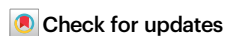


Asymmetric temperature responses to soil moisture drought legacies under anthropogenic forcing

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As anthropogenic forcing intensifies, soil moisture droughts have become prevalent across many regions. Soil moisture anomalies can influence near-surface air temperature by modulating both energy and carbon fluxes. However, the annual temperature response to soil moisture anomalies in the year following a soil moisture drought remains unclear. Here, we show that annual soil water loss in the year after soil moisture droughts induces a larger annual warming effect than the annual cooling from an equivalent water gain. This asymmetric response arises only when the effects of anthropogenic greenhouse gas forcing are included, where vegetation is more strongly suppressed after soil moisture drought than enhanced after soil moisture gaining, contributing to reduced transpiration and carbon uptake that collectively amplify warming. More than 60% of global land areas exhibit the asymmetry. Our findings suggest that past climate extremes may accelerate future warming by impairing vegetation's capacity to regulate biophysical and biogeochemical processes.

Soil moisture droughts have become more frequent and severe across many regions^{1–3}, exerting profound impacts on temperature^{4,5}. For instance, through land-atmosphere interactions, soil moisture anomalies can modulate surface energy fluxes and thereby influence temperature⁵. With soil moisture droughts projected to intensify under continued anthropogenic forcing^{6,7}, their impacts on temperature are expected to strengthen⁸, potentially amplifying climate warming⁹. Such amplification could not only undermine progress toward climate targets but also increase risks to human well-being and societal resilience^{3,10}. Understanding how soil moisture droughts affect temperature is therefore crucial, both for refining climate projections and for informing mitigation and adaptation.

While the concurrent impacts of soil droughts on temperature have been documented^{8,11}, their legacy effects, particularly those linked to soil moisture dynamics after soil moisture droughts, have received limited attention. A major challenge in assessing these legacy effects is the divergent trajectories of post-drought soil moisture and associated energy and carbon fluxes. In some cases, soil moisture droughts may exhibit memory effects that persist into the following year¹². When soil moisture fails to fully recover^{13–15}—remaining below pre-drought conditions in the subsequent year (hereafter, water loss; Supplementary Fig. 1a)—carryover water stress can reduce vegetation growth and transpiration^{12,16}, thereby promoting warming. In contrast, rising hydroclimatic variability has increased the frequency of abrupt

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transitions from dry to wet extremes^{17,18}. In such cases, the post-drought recovery of soil moisture (hereafter, water gain; Supplementary Fig. 1b) could promote the timely recovery of vegetation growth, thereby compensating for soil moisture drought-induced negative impacts and enhancing evaporative cooling¹⁹. Beyond biophysical impacts, water loss may also reduce ecosystem net biome production, increasing carbon release from land to atmosphere and intensifying positive carbon-climate feedback. These divergent outcomes raise the question of whether post-drought water anomalies yield symmetric or asymmetric impacts on temperature under enhanced anthropogenic forcings. Assessing this asymmetry is essential for understanding how past climate extremes influence ongoing warming.

Here, we diagnostically investigate how annual temperature responds to soil moisture anomalies in the year following a soil moisture drought, using reanalysis and observation-driven products spanning 1980–2024 (“Methods”). Specifically, we quantify the general sensitivity of annual temperature to soil moisture anomalies by computing linear regression coefficients between temperature and either water loss or gain after one year of soil moisture drought. To assess the role of human influence, we next attribute differences in temperature responses to anthropogenic forcings using the outputs from the Coupled Model Intercomparison Project Phase 6 (CMIP6).

Results

We observe a significant ($P < 0.01$) asymmetric temperature response to post-drought soil moisture anomalies. Specifically, when water loss and gain are of comparable magnitude, post-drought water loss induces stronger warming than water gain induces cooling (Fig. 1). The general sensitivity of temperature to water loss is about -0.13 K, meaning that one standard deviation decrease in soil moisture would translate to a 0.13 K temperature increase (Fig. 1a). In contrast, the temperature sensitivity to an equivalent increase in soil water is only -0.06 K (Fig. 1b). This asymmetry is statistically significant ($P < 0.01$), as shown by Monte Carlo experiments and the Mann-Whitney U-test (Fig. 2a). Additionally, our extensive sensitivity analyses support the robustness of this asymmetric temperature response, including analyses with individual reanalysis products, alternative study periods, and different for identification thresholds soil moisture drought (Supplementary Fig. 2). Moreover, temperature deviations at high water anomaly levels may reflect possible enhanced land-atmosphere interactions, where extreme soil moisture anomalies yield disproportionate temperature responses^{5,8}. To analyze the possible effects of nonlinear trends on the asymmetry, we remove nonlinear trends from both soil moisture and temperature anomalies using smoothing splines prior to identifying soil moisture drought. We again observe an asymmetric response (Supplementary Fig. 3). The absolute magnitude of the sensitivity is somewhat reduced relative to the analysis using non-detrending data, indicating that long-term climate trends may amplify the overall strength of the response. Moreover, we randomly select a series and calculate the temperature sensitivity. We find that the asymmetric pattern does not appear in the random-year experiment (Supplementary Fig. 4), indicating that the observed asymmetry is associated with post-drought conditions. Together, these analyses support the robustness of this asymmetric temperature response.

Shifts in soil moisture regimes reflect the growing influence of human activities, with anthropogenic forcing driving the increasing prevalence of soil moisture drought globally^{1,20}. Under more frequent soil moisture droughts, water loss could be more effective at enhancing surface warming, whereas the cooling effect of water gain could be constrained. The observed asymmetry may therefore be linked to anthropogenic forcing. To examine the possible necessity of anthropogenic forcing, we analyze simulations under different forcing scenarios, including both anthropogenic and natural (ALL), only greenhouse gas (GHG), natural (NAT), and aerosol (AER) forcings (Fig. 2). GHG simulations well reproduce the observed asymmetric

pattern, with magnitudes comparable to both ALL simulations and observations. In contrast, NAT and AER simulations show the opposite pattern, with stronger cooling from water gain than warming from water loss (Fig. 2a, Supplementary Fig. 5). These findings suggest that the observed asymmetric temperature response to post-drought water anomalies can be explained only when the effects of anthropogenic greenhouse gas forcing are included, but cannot be explained by natural climate variability or aerosol forcing alone. Notably, the warming sensitivity to water loss in ALL and GHG simulations is slightly weaker than in observation, possibly reflecting an underestimation of the temperature response to extreme negative soil moisture anomalies by Earth System Models (ESMs) (Fig. 1a, Supplementary Fig. 5a). The opposite asymmetry in NAT and AER simulations likely arises from the absence of greenhouse-gas-driven soil moisture depletion²¹ and the cooling influence of aerosols²², which could dampen the warming effect of water loss while amplifying the cooling effect of water gain. Moreover, sensitivities under ALL simulations differ from the combination of single-forcing experiments, as the standard deviation used in calculating the linear slope is not linearly additive. Most models across ALL simulations reproduce the asymmetric temperature response, while a subset of models do not (Supplementary Table S1). This difference appears to depend on how vegetation stress, recovery processes, and subsurface soil water buffering are represented and coupled to transpiration and surface energy partitioning within the models.

Since frequent soil moisture droughts under anthropogenic greenhouse-gas forcing influences vegetation activity—which regulates both energy and carbon fluxes^{5,23,24}—the observed asymmetry may be mediated by vegetation dynamics. To examine this possibility, we calculate the sensitivities of leaf area index (LAI) and related variables to post-drought soil moisture anomalies. These sensitivities are derived from ESMs under ALL simulations, which provide sufficient variables to explore potential mechanisms by incorporating current knowledge of various land surface processes and their interactions with the atmosphere²⁵. We also find significant ($P < 0.01$) asymmetries in their response to post-drought water anomalies (Fig. 2b–e). As expected, water loss reduces LAI, while water gain increases it, with both sensitivities being positive (Fig. 2b). However, the sensitivity to water loss is substantially stronger than to water gain. This aligns with the observed declines in forest recovery rates under rising temperatures and water stress²⁶. Transpiration and sensible heat, as key biophysical mediators of land surface energy balance, are tightly linked to vegetation activity and thus modulate temperature. Similar to LAI, the sensitivities for both transpiration and sensible heat show marked asymmetries (Fig. 2c, d). Furthermore, net biome production, representing the net carbon exchange between terrestrial ecosystems and the atmosphere, is more strongly reduced by water loss (less CO_2 absorption) than it is enhanced by water gain (Fig. 2e), implying that prior water stress could continue to suppress ecosystem carbon uptake even after soil moisture begins to recover²³.

Satellite LAI further suggests this asymmetric vegetation response (Supplementary Fig. 6a–d). LAI and related variables also exhibit consistent asymmetries under GHG forcings (Supplementary Fig. 7). Moreover, we analyze two ESMs, which do not simulate interannual variations in LAI. Notably, the asymmetric temperature response to post-drought soil moisture anomalies becomes weak in these simulations (Supplementary Fig. 8). These results are consistent with the role of vegetation dynamics in mediating the observed asymmetry. However, the analysis based on the two ESMs provides only suggestive support that vegetation dynamics can contribute to the asymmetry, rather than constituting a controlled attribution experiment. Together, the asymmetric temperature response manifests only under the influence of anthropogenic greenhouse gas forcing, stemming from the asymmetric vegetation response to post-drought water anomalies. Soil moisture drought legacy effects suppress vegetation growth,

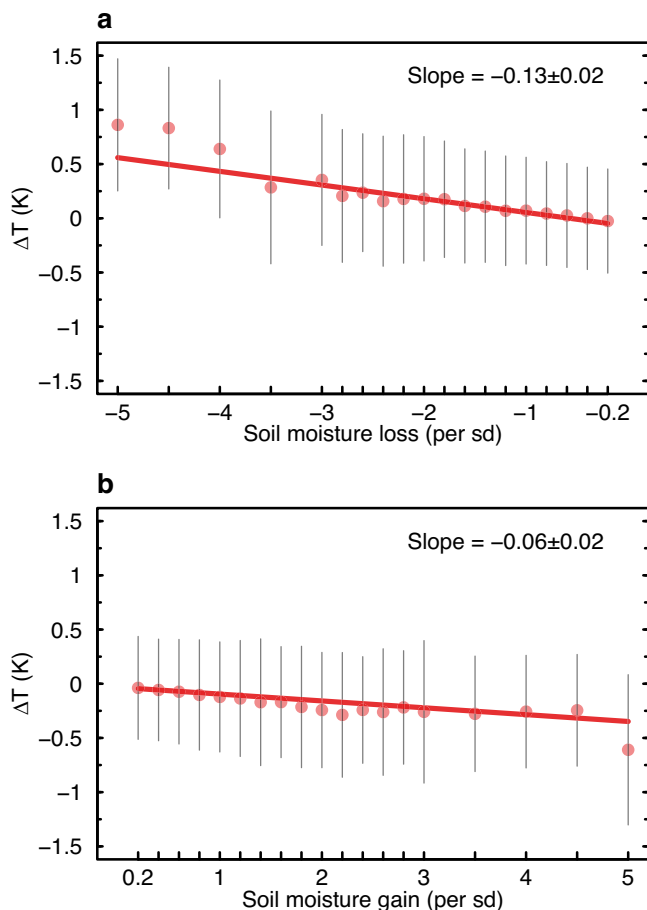


Fig. 1 | Temperature responses to soil moisture anomalies after one year of soil moisture droughts from 1980–2024. a Observed impacts of post-drought water loss on temperature. The y-axis indicates temperature change associated with water loss, based on reanalysis and observational products, while the x-axis represents changes in standardized soil moisture. Light red dots show the median temperature responses of all samples within each bin, and error bars indicate one standard deviation. Bin widths are 0.2 for soil moisture anomalies above -3 , and 0.5 for anomalies below -3 . The red line shows the linear regression between binned median temperature change (light red dots) and water loss. The upper right corner shows the linear regression coefficient and the mean standard deviation, representing the general sensitivity of temperature to post-drought water loss (“Methods”). **b** Same as (a), but for observed impacts of post-drought water gain on temperature.

linking to reduced transpiration and weakened carbon uptake, both of which exacerbate warming. In contrast, the beneficial effects of water gain are modest, likely limited by the time needed for vegetation recovery and impaired physiological functions^{15,26,27}.

Regarding spatial patterns, we further analyze the spatial distribution of temperature sensitivities to post-drought soil moisture anomalies. The results show that the asymmetric response is spatially widespread. Specifically, more than 60% of global areas exhibit stronger warming associated with post-drought water loss than cooling associated with comparable water gain (Fig. 3a). Pronounced hot-spots of asymmetry are observed in North America, eastern China, Africa, and the western Amazon, regions where land-atmosphere coupling and vegetation-water interactions can be particularly strong. These patterns indicate that the asymmetric temperature response emerges across diverse hydroclimatic regimes rather than being dominated by a few localized regions. The spatial distributions of asymmetry simulated under the ALL and GHG forcing experiments generally resemble those derived from observations (Fig. 3b, c). In

contrast, simulations driven solely by AER show a much weaker ability to reproduce the observed spatial structure of asymmetry (Fig. 3e). This discrepancy suggests that anthropogenic greenhouse-gas forcing plays a key role not only in shaping the global mean asymmetry but also in modulating its regional expression. Meanwhile, some regional asymmetries appearing in NAT simulations indicate that natural climate variability may contribute to local patterns, while it cannot explain the globally coherent signal observed in the data. Consistent spatial asymmetries are also found in vegetation-related variables, including LAI, transpiration, and sensible heat flux, which display similar spatial patterns. These results suggest that vegetation dynamics act as an important mediator linking post-drought soil moisture anomalies to regional temperature responses. Together, the spatial analysis confirms that the asymmetric temperature response is a spatially widespread phenomenon.

Discussion

Our analyses demonstrate that post-drought water loss induces a more substantial warming effect than the cooling from an equivalent water gain (Supplementary Fig. 13). This imbalance arises only under the influence of anthropogenic greenhouse gas forcing, driven by vegetation’s asymmetric response to water anomalies following soil moisture drought. Specifically, soil moisture drought legacies suppress vegetation growth more than water gain enhances it, contributing to reduced transpiration and weakened vegetation carbon uptakes that collectively amplify warming. This asymmetry suggests that past climate extremes accelerate future warming by impairing vegetation’s ability to regulate biophysical and biogeochemical processes under anthropogenic forcing. Because global temperature targets correspond to a finite cumulative carbon budget, such legacies could add uncertainty in estimating the near-term carbon budget and meeting climate targets, highlighting the need for mitigation policies that account for warming amplified by past climate extremes. Since limiting climate warming to specific levels implies a finite carbon budget²⁸, the legacy-induced warming identified here may shrink the remaining allowable emissions, reinforcing the urgency of stringent emissions cuts.

The contrasting responses across forcing experiments suggest that the asymmetric temperature response may not simply arise from generic drought legacy effects, but from their amplification under GHG simulations. While soil moisture drought legacies can occur under different climate states, GHG forcing can alter the background hydroclimatic conditions by increasing temperature, atmospheric evaporative demand, and vegetation water stress. Under these warmer and drier conditions, ecosystems can cross the threshold of hydraulic physiological failure, making vegetation recovery following soil moisture drought increasingly constrained and asymmetric. In contrast, under NAT forcing, background evaporative demand is lower, and vegetation generally operates farther from hydraulic stress thresholds, allowing more symmetric ecosystem recovery. Under AER forcing, aerosol-induced radiative cooling²⁹ can reduce surface energy input and evaporative demand, dampening warming responses to water loss while enhancing cooling associated with water gain. Together, these results suggest that increasing GHG forcing can intensify soil moisture drought legacies and strengthen the asymmetric temperature response.

Consistently, soil moisture droughts under GHG forcings can affect vegetation physiological and ecological processes—including hydraulic stress, carbon allocation shifts, phenological delays, and altered resilience trajectories—which may contribute to the asymmetric temperature responses. Severe soil moisture drought can push vegetation toward hydraulic failure thresholds, reducing hydraulic safety margins. Even when soil moisture partially recovers, hydraulic repair is often slow and incomplete^{30,31}. As a result, stomatal conductance may remain constrained, suppressing transpiration during

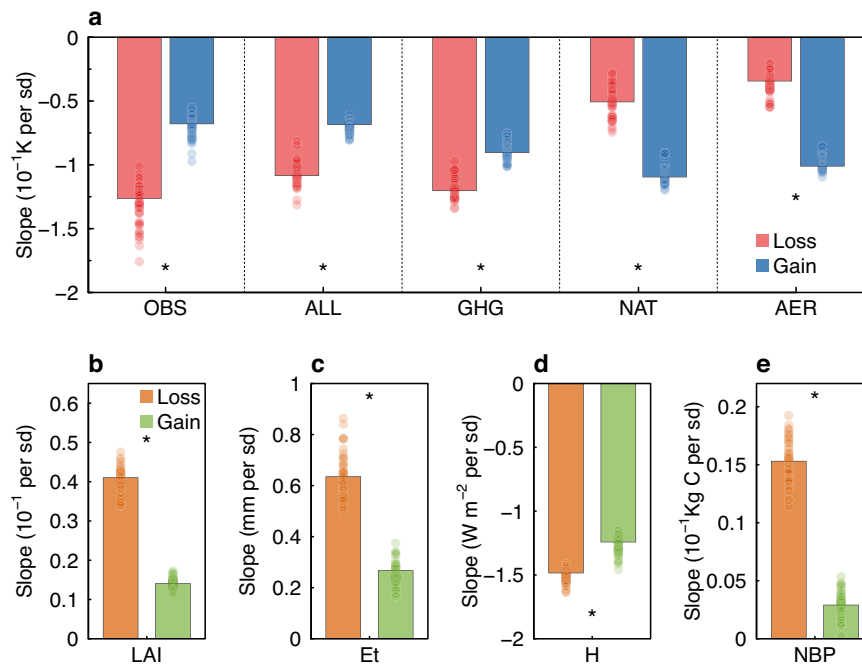


Fig. 2 | Attribution for the asymmetric temperature response to post-drought soil moisture anomalies based on CMIP6. **a** Sensitivities of temperature to post-drought soil moisture anomalies, represented as the linear regression slopes (same values as shown in the upper right corner of Fig. 1). Results are shown for reanalysis and observation-driven products (OBS), and for CMIP6 simulations: anthropogenic and natural (ALL), greenhouse gas only (GHG), natural only (NAT), and aerosol only (AER). OBS and ALL cover 1980–2024, and the time series length for GHG, NAT, and AER is the same. **b–e** Same as (a), but for leaf area index (b), transpiration (c),

sensible heat (d), and net primary production (e). All results are based on ALL simulations. LAI – leaf area index, Et – transpiration, H – sensible heat, and NBP – net biome productivity. In (a–e), the bars indicate the mean values across all datasets. Light-colored dots show the uncertainty ranges estimated using the Monte Carlo method, with 50 points. The two-sided Mann-Whitney *U*-test is used to compare sensitivities associated with water gain and loss; asterisks denote statistical significance at $P < 0.01$.

the post-drought year. Soil moisture drought can also alter internal carbon allocation³². Following severe water stress, carbon is preferentially allocated to maintenance respiration, hydraulic repair, and the replenishment of non-structural carbohydrate reserves. Consequently, stem growth and canopy rebuilding are delayed even when gross primary productivity partially rebounds³³. This decoupling between soil moisture recovery and canopy structure implies that structural suppression can amplify warming during continued water loss, whereas delayed canopy recovery may dampen cooling associated with water gain.

In addition, soil moisture drought legacies can delay vegetation seasonal development trajectories¹². When soil moisture recovery is delayed, green-up may be postponed and leaf area development reduced, suppressing seasonal transpiration and evaporative cooling. Even when soil moisture recovers, biological constraints on canopy development can therefore prevent a symmetric restoration of cooling. Overall, while soil moisture drought can influence multiple physiological and ecological processes in vegetation, its temperature impacts are primarily expressed through changes in land-atmosphere energy partitioning. Consistent with this mechanism, our results show that the asymmetric temperature response is closely associated with vegetation-mediated changes in sensible heat flux and transpiration.

Methods

Data

We collect monthly near-surface air temperature and soil moisture across global land regions from three reanalysis products spanning 1980–2024, namely GLDAS/CLSM (GLDAS_VIC10_3H version), ERA5-land, and MERRA2. GLDAS/CLSM simulates soil moisture using the Catchment Land Surface Model, driven by observed or reanalyzed meteorological forcing. ERA5-Land is produced by replaying the land component of the ECMWF ERA5 climate reanalysis, which combines a

coupled atmosphere-ocean-land model with abundant observations, constrained by physical laws, to produce a consistent and spatially complete dataset. MERRA-2 is generated by the NASA Global Modeling and Assimilation Office using the Goddard Earth Observing System. It assimilates a wide range of observations, including in situ measurements, satellite remote sensing data, microwave sounders, and hyperspectral infrared radiances. In addition, we use satellite-derived root-zone soil moisture from GLEAM v4.2, which integrates remotely sensed surface soil moisture and vegetation optical depth. The dataset is available for the period 1980–2023.

We also use raw monthly outputs from 17 Earth system models (ESMs) participating in CMIP6. All models used in this study are presented in Supplementary Table S2. The CMIP6 experiments included in our study are: historical anthropogenic and natural forcing (ALL), only anthropogenic greenhouse-gas forcing (GHG), only natural forcing (NAT), only aerosol forcing (AER), and the Shared Socioeconomic Pathway 5 Representative Concentration Pathway 8.5 (SSP5-8.5). These ESMs are chosen as they provide available datasets and incorporate dynamic responses of stomatal conductance and vegetation cover to elevated CO_2 and climate change. Moreover, we exclude a small subset of ALL-forcing models from the primary attribution analysis in the main text since they exhibit largely reversed asymmetry in temperature sensitivity to post-drought soil moisture anomalies relative to the observed response. Importantly, including these models in the ALL simulations yields similar results, but the magnitude of the asymmetry is reduced (Supplementary Fig. 9). We use one ensemble member per climate model. First, for each model, we select the r1i1p1f1 ensemble member if available; otherwise, we select the first available ripf member in alphabetic order with data available. To match the temporal range of reanalysis data, the historical ALL simulations are extended to 2024 using the SSP5-8.5 scenario. The CMIP6 simulations used in this study are

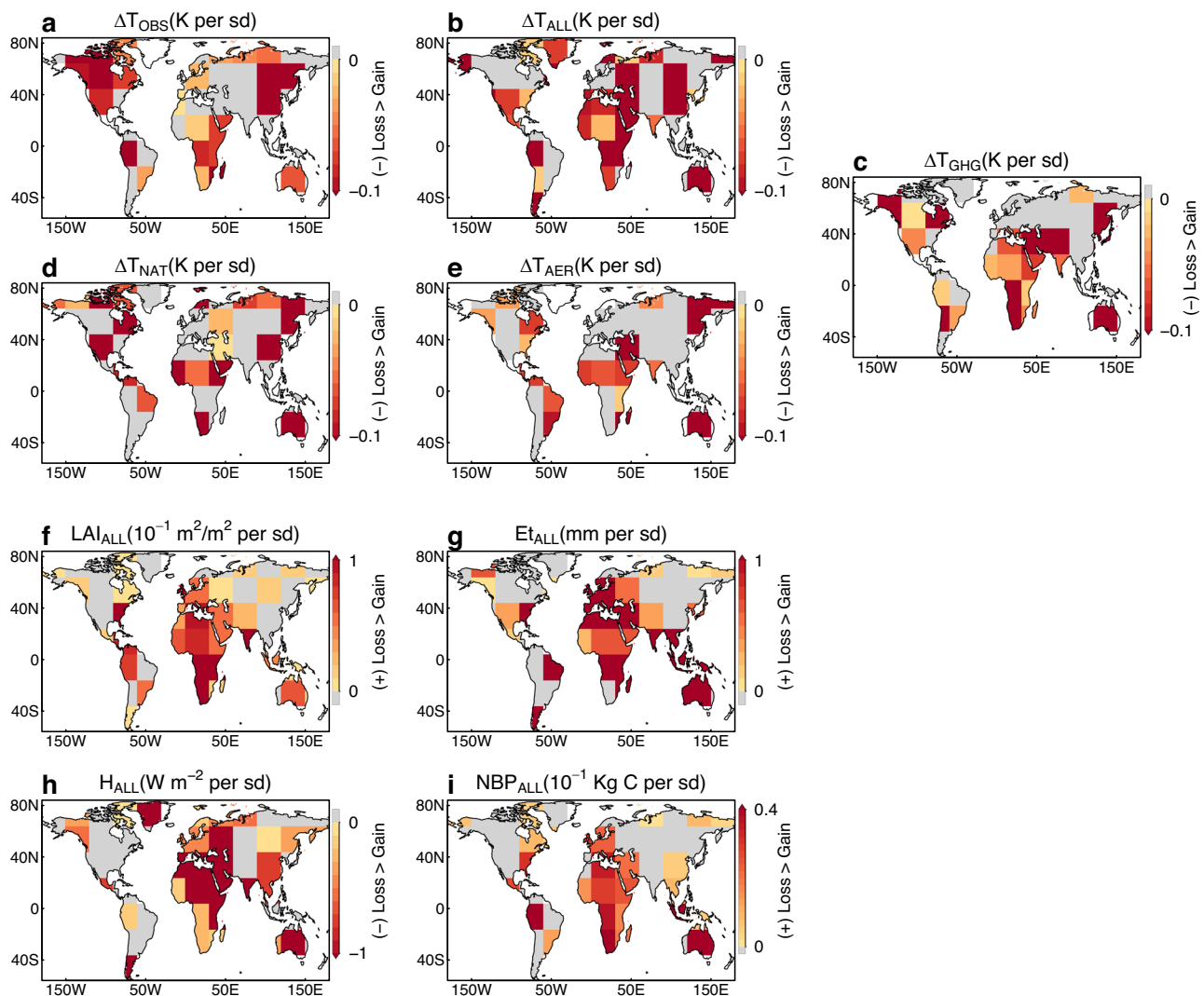


Fig. 3 | Spatial difference in sensitivity to post-drought water loss versus gain. **a** Spatial distribution of the difference in temperature sensitivity to post-drought water loss versus gain, based on reanalysis and observation-driven products (ΔT_{OBS}). Negative values indicate greater temperature sensitivity to water loss than to water gain. The anomalies in temperature and post-drought soil moisture are quantified at each grid cell, and the temperature sensitivity is then evaluated at the regional scale by aggregating grid-cell information within each region. **b–e** Same as

(**a**), but for ALL (**b**), ΔT_{ALL} , GHG (**c**), ΔT_{GHG} , NAT (**d**), ΔT_{NAT} , and AER (**e**), ΔT_{AER} simulations. In (**a–e**), the unit is K per sd. **f–i** Spatial distribution of the difference in sensitivity of LAI (**f**), $10^{-1} \text{ m}^2/\text{m}^2$ per sd, E_t (**g**), mm per sd, H (**h**), W m^{-2} per sd, and NBP (**i**), 10^{-1} Kg C per sd) to post-drought water loss versus gain, based on ALL simulations. In (**f, g, i**), positive values indicate stronger sensitivity to water loss than to water gain; the opposite is true in (**h**). The basemap data used in these plots is from Natural Earth (<https://www.naturalearthdata.com>).

concentration-driven rather than emission-driven. During the relatively short extension period (2015–2024), atmospheric CO_2 concentrations between observation and SSP5-8.5 simulations remain very similar (Supplementary Fig. 10). The climate response over this near-term period is not strongly sensitive to the specific SSP pathway used for extension. We also extend the historical ALL simulations to 2024 using the SSP3-7.0 scenario; the results are consistent with those obtained using SSP5-8.5 as the extension (Supplementary Fig. 11). Considering the different depths of ALL simulations, we linearly interpolate the profiles of each model and then extract the top-1m soil moisture, which is more closely associated with root-zone soil moisture³⁴.

We use observed temperature data from the Climatic Research Unit v4.08 for 1980–2023, paired with GLEAM soil moisture. CRU provides a widely used gridded temperature dataset derived from station measurements (Harris et al., 2020), while GLEAM estimates root-zone soil moisture using satellite observations and land-surface modeling constrained by observations. This combination, therefore, represents an observation-based estimate that is largely independent

of reanalysis systems. Additionally, we also collect four long-term satellite LAI products, including GIMMS LAI3g³⁵ (1982–2018), GLASS³⁶ (1982–2018), GLOBMAP³⁷ (1982–2020), and GIMMS LAI4g³⁸ (1982–2020). All datasets used in this study are resampled using bilinear interpolation to a common spatial resolution of $1^\circ \times 1^\circ$.

Quantifying the general sensitivity of temperature to post-drought soil moisture

To evaluate the biophysical impacts of post-droughts on temperature, we analyze temperature changes in the year following each soil moisture drought event. The annual mean values of the variables of interest are calculated, followed by the application of the space-and-time approach^{39–41} to derive post-drought impacts. The post-drought impacts on temperature are quantified as the difference in temperature between the year after and the year before each soil moisture drought event. Using a one-year interval allows us to capture the immediate post-drought legacy effects while minimizing contamination from the direct impacts of the drought itself. In addition, an annual framework can ensure the consistency across datasets and

climate regions in our global analysis, reducing the influence of seasonal sampling biases while retaining the dominant signal of ecosystem changes.

To facilitate soil moisture drought identifications and ensure consistency across multiple soil moisture products, each soil moisture product is standardized using a Z-score at each pixel across study periods for individual datasets. This standardization does not assume that absolute soil moisture deficits are physically equivalent across different climate regions. Instead, it defines soil moisture drought relative to the local climatological variability at each location. As a result, a Z-score threshold (e.g., -1) represents a comparable departure from typical conditions across different climate regimes, rather than a uniform physical deficit. This methodology has been widely adopted in soil moisture drought^{42,43} and impact studies^{12,44,45}. It enables consistent detection and comparison across diverse hydroclimatic regimes and data products, while avoiding biases arising from absolute thresholds that would disproportionately select certain regions. Importantly, our analysis focuses on anomalies relative to local climatology, not on absolute soil moisture levels. The temperature response is therefore interpreted as the sensitivity of temperature to departures from locally normal soil moisture conditions.

Soil moisture droughts are identified when the standardized soil moisture falls below a threshold of -1 , a common criterion that highlights moderate to severe soil moisture drought conditions³. For sensitive analysis, we also apply more stringent soil moisture thresholds of -1.2 and -1.5 , yielding similar results (Supplementary Fig. 2). The use of Z-score-based soil moisture anomalies is conceptually consistent with the construction of widely used drought indices such as the Standardized Precipitation Evapotranspiration Index⁴⁶. We additionally derive a standardized soil moisture index using a cumulative distribution function⁴⁷ and reanalyze the relationship between post-drought soil moisture anomalies and temperature. The resulting temperature responses are consistent with those obtained using Z-score-based anomalies (Supplementary Fig. 12), supporting the robustness and appropriateness of using standardized soil moisture anomalies to characterize soil moisture drought conditions in this study. Soil moisture Z-scores used to identify droughts are non-detrended in the main text. Moreover, to better represent temperature anomalies and enable more consistent comparisons across time and space, we subtract the multi-year mean temperature (over the entire study period) at each pixel from the annual values, thereby focusing on temperature deviations relative to climatology.

A space-and-time approach is used to isolate post-drought impacts by subtracting the background interannual variability from the gross change over time. This method has been demonstrated to be effective in quantifying changes in temperature due to climate-driven extremes³⁹ or forest cover change^{40,41}. This method assumes that for a given soil moisture drought pixel, the total change in temperature between the years after and before soil moisture drought (ΔT) can be decomposed into the temperature change induced by soil moisture drought (ΔT_D) and a residual change (ΔT_{res}) driven by natural climate variability:

$$\Delta T = \Delta T_D + \Delta T_{res} \quad (1)$$

From this, post-drought-induced temperature change is isolated as:

$$\Delta T_D = \Delta T - \Delta T_{res} \quad (2)$$

Here, ΔT is directly computed for each soil moisture drought pixel. ΔT_{res} can be approximated as the mean ΔT of the non-drought control pixels and neighboring pixels, and is considered comparable to the soil moisture drought pixel. Candidate control pixels are selected based on the following criteria: (1) they must fall within a

10×10 pixels window (approximately to $1000 \times 1000 \text{ km}^2$) surrounding the target soil moisture drought pixel; (2) they must not experience soil moisture droughts during the same year as the target soil moisture drought pixel nor in the following year; and (3) to exclude potential disturbances beyond soil moisture drought, such as wet extremes, their standardized soil moisture values during this period must range between -1 and 1 , and they are not affected by hot events in the following year. Because long-term anthropogenic warming and large-scale internal variability (e.g., decadal oscillations) operate coherently over spatial scales much larger than individual soil moisture drought pixels, these background signals are largely shared between the soil moisture drought pixel and its surrounding control pixels. Subtracting ΔT_{res} therefore removes a substantial fraction of the common background signal, including much of the long-term warming trend and low-frequency variability. Additionally, a multi-year soil moisture drought is treated as a single continuous event rather than independent events. However, considering the potential influence of long-term ecological or hydrological legacies associated with earlier droughts, the space-and-time approach may not completely exclude residual long-term drought legacy effects that may persist over multi-year timescales. If soil moisture drought conditions persist into the following year, that year is not treated as a post-drought recovery year, and such pixels are not considered eligible as controls. The size of the local window (10×10 pixels) used above is determined by the broadly similar background climate conditions (Supplementary Fig. 6f) and provides a valid reference. Alternative search windows, such as 10×6 , 6×10 , and 8×8 pixels, are also used, all of which yield consistent outcomes (Supplementary Fig. 2).

Soil moisture drought-induced changes in soil moisture (ΔSM_D) are calculated using the same space-and-time approach described above. A negative ΔSM_D is defined as post-drought water loss, while a positive ΔSM_D refers to post-drought water gain. To obtain an overview of the relationship between changes in temperature and soil moisture after soil moisture droughts, we group the data into bins based on ΔSM_D and calculate the median ΔT_D within each bin. Separate linear regression models ($y = ax + b$) are then fitted for the post-drought water loss and gain regimes, where the slope (a) quantifies the general sensitivity of temperature responses to soil moisture anomalies after one year of soil moisture drought, as shown in Fig. 1. Consistent refs. 41,48,49, such bin-based analysis can help reveal overall patterns by reducing the influence of outliers and random noise, thereby enabling a clear estimation of the general sensitivity of response variables to their drivers. Bin widths are 0.2 for soil moisture anomalies above -3 , and 0.5 for anomalies below -3 . Extremely low post-drought soil moisture anomalies (less than -3) can occur relatively infrequently. To ensure sufficient sample size per bin, we apply a wider bin width (0.5) for soil moisture anomalies below -3 . We also apply a constant bin width of 0.2 throughout, as well as alternative widths of 0.15 and 0.25 within the range -3 to 3 , yielding consistent results.

We note that ΔT_D and ΔSM_D are calculated independently using the space-and-time approach. This approach isolates drought-related temperature changes while removing background interannual variability. As a result, the temperature response is not directly regressed against the standardized soil moisture time series used to define soil moisture drought, but against independently ΔSM_D . This separation helps reduce potential self-referential bias and preserves a clear interpretation of soil moisture drought legacy effects. To estimate uncertainties, we use the Monte Carlo method by randomly selecting 80% of the total samples and repeating the regression. Bootstrap samples are drawn without replacement. For sensitivity analysis, we also test variations in bins, apply a two-segment regression, examine different study periods, and calculate mean ΔT_D within each bin (Supplementary Fig. 2). In all cases, the results remain consistent, supporting the robustness of our findings.

Post-drought effects in each CMIP6 experiment and regional analysis are calculated following the same approach. Our study follows the detection-and-attribution framework widely used in CMIP-based studies, employing factorial experiments under ALL, GHG-only, AER-only, and NAT-only forcings^{50–53}. Therefore, the current analysis demonstrates that GHG forcing is a necessary condition, but does not attempt to quantify the fractional contribution of GHG forcing relative to other forcings. We note that such experiments do not permit a clean quantitative decomposition of relative contributions across forcings. Since responses to external forcings may be nonlinear, interactions among GHG forcing, aerosols, and internal variability are not necessarily additive^{50–53}. Consequently, responses in ALL simulations cannot, in general, be expressed as a linear sum of single-forcing experiments, precluding a straightforward quantitative attribution. Moreover, the regional analysis in each panel is based on all samples across all products, ensuring a sufficiently large sample size for analysis.

Data availability

GLDAS CLSM and MERRA2 can be retrieved at <https://search.earthdata.nasa.gov>; ERA5-land is available at <https://cds.climate.copernicus.eu/datasets>; GLEAM soil moisture is available at <https://www.gleam.eu/>; CRU v4.08 is available at https://crudata.uea.ac.uk/cru/data/hrg/cru_ts_4.08/; GIMMS 4g LAI is available at <https://zenodo.org/records/8281930>; GLOBMAP LAI is available at <https://zenodo.org/records/4700264>; GLASS LAI is available at <http://www.glass.umd.edu/Download.html>; CMIP6 data can be retrieved at <https://aims2.llnl.gov/search>. The data generated in this study have been deposited in the Zenodo database, <https://doi.org/10.5281/zenodo.20810502>.

Code availability

All codes are direct implementations of standard methods and techniques described in detail in Methods and executed via R and Bash scripts. Codes used in this study are available in the Zenodo database, <https://doi.org/10.5281/zenodo.20810502>.

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J.L. conceived and designed the study. J.L. performed the analysis. J.L., M.W., X.L., W.L., X.T., X.W., C.L., and J.P. wrote the manuscript. X.L., Z.W., X.C., and H.T. supervised and funded the study. All authors contributed to the interpretation of the results and the revisions of the text.

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