

REVIEW

On the essentials of drought in a changing climate

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Droughts of the future are likely to be more frequent, severe, and longer lasting than they have been in recent decades, but drought risks will be lower if greenhouse gas emissions are cut aggressively. This review presents a synopsis of the tools required for understanding the statistics, physics, and dynamics of drought and its causes in a historical context. Although these tools have been applied most extensively in the United States, Europe, and the Amazon region, they have not been as widely used in other drought-prone regions throughout the rest of the world, presenting opportunities for future research. Water resource managers, early career scientists, and veteran drought researchers will likely see opportunities to improve our understanding of drought.

Unlike most natural disasters, but like a disease, a drought begins before it presents any symptoms (1). To understand this, imagine that it is May of 2013 and that you are a farmer in the Caribbean. It has been a little dry recently but otherwise all seems well ahead of the summer rains. The weather is warm, the skies are clear, and the horizon has a yellowish hue from dust carried across the Atlantic from the far-off Sahel (2). Although you do not know it yet, the worst drought in at least half a century has already begun (2). Before it is over, it will persist for 3 years, push 2 million people into food insecurity, and affect nearly every island in the Caribbean (2).

In the United States, drought cost \$250 billion in damages and killed nearly 3000 people between 1980 and 2020, making it the costliest natural disaster and the second most deadly one (3). Over the last 12 centuries of human civilization, multidecadal megadroughts contributed to the demise of some of the most complex societies of the preindustrial era, including the Khmer and Mayan Empires, the Puebloan cliff dwellers of the southwestern United States, and the Yuan Dynasty of China (4). The Old Testament vividly describes drought as a punishment from God that left “Judah wailing, her cities languishing, the land cracked, and wild donkeys standing on barren heights, panting like jackals.” Adding, “Even the doe in the field deserts her newborn fawn because there is no grass” (Jeremiah 14).

Droughts of the future may eclipse those of past centuries in their duration, severity, and frequency (5, 6). Although aggressively cutting greenhouse gas emissions reduces these risks, even low levels of warming could amplify drought hazards across much of the world, including the Caribbean, Central America, Brazil, western Europe, central Africa, Southeast Asia, and Australia (6, 7).

Defining drought

Although the crisis of drought is easily recognized, there is no universally accepted criterion for what constitutes one (4, 8–10). Instead, multiple definitions, indices, and metrics exist to meet the particular needs of different research communities or applications (10). What they have in common was adroitly articulated by the late Kelly Redmond: They are intervals of time when “the supply of moisture fails to meet its demand” (9). Whereas the atmosphere delivers the supply of moisture, the demand for it arises from countless sources—a hot, dry atmosphere demands water vapor from the surface; plants demand water for transpiration; and our infrastructure demands water resources for irrigation, municipal water supply, and hydroelectric power generation, among many other uses.

Droughts are classified according to their impact (8, 10), which imposes an approximate time scale for each type. A meteorological drought stems from rainfall shortages over a period of weeks, whereas an agricultural drought exacts crop losses and may linger for months. A hydrological drought develops on seasonal to interannual time horizons by depleting streamflow or reservoir levels.

Socioeconomic droughts, which affect water resources required for human applications (e.g., municipal drinking water), arise from either a shortage of supply or an excess of demand (10). Although the rest of this review will focus on the physics of meteorological and agricultural drought in a changing climate, the basic ideas are broadly relevant to other types of droughts.

An analytical arsenal for drought research

A simple “bucket” model (Eq. 1) builds on the concept of drought as a phenomenon that arises from either a shortage of precipitation supply (P) or an excess of evapotranspiration demand (E) [e.g., (11) and references therein]:

$$P - E = \frac{dS}{dt} + R_o + G_w \quad (1)$$

where the terms on the right are changes in soil moisture storage (dS/dt), runoff (R_o), and groundwater flow (G_w) (11).

In principle, if we had observations of precipitation minus evapotranspiration ($P - E$), dS/dt , and R_o going back at least a century, then we could readily characterize drought variability on intraseasonal to multidecadal time horizons. In practice, only precipitation measurements are available from the past few decades, and those records are subject to large uncertainties that affect our understanding of drought (12). Measuring E and dS/dt accurately and consistently across space and through time has vexed drought scientists for generations (8, 13).

Drought indices

As an alternative to measuring soil moisture directly, drought indices track relative departures from normal conditions (14, 15). The full palette of drought indices available for researchers and water resource managers is described in other reviews (8, 10), and new indices are routinely added to this collection (16). Broadly, they fall into two categories: indices that track the supply of moisture from precipitation alone (17) and those that approximate the balance of moisture arising from the combined effects of precipitation, evapotranspiration, and, sometimes, storage (14, 15).

The Standardized Precipitation Index (SPI) (17) is designed to track precipitation deficits and surpluses across multiple time scales (e.g., 1, 3, or 12 months), making it ideal for differentiating between different types of drought (e.g., meteorological versus agricultural). However, the SPI’s exclusion of evapotranspiration limits its usefulness for some applications and research questions (15). The Standardized Precipitation Evapotranspiration Index (SPEI) (15) was developed to address this limitation while preserving the robust statistical features of the SPI.

Both the SPI and the SPEI emerged to fill a need for drought indices that was imperfectly carved out by the Palmer Drought Severity Index (PDSI) several decades earlier (14). Like the SPEI, PDSI approximates evapotranspiration demand, but it also accounts for moisture storage by different types of soils (14). The “self-calibrating” PDSI (18) is most appropriate for large-scale studies of drought variability and long-term change (12, 19–21). Even so, the magnitude of future change expected from the PDSI depends strongly on its formulation and the historical data used to calibrate it (21).

The SPEI and PDSI depend on simplified estimates of potential evapotranspiration (PET) that must be parameterized, and doing so accurately requires meteorological variables beyond precipitation (12, 21). Consider the widely used, physically based Penman-Montieth equation, which approximates PET as a function of net surface radiation (R_n), soil heat flux (G), water vapor pressure deficit ($e_s - e_a$), slope of the temperature-saturation vapor

pressure relationship (Δ), psychrometric constant (γ), and two resistance terms (r_s , for surface resistance, and r_a , for atmospheric resistance) (22):

$$ET = \frac{\Delta(R_n - G) + \rho_a c_p \frac{e_a - e_u}{r_a}}{\Delta + \gamma(1 + \frac{r_s}{r_a})} \quad (2)$$

where ρ_a is the density of air and c_p is the heat capacity of dry air.

Use of Eq. 2 requires temperature, humidity, surface pressure, net radiation, and wind speed data (22). Of these, only temperature is widely available across large spatial scales and going back more than a few decades (12). Using the Penman-Montieth equation (Eq. 2) to study drought at continental scales therefore usually entails merging gridded observational datasets with reanalysis products (12); errors in these observational fields will introduce uncertainties into drought indices computed from them (12).

Nevertheless, PDSI and SPEI (as well as others) can be computed from observational and model output alike, which, ostensibly, allows projections of the future to be compared against historical conditions using the same indices for both data products (5, 12, 20, 21).

piration, including lateral flow and subsoil storage of moisture in the rock layer (23, 24).

More sophisticated land surface models (LSMs) assimilate data from multiple sources to estimate historical variations in land surface hydrology (25). However, as with drought indices, observational uncertainties affect the quality of soil moisture data in LSMS (26), and appropriate observational boundary conditions only span 1979 to the present (25). Consequently, LSM output covers a short and heavily forced period of the recent past, which presents a challenge for detecting and attributing the imprint of climate change in soil moisture (27).

An advantage of LSMS, however, is that they simulate the moisture, energy, and biogeochemical fluxes between the atmosphere and the land surface, just as the land surface components of general circulation models (GCMs) do. LSMS therefore also serve as an important bridge between observational data and climate model simulations of the past, present, and future.

Finally, over the past decade, observations of soil moisture from either in situ measurements (28) or remote sensing (29) have emerged as invaluable tools for validating LSMS and monitoring drought. However, these products cover

$$P - E = -\frac{1}{g} \left(\frac{\partial \phi}{\partial t} + \nabla \cdot \int_0^{p_s} q \vec{V} dp \right) \quad (3)$$

with total precipitable water, ϕ , defined as the vertical integral of water vapor:

$$\phi = \int_0^{p_s} q dp \quad (4)$$

Changes to the $P - E$ balance of Eq. 3 must originate from one of two sources: (i) localized fluxes of precipitation or evaporation (i.e., the first term inside the parentheses on the right) or (ii) the convergence or divergence of vertically integrated moisture flux (i.e., the second term inside the parentheses). This second term can be further decomposed into separate changes originating from the mean flow, transport by transient eddies, divergence of the high-level winds, and advection of moisture gradients by the lower atmosphere (30).

In addition to atmospheric moisture budgets, idealized numerical modeling experiments serve as invaluable tools for investigating the origins of drought (31). These experiments typically force a free-running atmosphere with prescribed sea surface temperature (SST) anomalies that are hypothesized to cause drought (31). Running multiple atmospheric simulations, all of which are forced with the same SST field, and then averaging these simulations together disentangles the SST “signal” in droughts from the atmospheric “noise.”

Causes of drought

The general circulation of the atmosphere delivers moisture from the world's oceans to its continents. Some of that moisture becomes trapped in glaciers, aquifers, and lakes; the rest flows through soils, plants, and rivers. Drought occurs from aberrations to the flow of moisture through these terrestrial systems.

The largest disruptions to the global hydrological cycle occur during the El Niño and La Niña events (Fig. 1) (31–33). For example, El Niño displaces tropical rainfall in northeast Brazil, Central America, and the Caribbean, causing drought in those regions (32). Meanwhile, the areas that normally see strong convection, such as Indonesia and northern Australia, also experience rainfall shortages, crop losses, and wildfires (34, 35).

Although the El Niño–Southern Oscillation's (ENSO's) impacts on the global climate were recognized decades ago, moisture budgets and idealized SST-forcing experiments have now revealed key details of the dynamical processes responsible for those teleconnections (31, 33). Winter storms shift equatorward during El Niño years (36) because deep convection modifies the structure and flow of the storm tracks and hence the transport of moisture (36). This in turn can trigger drought in the Pacific Northwest and the

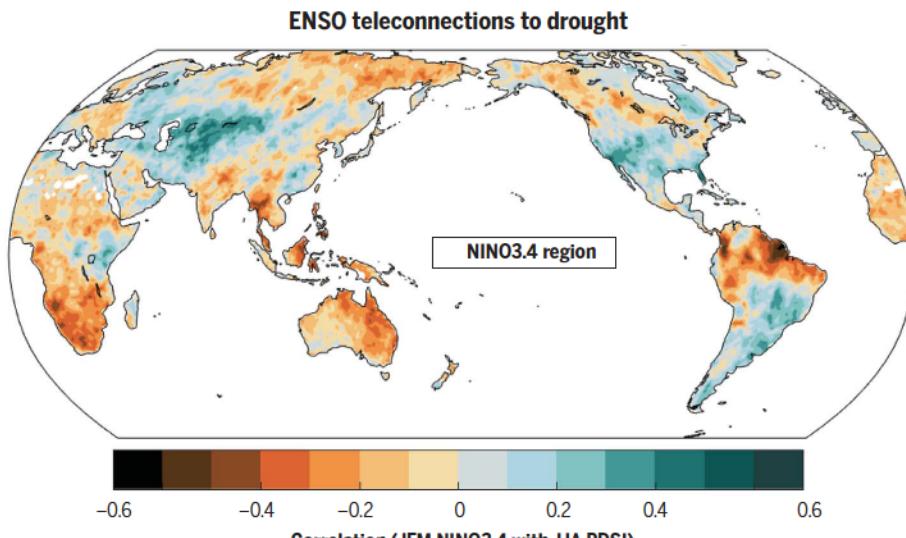


Fig. 1. Correlation coefficients between NINO3.4 SST (ERSSTv3b) and self-calibrating PDSI (30). All correlations are computed between boreal winter [January–February–March (JFM)] SSTs with PDSI during the following boreal summer [June–July–August (JJA)].

Modeling soil moisture

Given the apparent simplicity of Eq. 1, one might be tempted to model soil moisture directly using meteorological variables as boundary conditions, thus circumventing the need for drought indices (11). For example, the simplified bucket model extends global soil moisture estimates back to 1948 (11), but it lacks a number of critical processes that affect evapotrans-

a relatively short time period; they do not provide much information about interannual, let alone decadal, variations during the historical period.

Diagnosing drought dynamics

Atmospheric moisture budgets express the local balance of $P - E$ as a function of specific humidity (q) and horizontal winds (\vec{V}) (30):

southeastern United States owing to additional downstream effects (36).

On shorter time scales, seasonal modes of variability such as the North Atlantic Oscillation (NAO) can modify storm tracks crossing the Atlantic (37). During the positive phase of the NAO, winter storms crossing the Atlantic

tend to make landfall at higher latitudes (e.g., the United Kingdom and Scandinavia), which in turn favors drier conditions across France, Spain, Italy, and the Mediterranean region in general (37, 38).

Over longer time scales, decadal SST variability appears to be connected to drought (31),

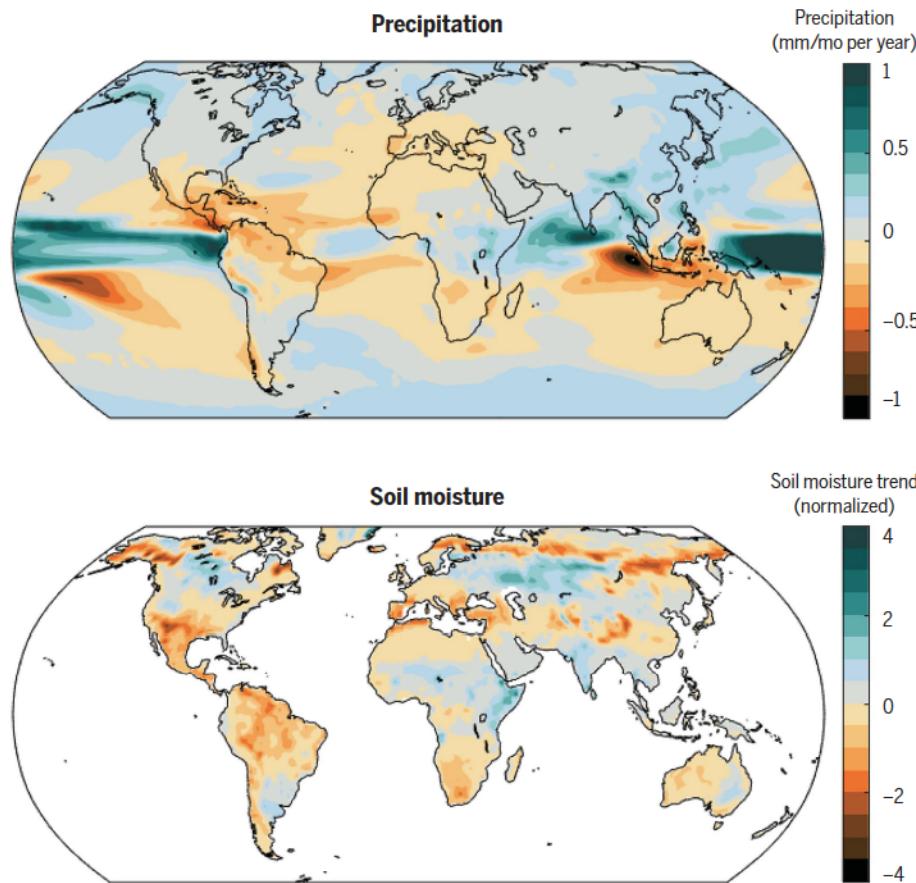


Fig. 2. Annual precipitation totals (blue) and JJA volumetric soil moisture averages (brown) from the CESM large ensemble (50).

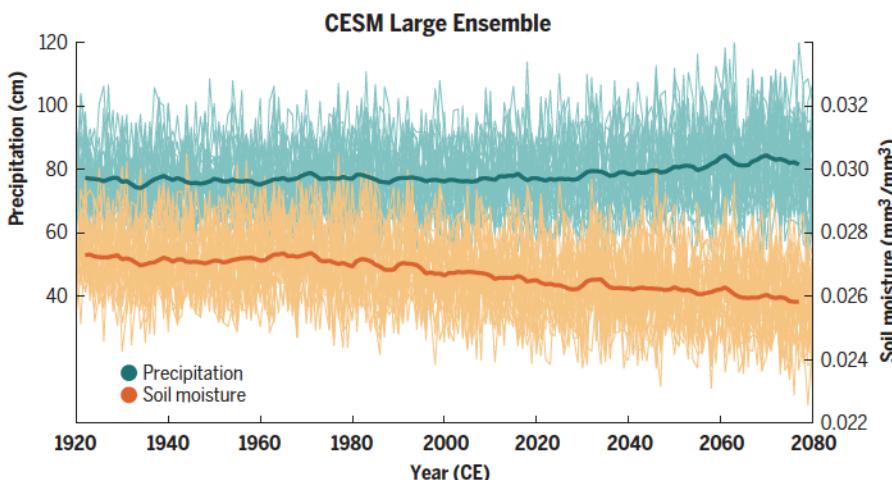


Fig. 3. Ensemble-averaged 21st-century Climate Model Intercomparison Project V (CMIP5) trends computed from annual precipitation (top) and column-integrated soil moisture (bottom).

although it can be difficult to disentangle such long-term effects from anthropogenic forcing (which may also affect decadal SST variations) (31).

Atmospheric moisture budgets also serve as invaluable tools for evaluating the realism of GCMs and for diagnosing their predictions of future aridity (39), although this remains a relatively underexplored area for future research.

Back to the future

If you are a water resource manager and you remember just one thing from this review, it should be this: Cutting CO₂ emissions reduces drought risk (6, 7). In many regions, including Central America, the Caribbean, the Amazon, Western Europe, and southern Africa, avoiding even just half a degree of warming makes a difference: Regional drying is more severe if global warming reaches 2.0°C than if it is curtailed at 1.5°C (6, 7).

Climate change alters the balance of moisture throughout the world by disrupting its supply through changes in the general circulation (39, 40). Meanwhile, higher temperatures can increase moisture demand from the land surface (12, 41) for the same reason that a sauna will dry out a towel faster than a steam room (see Eq. 2). Accordingly, regions seeing both a decrease in supply and an increase in demand are very sensitive to even low levels of warming (6).

Plants, however, may use water more efficiently as CO₂ concentrations increase in the atmosphere (42), and this “CO₂ fertilization effect” might partially offset a portion of future drying predicted for some regions (42–44). Nevertheless, there are several examples of models that predict reductions in soil moisture despite increases in overall precipitation and an increase in water use efficiency by plants (5, 42, 44). That is, the improvements in efficiency from higher CO₂ concentrations reduce the total amount of drying, which is substantial, but they do not reverse it (Fig. 2).

Finally, ENSO will likely continue to disrupt hydroclimate across vast spatial scales (32). When it does, the impacts of El Niño on drought could be even more severe than they are today for two reasons: (i) We expect climate change to strengthen ENSO events (45, 46), and (ii) a hotter atmosphere demands more moisture from the land surface when droughts occur (41, 47). Even now, higher temperatures may already be worsening aridity beyond anything seen in the past few centuries (27).

The future of drought research

Legions of studies have used the analytical arsenal described earlier to confront fundamental questions about the physics, dynamics, and risks of drought in a changing climate. For example, they have asked:

- 1) How do future droughts and long-term changes in aridity compare with modern-day

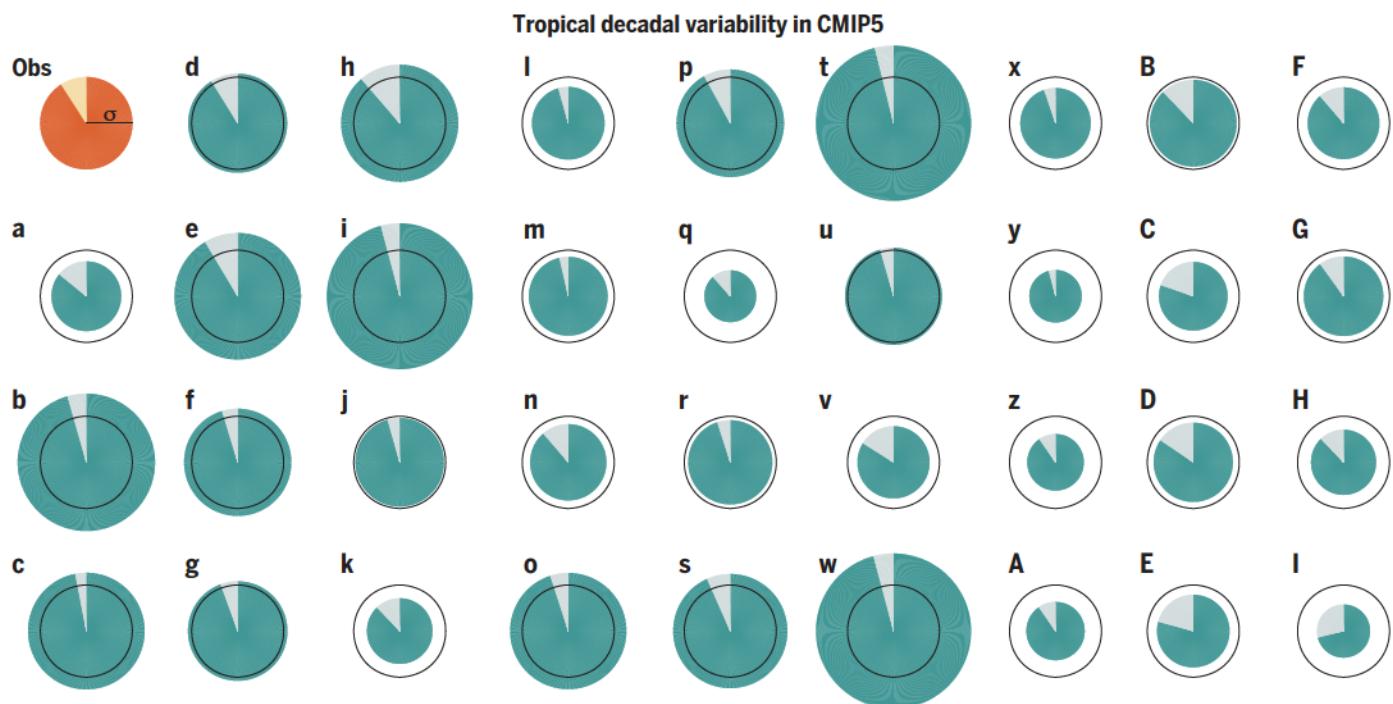


Fig. 4. Summary of tropical Pacific variability in observations and models.

The circle in the topmost left is derived from observational data, with the radius being proportional to the standard deviation ($\pm 1.5^{\circ}\text{C}$) of NINO3.4 and the lighter color representing the fraction of the total variance that occurs on decadal time scales in the NINO3.4 region ($\sim 8.5\%$). The remaining charts summarize this same information for individual members of the CMIP5. Again, the radii of each circle are proportional to the standard deviation of a given model, and the lighter color represents the fraction of total variance occurring on decadal time horizons. For reference, the outline of the observational NINO3.4 chart is included on each model diagram. Variability in circle size emphasizes the now well-known differences in El Niño and Southern Oscillation amplitudes across the CMIP5 archive. Differences in the fraction of variance occurring

on decadal time scales, in conjunction with differences in ENSO amplitudes, have received less attention. The models are identified by a single letter as follows:

(a) ACCESS1.3; (b) BNU-ESM; (c) CCSM4; (d) CESM1-BGC; (e) CESM1(CAM5-FV2); (f) CESM1(CAM5); (g) CESM1(FASTCHEM); (h) CESM1(WACCM); (i) CMCC-CESM; (j) CMCC-CMS; (k) CMCC-CM; (l) CNRM-CM5-2; (m) CNRM-CM5; (n) CSIRO-Mk3.6.0; (o) CanCM4; (p) CanESM2; (q) EC-EARTH; (r) FGOALS-g2; (s) FIO-ESM; (t) GFDL-CM2.1; (u) GFDL-CM3; (v) GFDL-ESM2G; (w) GFDL-ESM2M; (x) GISS-E2-H-CC; (y) GISS-E2-H; (z) GISS-E2-R-CC; (A) GISS-E2-R; (B) HadCM3; (C) HadGEM2-AO; (D) HadGEM2-CC; (E) HadGEM2-ES; (F) IPSL-CM5A-LR; (G) IPSL-CM5A-MR; (H) IPSL-CM5B-LR; and (I) MIROC-ESM-CHEM. When multiple realizations were available, only the first simulation was used.

conditions (19–21, 41)? Which indices and models should be used to characterize future droughts (21, 42, 43)? Is there already a *detectable* imprint of anthropogenic climate change on global drought (19, 20, 27)?

2) How will regional changes in temperature affect moisture demand from the atmosphere through evapotranspiration (42, 43)? What role does vegetation play in coupling the land surface to the atmosphere (42–44)?

3) How will the supply of moisture to land evolve in response to large-scale changes in the general circulation (39)? How will ENSO and other seasonal variations influence drought in the future (46, 48)?

In addressing these questions, researchers have begun assembling the puzzle of drought risks in a changing climate. Many regions may face events that are more severe, frequent, and long-lasting than those of the recent past (13, 21, 26) or even the last millennium (4, 5, 27). However, not all of the pieces fit together.

Wet, hot American summer drought

Perhaps the most contentious debate among drought researchers stems from differences

between drought indices (as described above) computed from GCMs and soil moisture simulated by those same models. Drought indices depict unprecedented drying throughout much of the United States (5, 20, 27), but these indices do not account for biological processes (such as CO_2 uptake) that may alter the surface moisture balance in the future (42, 43). They are also sensitive to the length and quality of historical data used to calibrate them (12, 21) and may distort the magnitude of future changes if they are not calibrated appropriately (12, 19–21). Finally, their reliance on the Penman-Montieth equation might overestimate future PET rates (43).

Soil moisture projections from LSMs help to characterize some limitations of drought indices, although they have their own pitfalls. For example, soil moisture data are not widely available in most regions, making it difficult to directly compare LSM output against the historical record (26). LSMs typically overestimate evapotranspiration rates (49), which in turn makes them too strongly coupled to the atmosphere, and artificially enhance precipitation in some regions (49). Although they can simulate CO_2 “fertilization” in plants, their mod-

ules for representing ecological interactions among plants, soil moisture, and runoff all introduce new uncertainties that propagate into their projections of the future (26, 44).

Although GCM-based drought indices and soil moisture variables do not paint an entirely consistent picture of future drying, their differences may be superficial (44) (Fig. 2). In the case of the Community Earth System Model (CESM) “large ensemble” (50), the apparent paradox of increased drought risk in a wetter climate is easy to reconcile from the perspective of soil moisture balance: The increase in demand for evaporation from higher temperatures exceeds the increase in supply from precipitation. Future research could elaborate on these details in other models and other parts of the world.

Expanding outward

Quantifying how uncertainties in the large-scale circulation of GCMs are manifest in regional predictions of drought presents a harder problem for researchers (51). For example, GCM simulations of the 21st century depict a scenario in which the subtropics become drier

but wet equatorial regions become rainier (40) (Fig. 3). In a general sense, this subtropical drying is a robust thermodynamic response to higher temperatures: A warmer atmosphere can “hold” more water vapor, yet the rate at which water vapor increases in the atmosphere outpaces the rate of precipitation increase (39, 40). Accordingly, less moisture evaporates from the ocean to meet the demand for precipitation, which slows tropical circulation (40).

Most of the slowdown in tropical circulation occurs in the meridional Hadley cells, causing them to widen (40), which dries the subtropics. GCMs predict a similar outcome over the tropical Pacific Ocean because the east-west Walker circulation should also slow in conjunction with the Hadley cells (52). However, this is not happening (53)—or if it is happening, recent trends in the historical record are being dominated by other processes. One possible cause for this discrepancy is that the Walker circulation is responding differently in reality than it does in models to greenhouse gas forcings (48). That is, the recent observed changes are a forced dynamical adjustment in the coupled ocean-atmosphere system that the GCMs do not capture (48).

Alternatively, substantial internal *decadal* variability in the equatorial Pacific Ocean may be overshadowing the forced response of the Walker circulation (54). On this point, models do not agree with each other, let alone with the observations, on the relative importance of decadal variability in the tropical Pacific (Fig. 4).

For the time being, the issue must be regarded as unresolved. However, its resolution is vital to our portrait of 21st-century drought risk because the structure of the Walker circulation affects rainfall throughout the world (39).

The issues described above will manifest in the mean moisture balance of the tropics and subtropics, but droughts of the future will be caused by both the long-term changes in the general circulation and short-term deviations during El Niño and La Niña events (in addition to other modes of climate variability). Again, GCMs do not agree with one another, nor the historical record, on the amplitude of ENSO fluctuations and the relative importance of decadal variability (Fig. 4) (55, 56). Because the tropical Pacific exerts a major influence on global precipitation patterns (32) (Fig. 1), frequency biases in this region likely affect the statistics of precipitation in regions with strong ENSO teleconnections. Quantifying the relationship between ENSO frequency biases in GCMs (as well as potential changes in ENSO frequency) and drought presents an important area for future research.

New additions to the analytical arsenal

During the past 30 years, intellectual and technological breakthroughs accelerated the pace of drought research. In the 1990s, personal com-

puters enabled scientists to develop, analyze, and deploy our current generation of drought indices. In the early 2000s, investments in high-performance computing and land surface models helped lay the foundation for the sophisticated LSMs used today. In the 2010s, satellites began making unprecedented global measurements of surface soil moisture (29). Although all these technologies brought powerful tools into our analytical arsenal, they are not very egalitarian. Most farmers living in the Majority World must confront the hazards of drought in a changing climate with little, if any, access to the technological advancements of recent decades.

Encouragingly, the late 2010s also introduced very low-cost soil moisture sensors, which are already being deployed through public partnerships with local communities (57). These sensors transmit information about the state of the land surface continuously and nearly instantaneously, and researchers can use this data to validate satellite retrievals, initialize near-term predictions, or study the flow of moisture through the land surface with an unprecedented density of in situ measurements. At the same time, local communities are able to use data from those devices to gain insight into current conditions. During the 2020s, this emerging “Internet-of-things” technology could become the new frontier of drought monitoring and modeling.

Imagine, again, that you are a farmer in the Caribbean during a drought, but this time, the year is 2035. It is exceptionally hot (7), aquifers are depleted (58), and there are frequent blackouts because reservoir levels are so low at the hydroelectric power plant (59). What would you ask us—the people alive today—to do now to ensure that you are resilient in the face of drought in a changing climate?

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