantially adds to the energy available for evapotranspiration. This makes temperature a simple index for framing both the energy available for evapotranspiration and the capacity of air to hold additional moisture. For example, temperature is a variable in the Thornthwaite equation (Thornthwaite 1948), which is used in calculation of the PDSI. When there is less water available (due to a lack of antecedent precipitation), more of the incoming solar energy goes into raising temperature; however, when conditions are dry, the increase in temperature is at the expense of evapotranspiration. Energy goes into warming, not evaporating. Under dry conditions, temperature becomes an index of the lack of water available for evapotranspiration. Thus, within historic contexts, temperature can be used as an index of dryness: dry is warm, and warm is dry.

Under a changing climate, in contrast, temperature will rise independently of aridity. Warmer temperatures will result from increased atmospheric emissivity and a changed energy balance of the air. Net radiation will increase, slightly increasing the energy available for evaporation, but the relationship between relative humidity (or vapor pressure deficit) and temperature for warming on that slow of a scale will not be the same as it is in current empirical contexts. In short, the correlation between air temperature and drought will still exist, but its value will not be in indexing drought conditions in relation to current conditions.

Instead, warmer temperatures paired to drought will represent a greater challenge for plants to cope with drought (Adams and others 2009). This context of drought being warmer has been termed "global-change-type drought" (Breshears and others 2005) to reflect the idea that whether or not "drought" is worse under a changing climate, the consequences of a given level of drought could be worse in combination with warmer temperatures. Because warmer temperatures elevate metabolism and respiration, a higher productivity will be required to match the demand. As plants shut down during moisture stress, they will exhaust carbon stores more quickly, and survival times between wetting events will shorten. The physiological mechanisms underlying drought mortality are discussed in chapter 3.

The correlation of drought to the season in which it occurs is another important characteristic that may also shift with warming. For example, if the period of common drought corresponds to the warm season (growing season), drought effects will be more severe, in a broad sense. Expansion of the growing season as temperatures warm could draw soil water down sooner as well, yielding deeper dryness in the soil earlier in the growing season; this would effectively increase the duration of the summer drought and have potential consequences for wildfire (e.g., Westerling and others 2006). Where precipitation is more abundant during the growing season, the susceptibility to drought is related to interruptions in that precipitation supply.

Measures of Drought

The drought types mentioned earlier offer some distinction for classifying drought prediction tools. Some measures target characterizing drought for broad purpose based primarily on weather measurements, while others target understanding hydrologic drought. Given the general interest in forecasting drought, both approaches are based on meteorological measurements and meteorological model outputs. Meteorological drought indices are either directly tied to the measurements or placed in a context for convoluting precipitation and temperature with time to give a mixed drought and exposure index. Hydrologic drought is necessarily more local in nature as it depends on characteristics of the basin of concern. Some reporting tools, such as the U.S. Drought Monitor, use a blend of hydrologic and meteorological drought metrics along with expert guidance.

Fundamentally, every kind of drought is meteorological in nature, but outcomes are shaped through local topographic, geologic, and biotic filters. In so far as models correctly reflect the hydrologic processes of interest, weather forecasts or stochastic simulations contingent on seasonal climatological forecasts can provide useful information on potential outcomes. We briefly discuss the most common ones here (see Hayes and others 2011 for a more thorough listing).

Meteorological Drought Metrics

Three of the most common metrics in this class are the Standardized Precipitation Index (SPI) (McKee and others 1993), the Palmer Drought Severity Index (PDSI) (Palmer 1965), and the Keetch-Byram Drought Index (Keetch and Byram 1968). Although these metrics theoretically address potential outcomes of drought (such as soil moisture, streamflow, or crop productivity and mortality), they are approximate models.

The SPI is a relatively simple approach to characterizing a precipitation anomaly (McKee and others 1993). Precipitation totals are calculated for a given window of specific dates over a series of years. A cumulative density function (CDF) of the values is computed and the quantiles are mapped to a normal CDF, allowing the probability of exceedance above and below the mean to be translated into approximate standard deviations of departure. The index is only applicable locally, and it is relative to the mean precipitation over the period of interest.

The PDSI and related derivatives are conceptually related to the water balance of a relatively thin, two-layer soil incorporating estimates of evaporation and runoff. The precipitation component is relatively straightforward, and runoff is computed based on the water-holding capacity of the soil column (porosity at field capacity). Evapotranspiration (ET) is drawn from the two layers of soil independently, with the thin top layer being available for direct evaporation, while the second soil layer retains water for transpiration. ET is calculated based on air temperature using the Thornthwaite method (Thornthwaite 1948). Because drier soils evapotranspire less than moist soils, there is a seasonal adjustment to the Thornthwaite estimate for well-watered soils. This seasonal adjustment does not respond to actual evolution of conditions, however, and close inspection reveals the adjustments to be relatively minor. A limit is ultimately placed on ET to prevent violation of the mass balance. To calculate the index, a convolution of the soil moisture estimated using this method over time provides a metric that combines both dryness and the length of time plants are exposed to the dryness. These calculations are then treated similarly to the SPI to derive a local dryness index.

The Keetch-Byram Index (Keetch and Byram 1968) is absolute in nature and not locally indexed. It is based on an exponential decay conceptualization of soil moisture in fuels. Precipitation less interception (constant) rewets fuels, and loss each day is calculated (in tables) as a function of the daily air temperature and the previous day's drought index. As the index approaches its driest extreme, the effect of further drying is diminished asymptotically.

Hydrologic Drought Metrics

Hydrologic drought metrics are generally based upon measurements of streamflow. Hydrologic drought may be defined as uncharacteristically low streamflows; however, values depend heavily on the averaging period for the quantification of flows (Vidal and others 2010). For example, annual-scale hydrologic drought is quantified by comparing total annual flow values among a series of years and finding quantiles of the distribution. Because there is a pronounced annual cycle, this averaging scale can be useful for a range of purposes.

For consequences that manifest at shorter time scales, shorter averaging periods are necessary and can be characterized by one of many methods of quantifying how low flows are in a particular low-flow period. In places with noted seasonality in runoff, such as those associated with seasonal precipitation or snowmelt-dominated runoff, season-scale averages can suffice. In places where precipitation and runoff do not show pronounced seasonality, a shorter time-scale average has greater utility.

Differences in seasonality related to presence and absence of snow and temperature-precipitation correlation form the basis for a comprehensive classification of hydrologic drought types (Van Loon and Van Lanen 2012). One example dichotomy is flows that are low because the warm season has low precipitation versus flows that are low because the precipitation is frozen in winter. Ecological considerations of how aquatic species respond to low flows in warm-dry versus frozen seasons might require examining these two types of hydrologic drought separately.

Common formal quantification of low flows occurs in two metrics: the 7Q10 metric and the return intervals of given percentile on the flow duration curve. The 7Q10 is the minimum weekly flow (drawn from the period of interest, such as winter or summer) in each year with a return period of 10 years. For example, with a record length of 20 years, the second to the lowest of the observed annual weekly minimum flow would be the 7Q10. Flow duration curves define the quantiles of flow in each year, and the 20th percentile daily flow would be that flow that is exceeded in 80 percent of the year (i.e. 292 days have higher flows, and 72 days have lower flows in that year). That value can be sampled from each year, and comparison can be made by way of quantiles of the distribution again (the 5- or 10-year return period, for example).

Physically Based Hydrologic Models as Metrics

Advances in computing have resulted in development of more computationally intensive, physically based hydrological models, which, like their computationally simpler predecessors, estimate soil moisture content. Although there is greater expectation or promise for precision and transferability, some of the promise of such models has yet to be realized (Blöschl and others 2013). Despite the availability of such models and example applications (e.g., Sheffield and Wood 2007), much work is still focused on use of common (less computationally intensive) drought metrics. Although there is a range of reasons for such choices, a key point is that the basic inputs of precipitation and radiant energy inputs (or proxies) shared in common across various models and metrics means that the outputs are relatively well correlated (Williams and others 2014), making the indices a more practical and efficient approach.

This capacity to use indices in a correlative manner, however, may limit the advancement of causative modeling. Although short-term (within a year) drought forecasting is conceptually intended to look at potential future outcomes (e.g., river levels, soil moisture levels, crop mortality, or fire danger), the development and validation of the historical/conceptual drought forecasting models have not focused on prediction of these values, per se; rather, they have focused on correlation between the indices and local outcomes. This leaves us with the strongest predictive capacity for droughts that have been observed in the past, without necessarily providing the best information for predicting drought consequences under a changed climate. Identifying and revising those parameterizations within physically based models that intrinsically rely on empirical relationships could help advance understanding of drought risks in a changing climate.

Because of their more comprehensive set of calculations, more detailed water-balance models blur the line between meteorological drought forecasting and hydrologic drought forecasting. These models offer a more flexible approach in estimating persistence of moisture availability; therefore, they may eventually become more useful than current indices for characterizing forest-related drought.

Relationships Between Meteorological Drought and Hydrologic Drought

Although both are referred to as "drought," the relationship between meteorological drought and hydrologic drought is not linear. Complex interactions among changes in precipitation, evapotranspiration, and streamflow (particularly at the extremes) result in lags and nonlinearities in responses. Understanding this complexity has similarities to understanding ecological consequences of meteorological drought.

While the basics are fairly clear (less input generally means less output), there are a number of aspects that make the problem more complex, and defined relationships offer some use in prediction of places and times of relative sensitivity. A number of statistical and physically based modeling tools have been applied to the problem of relating weather to low streamflows (Laaha and others 2013). Analysis of the relationship is commonly framed within a sensitivity or similar modeling framework, such as the Budyko (1974) relationship between precipitation, net radiation (energy for evapotranspiration), and how precipitation is partitioned into evapotranspiration versus runoff, or simpler sensitivity/elasticity frameworks (sensu Schaake 1990) that empirically relate precipitation and temperature to streamflow.

The effects of precipitation variability on streamflow are fairly pronounced (Milly and Dunne 2002); generally, when there is less precipitation there is less runoff (e.g., fig. 2.1). Thus, meteorological drought results in hydrologic drought; however, the lag in streamflow response after precipitation stops means that hydrologic drought integrates precipitation inputs over time, and the lowest streamflows result from some combination of low initial soil/groundwater recharge, a long period since last precipitation, and evapotranspirative demand. Rapid evapotranspiration, at an event time scale, contributes to reduced flows in many circumstances, as it makes water unavailable for runoff. However, at long time scales (e.g., annual), evapotranspiration in many places is functionally limited by available soil moisture and tends to positively co-vary with streamflow (Milly and Dunne 2002). In other words, years with hydrologic drought may also experience reduced evapotranspiration.

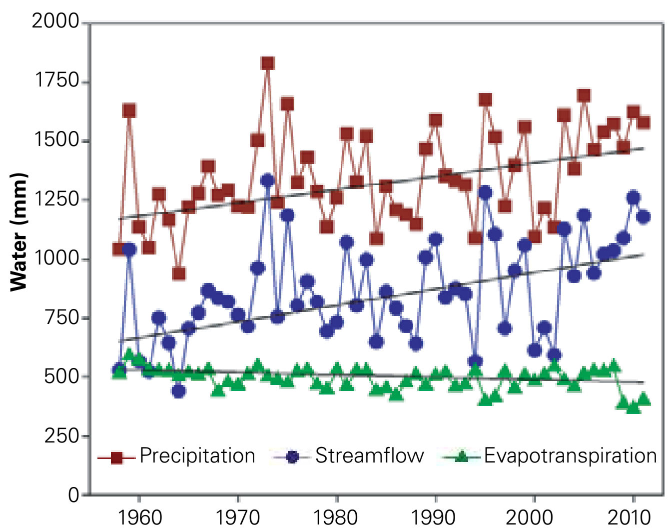

Figure 2.1—Streamflow variability as related to variations in precipitation and evapotranspiration at Hubbard Brook in the Northeastern United States.

One approach to understanding the relationships between meteorological drought and hydrologic drought is with direct correlations between flow data and indices such as PDSI, SPI, and variants (e.g., Haslinger and others 2014). These kinds of relationships often work best in rainfed agricultural catchments with limited snowpack, similar to validations where they were developed (Palmer 1965). Lag times from substantial snow accumulation or deep groundwater weaken relationships with these indices.

The water balance provides another approach, at least at interannual time scales. Using the water balance of soil (which is conceptually thin enough that we can ignore storage changes at interannual time scales) where Q represents both surface runoff and recharge to groundwater (from which it is assumed little evapotranspiration occurs, but which could store substantial recharge), we can write:

$$P = Q + ET \qquad (1)$$

Precipitation (P) is the largest term and is split into Q and ET. When P goes down, both Q and ET tend to decrease as well. However, we expect the fraction of precipitation allocated to Q versus ET to be a function of the ratio of net radiation (energy available for ET) to P (Budyko 1974, Milly and Dunne 2002). The general expectation is that as conditions become drier, a greater fraction of the P becomes ET (fig. 2.2), exacerbating hydrologic drought as meteorological drought progresses.

Figure 2.2 describes the Budyko relationship, where the red and blue lines represent physical limits related to energy and water availability for evaporation. The bottom axis is an aridity index, which is the ratio of the incoming energy (which can evaporate water if it is available) to precipitation (normalized by the latent heat of evaporation to put it in energy units). The left axis is the ratio of actual evapotranspiration to precipitation. Note that while the bottom axis can range to large values (lots of energy and little water), the vertical axis is constrained to be less than one because you cannot evaporate water you do not have. The curved line is Budyko's (1974) empirically derived relationship from a large number of rivers around the World. The black and gray points represent potential river basins one might find. Some rivers plot above the curve; others, particularly the gray ones, plot below. Points below the line partition more water into runoff than average, while points above the line evaporate more water. Variations above and below the line relate to factors, such as those discussed below, that enhance partitioning into runoff versus evapotranspiration.

The related climate elasticity of streamflow approach (Sankarasubramanian and others 2001, Schaake 1990) uses changes in precipitation or temperature to form estimates for annual streamflow changes. This method uses the relative (and interactive) roles of temperature and precipitation to understand variations in and mediation of streamflow (Fu and others 2007, Harman and others 2011, Potter and others 2011, Vano and others 2012).

In addition to aridity, a number of factors influence the partitioning of water into evapotranspiration rather than runoff/recharge (Woods 2003). A primary example is the soil depth or rooting depth. The more soil there is to hold water for trees to utilize, the greater the proportion that tends to evapotranspire. Similarly, coarse soils and steep slopes—both of which promote drainage of the soil profile—tend to reduce evapotranspiration. When more precipitation falls in cool months, interception losses are smaller than when it falls in warm months (Wolock and McCabe 1999). Runoff from basins where snowmelt dominates tends to be generated more efficiently than rain-related runoff (Berghuijs and others 2014). Forest cover and tree species may vary

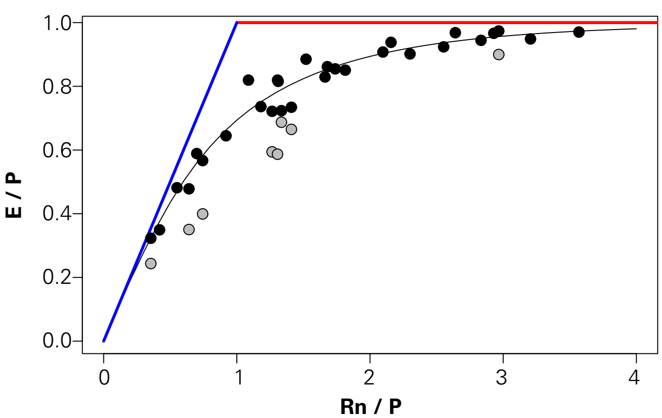

Figure 2.2—The Budyko (1974) relationship relating the aridity index given as the ratio of the net radiation (*Rn*, energy available for evapotranspiration) to precipitation (*P*) to the ratio of the actual evaporation (*E*) to precipitation. The blue line shows the energy limitation for potential evaporation where *Rn* = *P*. After there is sufficient net radiation to evaporate all precipitation, there is simply a limitation on evaporation given by the red line where *Ea* = *P*. Budyko's curve fits below these physical limits (black line) as an empirical relationship derived by plotting several of the World's rivers on these axes. The points (black and gray) represent some hypothetical streams. When equation 1 (see text) applies, the ratio of flow to precipitation (*Q* / *P*) is 1 – *E* / *P*, or the distance between the red line and the point.

independent of the slightly more static features, and they can exert substantial control on partitioning as well (e.g., Zhang and others 2001). Soil and vegetation controls are covered in more detail in chapter 10, but are mentioned here to illustrate potential controls on drought responses related to site characteristics.

Changes in precipitation characteristics with prolonged or extensive droughts could be important as well. Circulation changes associated with large-scale drought may promote variations in intensity, duration, and volume statistics for storms; these variations alter a storm's runoff generation potential (Potter and Chiew 2011). For example, if storms have similar frequency, but each delivers less water, a greater fraction may be lost to interception.

Spatial differences also play a role in the relationships between drought characteristics and streamflow characteristics, and these differences have been explored through hydrologic modeling approaches. For example, in the Western United States, framing the low-flow response in terms of both the annual recharge and the length of the dry season can highlight the influence of bedrock geology on late summer low flows (e.g., Tague and Grant 2009). The sensitivity of low flows to timing and amount of recharge is a function of how much groundwater levels are drawn down through a given volume of evapotranspiration or runoff. A range of recession analysis techniques have been applied for periods without precipitation (e.g., Kirchner 2009, Tallaksen 1995).

There are a number of ways in which meteorological extremes may ultimately manifest in low streamflows or similar hydrologic outcomes (Van Loon and Van Lanen 2012). Increasingly, physically based hydrologic models are being applied to evaluate historical and potential future droughts (e.g., Sheffield and others 2004) as opposed to more simple approaches related to indices such as SPI or PDSI. Unfortunately, not all models are equally suited for drought prediction, and scientists are actively testing across multiple models (e.g., Tallaksen and Stahl 2014).

Paleoclimatic Context for Evaluating Drought

Historic climate can be reconstructed through the use of many types of proxy variables. Interpretations of past climates, including drought, have been developed using tree rings; pollen, chironomids, diatoms, sediment, and isotopes taken from lake-sediment cores; records of Aeolian deposits in sand dunes; micro- and macrofossil remains in woodrat middens; speleothems; glacial and periglacial deposits; lake levels; and archeological records (Bradley 1999). These records are able to reveal small- to broad-scale changes in climate from decades to millennia. Each of these has certain biases and strengths. For example, plant macrofossils (leaves, fruit, etc.) can place a species at a specific location, while microfossils (pollen) have a range of uncertainty including source region and limited botanical identification. The most robust reconstructions derive from use of multiple proxies. Together with the continual development of varying climate proxies from sediment cores (e.g., Booth 2002), reconstructions of past climates will improve.

Tree rings are an important and common source for understanding late-Holocene droughts. Trees from mid to high latitudes produce annual growth rings, and the variation in ring width is influenced by a multitude of factors, including water availability, at various time scales. When aggregated from stand to continental scales, ring-width measurements can reflect these factors at various spatial and temporal scales. Fundamental to dendrochronology is the identification of the pattern of large and small rings through time in a collection of tree cores (i.e., cross-dating), which affords the ability to precisely date rings to a calendar year. Cross-dating (Douglass 1920) allows scientists retrospective estimates of past climatic and ecological change. Separating the climatic influence from the effects of genetic diversity and ecological effects, or, more realistically, reducing the nonclimatic input on growth from the climatic signal in tree rings, requires sophisticated standardization. The process involves transforming raw ring widths into time series of radial indices that are most likely to be a proxy of past climates. Advances in dendrochronology over the last 60 years (including techniques to interpret past climate through ring density and stable isotope composition as well as width) have enabled researchers to reconstruct climate from the stand and regional scale to continental and hemispheric scales (Cook and others 2004, Cook and others 2007, Cook and others 2010, Fritts 1976, Jacoby and D'Arrigo 1989, Meko and others 1993). Besides inferences about climate from growth increments, tree rings are used to date environmental disturbances tied to drought, like fire (Heyerdahl and others 2008, Whitlock and others 2003) and canopy disturbance (Lorimer 1985, Rubino and McCarthy 2004). These advances aid in the understanding of long-term spatiotemporal variations in drought and the atmospheric dynamics behind them.

The North American Drought Atlas (NADA) is the best source for understanding spatial and temporal patterns of tree-ring-reconstructed moisture variability over the last 500–1,000 years (Cook and Krusic 2004, Cook and others 2004). Chronology length and coverage varies in the atlas, with stronger fidelity in the Southwestern United States and weaker fidelity in New England and the Great Lakes region. The New England and Great Lakes regions generally show similar trends and patterns as other eastern regions, but more work is required to refine estimates of long-term drought severity and frequency. Another important note is that drought in the NADA is represented by different seasons in different regions. Winter precipitation is the primary signal reconstructed from tree rings from the Pacific Northwest to Northern Mexico and west Texas, while the eastern half of the atlas contains a summer moisture index (St. George and others 2010).

Continental and Subcontinental Patterns in Paleoclimate

Regional aridity and warmth was widely expressed during the middle Holocene. In many parts of southwestern and central North America, this was the warmest and driest interval of the past 10,000 years (Dean and others 1996, Spaulding 1991, Yu and others 1997). Proxy records from pollen, woodrat middens, tree-rings, lake levels, and Aeolian indicators document extensive warmth and aridity from 8 ka (kiloannus, or thousand years) to 3.8 ka, peaking at 6 ka (Benson and others 2002, Dean and others 1996, Holliday 1989, Mensing and others 2004). In contrast, the dry interval for the Pacific Northwest and parts of the northern Rocky Mountains was before 8ka (Brunelle and others 2005, Whitlock and others 2003).

In northeastern North America, various proxy records show that the mid-Holocene was a period of high aridity (Hardt and others 2010, Marsicek and others 2013, Menking and others 2012, Newby and others 2011, Newby and others 2014, Nichols and Huang 2012, Shuman and Donnelly 2006). Supporting earlier work by Webb and others (1993), these same studies indicate trends toward more moist conditions since the mid-Holocene aridity up to the 20th century. High hydroclimatic variability punctuated this trend, most notably between 3,900 and 5,200 years before present (e.g., Booth and others 2005, Foster and others 2006, Newby and others 2014).

Although pan-continental droughts were relatively rare during the last 1,000 years, the North American Drought Atlas indicates that they occurred more frequently during the Medieval Climate Anomaly (MCA) and could be related to defined climate modes (Cook and others 2014b). Notable pan-continental droughts include 1344–1353 C.E. (Common Era), 1661–1671 C.E., and 1818–1820 C.E. (Cook and others 2007). Cook and others (2014b) highlight the unusual nature of the 2012 pan-continental drought, but also indicate the potential for some predictability of these events as well as increased severity of pan-continental droughts as a result of greenhouse gas forcing.

Over the last 1,100 years, there have been several large-scale, hydroclimatic events in the conterminous United States. The MCA is widely recognized and regionally expressed as warm and/or dry (Bradley and others 2003, Lamb 1965, Mann and others 2009). Extensive periods of aridity characterized western North America during two centennial-scale arid periods, from 900 to 1100 C.E. and 1200 to 1350 C.E. (Cook and others 2010, MacDonald 2007, Stine 1994). In eastern North America, great droughts also characterize the MCA period, although the timing is shifted. Megadroughts (i.e., severe droughts of a decade or more) are seen in the extensive tree-ring record from the mid-Mississippi River Valley during the mid-900s C.E., 1100–1250 C.E., and then 1340–1400 C.E. (Cook and others 2010). Sediment core records document dry conditions during the MCA in the Eastern United States as well (Minckley and others 2011, Pederson and others 2005). These droughts come at the tail end of more-frequent drought conditions during the MCA across the Western United States (Cook and Krusic 2004, Cook and others 2004).

A persistent trend of increasing wetting was present across the Western United States during the Little Ice Age (Cook and others 2004). This trend was punctuated by severe droughts, such as the late 14th century and the 1805–1806 C.E. droughts centered on the Great American Desert (Cook and others 2007), the 1379–1388 C.E. drought centered on the Mississippi Valley, and the 16th century megadrought (Cook and others 2007, Stahle and others 2000). [The 16th century megadrought has been recently documented in the Northeastern United States (Ireland and Booth 2011) and is the most synchronous sustained drought in the Eastern United States of the last 500 years (Pederson and others 2013b).]

Regional Patterns in Paleoclimate

Eastern United States—As discussed previously, analyses of basin-scale streamflow records do not show an increase in drought frequency over the past several decades; however, there are several important aspects of the changing hydroclimate in the Eastern United States (figs. 2.3–2.6). Notably, there is a recent divergence in moisture conditions between the Northeastern and Southeastern United States (Melillo and others 2014). Meteorological drought has become more frequent and severe in the Southeastern United States since the 1980s drought (Laseter and others 2012, Melillo and others 2014, Pederson and others 2012, Seager and others 2009). While the 1980s drought was one of the more severe droughts since 1700 (Cook and others 1988), reconstructions of PDSI from tree rings indicates that prior centuries were generally drier and had more severe and extended drought (figs. 2.3 and 2.5). The most recent severe droughts fall short of the more severe droughts in the last millennium (Stahle and others 2013b, Stahle and Cleaveland 1992).

The Northeastern United States is currently experiencing one of the wettest growing-season pluvials since 1531 C.E. (figs. 2.3, 2.4, and 2.6) (Pederson and others 2013b). Variation in reconstructed PDSI occurs at decadal to multi-decadal time steps with a positive shift in increased moisture availability and extreme events since the early 2000s (Matonse and Frei 2013, Melillo and others 2014, Pederson and others 2013b). These shifts are observed across New England in rising streamflow and groundwater tables (Dudley and Hodgkins 2013, Weider and Boutt 2010). Varved sediments also indicate the Northeast is undergoing one of the wettest periods in the last 1,000 years (Hubeny and others 2011). These findings are supported at even longer time scales. Lake levels across Massachusetts indicate a positive trend in effective moisture over the last 3,000 years resulting in “exceptionally high” levels of water in the recent era (Newby and others 2014). The centennial trend of increased moisture in the tree-ring reconstruction and various hydrological measures in the Northeast generally follow the trajectory of the Massachusetts lakes indicating that the current period could be one of the wettest of the last 3,000 years.

Plains and Midwest—Reconstructions of moisture in the Plains and Midwest contain many of the same trends described above (fig. 2.3), although there is some spatial complexity in trends during the most recent century. The northwestern portion of the Prairie Pothole Region of the upper Great Plains has become drier over the past century while the southeastern portion, bordering on the western edge of the eastern

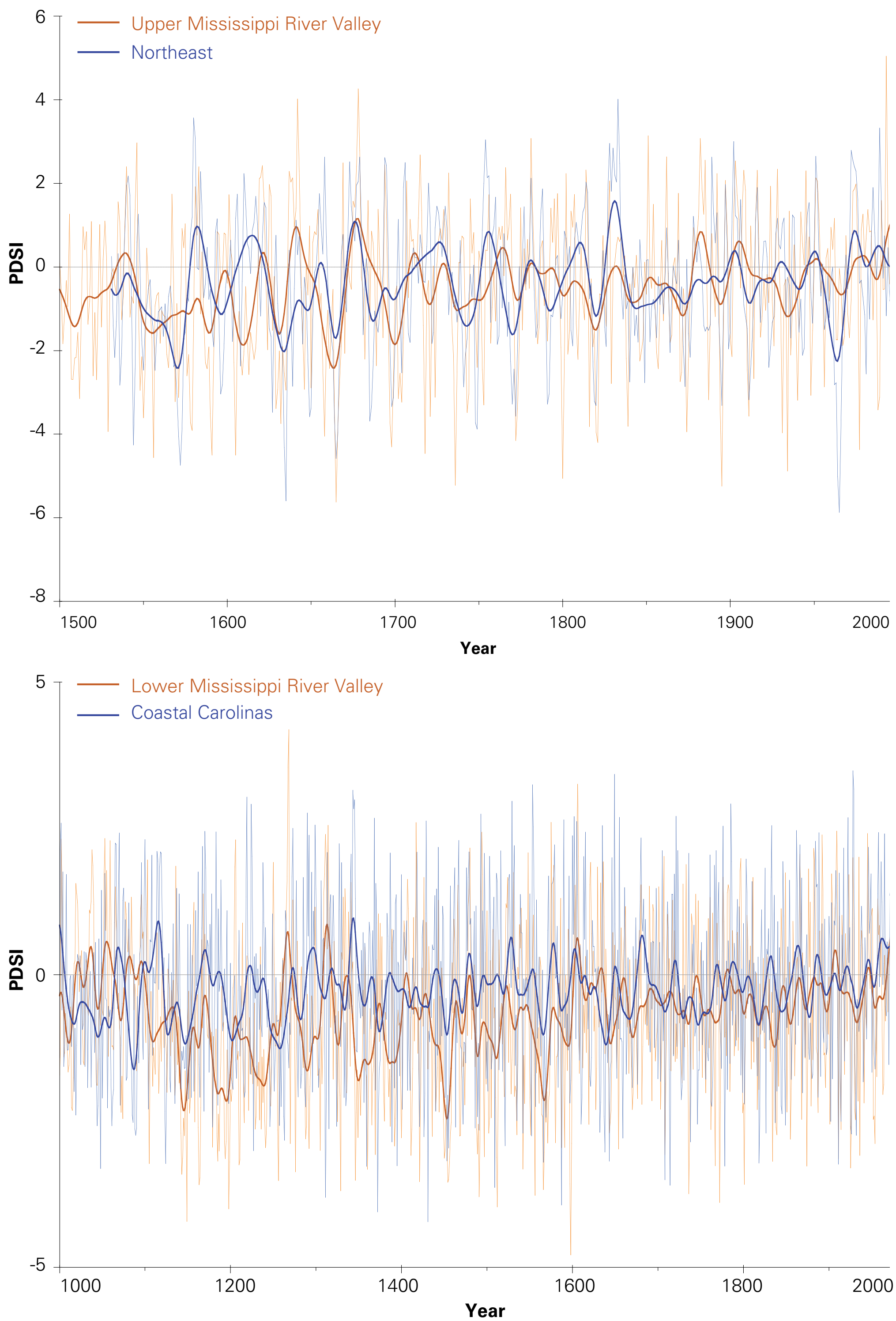

Figure 2.3—Time-series of Palmer Drought Severity Index (PDSI) over the last 500 years (Upper Mississippi River Valley and Northeast) and 1000 years (Lower Mississippi River Valley and Coastal Carolinas). Notable features in these series are the relative absence of annual values during the 16th century megadrought in the Upper Mississippi River Valley and Northeastern United States (upper panel) and similar features during the 12th and 13th centuries in regions to the south (lower panel). Annual values are shown in faint, thin lines, and 20-year smoothed data with the dark, thicker lines. The horizontal line at 0 represents the 1900–2005 mean. These data are drawn from reconstructions of PDSI from the North American Drought Atlas (Cook and others 2007) for the Upper Mississippi River Valley, Lower Mississippi River Valley, and Coastal Carolinas, and Pederson and others (2013b) for the Northeast.

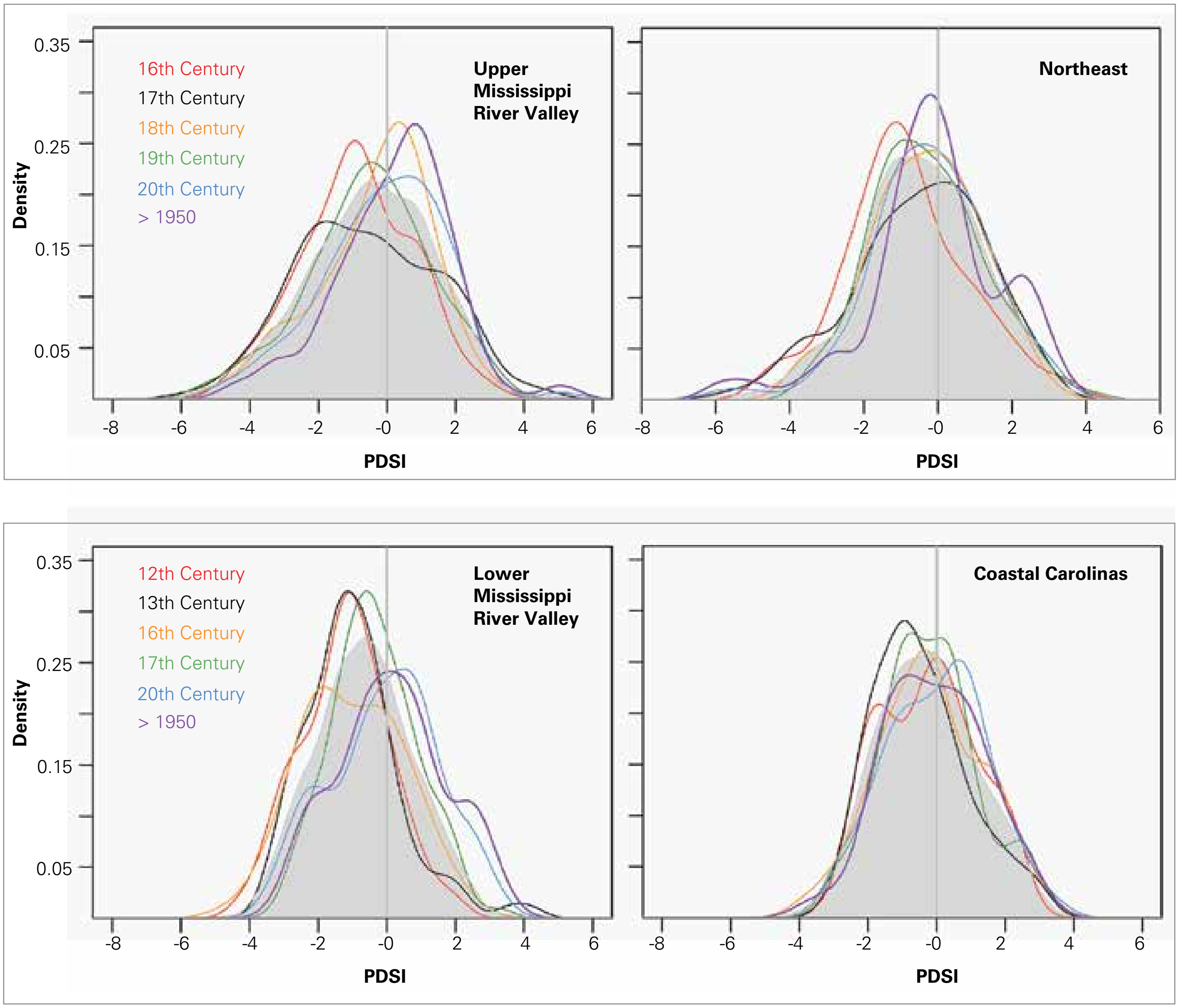

Figure 2.4—Density of Palmer Drought Severity Index (PDSI) over the last 500 years (Upper Mississippi River Valley and Northeast) and 1000 years (Lower Mississippi River Valley and Coastal Carolinas). The gray area represents the 1500–2005 and 1000–2005 distributions, respectively. The areas under the blue line and purple line represent the 1900–2000 and 1950–2005 periods, respectively. The end of the 20th century, a period of intense forest study, is generally one of the wettest periods in each region, indicating a general shift towards wetter conditions across the Eastern United States. The vertical line at 0 represents the 1900–2005 mean, so the values on the horizontal axis are departures from the mean. These data are drawn from reconstructions of PDSI from the North American Drought Atlas (Cook and others 2007) for the Upper Mississippi River Valley, Lower Mississippi River Valley, and Coastal Carolinas, and Pederson and others (2013b) for the Northeast.

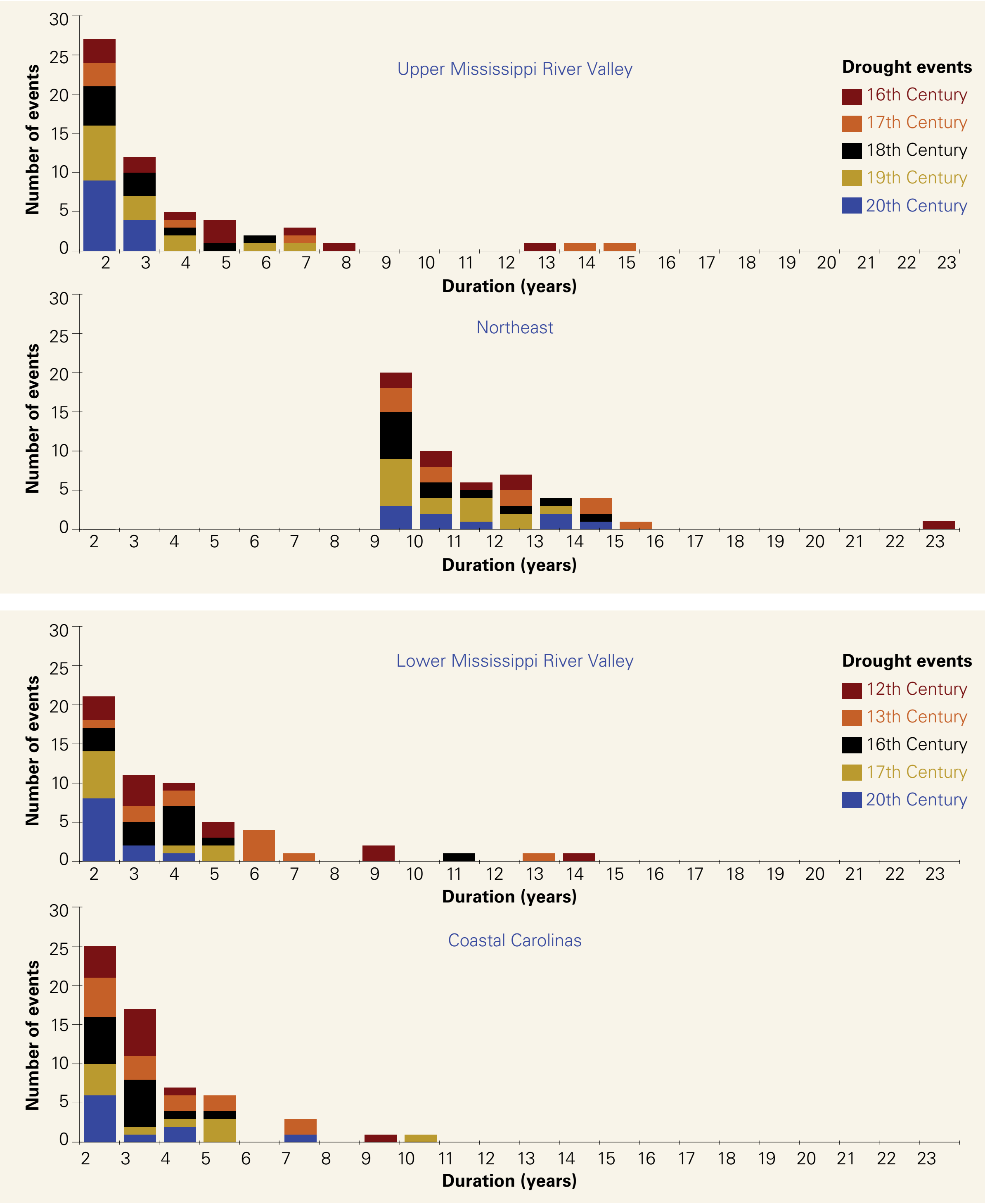

Figure 2.5—Frequency of discrete drought events of 2 years or more for key centuries over the last 500 years (Upper Mississippi River Valley and Northeast) and 1000 years (Lower Mississippi River Valley and Coastal Carolinas). In these four corners of the Eastern Deciduous Forest region, extended drought has been relatively unusual during the 20th century (blue). Droughts of the longest duration (>5 years) occur mostly in the 16th and 17th centuries in the north and the 12th and 13th centuries in the south. These data are drawn from reconstructions of PDSI from the North American Drought Atlas (Cook and others 2007) for the Upper Mississippi River Valley, Lower Mississippi River Valley, and Coastal Carolinas, and Pederson and others (2013b) for the Northeast.

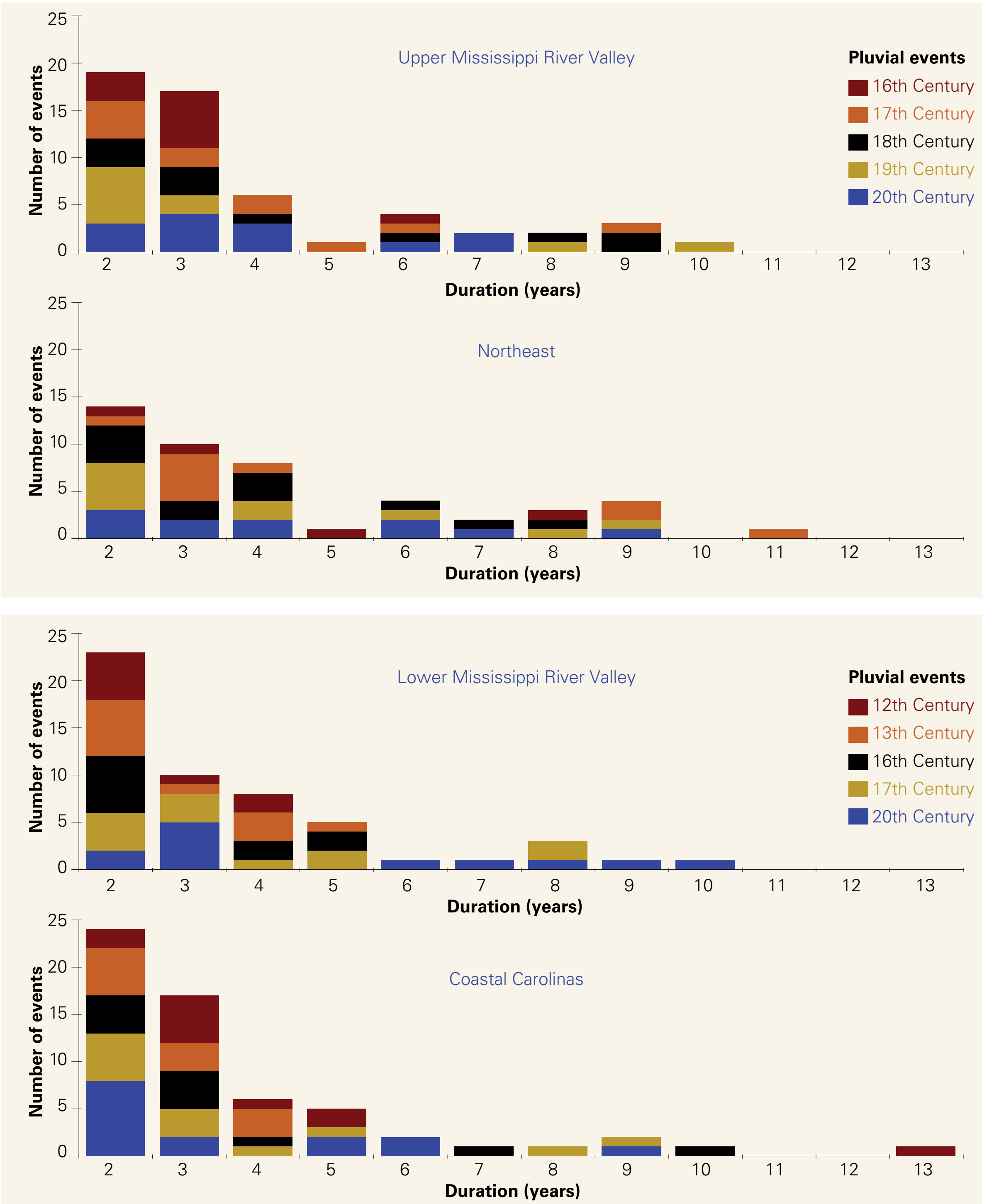

Figure 2.6—Frequency of discrete pluvial events of 2 years or more for key centuries over the last 500 years (Upper Mississippi River Valley and Northeast) and 1000 years (Lower Mississippi River Valley and Coastal Carolinas). In these four corners of the Eastern Deciduous Forest region, extended pluvials (>5 years) have been more frequent in the Lower Mississippi River Valley and the Northeast during the 20th century (blue). These data are drawn from reconstructions of PDSI from the North American Drought Atlas (Cook and others 2007) for the Upper Mississippi River Valley, Lower Mississippi River Valley, and Coastal Carolinas, and Pederson and others (2013b) for the Northeast.

U.S. forest, has become wetter (Millett and others 2009). Generally, droughts were more severe and of greater duration prior to the 20th century (figs. 2.5 and 2.6) (Cleaveland and Stahle 1989, Cook and others 2004, Cook and others 2010, Stahle and others 1985, Stahle and others 2007, Woodhouse and Overpeck 1998). Outstanding findings from reconstructed PDSI include more severe and extended droughts during the MCA (900s and 1100–1200), early Little Ice Age (late-1300s), and late 16th century (Cook and others 2010, Stahle and others 2007, Woodhouse and Overpeck 1998). During the past 500 years, however, the Dust Bowl drought was second only to the 16th century megadrought (Fye and others 2003). Since 1500 C.E., less extreme drought conditions (similar to the 1950s drought) have occurred 12 times; drought conditions that exceeded those of the 1950s drought have occurred 4 times during that same period (Fye and others 2003). Interestingly, there is a circa 600-year trend in wetting in the mid-Mississippi River Valley that peaks late in recent decades (fig. 2.3) (Cook and others 2010). Similar to the Northeastern United States, this trend towards increasing moisture is reflected in the southeastern Prairie Pothole Region, where a survey of ponds during the month of May suggests a substantial increase in the number of ponds since 1995 versus the period between 1974 and 1994 (Ballard and others 2014).

Southwestern United States Including California and the Great Basin—The Southwestern United States is an example of where understanding of long-term climate dynamics is crucial for sustainable management of environmental resources. Both an early and a recently updated reconstruction of streamflow in the Colorado River Basin indicate that water management agreements were developed during one of the wettest periods of the last 500 years (Stockton and Jacoby 1976, Woodhouse and others 2006). Creation of water distribution policies during such extreme periods (like the early 20th century pluvial) can lead to an over-expectation of water delivery when the climate system returns to "normal" or drought conditions. Since the early 20th century pluvial, there has been significant drying in the region.

Megadroughts were more common in the Southwest during the past several millennia. The most long-lasting and severe drought interval of the late Holocene is documented by multiple proxies across the central Great Basin extending to Great Salt Lake, and lasting from 2,800 to 1,850 years before present (Mensing and others 2013). During this period, wetlands and meadows shrunk in size, lake levels decreased, and waters became more saline. Drought was zonally limited, and areas of the northern Great Basin remained or grew wetter. This pattern of dry in the southwest and wet in the northwest across the Great Basin is supported by large-scale spatial climate pattern hypotheses involving ENSO (El Niño-Southern Oscillation), PDO (Pacific Decadal Oscillation), AMO (Atlantic Multidecadal Oscillation), and the position of the Aleutian Low and North Pacific High, particularly during winter (Mensing and others 2013).

As elsewhere in North America, droughts were most notable during the MCA (Cook and others 2004). Reconstructed droughts during this period were not only relatively more severe versus those during recent centuries (Cook and others 2010, Meko and others 2007), but drought during the MCA also covered a considerably greater portion of the Western United States. Approximately 30 percent of the Western United States experienced annual drought during the 20th century versus approximately 41.3 percent between 900 and 1300 C.E. (Cook and others 2004). Treelines were elevated in the Sierra Nevada and White Mountains during the MCA (LaMarche 1973, Lloyd and Graumlich 1997), and growth patterns indicate extended warm and dry conditions (Graumlich 1993, Scuderi 1993).

Reconstructions of hydroclimate covering the last 2,000 years indicate the occurrence of drought equal to or greater in severity than those during the MCA. A reconstruction of the San Juan River from bristlecone pine indicates a severe, 51-year drought from 122 to 172 C.E. (Routson and others 2011). [Although this record has low tree replication during this period, comparison to other records of drought in and adjacent to this region substantiates the severity of second century drought (Cook and others 2004, Cook and others 2007, Knight and others 2010).] The extent of this drought, from southern New Mexico to Idaho, is similar to the 12th century megadrought (Routson and others 2011), suggesting perhaps a recurring forcing.

Over the past 2,740 years, oscillations in hydrologic balance are documented throughout the region, with droughts occurring about every 150 years and intervals between droughts ranging from 20 to 100 years (Benson and others 2002). Similar patterns of synchronous drought exist from the Great Basin through New Mexico and coincide with Ancient Puebloan withdrawals and abandonments in Arizona and Colorado, documenting the likely wide extent and impact of these droughts (Benson and others 2002, Dean 1996, Grissino-Mayer 1996).

Insight into seasonal dynamics of precipitation delivery within the North American Monsoon region was recently revealed. Using indices of early-wood and late-wood widths from two tree species over a large region, Griffin and others (2013) reconstructed cool-season and monsoonal precipitation going back to 1539 C.E. The two records of reconstructed precipitation indicate periods of dry, cool-season conditions followed by a failure of the monsoon during the 1570s, 1660s, 1770s, and early 2000s. Similarly, the early 20th century pluvial was characterized by synchronous, dual-season surplus precipitation. Interestingly, much of the period during the instrumental record is characterized by a high frequency of asynchronous cool-season and monsoonal precipitation or "*opposing-sign precipitation anomalies*" (Griffin and others 2013). What this study reveals, however, is that this anti-phasing of cool-season and monsoonal precipitation is more of an exception rather than the rule during the 470-year reconstructions. Therefore, a fuller understanding of climate dynamics of the North American Monsoon region likely requires an understanding of the dynamics outside of the instrumental record.

More than half of California was in a state of moderate drought between June 2013 and June 2014, at which time 100 percent of the State reached that level of drought. The conditions in 2014 culminated in the driest 3-year period of the last century (Griffin and Anchukaitis 2014). To place this drought in a long-term context, tree-ring analysis of blue oak (*Quercus douglasii*) around the Central Valley of California was used to characterize annual variations of drought history over the past seven centuries (Meko and others 2001, Meko and others 2011, Stahle and others 2001, Stahle and others 2013a). Independent reconstructions of November–April precipitation for North and South Coast Ranges indicate strong annual correlations during the common period (1584–2003), although there were eras of asynchrony between the two regions during the 1770s, 1810s–1840s, and 1970s, for example (Stahle and others 2013a). Winter storm track position is thought to control the more extreme periods in the tree-ring records. An updated reconstruction of drought in this region using blue oak collected following the 2014 growing season indicates unprecedented drought conditions in 2014 compared to the prior 1,200 years (Griffin and Anchukaitis 2014).

Ecological Outcomes of Paleoclimate on Forests

Ecological responses at regional scales were widespread, with changes in ecosystem structure, composition, and range toward dry-adaptations and changes in disturbance patterns reflecting drier conditions. Forest response included, for example, conversion from mesic forest to prairie in the Midwest (Baker and others 1992), dominance of dry-adapted taxa (*Chrysolepis*, *Quercus*) at high elevations that are currently occupied by *Abies* and *Pinus* in the Sierra Nevada of California (Anderson and Smith 1994), and expansion of warm-adapted pines and oaks in the Colorado Plateau (Betancourt 1990). Regionally warm and dry climates evolved in the middle Holocene as a combined effect of insolation changes due to orbital relations, which affected the amount and seasonality of precipitation, and changes in the position of jet streams, storm tracks, and blocking of warm maritime air (Baker and others 1992, Dean and others 1996, Yu and others 1997).

The ecological impact of the Medieval Climate Anomaly droughts in the Western United States is reflected by the presence of in-situ stumps in present-day rivers and lakes (Kleppe and others 2011, Stine 1994); lowered lake levels (Stine 1990); decreased alpine treelines (LaMarche 1973, Lloyd and Graumlich 1997); decreased growth of trees (LaMarche 1974); and changes in species distributions, elevation zonation, and abundance (Lloyd and Graumlich 1997, Millar and others 2006). Shifts toward warm and dry conditions also altered fire regimes, such that fire activity was highest relative to the entire Holocene during the MCA in the Pacific Northwest and parts of the Northern Rocky Mountains (Brunelle and others 2005, Whitlock and others 2003), whereas changes in forest structure related to aridity (low fuel density, sparse stands) resulted in many small fires in the western Sierra Nevada relative to mesic intervals (Swetnam 1993). Mechanistic forcing for the dry centuries of the Medieval period has been related to changes in ocean circulation, in particular the development of persistent positive North Atlantic Oscillation conditions (Trouet and others 2009), and northward shifts in storm tracks across the eastern North Pacific with a contraction of the Aleutian Low (Graham and others 2007, MacDonald and Case 2005). More recent reconstructions suggest that the causes were more complex and that the MCA was characterized by an enhanced zonal Indo-Pacific Sea Surface Temperature (SST) gradient with resulting changes in Northern Hemisphere tropical and extra-tropical circulation patterns and hydroclimate regimes (Graham and others 2011).

An important hydroclimatic event with ecological implications in the Western United States is the early 20th century pluvial, one of the most contiguous pluvial events of the last 1,000 years (Cook and others 2011, Stahle and others 2007). This pluvial—a period of extended, above average moisture condition—covered about half of the continent and is thought to have been an important trigger in tree recruitment in the Southwestern United States (Savage and others 1996). Generally, the 20th century was a substantially wetter period across the United States versus the prior three to eight centuries (Cook and others 2004, Cook and others 2010, Stahle and Cleaveland 1992, Stahle and others 1988, Stahle and others 2007).

The ecological impact of historic droughts is well documented throughout the Holocene in the Eastern United States. Throughout the Holocene, drought has been an important contributor to forest change (Shuman and others 2009a). Drought variation during the MCA was an important factor contributing to the decline in American beech (*Fagus grandifolia*) in the Great Lakes region (Booth and others 2012). In a surprising contrast, drought is seen to be an important factor of deciduous forest expansion in Minnesota early in the Little Ice Age (Shuman and others 2009b). Abrupt hydroclimatic variability is attributed to peatland development in Pennsylvania (Ireland and Booth 2011). The 16th century megadrought likely influenced stand dynamics in a large wetland complex in the Mississippi Valley (Stahle and others 2006), while repeated droughts during the mid-18th century appear to be contributors of regional-scale canopy disturbance in the Southeastern United States (Pederson and others 2014). The latter finding follows observations of oak growth decline and mortality in the Midwestern United States during the late 20th century (Pedersen 1998).

Periods of abundant moisture could be key elements of forest dynamics in the mesic Eastern United States. An expansion of yellow birch (*Betula alleghaniensis*) occurs during a period of increased moisture and high lake levels (Jackson and Booth 2002). The rise of maple (*Acer* spp.) and other mesophytic species during the 20th century occurs towards the end of a centennial wetting trend (McEwan and others 2011, Pederson and others 2013b); however, given the complexity of factors in this region, it is yet to be determined whether this wetting is an important factor of this rise. However, if the trend in wetting over the last 2,000–3,000 years in the Northeastern United States (Newby and others 2014) is similar to the wetting trend over the Eastern United States from the end of the Little Ice Age into the 20th century (Pederson and others 2013a), there is much to investigate regarding drought-forest dynamics over this mesic region across spatial and temporal scales. Greater detail on the ecological impact of drought on forests is found in chapter 4.

Historical and Recent Drought

Approaches to quantifying changes in drought depend on how "drought" is defined (Trenberth and others 2014). In this context, we suggest two important observations: (1) climate variability occurs at multiple time scales (Hurst 1951), and (2) widespread drought-related forest mortality is occurring (Allen and others 2010, Breshears and others 2005, van Mantgem and others 2009). The first point is that many hydrologic phenomena show cycles at a large range of time scales, some tied to known climatic modes (e.g., El Niño, PDO, AMO), and others at longer scales. The implication is that the ability to detect trends in hydrologic phenomena may depend upon whether the time period of interest encompasses one or more of these cycles. The primary value in finding hydrologic trends, then, is in relating different co-occurring trends to one another to learn about potential causal relationships.

The second point is that, in many regions of the World, large numbers of long-lived plants seemed to be dying at rates unprecedented in historical times, indicating that hydrologic processes may be changing in ways that are ecologically meaningful. Because of this, we require an understanding of what aspects of changing drought regimes are causing these significant ecological impacts. The drought metrics and understanding of forest physiology and ecology outlined above indicate water balance, temperature (both as index to ET and as it affects physiology), and time are three key ways drought may be changing. This is a complex mix, particularly considering that the interactions among these three are of fundamental importance. A rich literature on historical (relying on the instrumental record) changes tying variability to known climate modes can help us interpret some of the trends and patterns of change. The recent literature is rich on the topic of recent trends for many regions in the United States, but only two regions are discussed here as examples.

Eastern United States

Compared to most other areas of the United States, the East receives large amounts of precipitation, with

30-year (1981–2010) normals ranging from 80 to 200 cm across the region (PRISM 2013). Precipitation is spatially and temporally variable in the East due to a combination of many factors, including proximity to the Atlantic Ocean and Gulf Stream, influence of the Great Lakes, and orographic effects associated mountains, including the Appalachian chain that extends from Alabama to north of Maine (Huntington and others 2009, Mulholland and others 1997). Broad-scale circulation patterns influence the eastern climate, contributing to patterns in precipitation. In the Southeast, precipitation is associated with El Niño/ Southern Oscillation (ENSO), with dry conditions occurring during La Niña events (Piechota and Dracup 1996, Ropelewski and Halpert 1986, Roswintiarti and others 1998). During the growing season, precipitation has been linked to the North Atlantic Subtropical High in the Southeast (Li and others 2011). Precipitation in the Northeast also suggests possible linkages with ENSO, as well as the North Atlantic Oscillation (NAO) (Bradbury and others 2003, Kingston and others 2007), with evidence of a relationship between negative NAO conditions and drought (Bradbury and others 2002, Seager and others 2012). A positive NAO promotes pluvial conditions (Seager and others 2012), although the ultimate causes of pluvials are complex (Ning and Bradley 2014, Seager and others 2012).

Despite the relatively abundant supply of precipitation in the East, droughts are a natural part of the climate system and occur with regularity. Over the last 100 years, eastern droughts generally last only 1–2 years, unlike arid regions of the Central and Western United States that experience more intense and prolonged meteorological droughts (Seager and others 2009). A notable example of a severe eastern drought is the drought of the early 1960s that lasted 4 years and affected a broad area, including New England, the Mid-Atlantic States, and parts of the Midwest (Namias 1966). The Southeastern United States has also experienced severe droughts over the last several decades, including one from 1984 to 1988 (Cook and others 1988) and more recently from 2006 to 2008 (Kam and others 2014).

Analyses of past occurrences of drought show no indication that they are becoming more frequent in the Eastern United States. In fact, precipitation has increased throughout the region, especially in the Northeast (Melillo and others 2014), which has resulted in an overall reduction in drought conditions (Andreadis and Lettenmaier 2006, Sheffield and Wood 2008a). These increases in precipitation have led to documented regional increases in streamflow and are consistent with patterns observed in many other regions of the United States (Groisman and others 2001, McCabe and Wolock 2002). Although there are no apparent increasing trends in drought during the recent past (Patterson and others 2013, Sheffield and Wood 2008a), most models indicate that drought frequency in the Eastern United States will increase by the end of the century, despite projected concurrent increases in precipitation (Hayhoe and others 2007, Sheffield and Wood 2008b). This paradox is largely explained by enhanced future evapotranspiration associated with longer growing seasons and warmer air temperatures. This may already be occurring in some regions. For example, recent analyses suggest that although there are no observable trends towards increasing drought frequency over the past 70+ years, overall streamflow has decreased by about 7 percent in the South Atlantic region of the Eastern United States (Patterson and others 2013).

Although droughts in the Eastern United States are not as severe as those in the Southwest and Great Plains, the affects can be acute because of the high population density and associated demands for water (Patterson and others 2013). The southeastern drought of 2006 to 2008 caused societal and economic hardships, with agricultural losses exceeding a billion dollars (Manuel 2008). The drought also led to interstate disputes over water use as supplies diminished. Recent evidence suggests that severe, short-duration droughts in the East can also have lasting effects on forest ecosystems (Klos and others 2009, Pederson and others 2014). These events can alter the structure of forests for centuries, particularly if they are exacerbated by other broad-scale disturbances, such as frost, ice storms, and insect outbreaks. Understanding the effects of extreme climatic events, such as droughts, will help determine how eastern forests will respond to future climate change.

Northwestern United States

Temperature recorded at long-term stations has increased in the Northwest since the early 20th century, with some of the cause attributed to increased carbon dioxide (CO_2) (Abatzoglou and others 2014). Less certain, however, are precipitation trends and their causes, as suggested by the long-term instrumental record and modeling studies. Most of the data for these long-term studies come from stations at lower elevations, where they can be collected more reliably. Because much of the water in the Northwestern United

States comes out of mountains, further data collection and analysis is needed at high elevations to gain a better understanding.

One of the most frequently discussed changes to Northwest hydroclimatology is trends in snowpacks (e.g., Mote 2003). Summers are very dry compared to winters in much of the Northwest, although there are June peaks in precipitation in the Northern Rockies and some summer monsoonal moisture over southeastern Oregon and southern Idaho. Because of the general lack of summer precipitation, snowmelt is a critical part of the water supply in forests and rivers during the Northwest's growing season. A lack of snow for melt means that the snowpack ablates earlier and stream runoff occurs earlier (Cayan and others 2001, Stewart and others 2005) because a shallower snowpack takes less time to melt. What has previously been unclear is the relative contributions of temperature and precipitation to these snowpack trends (Luce and others 2013, Mote and others 2005), which has important consequences for leveraging current trend information for application to future projections.

Streamflow has been declining in the Northwest since the late 1940s (Clark 2010, Dai and Trenberth 2002, Luce and Holden 2009), particularly drought-year flows. The declines have been driven primarily by winter precipitation declines (Luce and others 2013). Streamflow declines have shown a pattern where the driest years (25th percentile), in particular, are showing the strongest trends, approaching a 50-percent decline in drought-year flows over the last 60+ years (Luce and Holden 2009). This pattern is consistent with other observations of increasing variance in runoff in the Western United States (Pagano and Garen 2005) and with expectations for increasing variance in runoff ($P - E$) with warming (Seager and others 2012). Previous work missed detecting changes in precipitation because it was based on analysis of low-elevation stations (Mote and others 2005, Regonda and others 2005), while the decreases in runoff were related to decreased high-elevation precipitation driven by reduced westerly wind speeds over the region.

The twin contexts of decreasing precipitation and warming temperatures, with consequent changes in snowpack, make evaluation of drought effects in the region multifaceted. For example, earlier work showed that wildfire was more prevalent in the region in association with earlier snowmelt, presumed to be a function of warming temperatures alone (e.g., Westerling and others 2006). An awareness of the declining precipitation, though, has revealed that the precipitation variability has historically been a more important control on interannual variability in burned area (Abatzoglou and Kolden 2013, Holden and others 2012, Riley and others 2013), and increasing wildfire area is consistent with increasing drought severity as indexed by the lower streamflow quartile.

Low flows during the dry summer months are also showing declines (Dittmer 2013, Leppi and others 2011, Lins and Slack 2005). Some of this is related to earlier runoff timing, as the time between recharge of the soil mantle and groundwater and the runoff increases; but some is also related to declining recharge volumes (Safeeq and others 2013). Again, most of the effect has been attributed to earlier snowmelt and reduced accumulation as a function of increasing temperatures, but the precipitation variations affect both timing and total annual streamflow (Holden and others 2012, Luce and Holden 2009). More work in this area is needed to identify the relative contributions in the historical record.

Drought in the Pacific Northwest is strongly tied to several modes of climate variability, ENSO, the Pacific North America Pattern (PNA), and the Pacific Decadal Oscillation (PDO). The most severe droughts generally occur during El Niño events (Cayan and others 1999, Piechota and Dracup 1996, Redmond and Koch 1991, Ropelewski and Halpert 1987). ENSO-mediated precipitation variations may disproportionately affect mountain watersheds (Dettinger and others 2004). Precipitation variations are also well correlated to the PNA at intraseasonal to interannual time scales (Abatzoglou 2011) and the PDO (Mantua and others 1997) at decadal time scales. Trends in these indices are weakly related to the observed trends in westerly winds over the region, and consequently with trends in runoff and precipitation (Luce and others 2013). The related changes in pressure pattern driving the wind changes are consistent with increased CO_2 content, but the change in winds and pressure exceed the average expectation for wind decreases by 2080.

Climate Change Impacts on Drought Frequency and Severity

Will drought be more or less frequent in the future? Will it be more or less severe in the future? These questions are very difficult to answer with climate change projections from GCMs (Dai 2013, Hoerling and others 2012, Sheffield and Wood 2008b, Trenberth and others 2014). For example, some researchers have used the PDSI conceptualization of "drought," which in a warming future may index temperature more than it indexes water balances (Cook and others 2014a, Hoerling and others 2012, Sheffield and others 2012, Trenberth and others 2014), leading to suggestions of alternative models or indices of drought that may be more indicative. A more relevant question for this report is: will forests and rangelands be more severely impacted by drought in the future? Answers for this question are more complex, but potentially more informative and certain. For example, warming means that the droughts we have now are more likely to produce tree mortality for a given level of water deficit (e.g., Adams and others 2009). Beyond that, answers to the question of whether other aspects of drought will change in the future, particularly those related to precipitation, are much more challenging to elicit from GCMs. Despite these challenges, the recent Intergovernmental Panel on Climate Change (IPCC 2013) report provides some expectations of how climate change will impact drought frequency and severity in the United States.

General Drought Projection Information

A number of different projections are given in the IPCC (2013) Working Group 1 (WG1) report that relate to drought, such as changes in annual and seasonal precipitation (fig. 2.7) and associated changes in runoff and soil moisture (fig. 2.8). Except as specifically noted, in this chapter we discuss the multi-model means for RCP 8.5 2081–2100 compared to 1986–2005 averages from IPCC (2013) chapters 12 and 14. Generally, expected runoff patterns correlate strongly to expectations of future precipitation, but future soil moisture is expected to decline in most land areas. Theoretically, this is related to increased evapotranspiration, itself tied to increased downwelling longwave radiation. The combination of increased runoff, which is related to increased precipitation, and decreased soil moisture is intriguing. Usually soil

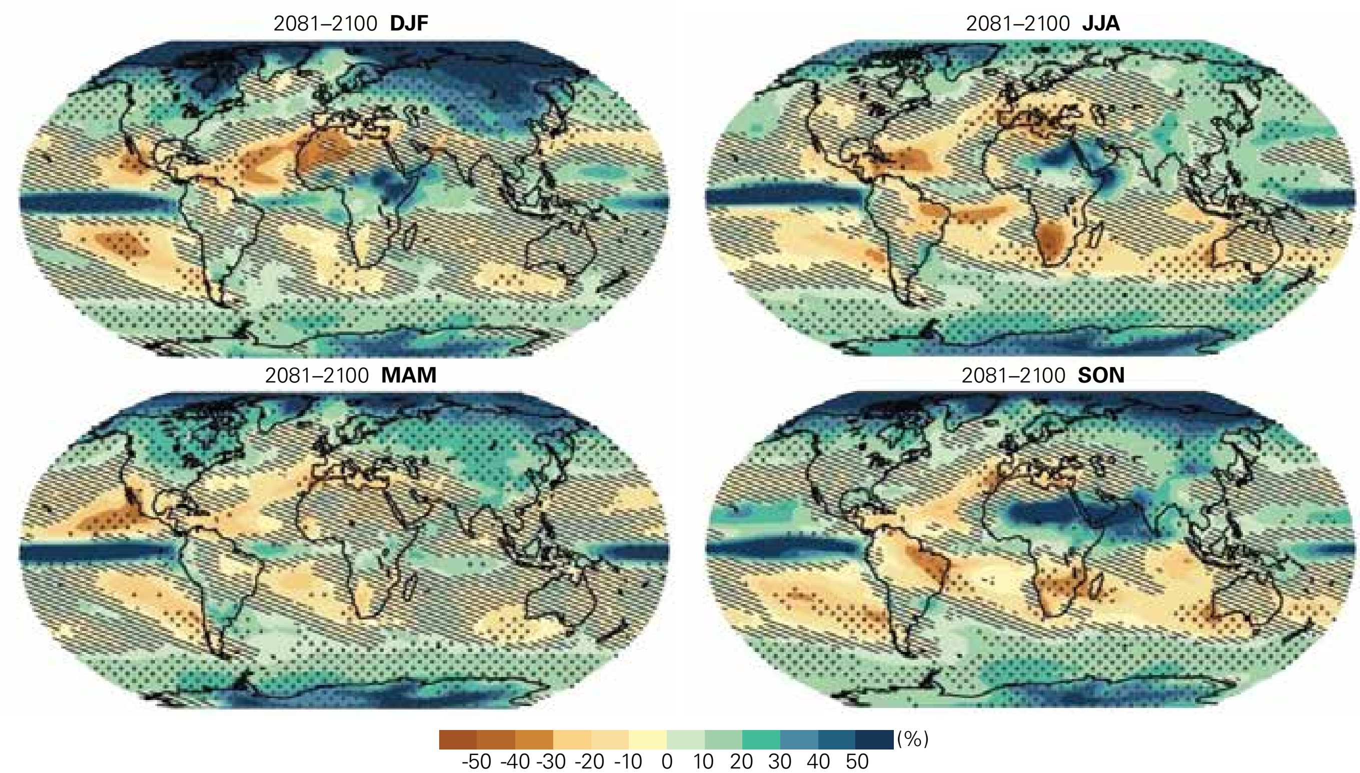

Figure 2.7—Coupled Model Intercomparison Project Phase 5 (CMIP5) multi-model average percentage change in seasonal mean precipitation relative to the reference period 1986–2005 averaged over the period 2081–2100 under the RCP 8.5 forcing scenario. Hatching indicates regions where the multi-model mean change is less than one standard deviation of internal variability. Stippling indicates regions where the multi-model mean change is greater than two standard deviations of internal variability and where at least 90 percent of models agree (from fig. 12.22 of IPCC 2013). Thirty-nine CMIP5 models were used for each panel. From top to bottom, the figures describe northern hemisphere winter (Dec.–Feb.), spring (Mar.–May), summer (Jun.–Aug.), and autumn (Sep.–Nov.). RCP means Representative Concentration Pathway.

moisture and runoff vary together in forested watersheds. The differential trends in soil moisture and runoff is explained conceptually by the fact that increased precipitation comes by way of increased precipitation intensity, meaning that more of the water might be expected to run off. It is not clear that this is an appropriate re-partitioning in forested landscapes, however, and might just be an artifact of the land surface model in the GCMs. An additional model output is also conceptually tied to increased precipitation intensity. The argument that similar or slightly increased total precipitation is delivered in higher intensity events yields an estimate that interstorm periods will be longer, increasing the number of consecutive dry days in most years, although the effect is minor across much of the continental United States (fig. 2.9).

Post-processing of GCM outputs has also been applied to examine how changes in precipitation amount, timing, and form (snow versus rain) interact with energy available for ET to estimate details of potential future conditions (e.g., Cook and others 2015, Elsner and others 2010, Hamlet and others 2013, Sheffield and others 2004, Vano and others 2012, Wood and others 2004). Some of these are essentially more-detailed versions of the land-surface models used to describe the lower boundary condition of the GCMs. These approaches relatively directly disentangle a range of drought definitions to describe how the components of drought are likely to change in the future. An important consideration in interpreting output of these simulations is that they can double count the effects of increased incoming longwave radiation on ET (Milly 1992).

Other approaches directly estimate PDSI values with projected temperature and precipitation changes (e.g., Dai 2013). Whether the calculated changes in PDSI reflect actual changes in drought severity is a subject of substantial debate (Cook and others 2014a, Hoerling and others 2012, IPCC 2013, Sheffield and others

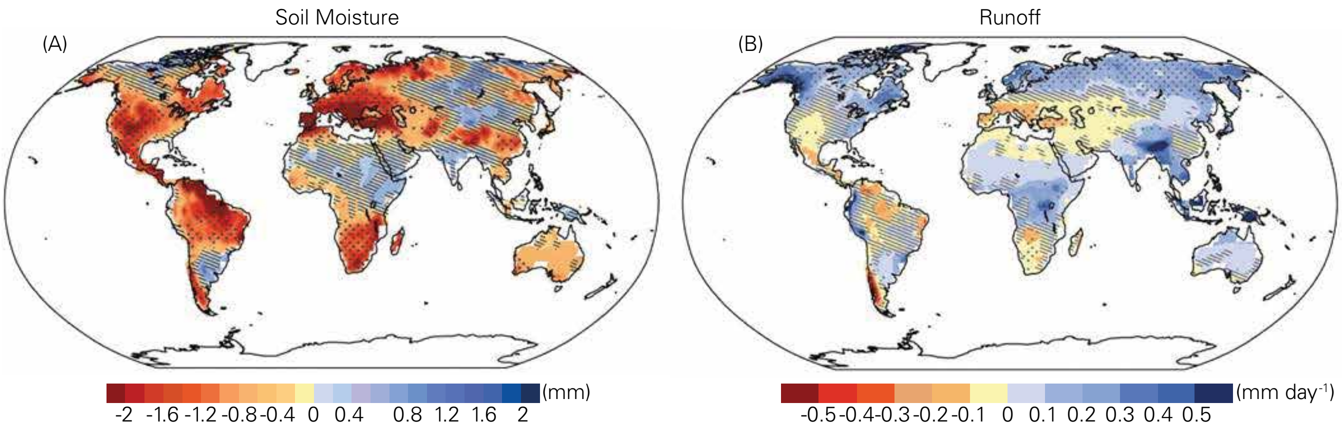

Figure 2.8—Change in annual mean (A) soil moisture (mass of water in all phases in the uppermost 10 cm of the soil) (mm), and (B) runoff relative to the reference period 1986–2005 projected for 2081–2100 under RCP 8.5 from the Coupled Model Intercomparison Project Phase 5 (CMIP5) ensemble. Hatching indicates regions where the multi-model mean change is less than one standard deviation of internal variability. Stippling indicates regions where the multi-model mean change is greater than two standard deviations of internal variability and where at least 90 percent of models agree. RCP means Representative Concentration Pathway. (From figs. 12.23 and 12.24 of IPCC 2013).

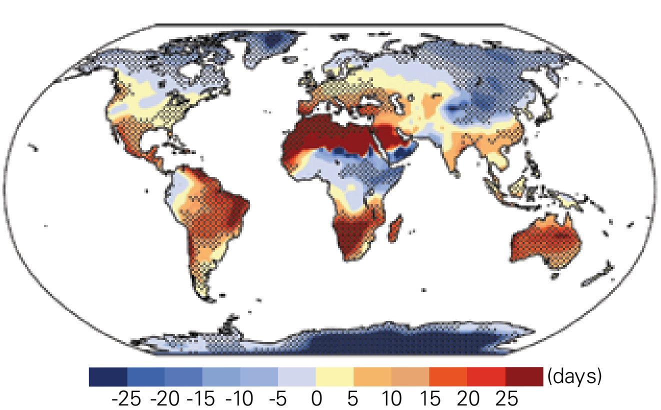

Figure 2.9—Projected change in the annual maximum number of consecutive dry days when precipitation is <1 mm, over the 2081–2100 period in the RCP 8.5 scenario (relative to the 1981–2000 reference period) from the Coupled Model Intercomparison Project Phase 5 (CMIP5) models. Stippling indicates gridpoints with changes that are significant at the 5-percent level using a Wilcoxon signed-ranked test. RCP means Representative Concentration Pathway. (From fig. 12.26 of IPCC 2013; updated from Sillmann and others 2013).

WHAT DOES RCP MEAN?

Representative concentration pathways (RCP) is a shorthand way of saying, *let us assume that the concentrations of anthropogenic greenhouse gases reach a certain level as indicated by the additional radiative forcing they cause by 2100*. Four levels were considered in the Intergovernmental Panel on Climate Change Fifth Assessment Report (IPCC 2013): 2.6, 4.5, 6, and 8.5 W/m^2 (Watts per square meter). Each "pathway" has a slightly different trajectory in greenhouse gas concentrations (and consequent forcing) over time, with the RCP 2.6 scenario, for example, showing higher forcing at mid-century and recovering toward the end of the century. The wattages associated with each RCP level are strictly related to the anthropogenic greenhouse gases, and increased water vapor that comes as feedback to the initial forcing may multiply the increased longwave radiation effect three- to four-fold. Thus, an RCP of 8.5 could yield a net increase in downwelling longwave radiation of 34 W/m^2.

2012, Trenberth and others 2014). Applicability of this index to future conditions and their consequences will depend in part on how the various components of the PDSI relate to the phenomenon of interest. For instance, applicability may vary depending upon whether forest die-off is more related to shallow soil versus how long it has been since precipitation has occurred. Most of the projections of increasing PDSI values in the United States are associated with increases in temperature, although an increase in the number of consecutive dry days during monsoon season may be important for areas affected by the North American Monsoon System (IPCC 2013). There is again the concern of double accounting of energy (first for heating and then for evapotranspiration) in using a temperature-based potential evaporation model (Milly 1992). This is increasingly well known for applications of the original formulation of the PDSI, which uses temperature explicitly through the Thornthwaite (1948) evaporation model. 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 (Cook and others 2014a, Sheffield and others 2012), implicitly carries a strong temperature dependence in the calculation of the vapor pressure deficit term as well. Distinguishing between contributions in future drought caused by increased evapotranspiration rates versus precipitation lapses may be helpful in interpreting this kind of work.

Projections of Relevant Factors

Temperatures are expected to increase 4–7 °C across the continental United States, with stronger increases in the interior than near the coasts. Summer relative humidity is expected to drop in the neighborhood of 4–8 percentage points, with weakest declines in the Southwest, where summer humidity is already low. 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.

Winter precipitation (DJF) increases on the order of 0–10 percent (with large differences among models) are expected over most of the continental United States, except for the Southwest where declines of 0–10 percent are projected. Alaska shows increases of 10–50 percent, increasing with latitude. Hawaii has a minor and uncertain decline. Summer precipitation (JJA) is projected to decline (0–20 percent) over most of the continental United States, except for the East and Gulf Coasts where 0–10 percent increases are projected. Alaska shows increases of 0–20 percent, increasing with latitude. Hawaii has a minor and uncertain increase.

Duration of dry spells is expected to increase in western U.S. mountains, where less snowpack accumulation and earlier melt combine to extend the dry summers (Barnett and others 2008, Westerling and others 2006). Some areas dependent on rainfall for moisture will also see increased dry-spell length. The maximum number of consecutive dry days (precipitation <1 mm) in a year is not projected to change substantially over most of the United States, except in the Southwestern United States (AZ, NM, TX) and Pacific Northwest (WA, OR, ID), which may see 5- to 10-day increases in dry-spell duration. In the Southwestern United States, the increases are expected to occur in summer months in association with changes in the North American Monsoon.

Teleconnections from tropical SST patterns are a primary control on drought occurrence in the United States (Dai 2011, Rajagopalan and others 2000). Despite substantial intermodel variability in projections of ENSO, it is expected to continue to be the dominant mode of climate variability. The interannual variability driven by ENSO provides some insights into future drought, insofar 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 and others 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 percent is expected across most of the continental United States, except the Southwestern United States where a decline in variance is expected (Seager and others 2012). Increases in interannual variability on the order of 30–40 percent are expected in Alaska.

Challenges for Interpreting Projections

Does a projected trend in precipitation or temperature portend a trend in drought? There are two important aspects of this question: (1) the time scale associated with the changes, and (2) whether the trends affect those extremes in weather that we term "drought." The first point is mostly one of an appropriate datum from which to measure a shift or departure. If the future of a location is drier, does it also represent an increased drought condition? If we are conceptually contrasting two time periods as two separate ecological equilibriums, simply being drier does not necessarily imply increased drought conditions or severity if the water "needs" of the new landscape shift in conjunction with the availability. However, during the transition, as natural and human communities adjust to the shifting dryness, increased "droughtiness" is a possibility. For example, one could ask, "Is Nevada a dry place, or has it been in a drought since the end of the Pleistocene?" Sagebrush and bristlecone pine researchers may, appropriately, have different answers. Similarly, increased moisture may not preclude a future with more severe drought episodes. For example, increased variance in precipitation may produce years that are extremely challenging to a moist-adapted landscape. While this question could be taken as one of semantics, it is more appropriately applied in a quantitative context by tying characteristic time scales of processes of interest (e.g., life spans, age to reproduction, seed durability) to drought as a transient phenomenon with its own time scales associated with duration and frequency (for a given intensity). Recognizing this aspect calls for greater precision in specifying the nature of the kind of drought of concern rather than just identifying a comparatively dry period in a time series as a drought without reference to time scales.

Coupled to this concern is the lack of knowledge of how interannual to interdecadal scale climate modes (e.g., teleconnections such as ENSO, PDO, AMO) might shift in response to increased atmospheric CO_2. There is evidence that the inability of GCMs to capture low-frequency modes of internal climate variability could lead to underestimation of risks of persistent drought (Ault and others 2014). Multiple years of drought may be more stressful to forests than single-year droughts, and the relative risks of reduced precipitation for several years to decades may not be well represented in GCMs.

The second point 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 and others 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 (Luce and others 2013, Luce and Holden 2009, Pagano and Garen 2005) (fig. 2.10). 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 (e.g., "Summary for Policy Makers" in Field and others 2012) (fig. 2.11). A shift in variance or mean could change the probability of exceeding a threshold or proceeding into novel weather (Field and others 2012, Jentsch and others

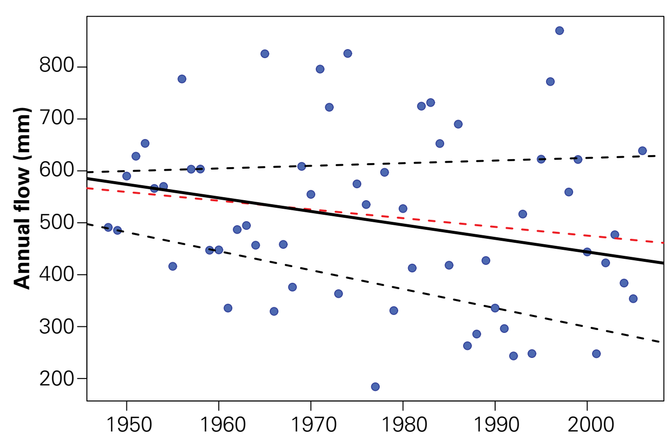

Figure 2.10—Annual runoff in the Boise River near Twin Springs, ID for water years 1948 to 2006. The dashed red line is the trend in, the mean annual flow, and the solid black line is the trend in median annual flow. The upper and lower black dashed lines are the 75th and 25th percentile annual flows, respectively. Note that while the wetter years show nearly no trend, drought years (the 25th percentile) have trended significantly, with about a 30-percent decline over the period of record. This shows a shift from a narrow distribution in the 1950s to a wide one in the 2000s (this would appear as (C) in fig. 2.11).

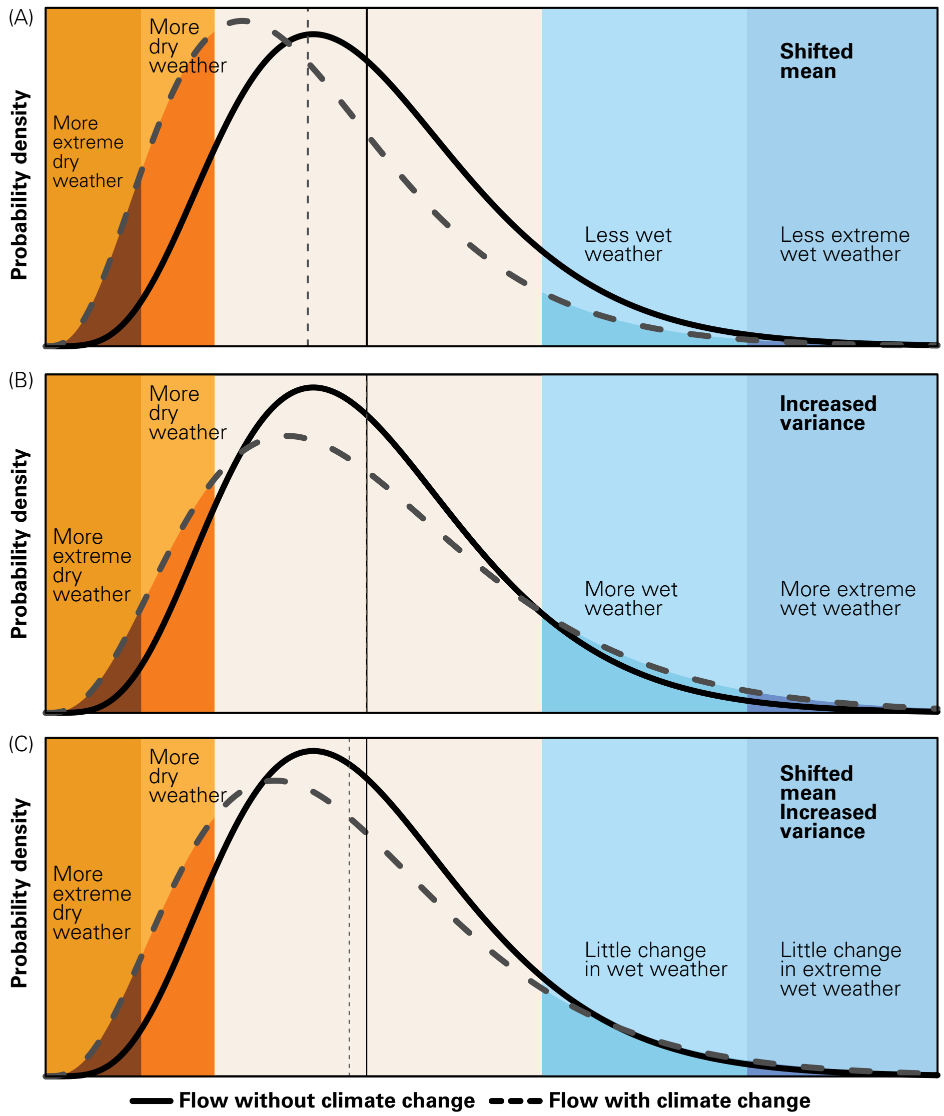

Figure 2.11—Different changes in runoff distributions between present and future climate and their effects on extreme values of the distributions: (A) effects of a simple shift of the entire distribution toward drier weather; (B) effects of an increase in runoff variability with no shift in the mean; (C) effects of an altered shape of the distribution, in this example a shift in mean weather toward drier conditions with an increase in variance. Vertical lines are means. [After fig. SPM.3 in Field and others (2012); also see Anderegg and others (2013)].

2007). 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 critical 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 (Dale and others 2001, Easterling and others 2000, Jentsch 2007).

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 and others 2010). Only a few GCMs accurately recreate the ENSO pattern, a driver of interannual scale variability in weather across much of the World (IPCC 2013, Seager and others 2012). 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 (e.g., figs. 2.7–2.9). 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 and others 2004). For example, interannual variability in Variable Infiltration Capacity (VIC) hydrologic projections (Vano and others 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 (Blöschl and Montanari 2010, Johnson and Sharma 2009). 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 and others 2011). While GCMs can model general circumstances of temperature, temperature stratification, and vapor 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 solution 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., Gao and Sorooshian 1994, Pitman and others 1990), reflecting the general scale-related issue that it is almost always raining somewhere within a GCM cell, but it is usually a fairly small proportion of the area experiencing precipitation. As an addition to the problem, 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 have higher spatial resolution (Rasmussen and others 2011, Salathé and others 2010).

Projections of the key climate phenomena feeding moisture to the continental United States in the summer—the North American Monsoon System (NAMS) and North Atlantic tropical cyclones—are uncertain (IPCC 2013); however, there appears to be a tendency toward drier conditions, according to the climate models with the strongest historical performance (Maloney and others 2014, Sheffield and others 2013). The most consistent projection for NAMS relevant to drought is an increase in the number of consecutive dry days by 15–40 percent (interquartile range). In the context of a warming future, an increase in the time between precipitation events could have substantial ecological importance (Adams and others 2009).

Applying Drought Projections To Predict Impacts

The overall consensus on drought projections is that there is a great deal of uncertainty, primarily because of uncertainty in projecting future precipitation; however, drought projections can be useful when placed in the proper context. Drought is a derived quantity with dimensions of severity, frequency, scale, and organization. It can be thought of as a collection of "extreme" events that must occur simultaneously to create a situation of concern, such as tree mortality or water scarcity. There is some sense that drought will intensify faster (or alternatively, drought effects will manifest more quickly during a dry spell) than it has in the past (Trenberth and others 2014). For example, "global-change-type drought" infers that drought is occurring in the context of much warmer weather (Adams and others 2009, Allen and others 2010) and, hence, the effects on forests are greater. When combined with projections of longer interstorm periods in some locations, impacts on terrestrial ecosystems could be substantial.

Summary

Historical and paleoclimatic evidence clearly shows that the nature of drought has always changed and continues to change. The direction of trends in recent history vary from region to region, with the western half of the United States indicating broadly drier conditions while the eastern regions show broadly wetter conditions. Although these patterns are often used to generalize about changing drought regimes (e.g., rarer, more frequent, more severe, or unchanging), individual findings can conflict with one another because the definitions or quantification methods for "drought" differ among many studies (Seneviratne and others 2012).

The relationship between forested landscapes and drought may differ substantially from that of the engineered or agricultural landscape. In particular, water supply thresholds may exist for particular crops or industries. In contrast, forest ecosystems often have mechanisms of resilience to drought, such that their ultimate response to drought may not relate to the *initial perturbation* after a drought event, but rather the context of that event and others like it in time and across landscapes and stream systems. While there is a well-developed science on how various biota respond to individual events, it is this longer term consequence that will have a lasting effect on species distributions.

Droughts severe enough to affect forest ecosystems may be driven primarily by periodic deficits in precipitation, as opposed to changes in potential evaporation (e.g., Abatzoglou and Kolden 2013, Holden and others 2012). Although incoming radiation is adequate in most places in the continental United States to evaporate a substantial fraction of the annual water balance (see fig. 2a in Sankarasubramanian and Vogel 2003), variation in annual precipitation is the dominant driver of variations in annual water balance (Milly and Dunne 2002). On longer time scales, however, the availability of warmer and drier air may more broadly shift many places to more consistently arid conditions. If drought is considered from the viewpoint of a global/absolute threshold, as might be the case for a particular plant or an engineered system, this broad shift could presage a drought from the point of view of system tolerance. However, if drought is defined in terms of variability over time (given a particular time-averaging window), which has substantial relevance to self-regulating or temporally dynamic systems, increasing potential evapotranspiration pressure primarily has relevance to the rapidity of drought onset after precipitation stops. This variation in framing has been a central issue in debates on the kinds of metrics to apply in evaluating trends in droughts and making projections.

The future of drought is uncertain. One aspect of uncertainty is that precipitation projections in GCMs are fairly poor. A second aspect of uncertainty is that persistently wetter or drier conditions do not necessarily reflect drought risks in natural systems the same way that they might be observed in engineered or agricultural systems. In natural systems, there is adjustment to gradual shifts in means or even regimes; however, variability on the order of a few years to a few decades can have substantial impacts on ecosystems. These intermediate time scales pose the greatest challenge for GCM projections. Improving our understanding of temporal trends in teleconnections and climate indices may be a fruitful alternative in understanding future variations at these intermediate time scales.

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