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CLIMATE CHANGE AND DROUGHT (Q FU, SECTION EDITOR)

Climate Change and Drought: From Past to Future

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Abstract

Drought is a complex and multivariate phenomenon influenced by diverse physical and biological processes. Such complexity precludes simplistic explanations of cause and effect, making investigations of climate change and drought a challenging task. Here, we review important recent advances in our understanding of drought dynamics, drawing from studies of paleoclimate, the historical record, and model simulations of the past and future. Paleoclimate studies of drought variability over the last two millennia have progressed considerably through the development of new reconstructions and analyses combining reconstructions with process-based models. This work has generated new evidence for tropical Pacific forcing of megadroughts in Southwest North America, provided additional constraints for interpreting climate change projections in poorly characterized regions like East Africa, and demonstrated the exceptional magnitude of many modern era droughts. Development of high resolution proxy networks has lagged in many regions (e.g., South America, Africa), however, and quantitative comparisons between the paleoclimate record, models, and observations remain challenging. Fingerprints of anthropogenic climate change consistent with long-term warming projections have been identified for droughts in California, the Pacific Northwest, Western North America, and the Mediterranean. In other regions (e.g., Southwest North America, Australia, Africa), however, the degree to which climate change has affected recent droughts is more uncertain. While climate change-forced declines in precipitation have been detected for the Mediterranean, in most regions, the climate change signal has manifested through warmer temperatures that have increased evaporative losses and reduced snowfall and snowpack levels, amplifying deficits in soil moisture and runoff despite uncertain precipitation changes. Over the next century, projections indicate that warming will increase drought risk and severity across much of the subtropics and mid-latitudes in both hemispheres, a consequence of regional precipitation declines and widespread warming. For many regions, however, the magnitude, robustness, and even direction of climate change-forced trends in drought depends on how drought is defined, with often large differences across indicators of precipitation, soil moisture, runoff, and vegetation health. Increasing confidence in climate change projections of drought and the associated impacts will likely depend on resolving uncertainties in processes that are currently poorly constrained (e.g., land-atmosphere interactions, terrestrial vegetation) and improved consideration of the role for human policies and management in ameliorating and adapting to changes in drought risk.

Keywords Drought · Climate change · Paleoclimate · Detection and attribution

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Introduction

Droughts are among the most expensive natural disasters in the world [1], with significant costs to ecosystems [2–4], agriculture [5, 6], and human societies [7–10]. Climate change is expected to increase drought frequency and severity over much of the globe [11–13], especially in semi-arid regions already experiencing significant water stress [14, 15]. Improving our understanding of drought dynamics and impacts in the context of climate change is thus a critical area of climate research, especially in light of the recent devastating droughts in California [16], the Mediterranean [10], and elsewhere [17, 18].

Comprehensive discussions of drought and climate change can be difficult, however, because drought is fundamentally a cross-disciplinary phenomenon extending across the fields of meteorology, climatology, hydrology, ecology, agronomy, and even sociology, economics, and anthropology. Drought is broadly defined as an *anomalous moisture deficit relative to some normal baseline*, but is more precisely classified based on where in the hydrologic cycle these moisture anomalies occur [19] (Fig. 1, adapted and modified from Van Loon [20]). Droughts often begin as precipitation deficits (*meteorological drought*), propagating over time (typically days to years) through the hydrologic cycle to affect soil moisture (*agricultural drought*) and then runoff, streamflow, and storage in aquifers and surface reservoirs (*hydrological drought*). For agricultural and hydrological drought, additional processes at the land surface and the land-atmosphere interface have the capacity to intensify or ameliorate precipitation-forced drought anomalies. Warmer temperatures, for example, can increase evaporative demand in the atmosphere and moisture losses from the surface, increase the fraction of precipitation falling as rain rather than snow, and advance the timing of the snow melt season in the spring. Vegetation (e.g., phenology, land cover) and land surface properties (e.g., soil type, topography) can also affect the manifestation of droughts through the soil water holding capacity, the efficiency of runoff generation, and the partitioning of energy and moisture fluxes at the surface. More recently, the influence of human institutions and behavior on both societal vulnerability and physical drought dynamics is being increasingly recognized [21]. Irrigation of croplands, groundwater withdrawals, and water conservation policies can all affect the cost of droughts to societies, how quickly droughts propagate through the hydrologic cycle, the severity of impacts on ecosystems and agriculture, and how quickly natural systems can recover [20–23].

There are large uncertainties, however, in how many drought processes will respond to climate change (e.g., precipitation [24]) and their importance for modulating drought variability (e.g., vegetation [25, 26]). Model projections can also be difficult to constrain because quantitative comparisons across models, historical observations, and the paleoclimate record are not always straightforward [27], and because of the paucity of direct long-term observations

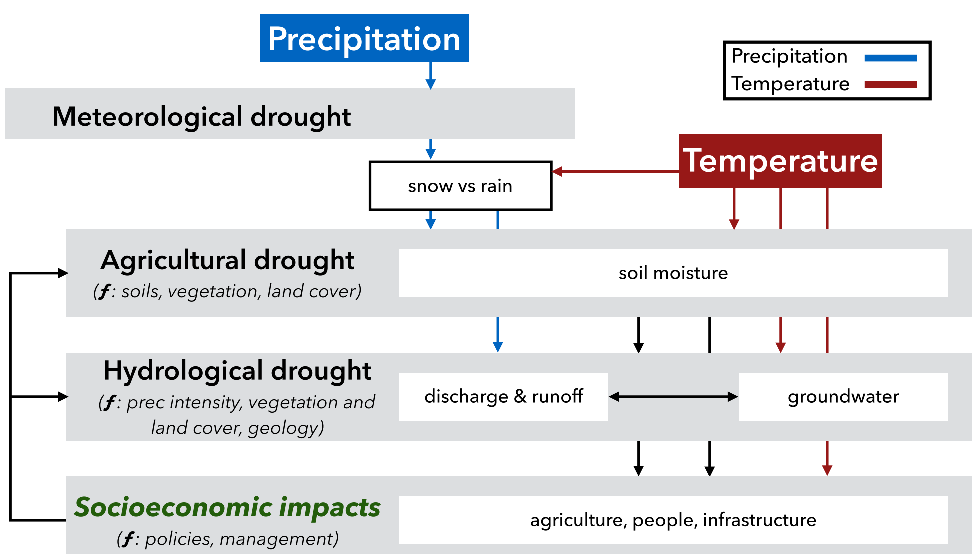

Fig. 1 Classical definitions of drought and the associated processes. Precipitation deficits are the ultimate driver of most drought events, with these deficits propagating over time through the hydrologic cycle. Other climate variables, however, can also affect both agricultural and hydrological drought. High temperatures, for example, can amplify soil moisture drought during the growing season by advancing snowmelt and increasing evapotranspiration. Non-climatic factors, such as land cover and soil type, can also influence drought by affecting the surface partitioning of precipitation (e.g., infiltration, runoff, interception) and modulating moisture fluxes between the surface and the atmosphere. Human activities (intentionally or not) can also exert a significant influence, either mitigating or exacerbating drought conditions and impacts (e.g., increasing hydrological drought through groundwater withdrawals, mitigating drought impacts by irrigating). Adapted and modified from Van Loon [20]

for many drought indicators (e.g., soil moisture, runoff). Further, the degree to which forced climate signals and trends in drought events can be detected is strongly influenced by natural climate variability, which in the short term can amplify or dampen forced trends and signals [28]. Reconciling across these disparate perspectives and uncertainties is challenging and in the past often resulted in divergent conclusions regarding climate change contributions to drought, even for the same event. The extent to which the recent California drought has been attributed to climate change, for example, depends largely on whether the studies in question focus on precipitation [29] or soil moisture [30–32].

In part due to these challenges and uncertainties, the Fifth Assessment Report of the Intergovernmental Panel on Climate Change (AR5) found only *low confidence* in the detection of global-scale trends toward increased drought and the attribution of said trends to anthropogenic climate change [33]. In the years since the AR5 was published, however, there have been steady advancements in our understanding of drought dynamics and the associated physical processes. These insights have been generated through further development of the paleoclimate record, new analyses of recent and historical drought events, and the widespread use and interrogation of climate models. Here, we review some of the major improvements in our understanding of drought dynamics since the AR5. We also highlight major remaining knowledge gaps and uncertainties that must be addressed to continue advancing our understanding of drought and climate change.

Drought in the Paleoclimate Record

The relatively short duration of the historical record (typically 150 years or less) severely limits our understanding of natural climate variability, especially for extreme events such as droughts which are by definition rare and therefore undersampled in modern observations. These constraints present challenges in our ability to attribute droughts to particular causes, including both internal variability and external forcing. Hydroclimate reconstructions developed from proxy records (e.g., tree-rings, sediment cores, speleothems) are critical tools for addressing this weakness by extending the historical record further back in time. Climate model simulations using paleoclimate and pre-industrial forcing histories (e.g., solar, volcanos, land cover) have also proven valuable for investigating the dynamics underlying paleodrought events, especially when these models are analyzed in tandem with empirical reconstructions. From a climate change perspective, reconstructions and model simulations of the last two thousand years (the Common Era) are particularly useful because this interval has some of the most detailed information on hydroclimate variability in the paleoclimate record and is most representative of modern era climate dynamics.

One major innovation emerging from tree-ring-based reconstructions has been the development of "drought atlases," annually resolved spatiotemporal reconstructions that target a soil moisture drought indicator (the Palmer Drought Severity Index; PDSI). The first two atlases were developed for North America in 2004 [34] and Monsoon Asia in 2010 [35], but more recently, new reconstructions for Europe and the Mediterranean [36], Australia and New Zealand [37], and Mexico [38] have become available. By constraining the spatial, as well as temporal, patterns of drought variability, the drought atlases independently capture the spatiotemporal fingerprints of the most important global teleconnection patterns with high fidelity [39, 40]. As an example, Fig. 2 shows composite average anomalies of PDSI from several drought atlases [35–37, 41] for strong phases (sea surface temperature anomalies in the NINO 3.4 region ±1 K) of the El Niño Southern Oscillation (ENSO) in the historical record. The expected global patterns of anomalous drought and wetness associated with ENSO are well-resolved (e.g., anomalous wetness associated with El Niño in Southwest

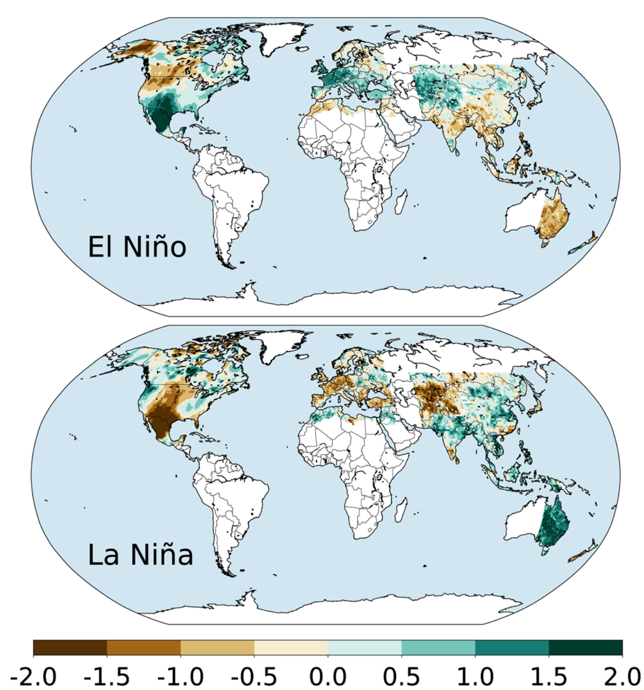

Fig. 2 Composite average drought anomalies (Palmer Drought Severity Index) from the North America, Old World, Monsoon Asia, and Australia-New Zealand drought atlases for strong El Niño and La Niña events. These highly resolved, spatiotemporal reconstructions of drought have proven to be extremely valuable for investigating drought and climate dynamics over the Common Era. Their utility is, however, limited by the targeted variable (a single soil moisture indicator) and their geographic extent, especially in the Southern Hemisphere

North America and La Niña in Eastern Australia), despite the fact that each regional reconstruction uses a mostly independent suite of tree-ring records. A robust signature of other climate modes is also apparent in the drought atlases [39, 40], including the North Atlantic Oscillation (NAO), Atlantic Multi-decadal Oscillation (AMO), and Pacific Decadal Oscillation (PDO), further highlighting the value of drought atlases for dynamical investigations.

Because they resolve these dynamics, the drought atlases have been especially valuable for investigations into the causes of the North American megadroughts. These multi-decadal drought events were clustered in time during the Medieval Climate Anomaly (MCA) and were more persistent than any drought of the last several hundred years [42, 43]. The megadroughts are clearly recorded within the North American Drought Atlas [41, 44], and they have been attributed to internal atmospheric variability [45–47], ocean forcing [48–52], and land-atmosphere interactions [53]. Recent studies using the drought atlases, however, have provided new evidence for persistent ocean forcing from the tropical Pacific and Atlantic ocean basins during some of the megadroughts. Coats et al. [54] demonstrated that coupled ocean-atmosphere model simulations of the last millennium from Phase 5 of the Coupled Model Intercomparison Project (CMIP5) were capable of producing megadroughts over the US Southwest. For models with stationary and realistic teleconnections between the tropical Pacific and North America, these megadrought periods were typically associated with decadal and longer-term shifts toward more frequent cold sea surface temperatures (SSTs) in the eastern tropical Pacific, a pattern associated with modern droughts in the region [55, 56] (e.g., bottom panel of Fig. 2). Additional support was found in a subsequent study using drought atlases from North America, Europe, and Monsoon Asia to infer the likely state of various climate modes (including ENSO) during megadroughts in the Southwest [40]. In that study, the authors found a consistent association between the occurrence of Southwest megadroughts and cold ENSO conditions, as well as modest evidence for a warm phase of the AMO during the 12th and 13th centuries, which also would have predisposed the region toward drought. Using the North American Drought Atlas and a time series modeling approach, Coats et al. [57] demonstrated that megadroughts can be generated for the western USA in the absence of any exogeneous forcing, arguing that they were likely a product of internal ocean-atmosphere variability. However, they noted that the mean drying shift during the Medieval Climate Anomaly was likely a necessary condition for the temporal clustering of megadroughts during this interval and hypothesized that such a shift could occur as a result of a centennial scale warm phase of the AMO. Additional support for this was demonstrated in Ault et al. [48], which used a linear inverse model to test the timing and characteristics of North American megadroughts against an internal climate variability null hypothesis conditioned on twentieth century observations. They found that they could not reject this unforced null hypothesis for duration, spatial scale, and magnitude of these megadroughts, but that the observed MCA clustering was not captured by their model. These findings suggest that while the characteristics of individual megadrought events could have arisen purely as a result of internal climate system variability, their clustering during the MCA may have been caused by external radiative forcing or another aspect of climate system variability not captured by the null generating process.

Tree-ring data are exactly dated and widespread over the midlatitudes, which allows them to be statistically compared and calibrated with the corresponding overlapping instrumental observations in both space and time. To date, however, the drought atlases have not been extended to South America or Africa, in large part due to the paucity of annually resolved and drought-sensitive tree-ring records available from these regions [27, 58]. With a few notable exceptions, tree-ring chronologies also only cover the last several hundred or thousand years. Even in those regions where moisture-sensitive tree-ring chronologies are abundant, drought atlases may have other limitations, including that they thus far target a single category (soil moisture) and metric (PDSI) of drought, and have seasonal biases imposed by the targeted variable (PDSI), local climate, and tree biology [59–61]. To extend drought reconstructions further into the past, target other hydroclimate variables, and expand reconstructions into regions where tree-ring chronologies are sparse or non-existent requires the use of additional and different paleoclimate archives. These include multiple proxy observations from lake and marine sediments, cave formations (speleothems), ice cores, and corals [27]. One region with a paucity of tree-ring records, but of particular concern, is east Africa, where historical drought is linked to severe disruptions of agriculture and food security but where models of future hydroclimate under anthropogenic greenhouse gases simulate wetter conditions [62]. Tierney et al. [63, 64] identified coherent but spatially heterogenous patterns of decadal-scale drought variability in eastern Africa using a diverse set of lake sediment proxies—organic and inorganic molecules, sediment layers, and charcoal—linking these modes of variability to low-frequency Indian Ocean sea surface temperature variability. Tierney et al. [62] combined marine sediment proxies for past rainfall and temperature over the Horn of Africa with climate model data to demonstrate that past warm periods were actually drier, raising questions about projections of wetter conditions in the Horn of Africa due to global warming.

As with tree-rings, however, these other proxies present their own challenges. While sediment and speleothem records potentially provide a multi-millennial perspective on hydroclimate, unlike tree-rings they rarely have annual resolution and their dating is subject to the precision of the age information from radiocarbon and other geochronological techniques. These features make it substantially more difficult to characterize consistent patterns of drought over a region or to calibrate proxies against hydroclimate metrics during the instrumental period as well as compare to model simulations. Anchukaitis and Tierney [63] used a combined age modeling and data reduction approach to identify large-scale and spatially coherent patterns of drought variability over east Africa, even in the presence of chronological uncertainty. Other approaches likewise make use of ensembles of age-modeled and time-uncertain proxy data to isolate consistent signals of hydroclimate anomalies in time and space [65]. An additional challenge to time uncertainty is that surface (lakes) and sub-surface (caves) hydrology integrates multiple climate and geological processes across a range of frequencies that may confound a simple interpretation of these records. For instance, lake levels or speleothem isotope series can show large-magnitude decadal- or centennial-scale variability resulting from simple interannual precipitation variability, as the lake or cave system "filters" the interannual hydroclimate signal through non-climatic physical or geochemical systems that lag or accumulate anomalies and that then suggest greater low-frequency power than is present in the climate system itself [66, 67]. The potential for non-climatic signals to confound paleoclimate interpretation has motivated the use of proxy-system or forward models to interpret these proxy records [68]. The models attempt to simulate the processes that translate multiple environmental signals through a "sensor" (e.g., the cambium of a tree or the polyps of reef-building corals), the formation of the archive that retains the sensor's response to the environment (e.g., sediments in a lake or the wood of a tree), and the reflection of those signals by the proxy measured in the lab (e.g., ring width or a carbonate stable isotope ratio). This approach can be used to understand spectral biases imparted by the proxy system (like the lake or speleothem systems discussed above) and identify potential non-linearities or multivariate signals [69]. One recent promising development has been the use of these models in data assimilation approaches to paleoclimate reconstruction [70], where climate model simulations, proxy-system models, and proxies themselves are combined to develop estimates of past multivariate climate variability that capture and account for the non-climatic influences of proxy data of all kinds.

Beyond helping to better characterize and constrain the magnitude and processes of natural drought variability and contextualizing recent events and trends, the paleoclimate record also provides a unique and critical "out of sample" test for climate models [71]. Such tests are crucial trials because these models are used for projections of a future where forcings and boundary conditions will be much different from today and waiting several decades for direct validation belies their utility. Combined model-data comparisons have already generated significant insights into the dynamics of drought variability in the past [54], in the modern era [31, 72], and in the future [62, 73]. The benefits of such model-data comparisons can be refined and expanded by further advancing proxy-system models [68], which better constrain some of the uncertainties in the proxy reconstructions themselves and facilitate the expansion of model-data comparisons to new regions and proxies.

The Role of Climate Change in Recent Drought Events

The science of detection and attribution is concerned with identifying statistically significant changes in the climate system (*detection*) and the underlying causes (*attribution*), especially regarding the role of anthropogenic climate change [74–77]. While traditionally employed to investigate trends in climate variables like temperature [77] and precipitation [78, 79], these analysis frameworks have been extended to investigate specific climate events and extremes [80, 81], typically focusing on two questions [82]. First, has anthropogenic climate change increased the probability that a given climate event would occur? Second, has anthropogenic climate change affected the magnitude or intensity of this event? For drought, detection and attribution can be difficult because of the lack of long-term and high quality instrumental data for many drought variables [82] (e.g., soil moisture, runoff, and streamflow) and large natural variability that can make detection of a climate change signal difficult (e.g., [83]). Despite these challenges, however, there is strong emerging evidence that climate change has already begun to influence recent drought events in several regions.

Climate change projections over the Mediterranean indicate robust future declines in precipitation with anthropogenic greenhouse warming [15, 84], and observations in this region show a negative precipitation trend over the twentieth century. Hoerling et al. [85] were the first to demonstrate that this long-term trend cannot be reconciled with natural climate variability alone, and is therefore likely forced by anthropogenic warming. Gudmundsson et al. [86] further concluded that this drying trend has already significantly increased the risk of meteorological drought in the Mediterranean. This was demonstrated more specifically for a drought in the eastern Mediterranean and Levant from 2007–2010 [10, 87], where Kelley et al. [10] concluded that a 3-year precipitation drought of equivalent magnitude to 2007–2010 in the region was three times as likely because of anthropogenic climate change. The paleoclimate record supports the exceptional nature of this recent drought, suggesting that the 1998–2012 period within which the Levant drought occurred was likely the driest 15-year period in the region of the last 500–900 years [72]. Kelley et al. [10] also concluded the 2007–2010 drought was a likely contributing factor to the Syrian civil war, though the exact role of climate change and the drought in this conflict remains an area of intense debate and uncertainty [88–91].

Evidence for a climate change signal can also be seen in the recent drought that affected California from 2011–2016 [30, 92], an event characterized by severe deficits across the hydrologic cycle in precipitation [93], snowpack [94], surface reservoir storage [16], and groundwater [95]. Most studies have concluded that the precipitation deficits were dominated by natural variability [29, 96]. Paleoclimate reconstructions do indicate, however, that cumulative precipitation deficits for some regions of California may be unprecedented relative to the last 400 years [97], and other studies have suggested that the frequency of occurrence of circulation patterns associated with drought in California [98] will increase with climate change [92, 99, 100]. Climate change connections to the precipitation deficits and associated circulation patterns for this specific drought, however, remain highly speculative. Instead, a climate change signal most clearly emerges through the direct impact of warming temperatures [30] on evaporative demand and snow cover. Griffin and Anchukaitis [31] investigated this by developing independent reconstructions of precipitation and soil moisture for California for the last 1200 years. They demonstrated that the accumulated (2012–2014) and single year (2014) soil moisture deficits (as indicated by PDSI) during this period were the most severe short-term droughts in the reconstruction. These soil moisture deficits were also significantly lower than would have been predicted from precipitation anomalies alone, pointing to warming as a likely important amplifier of the soil moisture drought. Additional evidence comes from Williams et al. [32], who estimated soil moisture variability back to 1895 using PDSI calculated from observations. As part of their analysis, they generated a counterfactual PDSI record with the anthropogenic trend in temperature removed. By comparing the two (PDSI calculated with and without observed warming), the authors concluded that anthropogenic warming significantly contributed to the drought by increasing evaporative losses, accounting for 8–27% of the observed drought anomaly in 2012–2014 and 5–18% of the anomaly in 2014. Warming also likely affected snow cover and melt in the Sierra Nevada Mountains, a critical source of recharge for surface reservoirs in California [101]. Spring snow water equivalent was at a record low across the Sierras in 2015 in both satellite [102] and surface [94] observations and may have been the lowest of the last 500 years [101]. Using regional climate model simulations alternatively including or excluding anthropogenic forcing, Berg and Hall [103] concluded that while snow cover would have been low regardless because of the precipitation deficits during the 2011–2015 drought, anthropogenic warming likely further reduced snowpack levels by 25% overall and by as much as 26–43% at middle and lower elevations in the Sierra.

Beyond California, other areas of Western North America have also experienced significant drought in recent years. Precipitation in the Southwest has been below average since the early twenty-first century [104]. As with California, these precipitation deficits appear most closely linked to natural variability, in this case because of an extended period of cold SSTs in the eastern tropical Pacific [28, 105]. In other regions, however, there is a clear warming signal in recent drought events. In 2015, the same snow drought that affected California extended across the Pacific Northwest and other parts of the west [106], with over 80% of observing stations west of 115° W reporting record low snowpack [94]. In the Pacific Northwest, this snow drought occurred despite near normal cold season precipitation [107]. Instead, it was driven almost entirely by record warm temperatures across the region [107], leading Mote et al. [94] to conclude (using an ensemble of regional climate model simulations) that anthropogenic warming was a significant contributor to the 2015 Pacific Northwest drought.

The Colorado River Basin has also experienced significant drought conditions since 2000 [108]. Focusing on Colorado River streamflow, Woodhouse et al. [109] compared this most recent drought period (2000–2012) against similar magnitude droughts in the 1950s (1950–1956) and 1960s (1959–1969). They found that while both the 1950s and 1960s droughts were linked to significant precipitation deficits, precipitation in the basin was near normal in the 2000s and this latest drought was likely driven by the much warmer temperatures. Udall et al. [110] subsequently concluded that historical warming of 0.9 °C has reduced Colorado River flow by 2.7–9%, which would account for roughly one third of flow losses during the 2000–2014 drought in the basin. McCabe et al. [111] found similar effects of warming on streamflow in the Upper Colorado River Basin, attributing reductions in streamflow of 7% over the last three decades to increased evapotranspiration and snowmelt from warming in the spring and summer (April–September).

The Millennium Drought affected Eastern Australia from the late 1990s through the early 2000s [112]. Precipitation anomalies during this event have been closely linked to natural SST forcing [113, 114], with El Niño accounting for about two thirds of the precipitation deficits in Eastern Australia [112]. There is some evidence that climate change may have enhanced the precipitation deficits through poleward storm track shifts and subtropical drying [115–118] and amplified surface drying through increased evaporative demand [113, 119, 120]. Others, however, have argued that high temperatures were a response—rather than contributing factor—to the drought because of reduced evaporative cooling [121]. The exact role of anthropogenically forced precipitation anomalies is also uncertain [112]. These uncertainties are further highlighted by disagreements regarding the ranking of the Millennium Drought in paleoclimate reconstructions over the region. Gallant and Gergis [122] concluded that streamflow in the Murray-Darling basin from 1998–2008 during the Millennium Drought was the lowest since 1783. Similarly, Gergis et al. [123] found that 1998–2008 was likely the driest decade in terms of precipitation deficits back to 1788. More recently, however, Cook et al. [124], using a reconstruction of summer season soil moisture, found the Millennium Drought to be well within the bounds of natural variability over the last 500 years. As in analyses of other regions (e.g., California), this reinforces how interpretations of drought and climate change can critically depend on the drought variables, time periods, and reconstructions being analyzed.

Various regions of Africa have also experienced severe droughts in recent decades. A drought in the Sahel region of West Africa persisted from the 1970s through the 1990s [125, 126], with significant impacts to people and ecosystems across the region [9]. While this drought was initially ascribed to desertification and poor land use practices [125], it has now been more definitively linked to anomalous SST forcing with an intrinsic anthropogenic component [127]. Cooler conditions in the North Atlantic (due to high levels of anthropogenic aerosol emissions [128, 129]) and a warmer Indian Ocean (attributed primarily to anthropogenic greenhouse warming [130]) both contributed to the Sahel drought [127]. Models forced with historical aerosol and greenhouse gas emissions or forcings faithfully reproduce the pattern of SST evolution and the resulting Sahel drought, though with a diminished magnitude compared to observations [131, 132]. East Africa is another drought-prone region, but one with complex topography and rainfall seasonality [133] and where many climate models perform poorly at reproducing observed hydroclimate [134]. While there is some evidence that warming conditions can exacerbate drought in the region [62, 135], to date, an anthropogenic signal cannot be attributed to recent droughts in the region with any confidence [83, 135].

In contrast to the long-standing literature on seasonal-to-multi-decadal drought, the concept of *flash drought* was introduced into the scientific literature relatively recently. This term is typically used to refer to soil moisture droughts that develop and intensify rapidly (especially over the summer), with often little or no advance warning [136, 137]. Despite their often short duration (subseasonal), such rapid onset and intensification can result in outsized negative impacts on agriculture and ecosystems, as was observed during the 2012 summer flash drought over the Central Plains of the USA [138, 139]. Flash droughts develop in response to both precipitation deficits [140] and high evaporative demand [141, 142], making them of particular interest from a climate change perspective. Over the twentieth century, the observed frequency of certain types of flash droughts has decreased over the USA [142]. In more recent decades, increased occurrence of flash droughts has been observed over China [143, 144] and Southern Africa [145], with the latter trend explicitly attributed to anthropogenic climate change. For most regions, however, additional work is needed to better quantify trends in flash drought risk and disentangle contributions from natural variability and anthropogenic forcing.

Uncertainties in Climate Change Projections of Drought Risk and Severity

Analyses of future drought risk and severity are based primarily on climate model simulations using different scenarios of anthropogenic forcing over the twenty-first century (e.g., the Representative Concentration Pathways, or RCPs, in the AR5). Climate change signals in recent drought events and trends are broadly consistent with model projections (e.g., [85, 107]), but many studies have arrived at divergent conclusions regarding the impact of climate change on future drought risk. Such disagreements have occurred even across studies analyzing the same set of model projections (e.g., the CMIP5 database). Here, we attempt to better reconcile these differing interpretations, including discussions of the most robust results and important underlying uncertainties.

One potential source of differences across studies, and one which is perhaps not highlighted enough in the literature, is the importance of how drought is defined when analyzing model projections. Figure 3 shows changes in a variety of drought indicators for the end of the twenty-first century (RCP 8.5 forcing scenario) from a 17-model ensemble drawn from the CMIP5 database. Clear differences in the robustness, magnitude, and even sign of projected changes across indicators are apparent, though some robust patterns clearly emerge. Foremost among the latter is that widespread surface drying in indicators of soil moisture (Fig. 3g, h) and runoff (Fig. 3b, f) cannot be predicted from much more localized precipitation declines (Fig. 3a) alone. This is perhaps most clearly demonstrated for Western North America, where significant precipitation declines are confined to Mexico and the extreme Southwest, but where robust patterns of surface drying affect a much

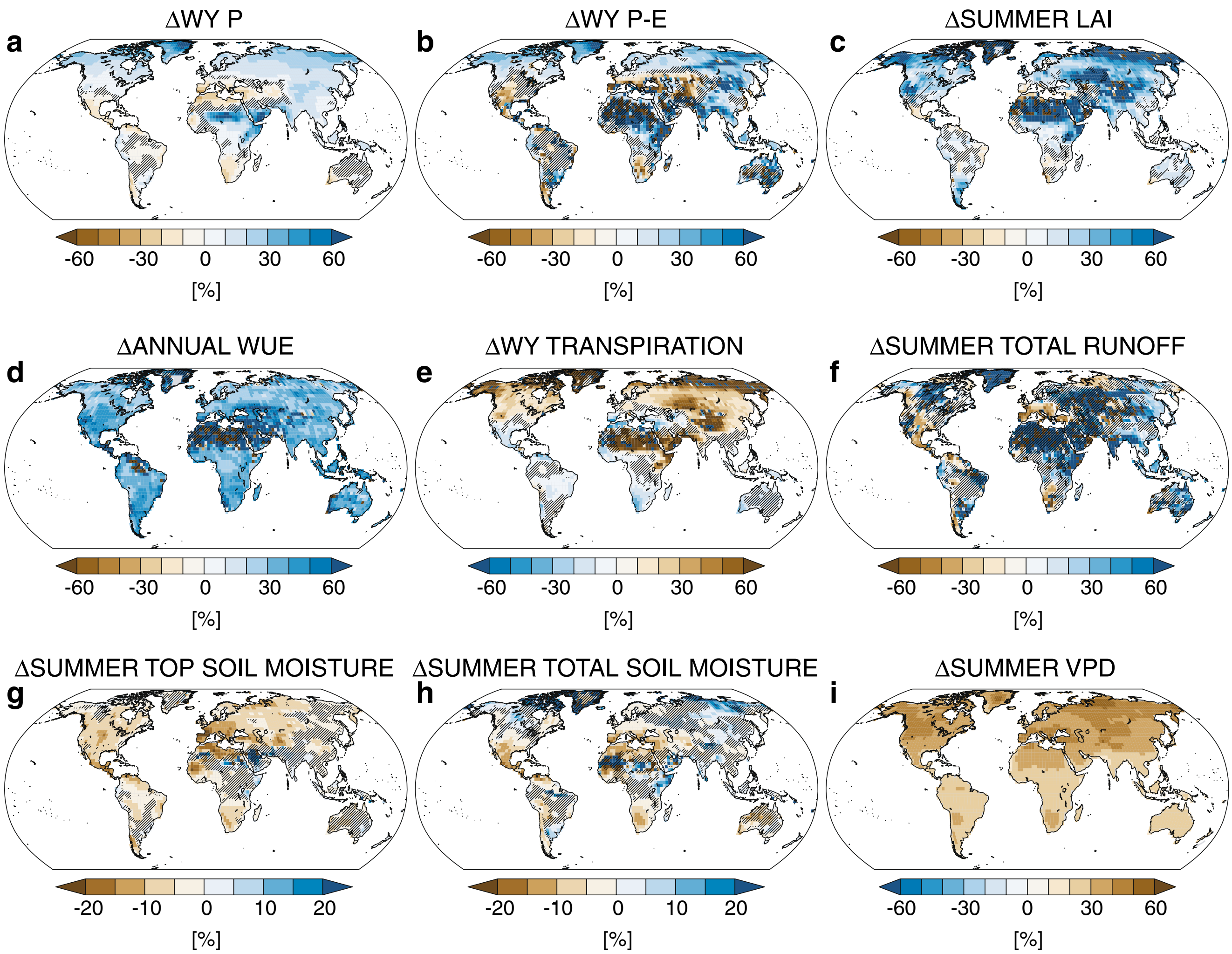

Fig. 3 End-of-century changes in hydroclimate variables from 17 models in the CMIP5 archive (2070–2099 minus 1976–2005) in **a** water-year (WY; October–September in the Northern Hemisphere and July–June in the Southern Hemisphere) precipitation (P), **b** WY precipitation minus evapotranspiration (P-E), **c** summer (June-July-August in the Northern Hemisphere; December-January-February in the Southern Hemisphere) leaf area index (LAI), **d** annual plant water use efficiency (WUE), **e** WY transpiration, **f** summer total runoff, **g** summer near surface soil moisture (0.1 m), **h** summer full-column soil moisture (note depth varies by model), and **i** summer vapor pressure deficit (VPD) all in percent (%). In all panels, drying tendencies are indicated in brown, wetting tendencies in blue (note the reverse color scales in **e** and **i**). Ensemble agreement is based on a pooled model-year K-S test (95%), with the additional requirement that at least two thirds of models agree with the direction of ensemble mean change. Hatched areas are insignificant

broader area that includes the Pacific Coast, the Montane West, and the Central Plains. These general patterns of widespread surface drying are broadly attributed to the more robust patterns of temperature versus precipitation change in the models. Warming in all seasons and regions amplifies surface drying by increasing evaporative losses as evaporative demand increases with the vapor pressure deficit (Fig. 3i) over land [11, 146]. Warming also reduces snow cover by increasing the fraction of precipitation falling as rain versus snow and increasing snow melt and sublimation [147–149].

Even at the surface, strong differences across drought indicators are also apparent. In the ensemble in Fig. 3, soil moisture drying (at the surface and throughout the soil column) is more robust and widespread compared to declines in hydrologic drought indicators represented by *P-E* and runoff. There are also clear differences in the pattern and intensity of near surface soil moisture drying versus the total soil column, with generally more robust and widespread drying in the former. This pattern has been noted previously and attributed to a greater sensitivity to warming-induced increases in evaporative demand near the surface versus deeper in the soil column [150, 151]. In some studies and regions, however, projected soil moisture drying at depth is amplified relative to the surface [73, 152] or near uniform throughout the soil column [25]. These differences between the response of various drought indicators to warming clearly highlights the complexity of future drought responses to climate change. Discussions of climate change and drought risk in future projections therefore need to carefully consider how drought is defined, and the degree to which any conclusions regarding future risk and severity depend on the drought indicators used.

The importance of indicators extends to another controversial topic within the climate change drought literature: the use of offline drought metrics calculated using output from the atmosphere component of the models. These include soil moisture indicators like PDSI and various metrics of aridity (e.g., the ratio of potential evapotranspiration to precipitation) often used to show increased drought risk and severity in the future [11, 153–156]. Such indicators are commonly employed in observational analyses because they allow for calculations of quantities for which direct observations are often not available (e.g., soil moisture) and have typically simpler data requirements compared to more sophisticated hydrologic models. One clear advantage of calculating such metrics from climate model projections is that it facilitates direct comparisons between observations, reconstructions, and models over past and future time intervals [54, 73]. There are two fundamental criticisms of these indicators, however, that have led some to conclude that using them may overestimate future drying. The first is that such offline calculations overestimate the direct impact of temperature on drought because embedded within temperature diagnostics is a temperature response to surface interactions [157, 158]. Temperatures are often warmer during droughts because of increased incoming solar radiation due to reduced cloud cover or increased sensible heating from declining soil moisture. Any temperature diagnostics used in the offline calculations thus conflate, to some degree, the temperature forcing and response and will likely overestimate the drying impact of externally forced warming trends. Second, such indicators have also been criticized for their simplicity compared to more complex, process-based models, including the typical land surface modules used in most climate models. This simplicity means that some important processes that may ameliorate surface drying (e.g., changes in plant water use efficiency with increased [CO_2]) are thus not represented.

While these criticisms are broadly valid, the actual degree to which these offline indices artificially amplify future drying trends, and are thus inferior to more process-based coupled models, is less clear. Milly et al. [159] found offline estimates of annual runoff declines were greater compared to runoff taken directly from climate models, though there was still good agreement between the two estimates for many of the main regions of projected drying (e.g., Southwest North America, Amazon, Europe, Southern Africa). Similarly, Swann et al. [26] concluded that PDSI overestimates drying for much of the world, in part because its simplified representation of vegetation processes does not account for increases in plant water use efficiency with increased atmospheric [CO_2]. This conclusion, however, was based on comparisons between PDSI (a soil moisture indicator) and a model-based runoff metric, *P-E*, two drought indicators that have already been shown to often respond quite differently in climate projections (e.g., Fig. 3). Indeed, while Cook et al. [73] found some differences between trends in model soil moisture and PDSI in model projections over Western North America, including a tendency for amplified drying in PDSI over the Central Plains, they found broad agreement between PDSI and soil moisture trends over much of the region. Further, the same study found stronger drying in model soil moisture versus PDSI for the Southwest, a result counter to the prevailing criticism of PDSI and other offline drought indicators. Similar results to Cook et al. [73] were found by Feng et al. [160], who also demonstrated strong agreement between PDSI and model projected soil moisture trends. Even for the observational record, Williams et al. [32] demonstrated that summer soil moisture calculations for the Sierra Nevada Mountains over the twentieth century using the simplified PDSI and a more physically based hydrologic model (the Variable Infiltration Capacity model) agreed remarkably well (Pearson's $r = 0.93$). Simplified drought indices like PDSI are therefore still useful, but should be evaluated thoughtfully when used in climate change projections.

Beyond drought definitions and indicators, there are also major process uncertainties that need to be resolved to increase confidence in model projections, especially regarding the role and response of vegetation. Because of direct CO_2 physiological effects and longer growing seasons, vegetation in model projections widely increases in terms of both total carbon assimilation (reflected in widespread increases in leaf area; Fig. 3c) and water use efficiency (the ratio of carbon assimilated to water losses from transpiration, WUE; Fig. 3d). These changes are important from a drought perspective because vegetation serves as the primary mediator of land-atmosphere exchanges across most land areas. Vegetation responses can affect a variety of important drought processes, including evapotranspiration [161, 162], runoff [163–165], and energy fluxes [166–169]. But despite vegetation's crucial importance for global water, carbon, and energy fluxes, and thus present and future droughts, observational constraints remain uncertain [162, 170–172], making model validation challenging over the present-day, let alone under high [CO_2]. Complicating data-model comparisons are the diversity of simplified vegetation assumptions within current generation Earth System Models (ESMs). Such diversity in process-representations can be seen in model choices related to plant hydraulic stress, canopy structures, stomatal conductance, carbon allocation, and more fundamentally, which ecosystem processes are prognostic versus prescribed.

One of the most important hypothesized processes is the impact on WUE, which has been shown in some analyses to counteract ET losses from a warmer atmosphere in observations and models [173–176]. The direct physiological impact of increased atmospheric [CO_2] is expected to ameliorate surface drying by increasing WUE, enabling plants to maintain the same (or higher) levels of carbon assimilation while using less water [26, 159, 166, 177, 178]. Observations over the last several decades strongly suggest that, globally, increased productivity has occurred without a commensal increase in water use [174, 175], especially over dryland areas where vegetation is most water limited [179, 180].

Notable, however, is remaining observational uncertainty about the WUE sensitivity to [CO_2], which may decrease with increasing scale (i.e., leaf to ecosystem) [176]. DeKauwe et al. [181] found increased water use efficiency at two long-term Free-Air CO_2 Enrichment (FACE) sites (Oak Ridge and Duke Forest), but this did not translate to an overall reduction in total transpiration at Duke. Similarly, in a remote sensing-based study over Eastern Australia, Ukkola et al. [182] found that [CO_2]-induced greening may have actually led to an increase in vegetation water consumption, contributing to a 24–28% reduction in streamflow from 1982–2010. Frank et al. [183], in a study incorporating tree-ring isotope records along with process-based vegetation modeling, found that while WUE of European forests increased over the twentieth century from 14–22%, warming temperatures and increased leaf area likely led to a net 5% increase in actual transpiration. Moreover, while stomatal conductance decreases in response to increasing [CO_2] can increase WUE, the associated transpiration decreases represent a physiological forcing on climate, increasing surface temperatures [173], diminishing rainfall [184], and affecting heatwave occurrence [185].

In model projections, widespread increases in WUE (Fig. 3d) also do not necessarily translate into reduced total surface water losses in the models (Fig. 3e). Increased leaf area (which increases the effective surface area from which water can be evapotranspired) and evaporative demand from a warmer atmosphere both favor increased evaporative losses, competing against WUE-induced water savings. Mankin et al. [25] analyzed drought projections for Western North America from a large ensemble of a single climate model. In this model, despite large precipitation increases over this region, widespread surface drying occurs (reflected in declines in snow cover, soil moisture, and runoff) colocated with significant increases in vegetation productivity (increases in leaf areas, photosynthesis, gross and net primary productivity). Such counterintuitive greening from both [CO_2] fertilization and warmer and longer growing seasons and soil drying is reconciled because the vegetation in the future is using up a much larger proportion of surface water compared to the past. In the same model, 40% of global land areas experience similar greening and drying, leading to a direct water trade-off between runoff and ecosystems [186]. This pattern ostensibly extends to the CMIP5 ensemble as well (Fig. 3), suggesting it is a near universal response in the model projections. The response of vegetation to climate and [CO_2] thus likely plays a critical role in projections of drought risk, and increasing confidence in these projections will require better constraining these uncertain processes.

Finally, uncertainties across climate model projections at the end of the twenty-first century are dominated by the greenhouse gas or radiative forcing scenarios used in the models [187]. Differences in warming across these scenarios project strongly onto changes in drought variability and risk, with more modest warming scenarios (e.g., RCP 2.6 or 4.5) typically resulting in less extreme drying compared to high forcing scenarios (e.g., RCP 8.5). Over western North America, for example, reduced warming significantly diminishes drying and megadrought risk at the end of the twenty-first century, a consequence primarily of reductions in evaporative demand with greenhouse gas mitigation [188]. Additional benefits of reduced drought risk have also been found in analyses of projections using the more aggressive 1.5° and 2° warming targets outlined by the Paris Agreement [189, 190]. Even at 2° of warming, however, some significant increases in drought risk for many regions are likely to occur, including the US Southwest, Central Plains, Mediterranean, and Central Europe [189]. More broadly, however, these studies strongly suggest that projected drought risk can be significantly reduced through climate mitigation.

Conclusions and Future Directions

Our knowledge of climate change and drought has advanced considerably since the publication of the AR5. This expanded body of knowledge includes numerous studies that more confidently attribute recent droughts to climate change [10, 32, 94, 103, 109, 111] and paleoclimate analyses that highlight the unusual severity of recent droughts in the long-term context of the Common Era [31, 72, 191]. These findings represent a marked shift from the much more conservative statements regarding drought and climate change in the AR5, which were appropriate for the time given the state of the science. Moreover, the research community has also developed a better appreciation for the complexity of drought responses across the hydrologic cycle in models and observations [16, 150] and begun to more explicitly address some of the most important process-level uncertainties in projections of drought risk and severity [25, 26]. Generating these new insights has involved scientists with a broad range of expertise, and such an interdisciplinary perspective will be required to continue advancing our understanding of climate change and drought in the coming decades.

The importance of direct temperature impacts on drought is one of the most critical remaining uncertainties. While the influence of warming temperatures on snow (and the resulting hydrology) is largely unambiguous [94, 103], the sensitivity of evapotranspiration (and associated moisture losses) to temperature is less clear. This uncertainty is centered primarily in the response of vegetation to both climate change and atmospheric [CO_2], responses which are mixed in terms of both magnitude and sign across models, experiments, and observations [25, 26, 181–183, 186]. Temperature effects are clearly important in determining the severity of recent droughts [31, 191] as well as projections from climate models, where significant surface drying (by a variety of indicators) extends over much broader geographic areas than would be predicted from precipitation trends alone [11, 25, 73, 150, 159] (Fig. 3). Interpretations of vegetation responses in climate model projections are further complicated by the likelihood that these models underestimate drought impacts on vegetation mortality and morbidity [192], overestimate the growth benefits of increasing atmospheric [CO_2] [193], and, by extension, changes in ecosystem water demands [25]. Reconciling the treatment of vegetation processes within these models with observational, paleoecological, and experimental evidence will be vital for improving these projections. Establishing covariability in precipitation and temperature as potentially having contributed to past severe droughts also remains a challenge for paleoclimatology, as independent and co-located reconstructions of the two variables remain rare. Nonetheless, such research would help establish the extent to which current and future “hot droughts” fall outside the range of late Holocene unforced variability. Paleoclimate records are an important but often underused resource, with clear applications for analyses relevant for questions regarding climate change and drought. A concerted effort to identify and fill existing data gaps in the paleohydroclimate record would improve the skill and utility of traditional climate field reconstructions and data assimilation approaches and expand the spatial and temporal scope of these efforts. A critical need in paleoclimate is the continued development of proxy-system models that will enable mechanistic understanding of how drought signals are imparted to different archives to ensure low-frequency variability is not spuriously detected [68].

While this review has focused primarily on physical and biological drought processes, it is increasingly apparent that societal water demands and human management of water resources cannot be treated as extraneous factors in analyses of climate change and drought risk [21, 22]. The manifestation of drought in the hydrologic cycle, and the associated impacts on people and ecosystems, is affected by both social and physical processes, and these dynamics are known to have occurred in the past [194–196] and can already be observed today. For example, the degree to which droughts can contribute to social disruptions, including conflicts and famines, depends critically on the response and effectiveness of local social and political systems [9, 197–199]. More broadly, human exposure to climate change-induced water stress and drought risk in the future is contingent on the greenhouse gas emissions pathways (a global policy decision) and changes in regional populations [198, 200]. We would therefore argue that some of the largest uncertainties for the future of climate change and drought may therefore lie in the social science realm, subject to decisions made at global (*How much will the world allow the climate to change?*) and regional (*How will we adapt to the impact of climate changes that do occur?*) levels. Addressing such questions will require close collaboration and depend on effective communication and knowledge transfer between the physical and social sciences [201], a challenging but necessary task.

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Compliance with Ethical Standards

Conflict of interest The authors declare they have no conflict of interests.

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