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Key Points:

- Idealized symmetric carbon dioxide forcing produces asymmetric precipitation variability between ramp up and ramp down
- Asymmetry spans daily to intraseasonal timescales and is pronounced over major land regions
- The asymmetry is sustained by persistent atmospheric moisture linked to delayed ocean cooling

Supporting Information:

Supporting Information may be found in the online version of this article.

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Asymmetric Response of Land Precipitation Variability Under Carbon Dioxide Removal

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Abstract Precipitation variability describes fluctuations between wet and dry conditions and plays an important role in shaping hydrological extremes and their societal impacts. Using idealized carbon dioxide removal experiments from CMIP6 models, we assess how precipitation variability responds to symmetric CO₂ ramp-up and ramp-down forcing across time scales from day-to-day to intraseasonal. Precipitation variability strengthens with increasing CO₂ and remains stronger during the ramp-down phase than during the ramp-up phase, even at comparable CO₂ concentrations. This asymmetry is evident across time scales and is particularly pronounced over several densely populated land regions, including North America, Europe, South Asia and East Asia. Diagnostics based on a simplified moisture-budget framework indicate that the asymmetry is primarily driven by thermodynamic moistening, which enhances variability in vertical moisture transport. These results demonstrate a pronounced asymmetric response of short-timescale precipitation variability to CO₂ forcing, highlighting the persistence of hydroclimatic variability beyond the period of CO₂ reduction.

Plain Language Summary Climate change is expected to increase fluctuations between wet and dry conditions, intensifying floods and droughts and affecting water resources, agriculture, and society. Climate targets require atmospheric carbon dioxide to peak and then decline, raising the question of whether rainfall variability can return to earlier conditions. Using climate model experiments in which carbon dioxide first rises and then declines, we find that rainfall variability increases during warming and does not reverse symmetrically. At similar carbon dioxide levels, variability remains higher during the decreasing phase than during the earlier increase, especially over major land regions. This asymmetric behavior is linked to the slow response of the ocean. Heat stored during warming is released gradually, keeping the atmosphere warmer and moister and sustaining stronger fluctuations between wet and dry conditions. These results suggest that even as carbon dioxide declines, enhanced rainfall variability may persist, posing challenges for managing hydroclimatic risks and adaptation.

1. Introduction

Precipitation variability is a key aspect of the hydrological response to climate change, reflecting fluctuations around the mean precipitation state (Pendergrass et al., 2017). Under global warming, precipitation variability is expected to increase (Akisanola et al., 2020; Alves et al., 2021; Brown et al., 2017; Pendergrass et al., 2017; Schwarzwald et al., 2021; W. Zhang et al., 2021, 2024), leading to larger fluctuations between wet and dry conditions. Such changes amplify the likelihood of extreme precipitation events (de Vries et al., 2024) and exert substantial impacts on human societies and ecosystems, including reduced water-resource reliability (Ficklin et al., 2022), altered agricultural productivity (Rowhani et al., 2011), disrupted ecosystem functioning (Gherardi & Sala, 2019) and economic growth (Kotz et al., 2022).

Precipitation variability occurs across a wide range of timescales (Akisanola et al., 2020; Alves et al., 2021; Brown et al., 2017; He & Li, 2019; Mo et al., 2025; Pendergrass et al., 2017; Schwarzwald et al., 2021; Wood et al., 2021; W. Zhang et al., 2021, 2024), and both climate model projections and observational evidence indicate that variability across these timescales, particularly from daily to intraseasonal ranges, has intensified under global warming (Akisanola et al., 2020; Pendergrass et al., 2017; W. Zhang et al., 2024). This amplification

tends to occur preferentially in climatologically wet regions, consistent with a “wet-get-more-variable” pattern reported in previous studies (Pendergrass et al., 2017; W. Zhang et al., 2021).

The amplification of precipitation variability is closely linked to increased atmospheric moisture under a warming climate (Akinsanola et al., 2020; He & Li, 2019; S. Zhang et al., 2024; W. Zhang et al., 2021, 2024). According to the Clausius–Clapeyron relationship, atmospheric moisture increases by approximately 6%–7% per degree of warming, providing a thermodynamic basis for enhanced precipitation variability (Allan & Soden, 2008; Fischer & Knutti, 2016; O’Gorman & Schneider, 2009; Trenberth, 1999). Meanwhile, changes in atmospheric circulation may also modulate precipitation variability. For example, weakened large-scale circulation variability under global warming may partly offset the thermodynamically driven increase in precipitation variability (He & Li, 2019; Held & Soden, 2006; Vecchi & Soden, 2007; W. Zhang et al., 2021; Zhou et al., 2024). Attribution studies further suggest that the observed increase in precipitation variability is primarily driven by rising greenhouse-gas concentrations, with anthropogenic aerosols exerting a partial offsetting effect (W. Zhang et al., 2024).

Anthropogenic emissions since the Industrial Revolution have increased atmospheric carbon dioxide (CO₂) concentrations, driving substantial changes in the climate system (Gan et al., 2023; Hu et al., 2021; Kotz et al., 2024; Lane et al., 2012; Q. Liu et al., 2026; Rahmstorf & Coumou, 2011; Yeh et al., 2021). To mitigate these impacts, the Paris Agreement aims to limit global warming to well below 2°C while pursuing efforts toward 1.5°C above preindustrial levels, a target likely to require carbon dioxide removal (CDR) strategies (Keller et al., 2018; Luderer et al., 2018; Peters et al., 2013; Realmonde et al., 2019; Rogelj et al., 2018; Sanderson et al., 2016; Tong et al., 2019). A central question is whether the climate system can recover once atmospheric CO₂ concentrations are reduced to preindustrial levels. Idealized climate model experiments have been widely used to investigate the response of the climate system to atmospheric CO₂ removal (Boucher et al., 2012; Bouttes et al., 2013; Chadwick et al., 2013; Gan et al., 2026; Garbe et al., 2020; Jackson et al., 2014; Li et al., 2025; Sgubin et al., 2015; Wu et al., 2015). Previous studies have shown that both global and regional climate responses can exhibit pronounced asymmetry and hysteresis between the CO₂ ramp-up and ramp-down phases (An et al., 2021; Gan et al., 2026; Hwang et al., 2024; S.-K. Kim et al., 2022; Kug et al., 2022; C. Liu et al., 2023; J.-H. Oh et al., 2022; Piao et al., 2025). Recent studies further suggest that such hysteresis extends to climate variability and weather systems, including temperature variability (Gan et al., 2026; S.-H. Kim et al., 2025) and western North Pacific tropical cyclone activity (Yoon et al., 2025). However, precipitation variability has received comparatively limited attention. In particular, mean precipitation has been found to respond asymmetrically under CDR scenarios (Cao et al., 2011; Chadwick et al., 2013; S.-K. Kim et al., 2022; Song et al., 2023; Wu et al., 2010). Similar hysteresis behavior has also been reported for regional precipitation changes in the tropical Pacific (Zhou et al., 2022), East Asia (Song et al., 2022) and South Asia (S. Zhang et al., 2023, 2024), as well as for ENSO-related precipitation (An et al., 2024; C. Liu et al., 2024). However, it remains unclear whether precipitation variability exhibits asymmetric responses between the CO₂ ramp-up and ramp-down phases, particularly across daily to intraseasonal timescales.

Here, we analyze daily precipitation outputs from eight models participating in Phase 6 of the Coupled Model Intercomparison Project (CMIP6) to investigate the response of precipitation variability under idealized CO₂ ramp-up and ramp-down forcing. The experiments include the standard 1pctCO₂ simulation, in which atmospheric CO₂ concentrations increase at a rate of 1% per year from pre-industrial levels, and the corresponding 1pctCO₂-cdr simulation from the Carbon Dioxide Removal Model Intercomparison Project (CDRMIP) (Keller et al., 2018), in which CO₂ concentrations subsequently decline at the same rate. By comparing periods with identical CO₂ concentrations along the ramp-up and ramp-down pathways, we assess whether precipitation variability responds asymmetrically to CO₂ forcing. We further diagnose the underlying mechanisms using a moisture-budget framework to separate the contributions from thermodynamic moistening and circulation changes.

2. Data and Methods

2.1. CMIP6 Experiments and Data Processing

Daily outputs from CMIP6 idealized CO₂ forcing experiments (Eyring et al., 2016), including 1pctCO₂, 1pctCO₂-cdr, and piControl simulations, are used. In the 1pctCO₂ experiment, atmospheric CO₂ increases by about 1% per year from the pre-industrial level to four times this value after 140 years. The 1pctCO₂-cdr experiment continues

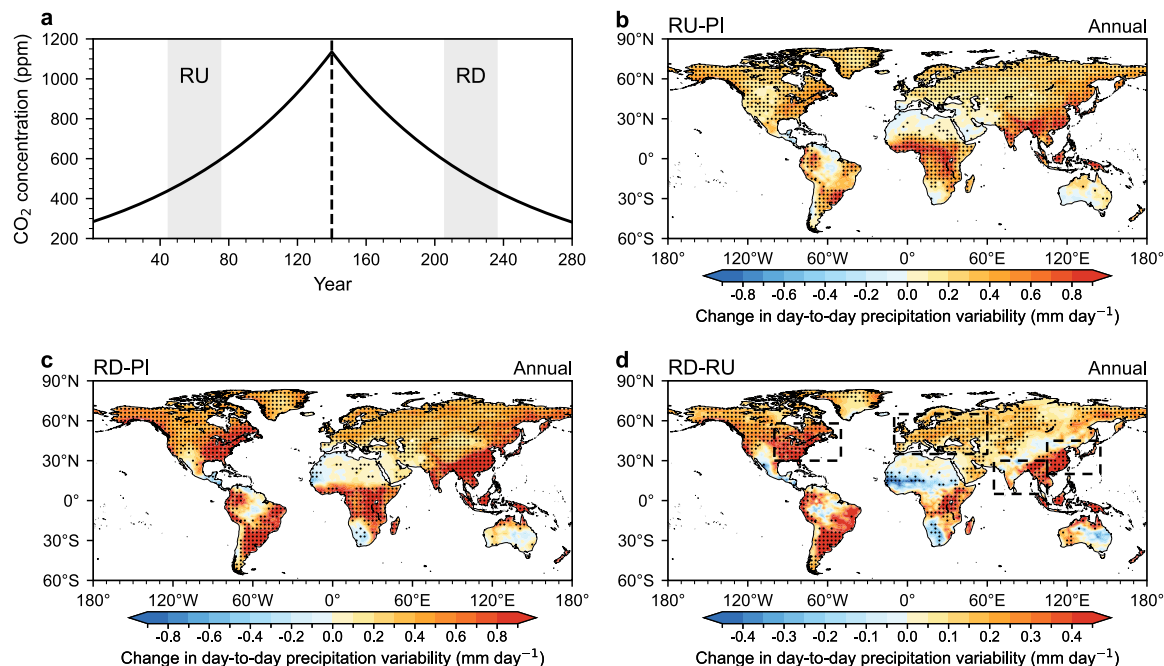


Figure 1. (a) Evolution of atmospheric CO₂ concentration (ppm) in the idealized carbon dioxide removal experiment, illustrating the ramp-up (RU) and ramp-down (RD) phases. (b–d) Changes in day-to-day precipitation variability during the RU (b) and RD (c) phases relative to the pre-industrial climatology, and the RD–RU difference (d). Black stippling indicates regions where the multimodel mean change is significant at the 90% confidence level and where at least six of eight models agree on the sign. Black boxes denote the analyzed subregions.

from this state and involves a symmetric decrease at the same rate for the following 140 years (Figure 1a). The piControl simulation maintains constant pre-industrial forcing and serves as a reference climate.

Years 1–140 of the 1pctCO₂ and 1pctCO₂-cdr experiments define the CO₂ ramp-up (RU) and ramp-down (RD) phases. The final 100 years of the piControl simulation are used to define the pre-industrial (PI) climatology. Eight CMIP6 models are analyzed: ACCESS-ESM1-5, CanESM5, CESM2, CNRM-ESM2-1, GFDL-ESM4, MIROC-ES2L, NorESM2-LM and UKESM1-0-LL. All model outputs are regridded to a common 1° × 1° grid using bilinear interpolation. Mechanistic diagnostics are performed using three models (CanESM5, CNRM-ESM2-1, and MIROC-ES2L) that provide the variables required for the moisture-budget analysis.

To examine the asymmetric response of precipitation variability, we select two periods with comparable CO₂ concentrations: the ramp-up (years 45–75; RU) and ramp-down (years 206–236; RD) phases. Using an alternative pair of CO₂-matched periods at the beginning of the ramp-up phase and the end of the ramp-down phase (years 3–33 and 248–278) yields consistent results (Figure S1 in Supporting Information S1), confirming the robustness of the asymmetric response of precipitation variability under CDR conditions.

2.2. Precipitation Variability Calculation

Precipitation variability is quantified at multiple time scales, including total day-to-day variability and variability associated with synoptic (3–10 days), monthly (25–30 days), and intraseasonal (30–80 days) fluctuations. Variability is calculated using a 5-year moving window applied to daily precipitation. The results are insensitive to the choice of window length, with similar results obtained using 3-, 7-, and 10-year windows (Figure S2 in Supporting Information S1).

Within each window, the daily precipitation series is linearly detrended and the mean seasonal cycle removed. Variability at different time scales is isolated using band-pass filters, except for total day-to-day variability, which is calculated from the unfiltered daily anomalies. The synoptic, monthly, and intraseasonal components correspond to 3–10-day, 25–30-day, and 30–80-day band-pass filters, respectively. Precipitation variability is defined as the temporal standard deviation of the filtered time series at each grid point (Brown et al., 2017; Mo et al., 2025; Pendergrass et al., 2017; W. Zhang et al., 2021, 2024).

2.3. Simplified Moisture-Budget Framework

To diagnose the mechanisms underlying changes in precipitation variability, we adopt a simplified moisture-budget framework following previous studies (Chou et al., 2009; Emori & Brown, 2005; W. Zhang et al., 2021, 2024). In the climatological mean, precipitation P is balanced by evaporation E and vertically integrated moisture advection,

$$P - E = -\langle \omega \partial_p q \rangle - \langle V \cdot \nabla q \rangle + \text{res} \quad (1)$$

where q is specific humidity, ω is vertical velocity, V denotes the horizontal wind vector, and res represents the residual. Angle brackets indicate vertical integration over the troposphere.

Previous studies have shown that, at a given time scale, precipitation variability can be well approximated by variability in vertical moisture advection, which largely captures both the phase and magnitude of precipitation variability (W. Zhang et al., 2021, 2024). Accordingly, the vertically integrated moisture-advection term can be represented using a simplified two-layer framework, in which lower-tropospheric specific humidity at 850 hPa characterizes moisture availability and mid-tropospheric vertical velocity at 500 hPa represents large-scale ascent associated with precipitation:

$$P_f \approx \left(-\frac{\omega_m q_l}{g} \right)_f \quad (2)$$

where the subscript f denotes variability at a given time scale, q_l represents lower-tropospheric specific humidity (850 hPa), and ω_m denotes mid-tropospheric vertical velocity (500 hPa).

Under this approximation, precipitation variability can be expressed as

$$\sigma(P_f) \approx \sigma \left(-\frac{\omega_m q_l}{g} \right)_f \quad (3)$$

where σ denotes the standard deviation operator. Because temporal variations in humidity are typically smaller than those in vertical motion, q_l can be approximated by its climatological mean ($\bar{q}_l$) when estimating precipitation variability (W. Zhang et al., 2021, 2024). Thus, Equation 3 can be simplified as:

$$\sigma(P_f) \approx \sigma \left(-\frac{\omega_m \bar{q}_l}{g} \right)_f \quad (4)$$

To further assess the applicability of this approximation in this study, we apply the framework to ERA5 reanalysis and compare the resulting vertical-moisture-advection variability with precipitation variability (Figure S3 in Supporting Information S1). The vertical-moisture-advection term reproduces both the climatological distribution and the long-term trend pattern of precipitation variability (spatial correlations of 0.87 and 0.70, respectively), supporting the use of Equation 4 as a diagnostic approximation.

Under this approximation, precipitation variability is controlled primarily by two factors: the variability of vertical motion and the background humidity field. The thermodynamic contribution (TH) is associated with changes in mean lower-tropospheric humidity and estimated as

$$\text{TH} \approx \frac{\bar{q}_l^{\text{RD}} - \bar{q}_l^{\text{RU}}}{\bar{q}_l^{\text{PI}}} \times 100\% \quad (5)$$

where $\bar{q}_l^{\text{RD}}$ and $\bar{q}_l^{\text{RU}}$ denote mean lower-tropospheric specific humidity over the corresponding RD and RU analysis periods, respectively, and $\bar{q}_l^{\text{PI}}$ denotes the baseline climatology derived from the piControl simulation.

The dynamic contribution is associated with changes in vertical-motion variability,

$$DY \approx \delta\sigma(-\omega_m) = \frac{\sigma(-\omega_m)^{RD} - \sigma(-\omega_m)^{RU}}{\sigma(-\omega_m)^{PI}} \times 100\% \quad (6)$$

where $\sigma(-\omega_m)^{RD}$ and $\sigma(-\omega_m)^{RU}$ are the vertical-motion variability over the corresponding RD and RU analysis periods, respectively, and $\sigma(-\omega_m)^{PI}$ denotes the baseline climatology derived from the piControl simulation.

The nonlinear interaction between humidity and circulation changes is diagnosed as the residual (NL),

$$NL \approx \delta\sigma\left[\left(-\frac{\omega_m q_l}{g}\right)_f\right] - TH - DY \quad (7)$$

where $\delta X = (X^{RD} - X^{RU})/X^{PI} \times 100\%$, with X^{PI} denoting the corresponding piControl climatology. The resulting terms therefore quantify the thermodynamic, dynamic and nonlinear contributions to the asymmetric change in precipitation variability within the simplified moisture-budget framework.

Although this simplified framework does not explicitly resolve the full vertical structure of atmospheric moisture, it provides a physically interpretable decomposition of precipitation-variability changes into thermodynamic and dynamic contributions, allowing the relative contributions of atmospheric moistening and circulation changes to be quantified independently. Specifically, the thermodynamic contribution isolates the effect of changes in mean lower-tropospheric humidity, whereas the dynamic contribution quantifies the effect of changes in vertical-motion variability. The nonlinear term represents the residual interaction between humidity and circulation changes that is not explained by the linear decomposition.

3. Results

3.1. Asymmetric Response of Precipitation Variability

Precipitation variability is defined as the standard deviation of daily precipitation after removing the long-term trend and seasonal cycle (Brown et al., 2017; Pendergrass et al., 2017; W. Zhang et al., 2021, 2024). Variability is estimated using a 5-year moving window (see Section 2.2). To examine whether precipitation variability exhibits asymmetric responses between the CO₂ ramp-up and ramp-down phases, we analyze variability across multiple time scales, including the day-to-day scale, synoptic scale (3–10 days), monthly scale (25–30 days), and intraseasonal scale (30–80 days). We select two 31-year periods with comparable atmospheric CO₂ concentrations to assess the asymmetry in precipitation variability: years 45–75 during the CO₂ RU and years 206–236 during the RD.

The reference climatology of day-to-day precipitation variability, defined as the last 100-year mean of the piControl simulation, is shown in Figure S4a in Supporting Information S1. Relatively large variability occurs over several climatologically wet and convectively active regions, including North America, Europe, South and East Asia, as well as parts of southern South America and Africa. Relative to this baseline, precipitation variability increases over most land areas during the CO₂ ramp-up phase (Figure 1b), in agreement with previous studies showing intensified day-to-day precipitation variability under global warming (Pendergrass et al., 2017; W. Zhang et al., 2021, 2024). During the ramp-down phase, the enhancement becomes stronger and more widespread (Figure 1c). Precipitation variability increases across more than 90% of global land areas during the ramp-down phase relative to the ramp-up phase, with particularly large changes over North America, Europe, South and East Asia, and parts of southern South America and Africa (Figure 1d).

In terms of percentage changes relative to the PI climatology, day-to-day precipitation variability increases by about 5%–15% across most land regions during RU (Figure 2a) and strengthens to roughly 10%–25% during RD (Figure 2b). This asymmetry is robust across models (Figure S5 in Supporting Information S1) and remains insensitive to the choice of CO₂-matched comparison periods (Figure S6 in Supporting Information S1).

Regional averages further illustrate the magnitude of these changes (Figure 2d). Over North America, precipitation variability increases by 7.5% (90% confidence interval: 6.4%–8.6%) during RU and by 14.4% (10.8%–18.4%) during RD. Europe shows increases from 4.7% (3.4%–6.1%) to 9.2% (6.5%–12.3%). Larger changes occur over South and East Asia. Over South Asia, variability increases by 11.4% (9.8%–12.9%) during

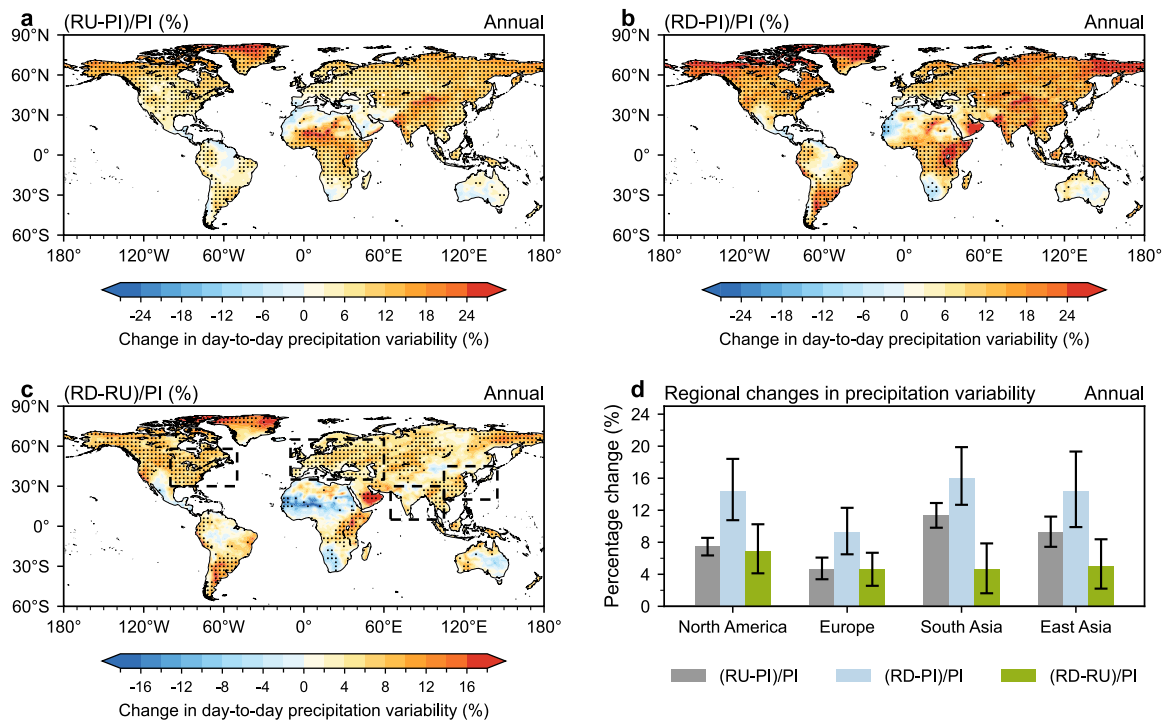


Figure 2. Relative changes in day-to-day precipitation variability under CO₂ forcing. (a–c) Changes in day-to-day precipitation variability during the RU (a) and RD (b) phases relative to the pre-industrial climatology, and the RD–RU difference (c). Black stippling indicates regions where the multimodel mean change is significant at the 90% confidence level and where at least six of eight models agree on the sign. Black boxes denote the analyzed subregions. (d) Regional percentage changes in precipitation variability over North America, Europe, South Asia and East Asia: $(RU - PI)/PI \times 100\%$, $(RD - PI)/PI \times 100\%$, and $(RD - RU)/PI \times 100\%$. Error bars denote the 10th to 90th percentile ensemble spread.

RU and by 16.1% (12.7%–19.9%) during RD, whereas over East Asia the increases reach 9.3% (7.4%–11.2%) and 14.4% (9.9%–19.3%), respectively.

The asymmetry is evident in all seasons (Figure 3). Although the magnitude and spatial pattern vary seasonally, precipitation variability generally remains stronger during RD than during RU. The largest increases occur during boreal summer (Figure 3b), particularly over monsoon regions such as South and East Asia. In boreal winter (Figure 3d), regional reductions emerge in parts of South Asia and other tropical regions, indicating a more heterogeneous seasonal response.

Precipitation variability at longer time scales shows a similar asymmetric behavior (Figure 4). Synoptic-scale variability (3–10 days) exhibits the largest changes (Figures 4a–4c), while monthly (Figures 4d–4f) and intra-seasonal (Figures 4g–4i) variability display comparable spatial patterns but smaller magnitudes. Regions with strong increases in day-to-day variability, including eastern North America, Europe, South Asia, East Asia, southern South America and parts of Africa, also show consistent increases across synoptic, monthly and intraseasonal variability. RD–RU differences remain positive across large areas, indicating that the asymmetry persists across multiple precipitation variability time scales (Figures 4c, 4f, and 4i).

3.2. Temporal Evolution and Hysteresis of Regional Precipitation Variability

Because the increases are particularly pronounced over densely populated regions such as eastern North America, Europe, South Asia and East Asia, we further examine the temporal evolution of precipitation variability in these regions. Figure 5 shows the evolution of regional-mean precipitation variability together with atmospheric CO₂ concentrations. In all four regions, day-to-day precipitation variability increases as CO₂ rises during the ramp-up phase and continues to increase for several years after atmospheric CO₂ reaches its peak. A 31-year centered running mean further indicates that precipitation variability does not decline immediately after atmospheric CO₂ reaches its maximum. Instead, variability begins to decrease approximately 10 years after peak CO₂ in North America and South Asia, around 15 years afterward in East Asia, and nearly 20 years afterward in Europe

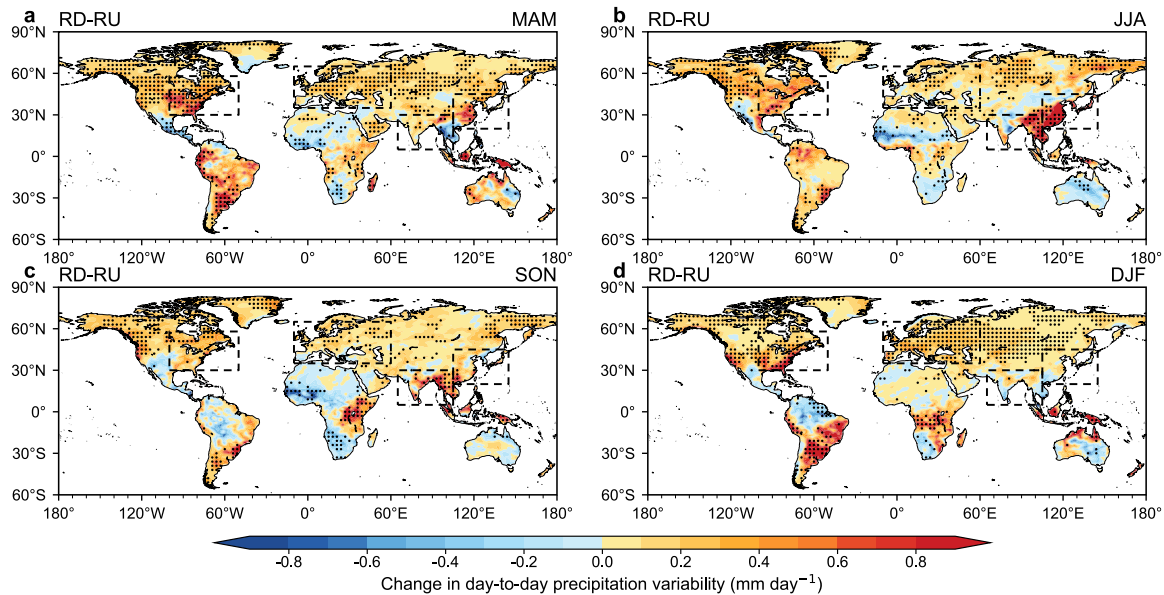


Figure 3. Seasonal differences in day-to-day precipitation variability between the CO₂ ramp-down and ramp-up phases. (a–d), Multi-model ensemble differences in day-to-day precipitation variability between the CO₂ ramp-down (RD) and ramp-up (RU) periods for (a) MAM, (b) JJA, (c) SON, and (d) DJF. Shading indicates changes in day-to-day precipitation variability (mm day^{−1}). Stippling denotes regions where at least six of the eight models agree on the sign of the change.

(Figure 5), highlighting a delayed adjustment of regional precipitation variability during the CO₂ removal phase. This temporal asymmetry is insensitive to the choice of moving-window length, with similar evolution and magnitudes obtained using alternative windows (Figure S2 in Supporting Information S1).

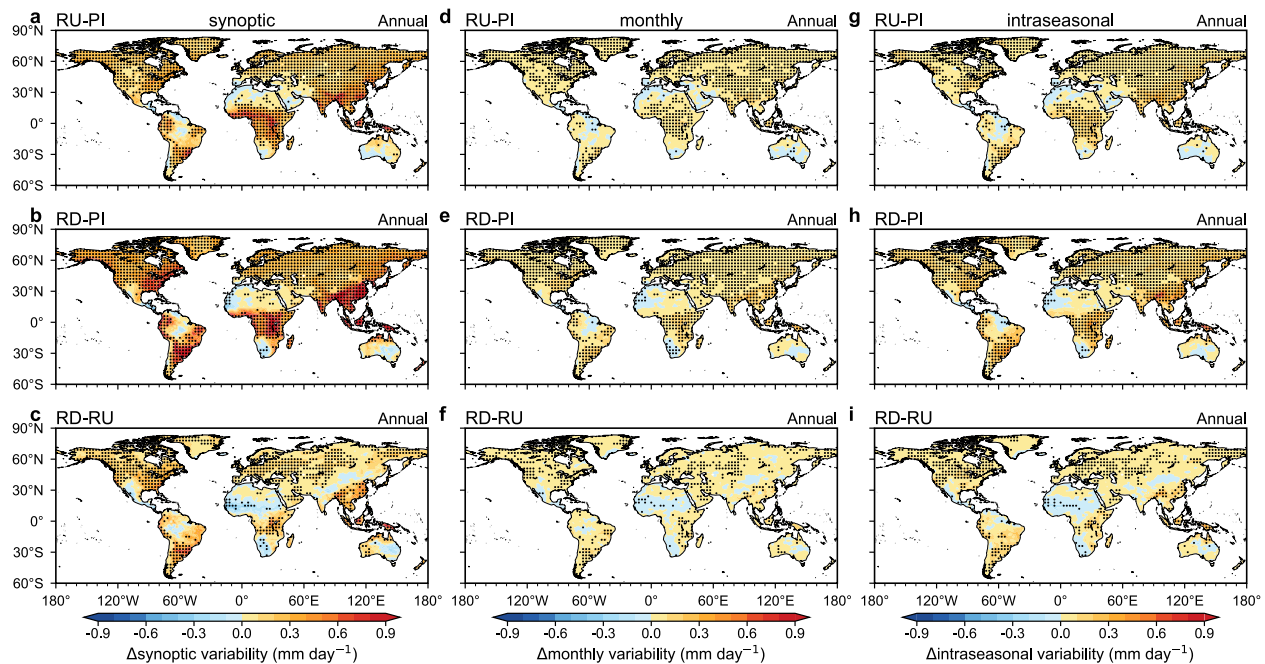


Figure 4. Multi-scale response of precipitation variability to CO₂ forcing. (a–c), Changes in synoptic-scale precipitation variability (3–10 days) for the annual mean during the CO₂ ramp-up phase (RU; years 45–75) relative to the pre-industrial (PI) climatology (a), during the ramp-down phase (RD; years 206–236) relative to PI (b), and the difference between the RD and RU phases (RD – RU) (c). (d–f), Same as (a–c) but for monthly-scale precipitation variability (25–30 days). (g–i), Same as (a–c) but for intraseasonal precipitation variability (30–80 days). Shading indicates changes in precipitation variability (mm day^{−1}). Stippling denotes regions where the Multi-model ensemble change is statistically significant at the 90% confidence level based on a two-tailed Student's *t*-test and where at least six of the eight models agree on the sign of the change.

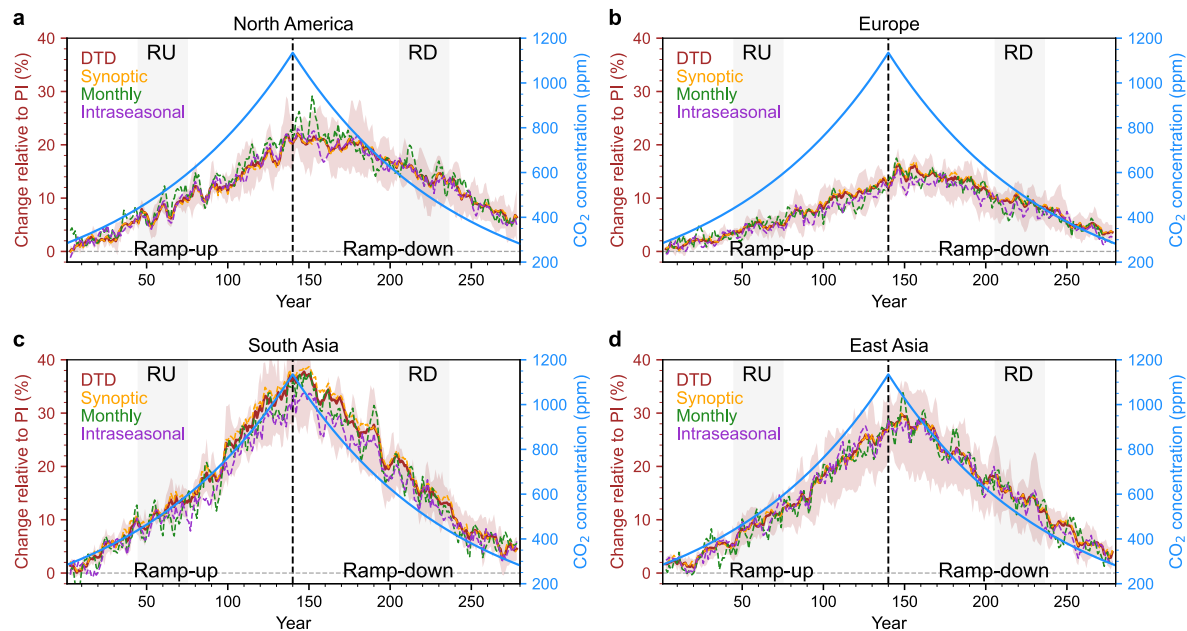


Figure 5. Time series of atmospheric CO₂ concentration (blue; ppm) and multi-model mean annual day-to-day precipitation variability (% relative to the pre-industrial climatology; red) over (a) North America, (b) Europe, (c) South Asia, and (d) East Asia. Pink shading indicates the intermodel spread (25th–75th percentile across the eight models). The vertical dashed line marks the year of peak CO₂ concentration. Gray shading highlights the two 31-year periods selected from the ramp-up (RU) and ramp-down (RD) phases for comparison at equivalent atmospheric CO₂ concentrations. Dashed lines show precipitation variability at synoptic (orange), monthly (green), and intraseasonal (purple) timescales. A 31-year centered running mean of day-to-day precipitation variability is used to identify the timing at which the smoothed variability begins to decline after the CO₂ peak.

This pattern indicates a clear asymmetry in the response of precipitation variability to CO₂ forcing, with variability at synoptic, monthly and intraseasonal time scales exhibiting similar temporal evolution and consistently stronger amplitudes during the ramp-down phase than during the ramp-up phase (Figure 5). Among the four regions, the response is most pronounced over South Asia and East Asia, where precipitation variability increases from 11.4% during RU to 16.1% during RD and from 9.3% to 14.4%, respectively. In contrast, North America and Europe exhibit more moderate changes, with variability increasing from 7.5% to 14.4% and from 4.7% to 9.2%, respectively (Figures 2d and 5). Individual models show qualitatively similar evolution, with stronger variability during RD than during RU at comparable CO₂ concentrations (Figure S7 in Supporting Information S1). A hemispheric analysis further shows that the delayed recovery of precipitation variability is more pronounced over Southern Hemisphere land than over Northern Hemisphere land (Figure S8 in Supporting Information S1), consistent with previous studies highlighting delayed Southern Ocean adjustment during CO₂ removal (e.g., Kug et al., 2022; J.-H. Oh et al., 2024).

3.3. Mechanisms Underlying the Asymmetric Precipitation Variability Response

To investigate the physical processes underlying this asymmetry, we diagnose precipitation variability changes using a simplified moisture-budget framework (see Section 2.3) following previous studies (W. Zhang et al., 2021, 2024). Because the required daily vertical velocity and humidity fields are available for only three models (CanESM5, CNRM-ESM2-1, and MIROC-ES2L), the mechanistic analysis is based on their multi-model mean.

We first calculate RU and RD epoch means of precipitation variability, vertical moisture advection variability, mean lower-tropospheric humidity and vertical velocity variability. Lower-tropospheric humidity exhibits widespread positive RD–RU anomalies across Northern Hemisphere land regions, whereas changes in vertical-velocity variability are generally weaker and spatially less coherent (Figure S9 in Supporting Information S1). For each field, changes are expressed as percentages relative to the corresponding climatology during the final 100 years of the piControl simulation. The asymmetric response is then defined as the RD-minus-RU difference in these percentage changes. Within this framework, the change in vertical moisture advection variability (ΔP^*) is

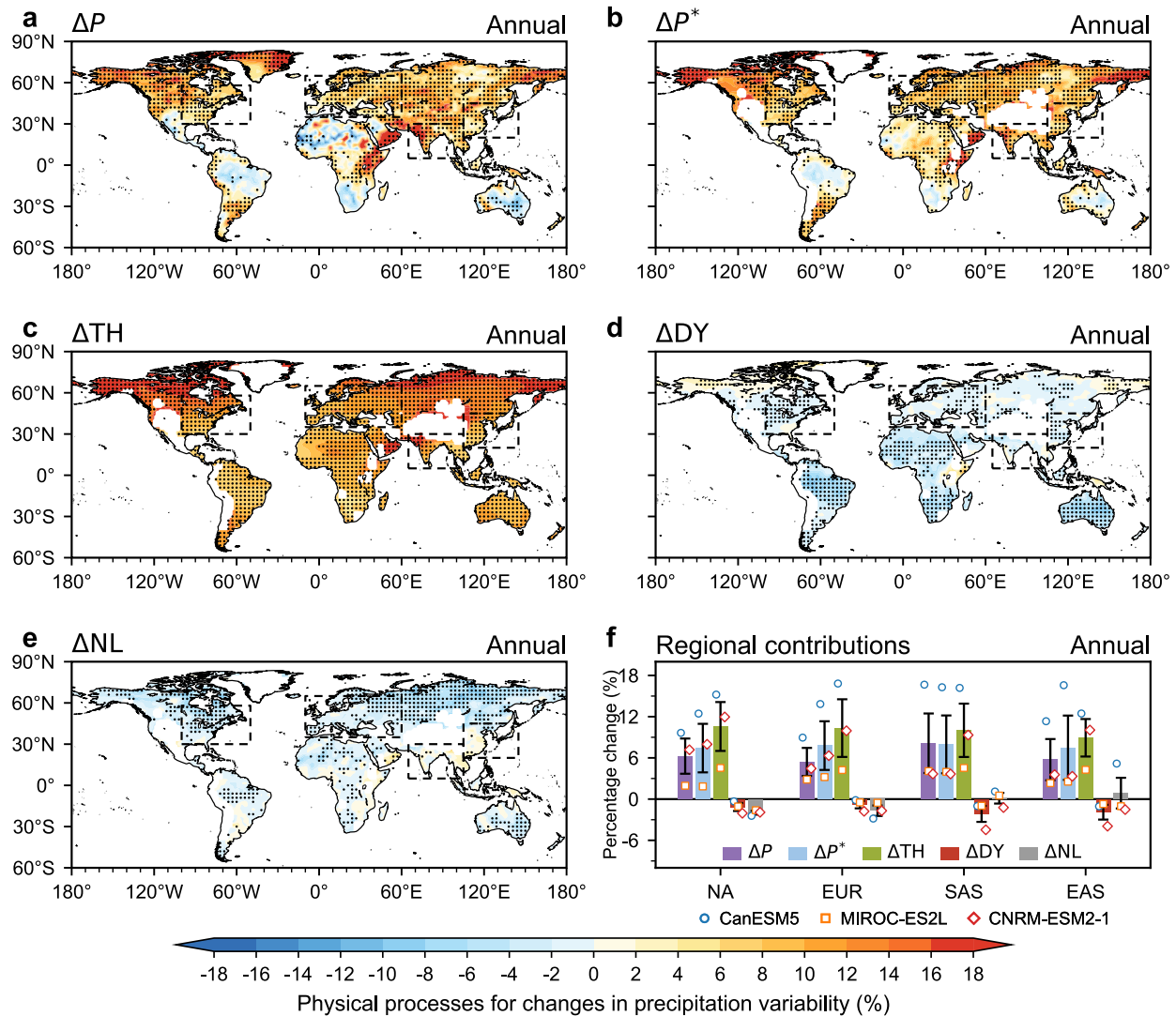


Figure 6. Physical decomposition of the asymmetric response of precipitation variability during CO₂ removal. (a–e) Relative differences in annual-mean precipitation variability between the ramp-down and ramp-up phases (RD–RU) for total variability (ΔP ; a), variability of vertical moisture advection (ΔP^* ; b), thermodynamic contribution (ΔTH ; c), dynamic contribution (ΔDY ; d), and nonlinear contribution (ΔNL ; e). Shading indicates the three-model mean (CanESM5, CNRM-ESM2-1, MIROC-ES2L), and stippling indicates agreement in the sign across all models. (f), regional contributions to the RD – RU difference over North America, Europe, South Asia, and East Asia. Bars denote the multi-model mean, markers indicate individual models, and error bars show the 10th–90th percentile bootstrap range.

further decomposed into a thermodynamic component associated with changes in mean lower-tropospheric moisture, a dynamic component associated with changes in vertical-motion variability, and a nonlinear residual term not explained by the thermodynamic and dynamic contributions.

Figure 6 shows the spatial patterns of the physical components contributing to the RD–RU asymmetry in day-to-day precipitation variability. The spatial distribution of total precipitation variability change (ΔP ; Figure 6a) closely resembles that of vertical moisture advection variability (ΔP^* ; Figure 6b). Positive anomalies dominate across much of the Northern Hemisphere land, particularly over North America, Europe, South Asia and East Asia, mirroring the regional hotspots identified in Figure 1. This correspondence indicates that changes in vertical moisture transport variability capture the spatial structure of the asymmetric response, providing further support for the simplified moisture-budget framework.

Decomposition of ΔP^* into thermodynamic and dynamic components further shows that the asymmetry is dominated by thermodynamic changes (Figures 6c and 6d). The thermodynamic contribution (ΔTH) is positive across most land regions and broadly follows the spatial distribution of total precipitation variability changes,

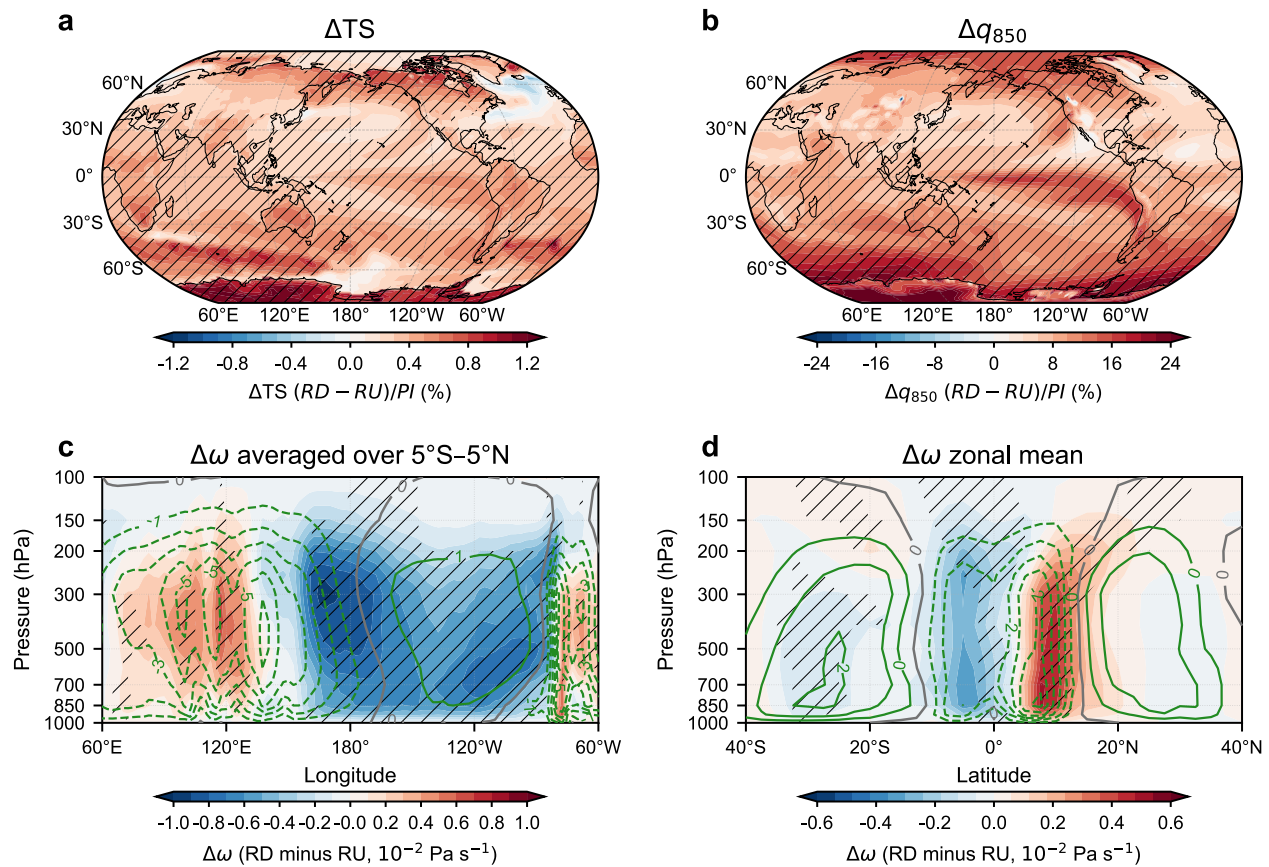


Figure 7. (a, b) Multi-model mean differences between the ramp-down (RD) and ramp-up (RU) phases at comparable atmospheric CO_2 concentrations in surface temperature (TS) (ΔTS ; TS over land and SST over ocean; a) and 850-hPa specific humidity (Δq_{850} ; b), expressed relative to the corresponding PI climatology as $(RD - RU)/PI \times 100\%$. (c, d) Changes in vertical velocity (ω) in the longitude–pressure cross-section averaged over $5^\circ S$ – $5^\circ N$ (c) and the latitude–pressure cross-section of the zonal mean (d). Shading denotes the RD–RU difference in vertical velocity ($\times 10^{-2} \text{ Pa s}^{-1}$), and green contours denote vertical velocity during the RU phase ($\times 10^{-2} \text{ Pa s}^{-1}$). Hatching indicates regions where the RD–RU difference is significant at the 5% level and at least six of the eight models exhibit the same sign of change.

with particularly strong signals over North America, Eurasia, and the monsoon regions of South and East Asia (Figure 6c). Consistent with this spatial pattern, ΔTH explains most of the regional enhancement in precipitation variability across all four study regions, whereas the dynamic and nonlinear contributions remain much smaller (Figure 6f). Individual models show consistent spatial patterns despite differences in magnitude (Figure S10 in Supporting Information S1), indicating that the thermodynamically dominated response is robust. Owing to the continued release of heat stored in the ocean after atmospheric CO_2 reaches its peak, surface temperatures respond asymmetrically during CO_2 removal (Figure 7a). Relative to the ramp-up phase, surface temperatures remain warmer across much of the globe during the ramp-down phase. These warmer surface conditions support a moister lower troposphere (Figure 7b), with inter-model differences in lower-tropospheric humidity closely scaling with regional surface warming (Figure S11 in Supporting Information S1). Within the simplified moisture-budget framework, the thermodynamic contribution is associated with changes in mean lower-tropospheric moisture, while atmospheric circulation is held fixed. Thus, increased atmospheric moisture provides a larger background moisture reservoir for vertical moisture advection, favoring a proportional intensification of precipitation events and consequently enhancing precipitation variability.

In contrast to the dominant thermodynamic contribution, ΔDY is weakly negative across many regions (Figures 6d and 6f), indicating that circulation-related changes partly offset the thermodynamically driven amplification of precipitation variability. To understand this dynamic response, we further examine the associated large-scale circulation changes (Figure 7). Over the tropics, the RD–RU SST pattern exhibits an El Niño-like warming pattern (Figure 7a), which reduces the zonal SST gradient across the tropical Pacific and weakens the

Walker circulation (Figure 7c). At the same time, persistent Southern Ocean warming enhances Southern Hemisphere warming relative to the Northern Hemisphere (Figure 7a), reducing the interhemispheric atmospheric energy transport and shifting the ITCZ centroid southward (Figure S12 in Supporting Information S1). The southward-displaced ITCZ is accompanied by a southward redistribution of tropical ascent (Figure 7d), consistent with a southward adjustment of the Hadley circulation (Broccoli et al., 2006; Byrne et al., 2018). Given the warmer mean climate state during the RD phase (Figure 7a), such changes are consistent with the expected weakening of atmospheric overturning circulation under greenhouse warming (Held & Soden, 2006; Zhou et al., 2024). This weakened overturning circulation may contribute to the reduced vertical-motion variability, which is consistent with the weakly negative dynamic contribution to precipitation variability. Outside the tropics, changes in circulation variability may be related to baroclinic eddy activity, which is sensitive to the meridional temperature gradient (Chang et al., 2012; Lehmann et al., 2014). Regional reductions in vertical-motion variability over parts of the Northern Hemisphere mid-to-high latitudes coincide with weaker meridional temperature gradients. However, these dynamical responses, particularly EKE, exhibit substantial inter-model differences across the mid-to-high latitudes (Figure S13 in Supporting Information S1). ΔDY likewise exhibits substantially weaker inter-model agreement than ΔTH , particularly over Europe and South Asia (Figure 6d and Figure S10 in Supporting Information S1), reflecting the larger uncertainty associated with circulation changes under warming owing to strong internal variability (Deser et al., 2012; Shepherd, 2014).

The nonlinear component (ΔNL) is defined as the residual change in vertical moisture-advection variability after removing the thermodynamic and dynamic contributions and represents the interaction between changes in atmospheric moisture and variations in vertical-motion variability. Compared with the thermodynamic contribution, ΔNL is generally weak and exhibits predominantly negative values across many land regions (Figure 6e). Over North America and Europe, its regional-mean magnitude slightly exceeds that of ΔDY , mainly because positive and negative dynamic anomalies partly cancel in the regional average, whereas ΔNL exhibits a more spatially coherent weak negative signal. This indicates that the interaction term contributes more coherently to the regional mean than ΔDY , although both remain substantially smaller than ΔTH . Overall, these results demonstrate that thermodynamic moistening is the dominant driver of the asymmetric response of precipitation variability during CO_2 removal, whereas dynamic and nonlinear contributions remain comparatively small (Figure 6f).

4. Discussion

The dominant thermodynamic contribution identified here suggests that the asymmetric response of precipitation variability arises primarily from the slow adjustment of the ocean following peak CO_2 . Continued release of ocean heat maintains warmer surface conditions during the ramp-down phase, increasing lower-tropospheric moisture and thereby strengthening the thermodynamic component of precipitation variability. By contrast, circulation-related changes are less robust and more uncertain.

Although our analyses highlight the dominant role of thermodynamic moistening in driving the asymmetric response of precipitation variability, delayed ocean adjustment may also influence regional precipitation variability through additional processes, including changes in the interhemispheric energy balance, the associated southward displacement of the ITCZ and large-scale moisture transport (Kug et al., 2022; H. Oh et al., 2022; Song et al., 2022). Persistent Southern Ocean warming during the ramp-down phase modifies the interhemispheric energy balance, leading to a delayed southward displacement of the ITCZ (Kug et al., 2022; Figure S12b in Supporting Information S1), which may influence regional precipitation variability across tropical monsoon regions (Byrne et al., 2018; Kug et al., 2022; H. Oh et al., 2022). During the ramp-down phase, delayed ocean cooling is also associated with enhanced horizontal moisture transport from the tropical oceans toward adjacent land regions (e.g., Kug et al., 2022; H. Oh et al., 2022). Lower-tropospheric horizontal moisture-advection variability exhibits widespread positive RD–RU changes, but its limited spatial correspondence with precipitation variability suggests a complementary contribution to the asymmetric response (Figure S14 in Supporting Information S1). A more comprehensive understanding of the regional characteristics of precipitation variability during CDR will therefore require complete moisture-budget diagnostics in future studies.

Interpretation of the dynamical response should nevertheless be treated with caution because circulation changes remain substantially more uncertain than thermodynamic changes. A limitation of this study is that the moisture-budget decomposition is based on only three CMIP6 models that provide all variables required for the analysis.

Although the thermodynamic response associated with lower-tropospheric moistening is highly consistent across models, the dynamical response exhibits substantially larger inter-model spread, indicating greater sensitivity to circulation uncertainty and internal variability. A recent study has suggested that midlatitude storm-track activity may remain enhanced during the ramp-down phase relative to the ramp-up phase under comparable CO₂ concentrations (Hwang et al., 2024). This apparent difference may partly reflect the different dynamical diagnostics used in the two studies, as recent studies suggest that storm-track intensification under warming is increasingly modulated by moist diabatic processes in addition to circulation changes (Auestad et al., 2025, 2026; Ma et al., 2020). Importantly, the principal conclusion of this study is based on the robustness of the thermodynamic response rather than the dynamical contribution. Given the limited number of available models and the resulting inter-model uncertainty, the dynamical contribution diagnosed here should be regarded as preliminary evidence for circulation-related changes. Future studies using large-ensemble simulations are needed to better investigate the dynamical response to CO₂ removal.

5. Conclusions

Using idealized CO₂ ramp-up and ramp-down experiments from eight CMIP6 models, we show that precipitation variability exhibits a pronounced asymmetric response under CDR. Even at comparable atmospheric CO₂ concentrations, precipitation variability remains systematically stronger during the ramp-down phase than during the ramp-up phase. This asymmetry is robust across models, seasons and time scales and is particularly pronounced over densely populated regions including North America, Europe, South Asia and East Asia.

The asymmetric response is primarily driven by delayed ocean adjustment, which maintains warmer surface conditions and enhanced lower-tropospheric moisture during the ramp-down phase. Within the simplified moisture-budget framework, this delayed atmospheric moistening produces a dominant thermodynamic contribution to precipitation variability, whereas circulation-related changes play a much smaller role.

Although our results show that precipitation variability responds asymmetrically to increasing and decreasing CO₂ forcing, with short-timescale variability remaining elevated during CO₂ removal, this delayed adjustment may hinder the recovery of hydroclimatic stability and increase the likelihood of rapid transitions between wet and dry extremes. Nevertheless, reductions in greenhouse-gas concentrations still induce cooling and reduced atmospheric moisture, which act to suppress precipitation variability relative to higher-emission states. These findings highlight that, despite the asymmetric response, emission reductions and CDR remain effective in mitigating precipitation variability, although full recovery may require extended timescales.

Conflict of Interest

The authors declare no conflicts of interest relevant to this study.

Availability Statement

The CMIP6 model data used in this study are freely available from the website, <https://metagrid.esgf-west.org/search/cmip6/>.

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