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REVIEW

Observed changes in hydroclimate attributed to human forcing

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Abstract

Observational and modeling studies indicate significant changes in the global hydroclimate in the twentieth and early twenty-first centuries due to anthropogenic climate change. In this review, we analyze the recent literature on the observed changes in hydroclimate attributable to anthropogenic forcing, the physical and biological mechanisms underlying those changes, and the advantages and limitations of current detection and attribution methods. Changes in the magnitude and spatial patterns of precipitation minus evaporation (*P–E*) are consistent with increased water vapor content driven by higher temperatures. While thermodynamics explains most of the observed changes, the contribution of dynamics is not yet well constrained, especially at regional and local scales, due to limitations in observations and climate models. Anthropogenic climate change has also increased the severity and likelihood of contemporaneous droughts in southwestern North America, southwestern South America, the Mediterranean, and the Caribbean. An increased frequency of extreme precipitation events and shifts in phenology has also been attributed to anthropogenic climate change. While considerable uncertainties persist on the role of plant physiology in modulating hydroclimate and vice versa, emerging evidence indicates that increased canopy water demand and longer growing seasons negate the water-saving effects from increased water-use efficiency.

1. Introduction

Water is essential for supporting life on Earth [1]. Water moves in the Earth system through the hydrologic (water) cycle maintaining the Earth's energy homeostasis [2]. Global estimates

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of annual total water exchanges between the atmosphere and surface (primarily precipitation and evapotranspiration) are on the order of 50 x 10^4 km^3 [2]. About 40 x 10^3 km^3 is transported from the ocean to land as precipitation and the same amount returns to the ocean as runoff [2]. This water exchange is critical for human well-being and development, as food security, electrical power generation, industry, and municipal water supply, amongst other socioenvironmental systems, depend on the availability and accessibility of water [3, 4]. Changes in hydroclimate (e.g., climate-driven changes in the water cycle) are one of the most impactful consequences of climate change on human society, given the importance of precipitation (seasonality and magnitude) on food security and economic development [3]. The atmospheric water balance can be expressed as the difference between precipitation (P) and evapotranspiration (E). That, in turn, equals the sum of vertically-integrated atmospheric humidity ("precipitable water," W) change (dW/dt) and horizontal moisture convergence ($\nabla \cdot \mathbf{Q}$). Since dW/dt is negligible on monthly and longer time scales [5]:

$$P - E \approx \nabla \cdot \mathbf{Q} \tag{1}$$

Overland, however, P–E is the sum of water storage (S) change (dS/dt) and runoff (R) [5]:

$$P - E \approx \nabla \cdot \mathbf{Q} \approx R + dS/dt \tag{2}$$

Climate models suggest significant changes in global hydroclimate due to the human-driven increases in greenhouse gas concentrations in the atmosphere through the twenty-first century [4, 6–8]. Simulated changes in precipitation include drier conditions in the subtropics and wetter conditions in the tropics and extratropics [7, 9]. However, this phenomenon occurs mostly over the oceans [10]. Regions that models robustly project will experience increased aridity arising from declines in precipitation, increased evapotranspiration, or both include the Mediterranean, southwestern South America, southern Australia, and southwestern North America [6, 8, 11, 12].

Observational and modeling studies find that anthropogenic climate change has already increased drought risk [13–19] and extreme precipitation events [20, 21]; reduced snowpack [22], ice sheets [23], and runoff [17]; and altered seasonal precipitation patterns [24] and phenology [25]. Many of these detection and attribution studies rely on climate models to identify and separate the anthropogenic signal from natural variability [12, 26]. A major limitation is that considerable biases exist in models [27], especially in simulating certain hydroclimate features at regional and local scales [28], as well as in the observational datasets used for evaluating models [26, 29, 30]. Novel approaches for detecting hydroclimatic changes include machine learning and other statistical tools to identify outliers in precipitation, temperature, and soil moisture from instrumental and reconstructed records and climate models [26].

In this Review, we provide a survey of the literature on observed changes in hydroclimate as a result of anthropogenic emissions and land use changes during the historical period (e.g., 1850–2020). We analyze the mechanisms underlying human-forced influences on hydroclimate and the advantages and limitations of the approaches to estimating the anthropogenic contributions to change, including detection and attribution methods. Given the role of increased global temperature in exacerbating contemporaneous droughts [14–16, 31], we focus on the effects of anthropogenic warming. Finally, we analyze studies on changes in extreme precipitation, wildfires, and the possible influence of plant physiology and phenology on observed hydroclimate changes and future drought risk.

2. Understanding the causes of the anthropogenic hydroclimate change

Large-scale changes in precipitation (intensity, duration, and frequency), soil moisture, runoff, and evapotranspiration are expected due to thermodynamic and dynamic drivers as the climate warms [7, 8] (Fig 1). Global thermodynamic changes are explained through the Clausius-Clapeyron relationship, which predicts an exponential increase in the water-holding capacity of the atmosphere and thus, a higher water vapor with temperature of ~7% K^{-1} warming, assuming a constant relative humidity [9]. Observations and simulations suggest an increase in water vapor of nearly 6% K^{-1}, slightly lower than predicted by Clausius-Clapeyron [32]. However, the distribution of the change is spatially complex, with increases of $\sim$4%–5% K^{-1} near the surface and 10%–15% K^{-1} at the Upper Troposphere and predominantly over the tropical Pacific Ocean [32–34]. In contrast, the observed increases in global mean precipitation and evaporation are roughly 1–3% K^{-1} [8, 9]. For precipitation alone, climate models indicate increases of 2.1–3.1% K^{-1} in global mean precipitation (e.g., [35, 36]) and 5.5% K^{-1} in the intensity of extreme precipitation (e.g., a high percentile of daily precipitation) [36]. That means that the observed and expected increases in global mean precipitation and evaporation, especially over land, falls below that predicted by Clausius-Clapeyron [9, 10]. An explanation for this discrepancy is that, at global scales, precipitation and evaporation are constrained by the atmospheric energy balance and dynamics [37, 38].

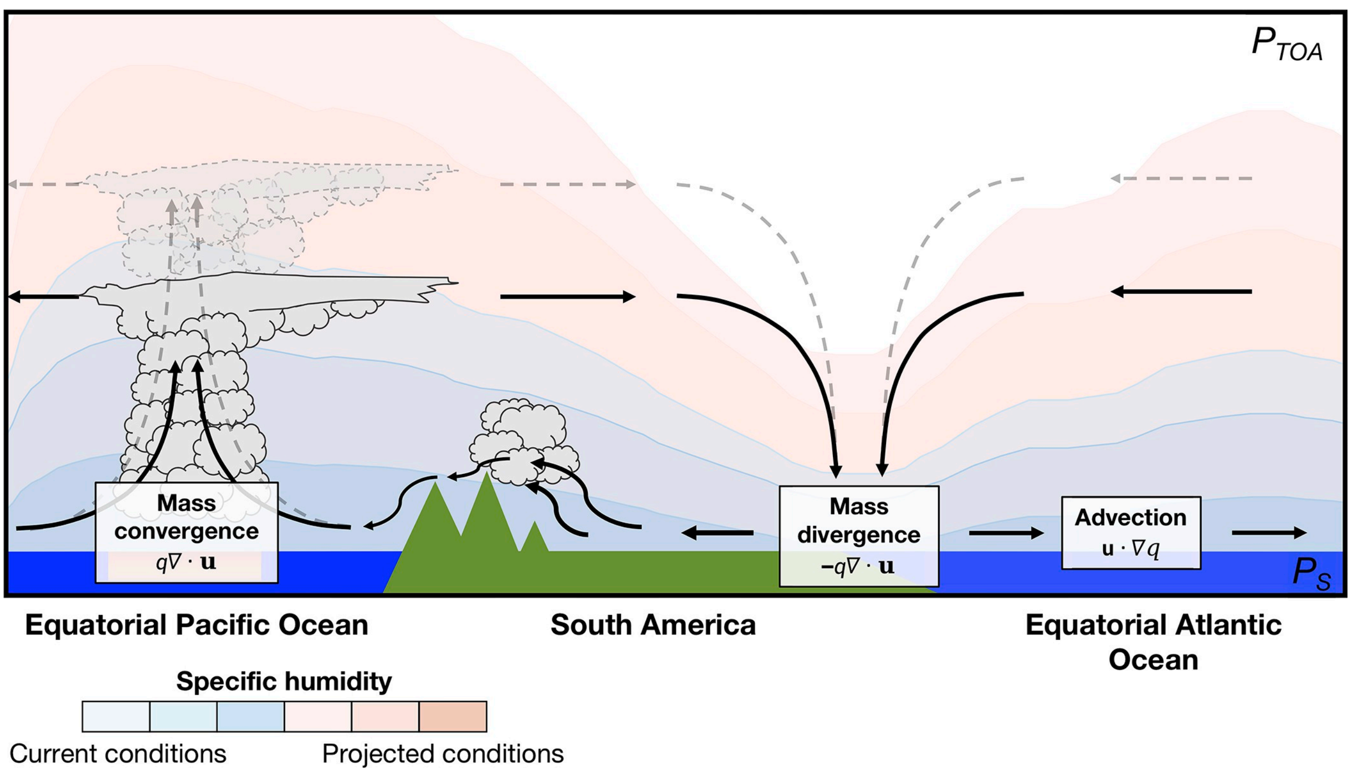

Fig 1. Representation of the dynamic and thermodynamic changes associated with the intensified water cycle. This diagram depicts changes with an El Niño-like pattern. As the climate warms, the atmosphere's water-holding capacity increases further (reddish colors). That causes higher mass convergence and lower mass divergence (dashed gray lines), increasing the contrast between areas dominated by a convergent flow (e.g., equatorial Pacific Ocean) and the opposite in areas dominated by a divergent flow (subtropics). That, in turn, may increase the intensity of hydroclimate extremes: first, higher mass convergence is associated with higher precipitable water (e.g., [13]), and second, lower relative humidity in a warmer climate could increase the risk of drought, especially flash droughts [112]. The implications of such changes are critical to understanding the effects and teleconnection patterns of climate modes of variability, such as El Niño-Southern Oscillation (ENSO) in a changing climate.

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Because the rate at which precipitation and evaporation increase as a function of temperature is smaller than for water vapor, a higher atmospheric water vapor residence time is expected [9, 39]. The estimated residence time of water vapor is ~8.5 days [40], which might increase by 3–6% K^{-1} due to the warming in the twenty-first century [39]. Higher water vapor concentration in the lower troposphere might lead to a slowdown in atmospheric circulation and a reduced mass exchange between the boundary layer and the free atmosphere [9, 41–43]. The weakened atmospheric circulation balances with the increased water vapor in the tropics, which is necessary to compensate for the smaller increase in global mean precipitation as compared to water vapor [41, 44].

Increased water vapor in the tropics is consistent with higher mass convergence and precipitation minus evaporation (*P–E*) rates in observations and simulations [34], along with higher divergence and lower *P–E* rates in the subtropics [9, 34]. These changes in global *P–E* patterns are often referred to as the "wet-get-wetter, dry-get-drier" or "wet-events-wetter, dry-events-drier" mechanism [9, 10], in allusion to increased precipitation in areas already humid (e.g., the tropics and high-latitudes), while the opposite occurs in arid and semi-arid regions (e.g., the subtropics) as the climate warms. However, the "wet-get-wetter, dry-get-drier" mechanism is only valid over the ocean and certain areas over land, where moisture is unlimited [10]. Over land, especially in arid and semi-arid regions where moisture is limited, this mechanism is inaccurate because, in many cases, evapotranspiration cannot exceed precipitation [10]. Another factor inhibiting precipitation over land in a warming climate is an increased land-ocean temperature gradient [10] since the ocean warms slower than the land surface. Though the increased gradient may support the development of a sea breeze [42] or monsoon [43] circulation, the cooler ocean temperatures can not provide the moisture supply needed to satisfy the moisture demand over land [36].

Changes in *P–E* patterns due to global and regional atmospheric warming patterns and dynamics are also expected in a warmer climate [45] (Fig 1). However, the mechanisms underlying these changes are not well-constrained from observations compared to the thermodynamic controls [28]. A slowdown in the global atmospheric circulation, especially in the tropics [41, 46] counteracts the thermodynamic intensification of the water cycle and, consequently, partially reduces the *P–E* gradients in the ocean [34]. A weakening of the tropical atmospheric overturning circulation (Hadley and Walker circulations) with warming is more robust across climate models, especially for the Walker Circulation [34, 41, 42]. This is coherent with a decrease in the East-West Pacific sea level pressure gradient in some observational studies [41, 46]. Nevertheless, more recent observational and modeling studies suggest a strengthening in the Walker Circulation in the historical period [47]. A plausible reason for this discrepancy is that climate models do not correctly simulate the response of the Walker Circulation to warming [28, 47]. For example, many state-of-the-art climate models have a cold bias in the tropical Pacific compared to observations [48]. The strengthening of the Walker Circulation might also be an initial response to the warming, followed by its subsequent weakening [47]. To date, what the response of the Walker Circulation to climate change might be is still unclear [4]. Observations have also shown a narrowing and strengthening of the intertropical convergence zone (ITCZ) and an insignificant change in its mean location [49, 50]. This is consistent with climate model projections for the ITCZ in the twenty-first century, which suggest a further narrowing of the ITCZ [45]. The current ITCZ width varies from 250 to 1500 km [49] and has narrowed from 20% K^{-1} in the Atlantic Ocean to 29% K^{-1} in the Pacific [50].

Regional and local scale hydroclimate changes in a warming climate might be enhanced by local feedback processes, including changes in moisture transport and mass convergence (Fig 1) and other land-atmosphere coupling controls, such as soil moisture [51, 52], land-use changes, and water management [34]. Modeling studies indicate that regional to local scale increases in precipitation are closer to Clausius-Clapeyron [9] and might be higher (>10% K^{-1}) in certain regions, especially considering extreme events alone [53]. A higher convective available potential energy (CAPE) has been associated with increased extreme precipitation in some regions over land [54]. CAPE is expected to increase further in a warmer climate [55] due to changes in lapse rates and increased low-level humidity, but mainly over the ocean [56]. More intense precipitation could increase the risk of flooding locally and alter runoff and recharge patterns [7, 57].

The projected shifts in *P–E* over land are associated with changes in other aspects of hydroclimate, including a decline in snowpack, runoff, and surface soil moisture [7]. Seasonal changes in runoff are expected, especially in snow-dependent regions [7, 58]. In these regions, simulations indicate that snowpacks would not only melt earlier and faster [58] but also an increase in the rain/snow ratio [57] would further inhibit snow accumulation. Consequently, an increased runoff in winter and spring due to a higher rain/snow ratio and faster snow melting is expected, and the opposite in the summer and autumn [7]. Soil moisture declines are more robust than precipitation across models in many subtropical regions [7, 57]. Significant declines in modeled soil moisture occur in regions even with small increases in precipitation [7, 59], which highlights the role of increased evaporative demand of the atmosphere over land, land-surface feedbacks, and changes in plant water use [7, 60, 61]. While some studies indicate that vegetation response to increased atmospheric CO_2 concentration could ameliorate drought risk by improving the water use efficiency of plants [11, 60], others, in contrast, suggest an amplification by increased plant growth and water use [61, 62].

Land-use changes from human activities and wildfires further contribute to shifts in *P–E* [34] and hydroclimate [63–65]. Irrigation increases evaporation and water vapor, which contribute to higher precipitation rates locally or in nearby areas, as dictated by moisture transport [63]. Deforestation and urbanization can also modify *P–E* by changing surface albedo, energy, and water budgets [34], while aerosols from wildfires increase cloud albedo and cool surface temperatures, changing global *P–E* patterns [64, 65]. More importantly, the direct impact of human water use on the water cycle is often misrepresented in many estimates and diagrams [66]. Current human freshwater use through agriculture, livestock, water withdrawals, and industry is 50% of global river discharge [66].

3. Observed changes in hydroclimate

Separating and estimating the anthropogenic contributions to observed hydroclimate changes from other external forcings and natural variability can be challenging [67]. Detection and attribution methods attempt to do so by combining statistical tools with instrumental and simulated climate data to detect changes in the climate system and quantify the causes of these observed changes [68]. Detection and attribution are also necessary to evaluate the skill of climate models in simulating crucial processes occurring in the climate system, both globally and regionally, which is inferred from their ability to replicate observations [68].

The traditional detection and attribution approach involves using climate models to assess the changes in the probability of occurrence of an event and if those changes are driven by anthropogenic-forced change [68]. This approach is well-suited to quantify the contribution of anthropogenic warming to hydroclimate extremes controlled by thermodynamics [67], including drought intensification and duration due to temperature-driven increase in evapotranspiration [13–16, 69]. However, it is limited to quantifying the anthropogenic contribution to extreme hydroclimate events further enhanced by global and local atmospheric dynamics [67]. Newer approaches like the optimal fingerprint method [12, 70] have been used recently in detection and attribution studies of human-forced hydroclimate change, including to quantify the contribution of climate change to drought risk [12], the occurrence of flash droughts [69], and changes in the mean and extreme precipitation [71]. Despite the advances in detection and attribution methods, limitations persist for several reasons [67]. Some of those limitations arise from systematic biases in climate models, deficiencies in observational data (including the quality, length, and spatial coverage of instrumental records), and shortcomings in the detection and attribution approaches themselves [67]. Nevertheless, a growing body of evidence suggests that anthropogenic climate change is already impacting global hydroclimate and is likely to do so as the climate warms further in the twenty-first century [11, 12, 57, 67, 71].

3.1 Observed changes in drought

Observations and simulations indicate an increase in the frequency, duration, and severity of drought in southwestern North America [15], the Mediterranean and southern Europe [72], Australia [73, 74], southwestern and central South America [18, 75, 76], Central America and the Caribbean [16, 17, 77], Africa [78, 79], and eastern Asia [18, 80] (Fig 2). These indications are drawn from a range of drought metrics such as the Palmer Drought Severity Index (PDSI) [12, 18, 70] (Fig 2A), the standardized precipitation index (SPI), standardized precipitation minus evaporation index (SPEI) [19, 81], and soil moisture [15].

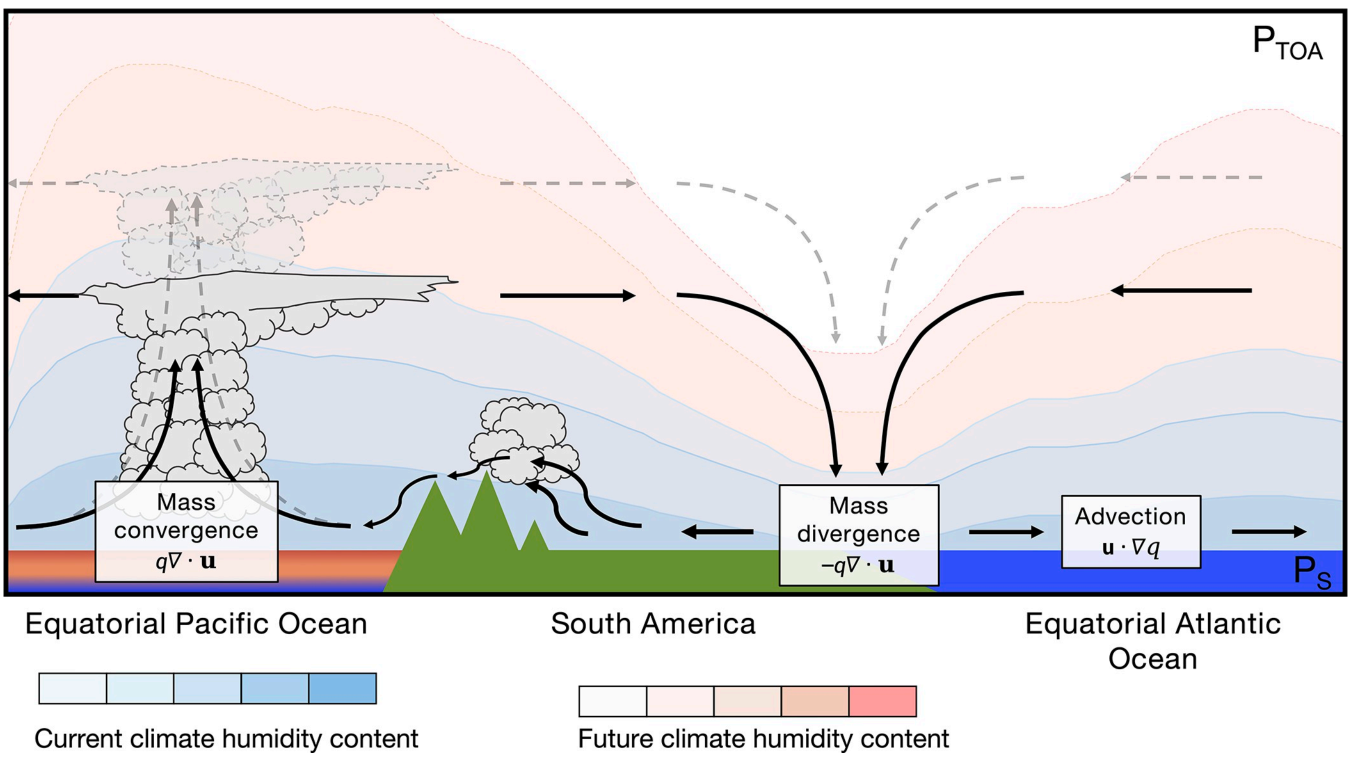

Fig 2. Trends in observed precipitation and Palmer Drought Severity Index (PDSI). (A) trends from version 2020 of the Global Precipitation Climatology Centre (GPCCv2020), (**B**) trends from the "self-calibrated" PDSI dataset from Dai [18], (**C**) trends from the Climatic Research Unit version TSv 4.01 (CRU TSv.4.01), and (**D**) "self-calibrated" PDSI dataset from van der Schrier et al. [147]. The trends were calculated over a common period from 1950–2018. While GPCC and CRU precipitation datasets indicate similar patterns in their trends, the PDSI datasets differ in their trends, but most importantly in their magnitudes. In addition to using different input climate data, those PDSI data sets use different calibration periods. For example, the CRU PDSI product uses the whole period (i.e., 1901–2021), and Dai PDSI uses 1950–2000. *This Figure was made with Natural Earth. Free vector and raster map data https://www.naturalearthdata.com.*

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In southwestern North America, the anthropogenic fingerprint on hydroclimate has been shown to be robust, as indicated by an array of detection and attribution studies using model simulations, proxy and observational records to quantify the contribution of human-forced change [11–15, 26, 31, 70]. A notable event in this region is the megadrought from 2000–2018 [15], ranked as one of the most severe droughts since at least 800 CE [15]. While natural variability has also influenced this multidecadal event, anthropogenic climate change accounted for 46% [15], largely by increasing atmospheric vapor pressure deficit [14, 15, 31]. Similarly, earlier studies indicate that anthropogenic warming increased the severity of the 2011–2016 California drought by 8–27% [14], making it the most severe drought since 1200 CE [31]. As in other parts of Western North America, precipitation deficits associated with the 2011–2016 California drought have been linked with natural variability but worsened by higher temperatures [14, 15, 31]. Decreased streamflow of the Colorado River in 2000–2014 is consistent with the droughts observed in southwestern North America in the early twenty-first century [82]. Reduced streamflow from 2.7 to 9% has been estimated as a result of a 0.9˚C increase in the mean temperature of the Upper Colorado River Basin [82].

Observed hydroclimate changes in southern Europe and the Mediterranean are also associated with increased aridity since at least 1960 [72, 84], which is consistent with climate model projections as a result of the warmth [72, 83]. Across the region, a trend toward aridity is observed in precipitation and soil moisture [72], including droughts like the severe Syrian drought in 2007–2010 [84]. It is estimated that anthropogenic climate change doubled or tripled the likelihood of the 2007–2010 Syrian drought [84]. Although previous studies have found an anthropogenic signal in precipitation reductions and warming temperatures [12], this is heterogeneous, and some parts of the Mediterranean are more strongly influenced by natural variability than others [81, 85]. For example, tree ring-reconstructed soil moisture indicates that droughts of similar duration and severity as the 2007–2010 Syrian drought have occurred across Mediterranean regions in the last 900 years [85]. However, the 1998–2012 Levant drought is likely the most severe since 1100 CE [85].

A significant increase in the duration, frequency, and severity of seasonal drought has been observed in portions of southwestern and southeastern Australia since the twentieth century, particularly in the autumn and winter [73, 74]. This change has been attributed to anthropogenic forcing, arising from greenhouse gas emissions and changes in atmospheric ozone levels [74]. The duration of seasonal droughts in parts of southeastern Australia has also increased between 10 and 69% since the mid-twentieth century [73]. The worst drought recorded in southeastern Australia, the 2001–2009 "Millenium Drought", severely affected the agricultural sector of the country [86]. Earlier studies argue that the cause of the Millennium Drought was linked to fewer La Niña events and a negative Indian Ocean Dipole [87], suggesting a strong contribution from natural variability [87, 88]. Also, while higher temperatures are associated with drought in Australia, it remains unclear whether warmer temperatures contributed to or resulted from the drought [88]. Reconstructed soil moisture data, for example, indicate that droughts like the Millennium Drought in eastern Australia are in the range of natural variability for at least the last 500 years [88, 89]. The anthropogenic contribution to the recent 2019–2020 drought and associated bushfires is complex, as the signal is present in some drivers (e.g., temperature) [90].

Trends toward increased aridity are observed in the central (Brazil) and southwestern (Chile and Argentina) areas of South America [18]. In central Chile, the drying trend since late 1970s [68, 69] is characterized by multiple droughts, including an ongoing multidecadal megadrought that began in 2010 [76, 91]. The megadrought has reduced snowpack in the Andes, streamflow of major rivers, vegetation productivity [75], and lake levels [92]. While previous studies have found an anthropogenic contribution of ~25% to the Chilean megadrought, natural variability has also contributed [76, 93]. Furthermore, significant uncertainties persist in modeled precipitation in South America [94], precluding more accurate estimates of the anthropogenic contribution to the recent drought. However, reconstructed soil moisture indicates that the ongoing megadrought is the most severe of the last millennium and that its occurrence is unexpected from natural variability alone [95]. In this context, the drying trend in central Chile and western Argentina is associated with changes in the Southern Annular Mode and the expansion of the Hadley cell, both of which are expected from anthropogenic climate change [76, 93, 95]. Land-use feedback in the Amazon is also critical in recent aridity in portions of South America [96]. Deforestation has contributed to 4% of recent drying in parts of the Amazon, while longer dry seasons are associated with higher deforestation rates [96].

In the Caribbean Islands and Central America, a drying trend observed since at least 1950 includes several short-term (up to a year-long) and multiyear droughts [16, 17, 77, 97, 98]. Although previous studies found a modest but statistically significant decline in precipitation in some parts of these regions [97], more recent studies indicate that rainfall has not markedly changed and has indeed slightly increased in certain areas [18, 99, 100]. Further, there is a strong influence of natural variability in precipitation in the Caribbean and Central America, modulated by the El Niño–Southern Oscillation, the North Atlantic Oscillation, and the Atlantic Meridional Mode [77, 99, 101, 102]. Some of the worst droughts in these regions, for example, occurred during El Niño events [98, 103]. Paleoclimate and modeling studies have further suggested an El Niño-like pattern associated with drought in Central America [104]. However, drought conditions in Caribbean Islands and Central America are likely exacerbated by an increased atmospheric evaporative demand [16], driven by higher temperatures [99, 100, 105].

Droughts have also intensified in portions of Africa, including the Sahel, northern, and southwestern South Africa in recent decades [11, 18, 57, 78, 79, 106–108]. Some of the worst African droughts resulted from a failure of the rainy season, associated with the Indo–Pacific internal variability [78], resulting in droughts in the Sahel [106], eastern [107, 108], and southern Africa [79]. Although some studies have attributed the drying over southwestern South Africa to anthropogenic climate change [79], others suggest that the recent droughts are possible due to natural variability [100]. For example, Pascale et al. [79] estimated anthropogenic climate change increased the 2015–2017 "Day Zero Drought" five to six-fold. The rationale of the human-driven hydroclimate change in Africa is an increased sea surface temperature over the Indian Ocean and the opposite in the North Atlantic, as a result of a higher atmospheric greenhouse gas concentration and aerosols, respectively, and the southward shift of the ITCZ [11, 57, 109]. Further, paleoclimate records indicate that the drying experienced in eastern Africa is uncommon from natural variability alone [110]. Given the limitations and contradictory results regarding the anthropogenic fingerprint in African drought, it is still difficult to quantify the contribution.

Other regions that have experienced a drying trend are southeast and eastern Asia, as observed in precipitation, streamflow of major river basins, and PDSI between 1950 and 2018 [18, 80]. Studies suggest that the drying in eastern Asia is due to increased temperatures and a significant decline in humidity and precipitation [18, 80]. From a long-term perspective, paleoclimate records indicate that the region has seen multidecadal or megadroughts in the Common Era [80, 91, 111], suggesting that major droughts are possible as a result of natural variability.

An increasingly frequent type of drought is "flash drought" (i.e., a drought that unfolds relatively fast [112]). Over 74% of the land surface has seen an increased frequency of flash droughts attributed to anthropogenic warmth [69]. However, Qing et al. [113] found that flash droughts did not increase in frequency in 2000–2020, but their onset speed was faster. The anthropogenic contribution to the increased flash/slow drought ratio is 48%, while the warming has increased the onset of flash droughts by 39% [69]. A flash drought results from a marked precipitation deficit and strong evapotranspiration associated with low humidities, high temperatures, reduced cloud cover, and strong winds [112]. In this context, an increased evaporative demand of the atmosphere driven by anthropogenic climate change might increase flash drought risk.

3.2 Observed changes in extreme precipitation events

Observational studies reveal an increase in extreme precipitation events in many land areas globally, including daily precipitation exceeding 50 mm and high percentiles [20, 21, 114]. A higher frequency of extreme precipitation events is observed in about two-thirds of the stations across the world, mainly in Asia, Europe, and North America [21]. More frequent extreme precipitation events are expected due to anthropogenic climate change [114] as part of the water cycle intensification [37] (Fig 1). In agreement with the Clausius-Clapeyron relationship, more intense precipitation is likely to occur as the lower troposphere warms because of the higher water-holding capacity of the atmosphere [9, 114]. Modeling studies have indicated an increased frequency of extreme precipitation events consistent with Clausius-Clapeyron, and that anthropogenic warming has contributed to 18% of the observed extreme events by 2015 [115]. Dong et al. [20] also found anthropogenic fingerprints on heavy precipitation events in the last few decades over most continents. However, large uncertainties in the observed anthropogenic contribution to extreme precipitation events prevail due to the strong influence of natural variability, the role of local and regional dynamics, and, most importantly, limitations in instrumental records [21, 114]. Using a combination of instrumental data, Sun et al. [21] found a median increase in extreme precipitation over land of 6.6% K^{-1} (5.1%–8.2% K^{-1}). That is similar to the change predicted by Clausius-Clapeyron [9]. However, the estimated sensitivity of extreme precipitation varies substantially, probably due to local and regional dynamics or because the global mean temperature differs considerably from the local mean temperature [21].

Extreme precipitation events have increased in Europe since at least 1901 [116] and even more since 1950 [21, 114]. Such changes include the 2021 extreme event that led to major flooding in Western Europe [117]. A detection and attribution analysis of that event estimated an anthropogenic contribution of 3–19% to its severity and increased its likelihood 1.2–9 fold compared to 1.2˚C colder conditions than today [117]. Similarly, most of the ground stations of Central Europe saw an increased frequency of heavy precipitation events from 1901–2013 that would not be possible from natural variability alone [116]. Based on instrumental data, Europe has the highest number of stations showing a trend toward more extreme precipitation events, partly due to its dense station network [21]. However, such a trend is seasonally and spatially heterogeneous [118]. For example, in northern Europe, extreme precipitation events are increasing in winter, but the opposite is happening in the southern part of the continent in summer [118].

The intensity of heavy precipitation has also increased significantly in North America since 1950, although there is substantial spatial variability across the region [21, 114, 119]. In many parts of the United States, Mexico, and Canada, there is a detectable anthropogenic fingerprint in the increased frequency and intensity of heavy precipitation events [20, 21, 120]. While most stations reveal an anthropogenic footprint on extreme precipitation, in the southern United States, the signal is virtually undetectable, probably due to the strong influence of natural variability [83]. However, climate models indicate that even areas currently without a detectable anthropogenic signal, like the southern United States, show significant increases in extreme precipitation events as the climate warms [119].

In Asia, a trend toward a higher intensity of heavy precipitation is attributed to anthropogenic forcing, especially over the monsoonal regions [20, 21]. The increased frequency is noticeable in one and five days-extreme precipitation events [21, 114], both of which are associated with flooding events [118]. In some regions, like southern Asia, these changes occur along with a higher frequency of dry spells [121]. While in the twentieth century, south and central Asia experienced a decline in heavy rainfall events, this decrease has been attributed to human-derived aerosols [122]. Although many areas of Asia are seeing an increased frequency and intensity of heavy rainfall events [21, 114], there is a large spatial variability probably driven by local factors and regional dynamics [122].

Changes in one and five-day precipitation extremes in Oceania (Australia, New Zealand, and other small insular countries) are insignificant across the region on average [114]. Most stations evaluated by Sun et al. [21] show a decreased frequency of heavy precipitation events between 1950–2018, consistent with previous studies, especially in southern Australia [123]. In contrast, regions like northwestern Australia have experienced increased extreme precipitation in the twentieth and early twenty-first centuries [21]. Similarly, trends in heavy precipitation events in New Zealand are not statistically significant, with a comparable number of stations indicating positive and negative trends [114].

The lack of observational data in many parts of Central and South America and Africa precludes a comprehensive analysis of the observed change in extreme precipitation, let alone detection and attribution analysis [21]. Where station data is available, most instrumental records indicate increased extreme precipitation events in parts of these regions [21, 114, 123]. In Africa, for example, over 64% of the available stations indicate a significant increase in daily extreme precipitation [21]. The positive trends are observed in southern and eastern Africa [21, 114, 124]. McBride et al. [124] found that the number of rainy days in South Africa has not changed in the period 1921–2020. However, the probability of heavy and very heavy rainfall (>75 mm and >115 mm, respectively) have increased significantly in South Africa [116]. Previous studies suggest an increased frequency of heavy precipitation in northern (Colombia, Venezuela, and Guyana) and southeastern (Brazil) South America, while a decrease in northeastern Brazil was detected [114]. An increase in heavy precipitation is also observed in the Amazon, although not statistically significant [106]. In Central America, a small but significant increase in heavy daily rainfall is observed in El Salvador, Guatemala, and Panama [17, 125], while a similar trend is observed in some Caribbean Islands for the periods 1986–2010 [100].

4. Plant physiology, phenology, and hydroclimate change

Vegetation and its effects (e.g., fire) are a critical component of the coupled climate system [25, 64, 126]. Many land-atmosphere processes are mediated by plants, including exchanges in energy and momentum [25, 126–129]. Physiological processes in plants, like photosynthesis, are affected by human-driven increases in atmospheric CO_2 [25, 126–129]. In addition, phenological changes also influence the Earth's energy balance by modulating the carbon cycle, energy and water fluxes, and changing surface albedo and roughness [126, 130]. Conversely, vegetation is affected by the climatic conditions and climate variability [25, 126]. For example, changes in plant phenology in a warming climate are a leading cause of the observed carbon uptake [25], while drought increases plant mortality and reduces gross primary productivity (GPP) [131]. Biomass burning from wildfires further affects global hydroclimate by altering surface temperature and precipitation patterns [64, 65].

4.1 Changes in plant physiology and hydroclimate change

As CO_2 concentrations in the atmosphere increase, mainly from human activities, Earth system models (ESMs) project rising plant water-use efficiency [11, 60]. Higher atmospheric CO_2

improves plant water-use efficiency by enhancing photosynthesis and reducing water loss through transpiration [11, 60]. Paleoclimate studies further confirm decadal to centennial changes in plant water-use efficiency, along with enhanced photosynthesis and stomatal reductions [132, 133]. Consistent with ESMs, observational experiments indicate a nearly proportional increase in plant water-use efficiency to atmospheric CO_2 concentrations and reduced water consumption by plants [133].

Some modeling studies suggest that an increased plant water-use efficiency would significantly reduce drought risk, counteracting the effects of warmth-driven atmospheric moisture demand [60, 133]. Other modeling studies, in contrast, indicate that increased canopy water demand and longer growing seasons negate the water saving effects from increased water-use efficiency [61, 62]. Therefore, the absolute contribution of higher atmospheric CO_2 concentrations to drought risk is still debatable [61, 134]. In addition, the ESMs used in these studies are subject to model biases in how they estimate evapotranspiration and represent the underlying processes [61, 134]. A recent observational study indicates that increased temperature and vapor pressure deficit might be more crucial than decreased stomatal conductance in regulating evapotranspiration [135]. Therefore, while plant physiology plays a critical role in modulating some features of the water cycle, it is still uncertain whether CO_2 fertilization would reduce future drought risk globally [8].

Conversely, drought, especially severe drought, can limit plant photosynthesis, affecting the global carbon cycle [8]. While plant water-use efficiency may reduce drought risk by modulating evapotranspiration, severe drought can offset such an effect by constraining photosynthesis–therefore, decreasing GPP and the ability of plants to capture atmospheric CO_2 [8]. Drought is also associated with wildfires, forest mortality, and decreased vegetation growth [131], all of which alter the global CO_2 sink [136]. Fires also play a central role in the climate system's carbon budget and biogenic aerosol emissions and are known to be growing in size and intensity as a result of anthropogenic climate change [90, 137]. Satellite-based studies indicate a correlation between global net land CO_2 sinks and terrestrial water storage [127] and considerable carbon release from the tropics and mid-latitudes during severe droughts [138]. Plant growth is also limited by reduced soil moisture and increased vapor pressure deficit, although the first is the dominant driver of plant stress [131].

Observational studies suggest that photosynthesis and carbon uptake have increased since at least 1981, as estimated from a higher GPP, leaf area index (LAI), and longer growing seasons in 82 ± 5% and 25–50% of the global vegetated area [127–129]. Measurable hydrologic changes are also observed from ground observations and remotely sensed data [128, 129]. Those changes are, in part, modulated by CO_2 fertilization [127], although regional-scale hydrologic changes depend on several factors, including an increase in vapor pressure deficit due to reduced transpiration/warmer temperatures and in LAI due to the fertilization effects on GPP [127]. In Australia, for example, the threshold for water limitation of vegetation cover decreased significantly from 1982–2010 [139]. In other words, plants are thriving with less precipitation and less water availability. Further, in semi-arid and sub-humid basins across Australia, a 24–28% decrease in streamflow is observed, along with an increased LAI, while changes in wet and arid basins are insignificant [139]. This is consistent with a significant decline in baseflow in many basins in eastern Australia, where baseflow decreased to -1 mm a^{-2} (i.e., mm/yr/yr) in 73% of the basins during the 1981–2013 period [129].

Studies using an array of observational data suggest that ESMs underestimate the role of terrestrial water availability on GPP and the carbon cycle [136, 140]. In water-limited areas and seasons, GPP is inversely proportional to vapor pressure deficit, even when the soil water content is high [140]. Humphrey et al. [136] further found that the rate at which atmospheric CO_2 varies is strongly associated with changes in terrestrial water storage (e.g., the total water content over land, including surface and groundwater [141]). That is consistent with recent modeling studies that suggest ESMs underestimate the effects of soil moisture–atmosphere feedback on global carbon uptake over land [142]. In fact, soil moisture is estimated to account for up to 90% of interannual variability of the global land carbon uptake by modulating photosynthesis [142]. Overall, while there is a growing body of evidence suggesting the critical role of CO_2 fertilization on drought risk, limitations in observations (e.g., relatively short time intervals) and the current ESMs preclude a more accurate characterization of these processes.

4.2 Changes in phenology and hydroclimate change

Over the last several decades, a significant trend toward earlier spring onset is observed in North America, Europe, and Eastern Asia [25]. In North America, data from the US National Phenology Network indicate an earlier spring onset trend of 0.9 days per decade, on average [25]. In Europe, previous studies suggest a trend of 4.2 days earlier per decade from 1982–2011 [25, 143]. Similarly, between 1982–2011, spring onset advanced by 5.5 days per decade in China, significantly faster than the change observed in Europe and North America [25]. These trends are consistent with the ones estimated from remotely sensed products [25]. However, satellite-based studies indicate that spring onset trends have decreased since 2000 during the warming hiatus [25]. Concurrent with early spring onset trends, a delay in autumn is also observed in Europe and China, although these phenological changes have not been extensively studied as have changes in the spring onset [25]. Overall, the combined trends toward early spring and late autumn onsets indicate a tendency for longer growing seasons [25, 61]—though for winter crops in some areas, the growing season is shrinking [144]. While temperature is recognized as the most critical factor in phenology changes, other factors, including water and nutrient availability, ecological interactions, and photoperiod, are also critical modulating factors [25].

As plants respond to variations in temperature, precipitation, and soil moisture through phenological changes, they can feed back to the climate system by modulating the energy budget [25, 126]. Observational and modeling studies indicate that the direct impacts of phenological changes in a warming climate are related to altered, usually longer, growing seasons [25]. Longer growing seasons often result in more dense vegetation, increased transpiration rates, surface roughness, and changes in the surface albedo [126]. Higher surface roughness increases the exchange of energy and momentum between the land surface and the atmosphere by promoting turbulence [25, 126]. Plant phenology changes have been further associated with alterations in the hydrologic cycle by modulating evapotranspiration, thus influencing runoff, soil moisture, and precipitation [25].

Idealized simulations indicate significant changes in the planetary boundary layer by shifting the timing of plant phenology in North America [130]. Early spring onsets, for example, are associated with a decreased Bowen ratio and increased low-cloud fraction. Therefore, the land atmospheric coupling is mainly controlled by soil moisture in arid areas, while in areas with relatively high soil moisture, temperature is the main contributing factor [130].

5. Summary and conclusions

The effects of anthropogenic climate change on hydroclimate are apparent, as suggested by recent detection and attribution studies [12, 14–16, 19, 21, 24, 34, 66, 72]. Climate change is expected to alter global hydroclimate further as atmospheric greenhouse gas concentrations continue to rise in the twenty-first century [6, 7, 19, 57, 61, 114]. In summary, the recent literature indicates that:

a. Global hydroclimate changes in a warmer climate are driven by thermodynamic and dynamic controls (Fig 1) [34, 37]. These controls are further modulated by regional and local scale factors and feedback mechanisms including topography, the response of vegetation to increased atmospheric CO_2 [60, 61, 133], and land-atmosphere interactions [51, 52, 114]. Expected changes in the magnitude and distribution of $P–E$ resembles a "wet-get-wetter dry-get-drier" or "wet-events-wetter, dry-events-drier" pattern [9] and, while this mechanism is observed over the ocean, it is not happening over land, especially in moisture-limited regions [10]. Observations and simulations also indicate an intensification of the water cycle in the historical period, as revealed by higher $P–E$ rates [34, 37]. This is consistent with Clausius–Clapeyron [9], although increases in global mean precipitation are smaller than water vapor [35] due to atmospheric energy balance and dynamic constraints [37, 38]. As a result of these different responses, a slowdown of the tropical overturning circulation [41, 44, 46] is predicted. While this is expected based on simulations, discrepancies prevail in observations on whether the Walker Circulation has slowed in the instrumental era [5, 28, 41, 47]. A possible explanation is that current climate models do not correctly simulate the response of the Walker Circulation to warming [28, 47].

b. An increased frequency of extreme precipitation events and drought observed and attributed to anthropogenic climate change have occurred in Europe and the Mediterranean, southwestern North America, the Amazon, and southwestern South America, South and Central Africa, Australia, Southeast Asia, and the Caribbean [11, 12, 14–16, 18, 21, 69, 114]. Although a warming climate may not directly cause a greater number of droughts, it does contribute to making drought onsets quicker and magnitudes more severe [13] (Fig 1). Nevertheless, uncertainties persist in the observational records used as input data to calculate drought indices, the indices themselves, and how drought is defined in the first place [10, 13]. For example, PDSI calculated using different precipitation data yields different results of drought severity, duration, and trends [13, 18] (Fig 2B and 2D). Different approaches to estimating the potential evapotranspiration (e.g., Penman-Monteith or Thornthwaite) further yields disparate results even using the same temperature data [13] due to the sensitivity to temperature variations of each approach [13, 145, 146]. Another source of uncertainty is the sensitivity of drought indices to variations in precipitation versus evapotranspiration, and this is critical to determine the accuracy of drought risk projections during climate change [146]. For example, PDSI is less sensitive than SPEI to variations in evapotranspiration [146]. The reference or baseline period is also critical for any drought metric because it should include the full range of natural and human-forced variability [13] (Fig 2B and 2D). Some previous studies advise using the full-time interval as the reference period [145, 147], while others recommend a period in which the anthropogenic signal is more pronounced [14, 16, 18, 59, 98].

c. Albeit an improved plant water-use efficiency is expected to ameliorate drought risk as the climate warms [60, 133], the net effect of plant water savings is not yet fully constrained [34]. Recent modeling studies suggest that increased canopy water demand and longer growing seasons negate the water saving effects from increased water-use efficiency [61, 62]. This is consistent with observational studies that have found reduced streamflow and base flow in Australia, synchronous with increased GPP [127–129], which is consistent with the results from Mankin et al. [61].

d. While many detection and attribution studies indicate that anthropogenic climate change is shifting global hydroclimate [12, 70], uncertainties prevail mainly for three reasons: first, constraints in the quality, spatial and temporal coverage of instrumental records [6] on which detection and attribution methods rely [29, 30]. Second, biases of current climate models [27, 28]. Third, the strong influence of natural variability [59, 148–150]. Despite these limitations, a growing body of evidence suggests that anthropogenic climate change is altering the global hydroclimate [7, 12, 34, 70]. An emerging concern regarding the biases in climate models is that about one-quarter of the models from phase six of the Coupled Model Intercomparison Project (CMIP6) have sensitivities of ~10% higher than those of the fifth phase (CMIP5) [27, 151] (Table 1; Fig 3A). Those “hot models” [27] also have higher biases in simulating global temperatures in the historical period [151] and past climates [27]. With that in mind, one may question whether higher sensitivity models also suggest more drastic changes in hydroclimate in the twenty-first century. For example, do hot models project a higher drought risk than those with lower sensitivity? As we show in Fig 3, the projected surface soil moisture from low and high Equilibrium Climate Sensitivity models are similar (r = 0.97), at least from the eight CMIP6 models we compared (Fig 3A and 3B).

In conclusion, the evidence indicates that anthropogenic climate change has altered the global hydroclimate across the twentieth Century with an influence that is expected to grow in the future. However, uncertainties in quantifying the anthropogenic contribution to these changes persist due to a broad range of challenges. Addressing these challenges is necessary to

Table 1. The CMIP6 models used in Fig 3. We classified as “hot models” those with Equilibrium Climate Sensitivity (ECS) below 4.7 K/4xCO_2.

Model	ECS	Horizontal resolution (lon/lat)
CAMS-CSM1-0*	2.2	1.1˚ × 1.1˚
MIROC6*	2.5	1.4˚ × 1.4˚
GISS-E2-1-G*	2.6	2.5˚ x 2.0˚
NorESM2-LM*	2.7	2.5˚ × 1.9˚
MIROC-ES2L*	2.7	2.8˚ × 2.8˚
BCC-CSM2-MR*	2.9	1.1˚ × 1.1˚
GISS-E2-1-H*	3.0	2.5˚ x 2.0˚
MRI-ESM2-0*	3.3	1.1˚ × 1.1˚
BCC-ESM1*	3.3	1.1˚ × 1.1˚
SAM0-UNICON*	3.7	0.95˚ × 1.25˚
GFDL-CM4*	3.8	1.3˚ × 1˚
EC-Earth3-Veg*	4.0	0.7˚ × 0.7˚
NESM3^	4.5	1.9˚ × 1.9˚
IPSL-CM6A-LR^	4.5	2.5˚ × 1.3˚
CESM2-WACCM^	4.6	1.3˚ × 0.9˚
CNRM-ESM2-1^	4.7	1.4˚ × 1.4˚
CNRM-CM6-1^	4.8	0.5˚ × 0.5˚
CESM2^	5.1	1.3˚ × 0.9˚
UKESM1-0-LL^	5.3	1.9˚ × 1.3˚
E3SM-1-0^	5.3	1.0˚ × 1.0˚
HadGEM3-GC31-LL^	5.4	1.88˚ ×1.25˚
CanESM5^	5.6	2.8˚ × 2.8˚

* Models with ECS below 4.5 K/4xCO_2

^Models with ECS at or above 4.5 K/4xCO_2

Source: Pendergrass [152]

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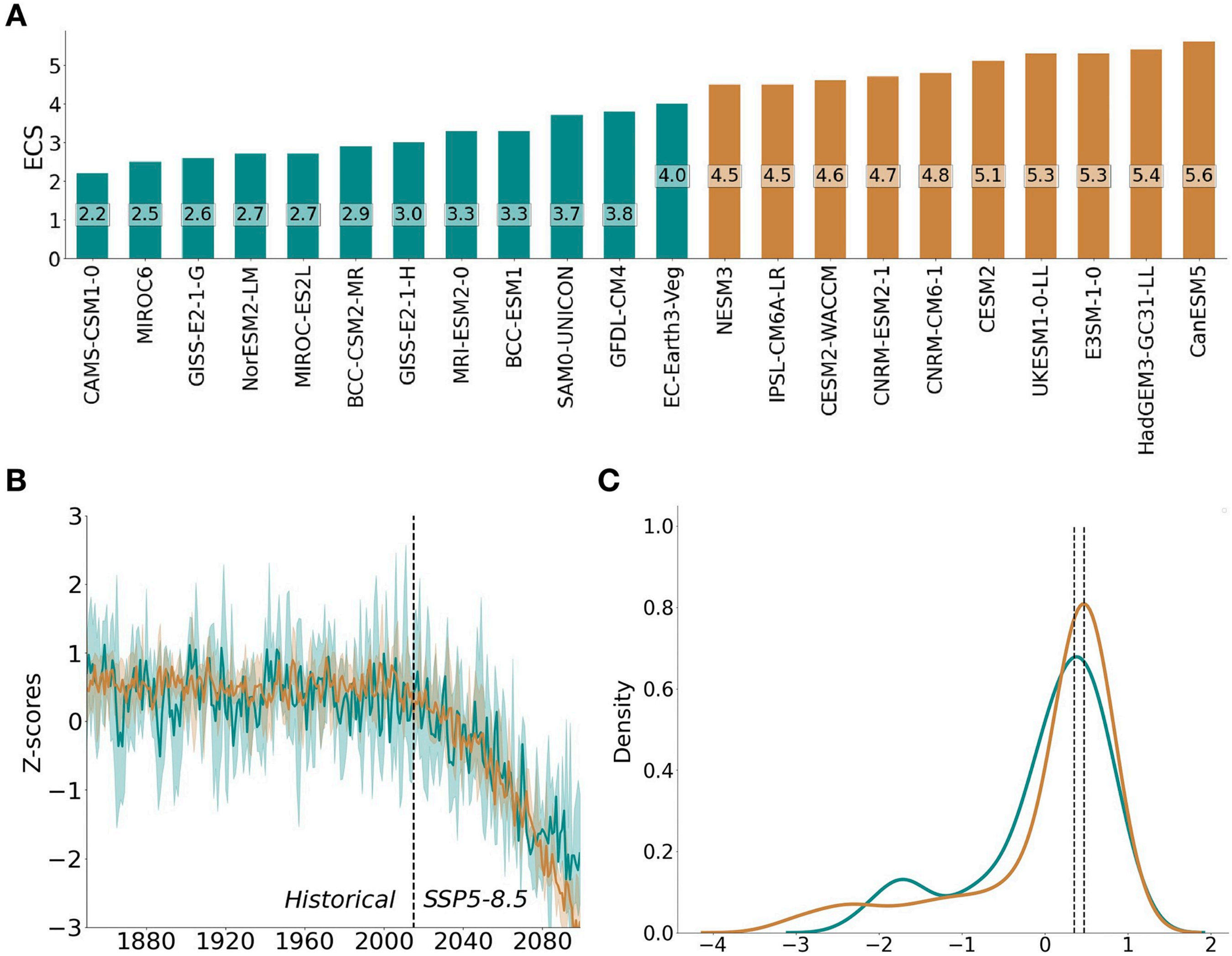

Fig 3. High and low Equilibrium Climate Sensitivity (ECS) models and projected soil moisture changes. **(A)** ECS of the phase 6th of the Coupled Project Intercomparison Project (CMIP6) from Table 1. (**B**) Historical surface soil moisture anomalies (baseline 1950–1980) appended to the SSP5-8.5 scenario from high and low ECS models. Simulated soil moisture anomalies are normalized as z-scores from CanESM5, CNRM-CM6-1, CESM2, and IPSL-CM6A-LR as high ECS models, and CAMS-CSM1-0, MIROC6, MRI-ESM2-0, and BCC-CSM2-MR as low ECS models. Shading represents the maximum and minimum z-scores in each of the CMIP6 models we used, while the dark lines represent the multimodel mean. **(C)** Kernel density estimates of z-scores from the high and low ECS models. Dashed lines represent the mean of each density curve.

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better anticipate changes in hydroclimate as CO_2 concentrations in the atmosphere continue to rise.

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