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6 **Northward Dust-Induced Shift of Tropical Atlantic Precipitation in** 7 **CMIP6 Models**

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18 **Abstract.** Mineral dust has substantially increased since the pre-industrial period, particularly over North Africa
19 and the Atlantic Ocean. However, the impacts of increased dust burden over the tropical Atlantic Ocean on the
20 position of the Intertropical Convergence Zone (ITCZ) and regional precipitation remain poorly resolved. Here, we
21 analyze seven CMIP6 models participating in the Aerosol Chemistry Model Intercomparison Project
22 (AerChemMIP) piClim-2xdust experiment to quantify the changes in precipitation and the ITCZ and isolate the
23 doubling of dust concentration over the tropical Atlantic. We find a robust northward shift of the ITCZ, with
24 corresponding increased precipitation on its northern flank and suppressed precipitation to the south, consistent
25 across five of seven models and strongest in boreal spring and summer. This precipitation dipole is driven by a
26 coherent radiative–dynamical mechanism whereby asymmetric dust absorption and rapid cloud adjustments warm
27 the northern tropical Atlantic troposphere and cool the southern side, thereby strengthening the interhemispheric
28 energy gradient. This heating enhances vertical moisture transport northward – controlled primarily by dynamic
29 rather than thermodynamic adjustments – producing a convective dipole that tracks the precipitation response. Two
30 models representing dust as cloud condensation nuclei and ice-nucleating particles deviate from this response,
31 underscoring the sensitivity of precipitation responses to aerosol-cloud microphysical representation. Overall, our

findings suggest that if dust over North Africa continues to increase in the future, the associated radiative and dynamical adjustments identified here could significantly influence tropical Atlantic precipitation, with consequences for the West African monsoon.

1 Introduction

Precipitation over the tropical Atlantic Ocean is a central component of the global hydrological cycle and a key regulator of tropical climate variability (Dommenges & Latif, 2000). Much of this precipitation is organized by the Intertropical Convergence Zone (ITCZ), a narrow band of deep convection that migrates seasonally in response to hemispheric energy imbalances (Schneider et al., 2014). The Atlantic ITCZ is particularly influential because its meridional position modulates sea surface temperature (SST) gradients, surface fluxes, and large-scale overturning circulations that extend into adjacent continental regions (Byrne et al., 2018). Specifically, the Atlantic ITCZ directly affects the West African monsoon (WAM), shaping rainfall across the Sahel, the Gulf of Guinea, and the broader African continent (Maloney & Shaman, 2008; Nicholson, 2018; Tompkins & Adebisi, 2012; Taguella et al., 2025). These precipitation patterns sustain agriculture, water resources, and ecosystems while also influencing dust emissions from North Africa through rainfall-driven changes in soil moisture, vegetation cover, and surface erodibility across the Sahara-Sahel dust source region (Pierre et al., 2012; Prospero & Lamb, 2003). Beyond Africa, the Atlantic ITCZ variability influences hurricane activity in the tropical Atlantic (Liao et al., 2023; Vimont & Kossin, 2007) and moisture transport to South America and resultant hydrology in the Amazon basin (Zhao & Capotondi, 2026). Given the wide-ranging importance of the ITCZ to the global hydrological cycle, assessing its sensitivity to external perturbations is essential for understanding the controls on the position and intensity of tropical Atlantic precipitation.

Aerosols can modulate tropical precipitation by influencing cloud microphysics, radiative fluxes, and convective organization through the large-scale circulation (Fan et al., 2016). Among aerosol species relevant to precipitation, mineral dust plays an important role due to its abundance in the atmosphere and its multiple pathways for interacting with radiation and clouds, thereby influencing tropical precipitation (Huang et al., 2019; Knippertz & Todd, 2012). For example, dust particles can act as cloud condensation nuclei (CCN) and ice-nucleating particles (INP), altering droplet number concentrations and glaciation

60 pathways, thus modifying precipitation efficiency (DeMott et al., 2003, 2015; Vergara-Temprado et al.,
61 2017). In addition to microphysical pathways, dust plumes over the tropical Atlantic can lead to
62 substantial radiative perturbations that alter the thermodynamic and dynamical environment and,
63 consequently, precipitation (Dunion & Velden, 2004; Wong et al., 2009). For example, dust can absorb
64 and scatter both shortwave and longwave radiation, thereby warming the lower troposphere, stabilizing
65 the atmospheric column, and suppressing local convection. Previous studies have attributed such rapid
66 adjustments in convection, which occur independently of SST changes, to fast precipitation responses to
67 dust radiative perturbations (Haugvaldstad et al., 2025). Beyond these fast responses, dust can also
68 influence large-scale circulation patterns. By altering meridional temperature gradients, dust can shift the
69 Hadley circulation, modify the latitude of the ITCZ, and change the locations of surface convergence
70 zones (Yoshioka et al., 2007). These circulation-driven responses can produce basin-scale precipitation
71 anomalies, including the position of the Atlantic ITCZ (Evans et al., 2020). These factors highlight the
72 need for a comprehensive, multi-model assessment of dust–precipitation interactions, particularly over
73 the tropical Atlantic Ocean.

74 A central challenge in quantifying dust-precipitation interactions is the substantial uncertainty associated
75 with simulating the dust cycle in Earth system models (Gliß et al., 2021; Huneus et al., 2011a; Sand et
76 al., 2021). This is despite dust constituting a substantial fraction of global aerosol mass, with the majority
77 originating from sources over the Sahara and Sahel regions of North Africa, where it can influence tropical
78 Atlantic precipitation (Kok et al., 2021; Adebisi et al., 2025). Climate models vary widely in their
79 treatment of dust emissions, including threshold wind speeds, soil erodibility, and land-surface conditions
80 (Kok et al., 2021). They also differ in their representation of dust transport and deposition processes,
81 which determine the vertical and horizontal distribution of dust across the Atlantic Ocean, as well as their
82 assumptions about dust size distribution and optical properties, which strongly influence radiative forcing
83 and cloud interactions (Adebisi et al., 2023; Adebisi & Kok, 2020). These uncertainties propagate into
84 model estimates of dust’s influence on precipitation, contributing to the wide range of dust–precipitation
85 sensitivities reported in the literature (Huang et al., 2014; Zhao et al., 2011). Compounding these
86 challenges, paleorecords suggest that global dust mass loading has increased by approximately $55 \pm 30\%$
87 since pre-industrial conditions (Kok et al. 2023), driven by land-use change (e.g., agriculture (Ginoux et

88 al., 2012), desertification, and climate variability, which are largely missing in climate models (Mahowald
89 et al., 2010; Webb & Pierre, 2018). In addition, future dust trends are similarly uncertain, with projections
90 ranging from substantial increases to significant declines depending on assumptions about land
91 management, vegetation, and climate forcing (Kok et al., 2023; Li et al., 2021; Pu & Ginoux, 2017; Tegen
92 et al., 2004).

93 To address these gaps, we leverage the coordinated climate model experiments available through the
94 Coupled Model Intercomparison Project Phase 6 (CMIP6) to provide additional insight into how dust
95 influences precipitation over the tropical Atlantic. Experiments in CMIP6, such as the Aerosol Chemistry
96 Model Intercomparison Project (AerChemMIP; Collins et al., 2017), provide a unique opportunity to
97 examine dust–precipitation interactions across a broad ensemble of models with diverse dust
98 representations (see Table 1). In particular, the piClim-2xdust experiment – featuring a prescribed
99 doubling of dust emissions – paired with its corresponding control simulation, piClim-control, provides
100 a framework for diagnosing dust-induced changes in atmospheric circulation, radiative fluxes, and
101 precipitation. Although recent studies have used these experiments to quantify global dust effective
102 radiative forcing (Haugvaldstad et al., 2025) and to assess dust impacts in other regions such as Asia
103 (Zhao et al., 2024), to our knowledge, no study has focused on the regional precipitation responses over
104 the tropical Atlantic region using CMIP6 models. By analyzing these experiments across multiple CMIP6
105 models, we can assess the robustness of dust-induced precipitation shifts, identify model-dependent
106 sensitivities, and evaluate the extent to which dust contributes to variability in the tropical Atlantic ITCZ.
107 We find that dust perturbation produces a robust northward shift of the Atlantic ITCZ by preferentially
108 warming the northern tropical Atlantic atmosphere, enhancing vertical moisture transport and
109 precipitation north of the equator while suppressing precipitation to the south.

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112 **2. Data and Methodology**

113 **2.1 Dust Perturbation CMIP6 Sensitivity experiments**

114 To investigate the impact of dust emissions on tropical precipitation changes, we leverage seven models
115 (see Table 1) from the AerChemMIP sensitivity experiment in CMIP6 (Eyring et al., 2016), which used

116 a 30yr fixed sea-surface temperature (SST) simulation with boundary conditions set to pre-industrial
117 values except for dust aerosol emissions, which have doubled from their pre-industrial values (piClim-
118 2Xdust). Details of the treatment of dust and other aerosols in individual participating models can be
119 found in Haugvaldstad et al. (2025), with additional information in Table 1. Generally, in piClim-2xDust
120 experiments, dust emissions are doubled using model-specific tuning factors in each model's dust
121 emission scheme. However, the resulting atmospheric dust optical depth (DOD) does not necessarily scale
122 linearly with emissions, because changes in transport, removal, and circulation can modify the amount
123 and distribution of dust retained in the atmosphere. Therefore, dust perturbation exhibits substantial
124 diversity in dust representation across models and regions (see Section 3.1).

125 We used the full 30-year length of the simulation for each model run to capture internal variability and to
126 robustly estimate the simulated climatology. We note that the key advantages of using these models
127 include their state-of-the-art representations of dust processes and the availability of dust- and radiative-
128 forcing diagnostics needed for our analysis. However, we acknowledge limitations in dust representation
129 across models, including potential underestimation of dust loading and biases in simulated dust size
130 distributions (Adebiyi & Kok, 2020; Kok, 2011; Meng et al., 2022). In addition, each model run represents
131 dust deposition processes differently (Huneus et al., 2011), which contributes to the observed inter-
132 model spread. Our analysis acknowledges these limitations by focusing on ensemble-mean responses and
133 highlighting areas of inter-model agreement and disagreement.

Table 1: Information on CMIP6 models used in this study, including their resolution, dust sizes and refractive indices, and whether microphysics is enabled in the model, along with model references. Table is adapted from (Haugvaldstad et al., 2025)

Model	Lat°/long°	Vert. Levs	Size char.	Dust refrac. index	LW Scatt.	Ref. Emission scheme	Model Ref
CNRM-ESM2-1	1.4×1.4	91	(**) 0.01, 1.0, 2.5, 20 μm	$1.52 + 0.008i$	N	(Séférian et al., 2019)	(Nabat et al., 2020; Séférian et al., 2019)
GFDL-ESM4	1.0×1.2	49	(**) 1 (0.05), 2 (0.15) 3 (0.30), 6 (0.27), 10 μm (0.23)	$1.49 + 0.00203i$	N	(Evans et al., 2016)	(Dunne et al., 2020)
GISS-E2-1-G	2×2.5	40	(**) 0.2, 0.5, 1, 2, 4, 8 μm	$1.564 + 0.002i$	N	(Miller et al., 2006)	(Bauer et al., 2020)
IPSL-CM6A-LR-INCA	1.25×2.5	79	(***) 2.50 (2.0) μm	$1.52 + 0.00147i$	N	(Schulz et al., 1998)	(Balkanski et al., 2007; Hauglustaine et al., 2014; Lurton et al., 2020)
MIROC6	1.4×1.4	81	(**) 0.22 (0.0045), 0.46 (0.029), 1 (0.1766), 2.15 (0.2633), 4.64 (0.2633) 10.0 (0.2633)	$1.53 + 0.002i$	Y	(Takemura et al., 2009; Tegen et al., 2002)	(Tatebe et al., 2019)
MPI-ESM1-2-HAM	1.9×1.9	47	(****) 0.05–0.5 (1.59), > 0.5 μm (2.0)	$1.52 + 0.0011i$	Y ^(a)	(Tegen et al., 2019)	(Tegen et al., 2019)
UKESM1-0-LL	1.25×1.88	85	(**) 0.06324, 0.2, 0.6324, 2.0, 6.324, 20.0 63.24 μm	$1.53 + 0.00148$	N	(Woodward et al., 2022)	(Mulcahy et al., 2020; Williams et al., 2018)

* Mode number mean radius, with the standard deviation shown in parentheses. ** Upper diameter of each size bin; where available, the emitted mass fraction is provided in parentheses. *** Mass median diameter, with the standard deviation shown in parentheses. **** Dry-radius interval, with the standard deviation shown in parentheses. ^a Dust particles may be converted from the insoluble mode to the soluble mode through heterogeneous chemical processing.

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2.2 Diagnostics changes of Precipitation shift to dust perturbation emissions

We diagnose regional precipitation changes in response to an idealized doubling of mineral dust for each of the seven CMIP6 Earth system models and their multi-model mean as the difference between the perturbed dust (piClim-2Xdust) and the pre-industrial control run (piClim-Control).

$$\Delta P_i = P_{i,\text{piClim2xdust}} - P_{i,\text{piClim-Control}} \quad (2.1)$$

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Seasonal dust-induced responses are computed for four seasons: December-February (DJF), March-May(MAM), June-August (JJA) and September-October (SON). All fields used in the study are regridded to a common $2^\circ \times 2^\circ$ grid using a conservative remapping approach (Jones et al 1999). In addition, to determine whether dust-induced precipitation responses are statistically significant, we conducted a two-sample t-test for each model by comparing seasonal precipitation between piClim-2xdust and piClim-Control. For each season, individual years are treated as samples of internal climate variability, and the test evaluates whether the mean precipitation differs between the two experiments. Statistical significance in the multi-model mean is then evaluated based on the agreement in the sign and significance of the individual model responses across the seven models.

To quantify the ITCZ shift in response to doubled dust, we calculated the precipitation centroid between $20S^\circ$ and $20N^\circ$. First, for each season, model, and experiment (control and 2xdust), we compute the zonal-mean precipitation over the Atlantic-Africa longitude band ($40^\circ W-5^\circ E$) given by:

$$\bar{P}(\phi) = \langle P(\phi, \lambda) \rangle_{\lambda_1 \leq \lambda \leq \lambda_2} \quad (2.2)$$

Where ϕ latitude, λ longitude, and angle brackets denote the longitude average. We restrict the tropical latitude for this study to $\phi \in [-20^\circ, 20^\circ]$. Second, we further define the ITCZ latitude as the precipitation-weighted mean latitude given by:

$$\phi_{ITCZ} = \frac{\int_{-20^\circ}^{20^\circ} \phi \bar{P}(\phi) \cos \phi d\phi}{\int_{-20^\circ}^{20^\circ} \bar{P}(\phi) \cos \phi d\phi} \quad (2.3)$$

We then computed ϕ_{ITCZ} for both control and piClim-2xdust experiments and defined the dust-induced ITCZ shift as

$$\Delta \phi_{ITCZ} = \phi_{ITCZ}^{(piClim2xdust)} - \phi_{ITCZ}^{(piClim-Control)} \quad (2.4)$$

Therefore, positive values of $\Delta \phi_{ITCZ}$ indicate a northward shift of the precipitation belt.

To relate the ITCZ shift to the north-south redistribution of precipitation, we calculated a precipitation-response contrast between the northern tropical Atlantic region (North TA; $40^\circ W-5^\circ E$, $0^\circ-10^\circ N$) and the southern tropical Atlantic (South TA; $40^\circ W-5^\circ E$, $10^\circ S-0^\circ$) region. For each model and season, the dust-induced precipitation response was first calculated using Eq. (2.1) and then area-averaged over the North TA and South TA Atlantic regions. The precipitation-response contrast is defined as;

$$\Delta P_{ITCZ} = \Delta P_{North\ TA} - \Delta P_{South\ TA} \quad (2.5)$$

In addition, annual and seasonal ITCZ shifts were calculated for each year, and statistical significance was assessed using a two-sided one-sample t-test against zero. The null hypothesis is that the mean dust-induced ITCZ shift is zero. Shifts with $p < 0.05$ were considered statistically significant.

2.3 Radiative Kernel

Mineral dust can influence tropical precipitation not only by changing how much radiation reaches the surface, but also by altering where the atmosphere is heated. This is important for the location of the ITCZ because tropical convection is strongly tied to atmospheric heating. While precipitation balances radiative cooling in a global-mean sense, regionally this balance does not necessarily hold (Muller & O’Gorman, 2011). This is especially relevant for the tropical Atlantic, where dust-induced radiative heating can reorganize the circulation and shift the regional precipitation pattern. Although radiative forcing diagnostics are commonly applied to top-of-atmosphere (TOA) radiative perturbations, we focus here on atmospheric radiative heating and cooling because these quantities are more directly connected to the fast precipitation response in fixed-SST simulations. To quantify these radiative effects, we use radiative-kernel diagnostics to estimate dust effective radiative forcing and its rapid adjustment components (Soden et al 2008).

Dust effective radiative forcing (ERF) comprises the instantaneous radiative forcing (IRF), defined as the direct radiative perturbation caused by the change in the forcing agent, and rapid adjustments.

$$ERF = IRF + A_x \rightarrow (2.6)$$

Consequently, the relative contribution of IRF to the total atmospheric radiative response, we calculated the fractional IRF contribution as $100 \times \frac{IRF}{ERF}$. The total rapid adjustment term, A_x can be defined as changes in atmospheric and surface conditions that occur in response to the IRF but are independent of global-mean surface temperature change. It is decomposed into contributions from atmospheric temperature (T), water vapor (q), surface albedo (a), and clouds (C).

$$A_x = A_T + A_q + A_a + A_c \rightarrow (2.7)$$

The IRF isolates the perturbation from aerosol-radiation interactions, corresponding to the direct radiative effect of dust, while radiative forcing associated with aerosol-cloud interactions is captured by the cloud

adjustment A_c . Since SSTs are fixed in these simulations, radiative feedback associated with SST changes are suppressed, allowing radiation changes to be attributed to the effective radiative forcing (ERF). We quantify these components as identified in equations 2.6 & 2.7 using the radiative kernel approach of Soden et al. 2008, following the ERF decomposition described by Myhre et al. 2018; Smith et al. 2018;2020; Thornhill et al. 2021. In this kernel, non-cloud A_x terms are defined as:

$$A_x = \frac{\partial R}{\partial x} \Delta x = K_x \Delta x \quad (2.8),$$

where $K_x = \partial R / \partial x$ is the radiative kernel, a precomputed diagnostic tool defined horizontally, vertically, and for each month in a year, and R is the radiative flux being diagnosed. The Δx term is the actual change in x , diagnosed as the time-mean difference in x between piClim-2xdust and piClim-control. In addition, cloud radiative kernels provide a more explicit estimate of cloud adjustments, but their use requires diagnostics that are not available from these simulations. We therefore define A_c using the conventional cloud radiative forcing residual, corrected for cloud-masking effects, following Soden et al. (2004).

$$A_c = (ERF - ERF^{clr}) - (IRF - IRF^{clr}) - \sum_{x \in \{T_s, T, q, a\}} (A_x - A_x^{clr}) \quad (2.9),$$

where *clr* refers to clear-sky conditions and A_x^{clr} are diagnosed with clear-sky radiative kernels. The clear-sky instantaneous radiative forcing IRF^{clr} is estimated as the residual between the clear-sky effective radiative forcing, ERF^{clr} , and the clear-sky rapid adjustment term, A_x^{clr} . To convert this clear-sky estimate to all-sky conditions, we apply separate cloud-masking factors to the shortwave and longwave components;

$$IRF_{SW} = \frac{IRF_{SW}^{clr}}{C_{SW}}, \quad IRF_{LW} = \frac{IRF_{LW}^{clr}}{C_{LW}} \quad (2.10)$$

Where $C_{SW} = 2.43$ and $C_{LW} = 1.24$. These factors account for the masking of radiative forcing by clouds and follow the kernel-IRF approach of Kramer et al. (2021). The SW factor is based on aerosol forcing, while the LW factor was derived from clear-sky and all-sky double-call radiative transfer calculations for CO_2 and has been used as a representative longwave cloud-masking factor. Although these factors are not dust-specific, they are comparable in magnitude with dust-specific clear-sky to all -

sky TOA forcing ratios reported by Zhu et al. (2007) over major dust outflow regions. We therefore treat the resulting all-sky IRF as a diagnostic estimate and acknowledge uncertainty associated with the cloud masking correction. In addition, to examine the role of radiative forcing on dust-induced precipitation response, we calculated the inter-model relationship between each radiative component (see 2.6 & 2.7) north-south contrast ($X_{North\ TA} - X_{South\ TA}$) and precipitation north-south contrast ($\Delta P_{North\ TA} - \Delta P_{South\ TA}$) using ordinary least-squares regression and Pearson correlation for each season and the annual mean. Correlations with $p < 0.05$ were considered statistically significant.

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244 **2.4 Moisture Budget decomposition**

To better understand the dynamical pathway through which dust influences precipitation, we analyze the atmospheric moisture budget and compare the piClim-2xDust experiment with the piClim-control simulation. The moisture budget describes the balance between precipitation, evaporation, and moisture transport within the atmospheric column (Akinsanola & Zhou, 2019).

$$249 \quad \Delta P = \Delta E - \Delta(\partial_t q) - \Delta\langle V_h \cdot \nabla q \rangle - \Delta\langle \omega \partial_p q \rangle \rightarrow (2.11)$$

Where P is precipitation, E represents surface evaporation, q is specific humidity, V_h is the horizontal wind vector, and ω is pressure velocity. The Δ denotes the dust-induced change, calculated as the difference between piClim-2xDust and piClim-Control. The terms $\Delta(\partial_t q)$, $\Delta\langle V_h \cdot \nabla q \rangle$, and $\Delta\langle \omega \partial_p q \rangle$ represent dust-induced changes in the atmospheric moisture storage tendency, horizontal moisture advection, and vertical moisture advection. The angle brackets $\langle \cdot \rangle$ denote the mass-weighted vertical integration of atmospheric variables over the tropospheric column (1000-100 hPa), defined for any variable, M , as

$$257 \quad \langle M \rangle = \frac{1}{g} \int_{p_t}^{p_s} M dp \rightarrow (2.12)$$

Where g is the gravitational acceleration, p_s is surface pressure, and p_t corresponds to the upper integration level of the troposphere (100 hPa). The above expression ensures that the vertically integrated moisture flux divergence is directly comparable to changes in surface precipitation, because in the tropics, precipitation variability is often strongly controlled by vertical moisture transport associated with large-scale ascent and subsidence.

263 To isolate the processes responsible for precipitation response to dust perturbation, we further
 264 decomposed the vertical moisture advection into three components (dynamic, thermodynamic, and
 265 nonlinear components) described as follows (Seager et al., 2010).

$$266 \quad -\Delta\langle\omega\partial_p q\rangle = -\langle\Delta\omega\partial_p \bar{q}\rangle - \langle\bar{\omega}\partial_p \Delta q\rangle - \langle\Delta\omega\partial_p \Delta q\rangle \rightarrow (2.13)$$

267 Here, the dynamic component $-\langle\Delta\omega\partial_p \bar{q}\rangle$ quantifies the influence of dust-induced circulation changes on
 268 vertical moisture transport. This term captures the effect of modifications in large-scale ascent and
 269 subsidence driven by dust-induced atmospheric heating and circulation adjustments. The thermodynamic
 270 component $-\langle\bar{\omega}\partial_p \Delta q\rangle$ represents changes arising from dust-induced modifications in atmospheric
 271 humidity under the background circulation. Lastly, the nonlinear component $-\langle\Delta\omega\partial_p \Delta q\rangle$ accounts for
 272 the interaction between circulation anomalies and moisture anomalies. This term is generally smaller than
 273 the dynamic and thermodynamic components but can contribute to the total moisture budget in regions
 274 with strong moisture-circulation coupling. By decomposing the moisture budget, we can attribute changes
 275 in precipitation to specific dynamical and thermodynamical processes. This process-based, mechanistic
 276 understanding provides an interpretation of how increased atmospheric dust loading alters moisture
 277 transport, atmospheric circulation, and directly shifts precipitation patterns over the tropical Atlantic and
 278 West African monsoon regions.

279

280 **3. Results**

281 **3.1 Diversity in simulated dust perturbation**

282 We examined dust optical depth, as a proxy for dust loading, in response to doubled emission perturbation
 283 and found substantial diversity in the simulated dust response over North Africa and the tropical Atlantic.
 284 This diversity is not unexpected, as previous evaluations of CMIP6 dust simulations have shown that
 285 models can reproduce the broad observed dust distribution but differ substantially in dust emissions,
 286 removal, optical properties, and dust optical depth (Zhao et al. 2022). We therefore interpret the piClim-
 287 2xDust response not only as a forced increase in dust, but also as a measure of how differently CMIP6
 288 models translate the same idealized emission perturbation into atmospheric dust optical depth.

289 First, while dust emissions are perturbed in the piClim-2xDust simulation, the multi-model mean (MME)
 290 shows broadly similar dust transport pathways over much of North Africa and the Atlantic Ocean (Figure

291 1a & b). This is expected because both experiments use pre-industrial meteorological conditions, so the
 292 circulation patterns influencing dust transport remain broadly similar within each model. However, these
 293 background conditions differ across models because of differences in their piControl meteorology,
 294 contributing to inter-model spread in dust distribution. In addition, different representations of dust
 295 processes at emission and during transport, along with different representations of optical properties
 296 across the models (Table 1 & Fig. S1) can result in differences in dust optical depth (DOD) in response
 297 to dust perturbation (Fig. 1). For example, across individual models, the difference in DOD between
 298 control and 2xdust experiments varies substantially over the tropical Atlantic (between 0.021-0.16 for
 299 north (green: 0-10°N) and 0.001-0.029 for south (red: 10°S – 0°); Fig. 1g). Notably, models such as
 300 IPSL-CM6A-LR-INCA and MPI-ESM-1-2-HAM produce relatively large increases in dust over the
 301 tropical Atlantic (Fig. 1g). One possible reason for this large difference could be because of the
 302 combination of low imaginary refractive index ($k < 0.00147i$) and the small particle size ($D < 2.50 \mu m$)
 303 producing more scattering at 550 nm (Table 1), unlike UKESM1-0-LL with similar imaginary refractive
 304 index but with larger particles size ($D_{max} = 63.24 \mu m$). In contrast, models such as MIROC6, GISS-
 305 E2-1-G, and GFDL-ESM4 simulate a more moderate response in DOD, in part because of the similarity
 306 in size distribution and refractive index (Table 1), although large differences in DOD over north tropical
 307 Atlantic can be associated with differences in dust transport (Fig. 1g). We also show differences in the
 308 DOD response to dust emission perturbation among the models by estimating the annual percentage
 309 change in DOD perturbation relative to the control simulation (Fig. 1h & S2). The results show less than
 310 100% change in DOD for the 2x-dust simulation over much of North Africa but more than 100% change
 311 over parts of the North Atlantic Ocean relative to the control simulation (Fig. S2a). This reflects that more
 312 of the finer dust particles are being transported downwind, even though a lot of the coarser dust particles
 313 are deposited shortly after emission. Over the tropical Atlantic Ocean (Fig. 1h), however, this percentage
 314 change in the DOD perturbation is largely less than 100 % across most of the models (MME is 89%),
 315 except for MIROC6 and MPI-ESM-1-2-HAM, which are decidedly above 100%.
 316 The differences in DOD perturbation across models are also reflected in the seasonal distribution (Fig.
 317 1d-f & Fig. S3). For example, maximum DOD typically occurs between 15-25°N during spring (Mar-
 318 May) and summer (Jun-Aug), corresponding to the pre-monsoon and early monsoon period over the West

319 and North African region (Fig. 2d-f & Fig. S3). Over the tropical Atlantic, the maximum DOD occurs
320 during winter (Dec-Feb) when the ITCZ is most at the southernmost point. Overall, our results show that
321 the seasonal latitudinal migration of the dust transport is consistent with the migration of the ITCZ (see
322 lines representing the location of the ITCZ in MME; Fig. 2), highlighting that dust transport over North
323 Africa and the Atlantic Ocean is intimately linked to the large-scale circulation and precipitation over the
324 tropical Atlantic Ocean.

325

326 **3.2 Dust-Induced Precipitation Response**

327 To understand changes in precipitation in response to dust perturbation, we estimate the change in
328 precipitation between the control and 2xdust experiments (Eqn. 2.1). We find that dust perturbation results
329 in precipitation response characterized by changes in intensity and shift in maximum precipitation over
330 the tropical Atlantic that varies across models (Fig. 2). Across most of North Africa and the North Atlantic
331 Ocean, there are small increases in precipitation due to dust perturbation in the MME (Fig. 2c & Fig. S4).
332 However, the largest change is in the tropical Atlantic, indicating increased precipitation on the north side
333 of the ITCZ and decreased precipitation on the south side (Fig. 2a-c). This change in precipitation is
334 largely consistent across at least 5 of 7 models (Fig. 2f & g). However, two models exhibit patterns that
335 differ from the MME. In contrast to other models, MIROC6 shows that the precipitation response to dust
336 perturbation results in increases in the south and decreases in the north of the ITCZ, whereas the MPI-
337 ESM-1-2-HAM model shows a decrease in precipitation over both the north and south regions in response
338 to dust perturbation (Fig. 2g & Fig. S5). The reason for these different precipitation responses is likely
339 due to the different representation of dust as CCN and INP in MIROC6 and MPI-ESM-1-2-HAM
340 compared to other models, thereby inducing substantial cloud microphysical responses and resultant
341 precipitation distribution (see Table 1).

342 In addition, these north-south precipitation responses to dust perturbation are also consistent across
343 seasons, as the ITCZ moves latitudinally over the tropical Atlantic (Fig. 2d-f). Specifically, on the north
344 side of the ITCZ, precipitation increases mostly occur during spring and summer when dust loading is at
345 its maximum over North Africa and the Atlantic Ocean (compare Fig. 2f to 1f). In contrast, south of the
346 ITCZ, precipitation decreases are predominantly in winter and spring. These seasonal changes in

347 latitudinal precipitation response also vary widely across models (Fig. S5). The inter-model differences
 348 seem to depend on the baseline distribution of simulated precipitation in the control, resulting in different
 349 precipitation responses to dust perturbations across locations and seasons across the models (Fig. S5).
 350 Overall, dust-induced precipitation response to dust perturbation over the north side of the tropical
 351 Atlantic accounts for 0.78%, and the decrease over the south side of the tropical Atlantic accounts for
 352 1.76% of the mean precipitation in the control simulation (Fig. 2h).
 353 Furthermore, we also find that the north-south precipitation response results in an annual, multi-model
 354 mean 0.082° northward shift in the ITCZ location over the tropical Atlantic (Fig. 3). Using an alternative
 355 metric based on the median latitude of the precipitation distribution, rather than the precipitation centroid,
 356 yields a comparable MME northward shift of 0.11° (Fig. S6). Although the ITCZ shift is northward for
 357 MME, its northward extent varies depending on model, location, and season (Fig. 3a & S4). Specifically,
 358 5 out of 7 models show a mean northward shift (ranging between 0.16° and 0.36°), while two models
 359 (MIROC6 and MPI-ESM-1-2-HAM) show a mean southward shift in the ITCZ location. Annual dust-
 360 induced ITCZ shifts are statistically significant in five models (CNRM-ESM2-1, GFDL-ESM4, IPSL-
 361 CM6A-LR-INCA, MIROC6, and UKESM1-0-LL). In addition, as a function of seasons, the relationship
 362 between precipitation response around the ITCZ (calculated as $\Delta P_{ITCZ} = \Delta P_{north} - \Delta P_{south}$) and the
 363 ITCZ shift varies seasonally (Fig. 3b), with spring showing the strongest precipitation-response sensitivity
 364 to ITCZ and the weakest in winter (Fig. 3b). On annual timescales, this relationship remains positive
 365 across all 7 models ($r = 0.74$). Overall, dust perturbation results in a significant precipitation response in
 366 the tropical Atlantic Ocean, with changes in magnitude and a shift in the maximum precipitation, leading
 367 to a significant northward shift in the location of the ITCZ.

368

369 **3.3. Influence of radiative forcing on dust-induced precipitation response**

370 To understand the mechanism responsible for the dust-induced precipitation response and the shift in the
 371 ITCZ location over the tropical Atlantic, we first examine the change in the effective radiative forcing
 372 induced by the dust perturbation in the atmosphere. Previous studies have suggested that the radiative
 373 effects of dust, in cases of high dust loading, often dominate microphysical changes where dust particles
 374 act as CCN or INP (Bangert et al., 2012; McGraw et al., 2020; Pan et al., 2024). Moreover, because 5 out

of 7 models represented here do not include dust microphysical changes, this suggests that the simulated precipitation response is strongly associated with the response in the radiative forcing to dust perturbations (Haugvaldstad et al., 2025; Zhao et al., 2024). Consequently, the spatial distribution of dust radiative absorption due to dust perturbations tracks changes in DOD (compare Figure 4a and Fig. 1c), with cooling in the longwave dominated by warming in the shortwave (Fig. S7). Note that the difference in the absolute magnitude of the shortwave warming and longwave warming over the Atlantic Ocean is, in part, because of the preferential long-range transport of finer particles which interact more strongly in shortwave than coarser particles in the longwave that settles much faster after emission. Although the largest net radiative heating occurs offshore of North Africa, a substantial fraction of the warming also extends over the tropical Atlantic region north of the ITCZ (Fig. 4a, d, & g). This warming on the northern side of the ITCZ, along with the associated cooling south of the ITCZ, is represented in 5 of 7 models (Fig. 4g & Fig S8). The remaining two odd models are MIROC6 and MPI-ESM-1-2-HAM, which show some cooling over both regions (Fig. 4g & Fig S8).

These atmospheric radiative responses broadly scale with the instantaneous radiative forcing (IRF) across models and seasons. We find that IRF explains approximately 72% of the variance in total radiative forcing induced by dust perturbation, dR_{atm} , in the atmosphere (Fig. 4a, b, d, & e; Fig. S7 & S8). In addition, because the radiative heating response to dust perturbations can arise from rapid adjustments in the atmospheric layer, we estimate the adjustment due to changes in land surface temperature, atmospheric temperature, stratospheric temperature, water vapor, surface albedo, and clouds (see methods). Among these adjustments, the change in radiative heating due to cloud adjustment is about an order of magnitude larger than the other adjustments (Fig. 4c & S8). Similar to IRF, cloud adjustments predominantly result in heating on the northern side of the ITCZ and cooling on the southern side (Fig. 4c, f, i). The cooling south of the ITCZ highlights that the regional radiative response is not controlled solely by dust direct effects. South of the ITCZ, where dust perturbation is weaker, rapid cloud adjustments appear to dominate and locally offset the direct dust heating. This suggests that the southern cooling is likely associated with dust-induced circulation changes (see section 3.4), rather than a purely local response to dust loading.

Through changes in instantaneous radiative forcing and rapid adjustment due to changes in cloud cover, we find that the radiative heating response to dust perturbation strongly correlates with the precipitation response over the tropical Atlantic (Fig. 5). Figure 5 shows scatterplots between precipitation response around the ITCZ ($\Delta P_{north} - \Delta P_{south}$) and changes in radiative forcings ($\Delta X_{north} - \Delta X_{south}$; where X are the radiative forcing terms). The strong association between the precipitation response and the radiative forcing response yields a high annual correlation value of 0.84 across the 7 Earth System Models, and a seasonal value as high as 0.97 for MAM that is statistically significant ($p < 0.05$) (Fig. 5a). Similar statistically significant ($p < 0.05$) high correlation values for IRF and cloud adjustments, of 0.77 and 0.88, respectively (Fig. 5b & c). With the relatively small magnitude of the other adjustment terms, their relationships with the precipitation responses are much weaker, suggesting smaller influence on the radiative forcing from adjustments due to changes in temperature, specific humidity and albedo, over the tropical Atlantic (Fig. 5e-f & Fig. S8). Overall, our results shows that precipitation response and the associated shift in ITCZ over the tropical Atlantic is strongly associated with radiative forcing and associated cloud adjustment induced by the dust perturbation over the region.

3.4 Influence of dynamics on dust-induced precipitation response

Although the radiative forcing response is associated with cloud adjustments that can directly influence precipitation (Fig. 5), it can also induce dynamical responses that indirectly affect precipitation over the tropical Atlantic. To understand this dynamical response to dust perturbations, we conduct a moisture budget analysis to estimate the zonal, meridional, and vertical moisture transport (Fig. 6 and Eqn. 2.11). Our analysis reveals that the vertical advection component dominates the moisture response to dust perturbations across most seasons in the multi-model mean ensemble (Fig 6a-f & Fig. S9). This vertical advection term represents the tendency of upward transport of moisture through the troposphere, which is crucial for sustaining mid-level moisture and convective instability (Tomassini, 2020) . The largest change in the vertical moisture advection occurs broadly across the North Atlantic Ocean, primarily during spring and summer (Fig. S9). During this period, there are some contributions from the meridional advection term (Fig. S9), but such contributions are primarily from two models (GISS-E2-1-G and MIROC6; Fig. S10). In contrast to the meridional advection term, the dominant contribution of the vertical advection accounts for about 76.3% of the total moisture advection over the tropical Atlantic.

Specifically, we found strong vertical moisture transport over the northern tropical Atlantic (0-10°N), which could strengthen mid-tropospheric moisture and increase convective activity over this region. In contrast, the southern equatorial region (10°S – 0°) exhibited weaker vertical moisture transport, indicating suppressed convective activity. The modelled dipole pattern suggests a northward shift in ascending motion and corresponds to a northward shift in the precipitation pattern (compare Fig. 2 & 6). For 5 of the 7 models, the dynamical response consistently shifts the ascending region northward, resulting in enhanced (reduced) vertical moisture transport in the northern (southern) parts of the tropical Atlantic (Fig. S10). The exceptions are MIROC6 and MPI-ESM-1-2-HAM, which, like the precipitation response, result in the southward shift in the ascending region (Fig. 6h & S10).

To better understand what controls the northward shift in the ascending region over the tropical Atlantic, we further decompose the vertical moisture transport term $\Delta\langle\omega\partial_p q\rangle$ into dynamic $-\langle\Delta\omega\partial_p \bar{q}\rangle$ and thermodynamic $-\langle\bar{\omega}\partial_p \Delta q\rangle$ contributions. Our result indicates that the dynamic component dominates the response of the vertical moisture transport to dust perturbations across the tropical Atlantic domain in all seasons (Fig. 7 & Fig. S11 & S12). Similarly, over the tropical Atlantic, the decomposition of the terms shows that the dynamical components of vertical moisture transport tracks and dominate the total vertical moisture transport across the seasons for each model (Fig. S13 & S14). In contrast, the thermodynamic component related to changes in atmospheric humidity is comparatively small. This result suggests that dust perturbation primarily modifies moisture transport through circulation adjustment rather than through changes in atmospheric moisture content alone over the tropical Atlantic. This defined mechanism by which the precipitation responds to dust perturbation shows in the strong relationship between precipitation response around the ITCZ and the vertical moisture transport terms (Fig. 8). Specifically, the correlation between the precipitation response and the vertical moisture transport shows 0.95, significantly strong in all seasons (Fig. 8 & Fig. S15). Overall, our results show that dust perturbation strongly modulates precipitation over the tropical Atlantic, shifting precipitation and the ITCZ northward, with a mechanism primarily driven by cloud and dynamical adjustments and controlled by changes in the vertical moisture transport.

458 **4. Discussion and Conclusion**

459 Understanding how mineral dust influences tropical Atlantic precipitation is critical for constraining the
460 sensitivity of the ITCZ to aerosol forcing, with direct implications for the West African Monsoon, Sahel
461 rainfall, and broader tropical circulation. Using seven CMIP6 models participating in AerChemMIP, we
462 isolated the fast atmospheric response to a doubling of dust emissions by comparing 30-year fixed-SST
463 piClim-2xdust simulations with their piClim-control counterparts and diagnosing changes in
464 precipitation, ITCZ position, radiative forcing, and moisture budget components. We find that this dust
465 perturbation produces a robust northward shift in the Atlantic ITCZ, with increased precipitation on the
466 northern flank and suppressed precipitation on the southern flank. This pattern is consistent across five of
467 seven models and across all seasons, though its magnitude varies substantially with model, most strongly
468 in boreal spring and summer when dust loading over North Africa and the Atlantic is near its seasonal
469 peak.

470 In addition, we find that this precipitation dipole emerges from a coherent radiative–dynamical
471 mechanism, whereby the absorption of radiation by dust drives a circulation pattern that results in a
472 precipitation response. Specifically, the preferential loading of dust north of the ITCZ generates
473 asymmetric atmospheric absorption, warming the northern tropical Atlantic troposphere while leaving the
474 southern side comparatively cooler, thereby strengthening the interhemispheric energy gradient that
475 controls ITCZ position. Among the components of effective radiative forcing, cloud rapid adjustments
476 emerge as the dominant modulator, amplifying the meridional heating contrast and contributing an order
477 of magnitude more than adjustments in temperature, water vapor, or surface albedo. This radiative heating
478 then drives a dynamical response that is the proximate cause of the shift in precipitation. Specifically,
479 dust-induced circulation adjustments enhance upward moisture transport over the northern tropical
480 Atlantic while suppressing it over the southern tropical Atlantic, producing a convective dipole that
481 closely tracks the precipitation anomaly. Critically, this dynamical response is shown to be driven
482 primarily by changes in large-scale circulation — the dynamic component of vertical moisture advection
483 — rather than by changes in atmospheric moisture content alone, indicating that dust reshapes tropical
484 Atlantic precipitation principally by reorganizing the atmospheric circulation rather than by modifying
485 the thermodynamic environment.

486 Furthermore, our results also highlight a potential contrasting role of aerosol-cloud microphysics to the
487 radiative–dynamical mechanism in dust-precipitation interaction over the tropical Atlantic. The two
488 models that deviate from the above multi-model consensus – MIROC6 and MPI-ESM-1-2-HAM – also
489 represent dust as cloud condensation nuclei and ice-nucleating particles, introducing microphysical
490 pathways that substantially alter cloud and precipitation responses and, in some cases, reverse the sign of
491 the ITCZ shift, underscoring the sensitivity of dust-precipitation interactions to the representation of
492 aerosol-cloud microphysics. We do not interpret this deviation as evidence that either group of models is
493 necessarily wrong. Rather, it suggests that models without explicit dust-cloud microphysical effects
494 mainly capture the radiative-dynamical pathway, whereas models that include dust as CCN and INP allow
495 additional microphysical pathways that may amplify, offset, or even oppose the radiative response.
496 Because only two models in this ensemble include these processes their behaviour should be interpreted
497 as an important source of structural uncertainty rather than a definitive constraint on the real-world sign
498 of the response.

499 This behaviour is broadly consistent with previous studies showing that dust acting as CCN and INP can
500 modify droplet number, glaciation pathways, and precipitation efficiency in deep convective clouds,
501 sometimes invigorating convection and sometimes suppressing rainfall depending on environmental
502 factors (DeMott et al., 2003, 2015; Min et al., 2009; Rosenfeld et al., 2001). While our results highlight
503 the radiative–dynamical mechanism, future precipitation response to dust perturbation would depend on
504 the balance between radiative and microphysical effects of dust over the tropical Atlantic, suggesting the
505 need for better constraints on aerosol-cloud interactions for future projection of precipitation.

506 In addition, our findings have important implications for dust impacts on future projections of
507 precipitation, where model representation of dust properties for fast atmospheric response and the role of
508 sea surface temperature will be crucial on short- and long-term scales, respectively. On short timescales,
509 the fast precipitation response identified here – driven by radiative absorption and cloud adjustments – is
510 highly sensitive to how models represent dust optical properties, particularly particle size distribution and
511 the complex refractive index (e.g., Balkanski et al., 2021). Climate models consistently underestimate
512 coarse dust by up to a factor of four relative to observations, missing the substantial longwave absorption
513 and additional shortwave warming that coarse particles provide (Adebiyi & Kok, 2020). Since our results

show that radiative absorption is the dominant driver of the northward ITCZ shift, this systematic underestimation of coarse dust implies that models likely underestimate the magnitude of the dust-induced precipitation response over the tropical Atlantic. Compounding this, most models commonly overestimate the imaginary part of the dust shortwave refractive index, particularly over North Africa (Adebisi et al., 2023). This can result in misrepresentation of the balance between dust scattering and absorption, which directly controls the sign and magnitude of atmospheric heating (Kok et al., 2023), and thereby affects both the interhemispheric energy contrast and the resulting ITCZ shift. On longer timescales, the fixed-SST framework used here captures only the fast atmospheric adjustment and excludes ocean feedback that could substantially amplify the precipitation response. In a coupled system, dust-induced asymmetric radiative heating north of the ITCZ would cool tropical North Atlantic SSTs through reduced surface insolation while simultaneously driving wind-evaporation-SST feedback that reinforce the interhemispheric SST gradient and further displace the ITCZ northward (Strong et al., 2015). These dust-SST feedback have been shown to modulate the Atlantic Meridional Mode on interannual to decadal timescales, with cascading impacts on Sahel rainfall and West African monsoon precipitation (Williams et al., 2016). Overall, the precipitation shifts identified here are likely a conservative lower bound on the full dust-climate response, and coupled model experiments with improved dust optical properties and size distributions are needed to better constrain the total impact of dust on tropical Atlantic hydroclimate.

532

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541

542 **Competing interests:**

543 The authors declare that they have no competing interests.

544

545 **Code and Data availability**

546 All data used in this study are publicly available. The Aerosols and Chemistry Model Intercomparison
547 Project (AerChemMIP) data used is available from World Climate Research Programme and the Earth
548 System Grid Federation for open access to the data of AerChemMIP
549 (https://aims2.llnl.gov/search/cmip6/?mip_era=CMIP6&activity_id=AerChemMIP,WCRP,2024a, last
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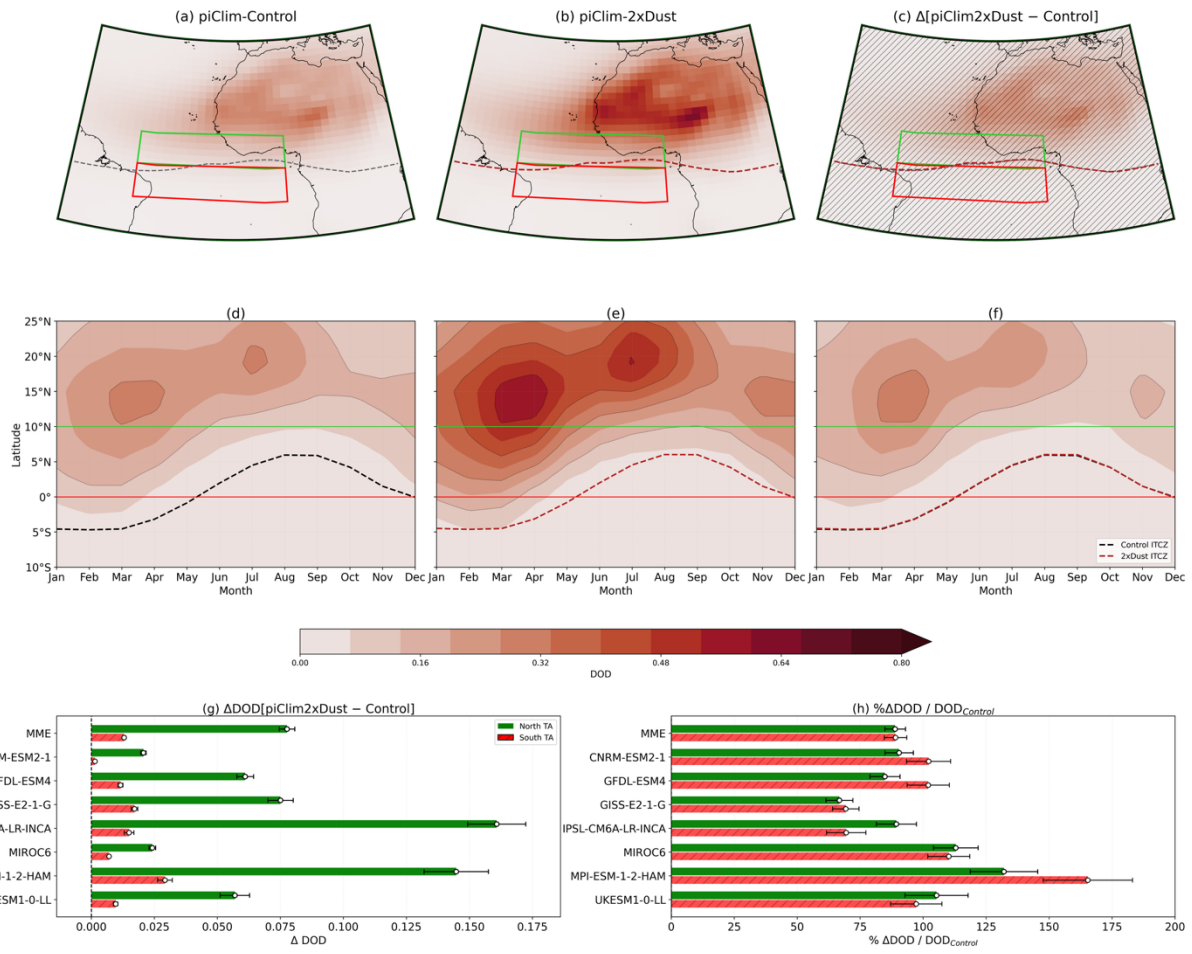
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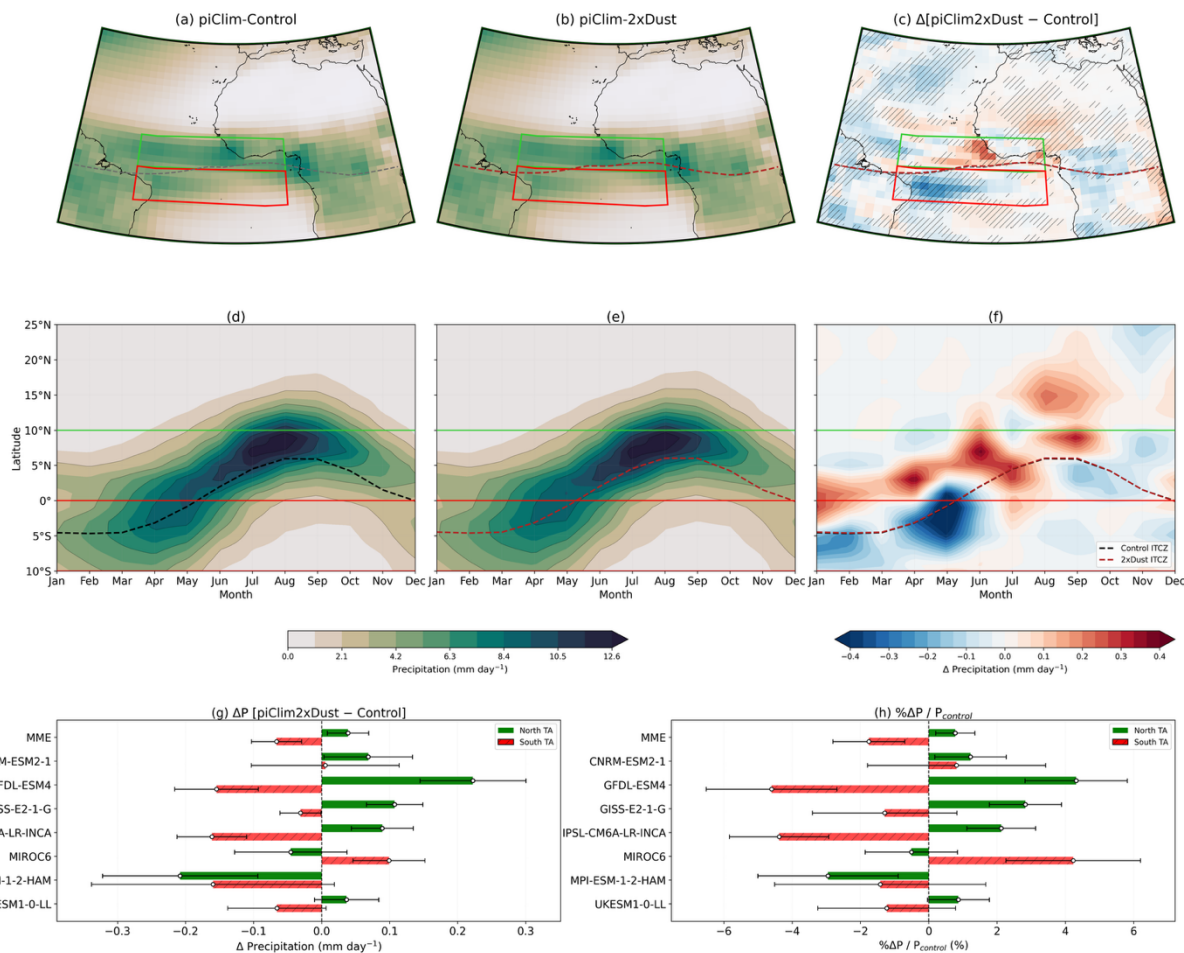
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843 Figure 1. Dust optical depth (DOD) response to dust emission perturbation. Annual mean dust optical depth (DOD)
844 in (a) piClim-Control and (b) piClim-2xDust, and (c) DOD difference relative to control. Hatching indicates grid
845 cells where at least 70% of models agree on the sign of the DOD change. Green and red boxes denote the North
846 Tropical Atlantic (North TA; 40°W–5°E, 0°–10°N) and South Tropical Atlantic (South TA; 40°W–5°E, 10°S–0°),
847 respectively. Panels (d–f) show the seasonal zonal mean DOD averaged over 40°W–5°E for piClim-Control,
848 piClim-2xDust, and their difference. Dashed curves show the precipitation-centroid ITCZ position in the control
849 (black) and 2xDust (red). Panels (g, h) show the individual model and regional mean response for North TA and
850 South TA, expressed as (g) DOD change to double dust and (h) percentage change relative to the control. Bars
851 show annual means and error bars represent the standard error.



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858 Figure 2. Precipitation response to dust emission perturbation. Annual mean precipitation in (a) piClim-Control
859 and (b) piClim-2xDust, and (c) difference relative to the control. Hatching shows grid cells where at least 70% of
860 models agree on the sign of precipitation change. Green and red boxes denote the North Tropical Atlantic (North
861 TA; 40°W–5°E, 0°–10°N) and South Tropical Atlantic (South TA; 40°W–5°E, 10°S–0°), respectively. Panels (d–
862 f) show the seasonal zonal mean precipitation averaged over 40°W–5°E for piClim-Control, piClim-2xDust, and
863 their difference. Dashed curves show the precipitation-centroid ITCZ position in the control (black) and 2xDust
864 (red). Panels (g, h) show the individual model and regional mean response for North TA and South TA, expressed
865 as (g) precipitation change to double dust and (h) percentage change relative to the control. Bars show annual means
866 and error bars represent the standard error.

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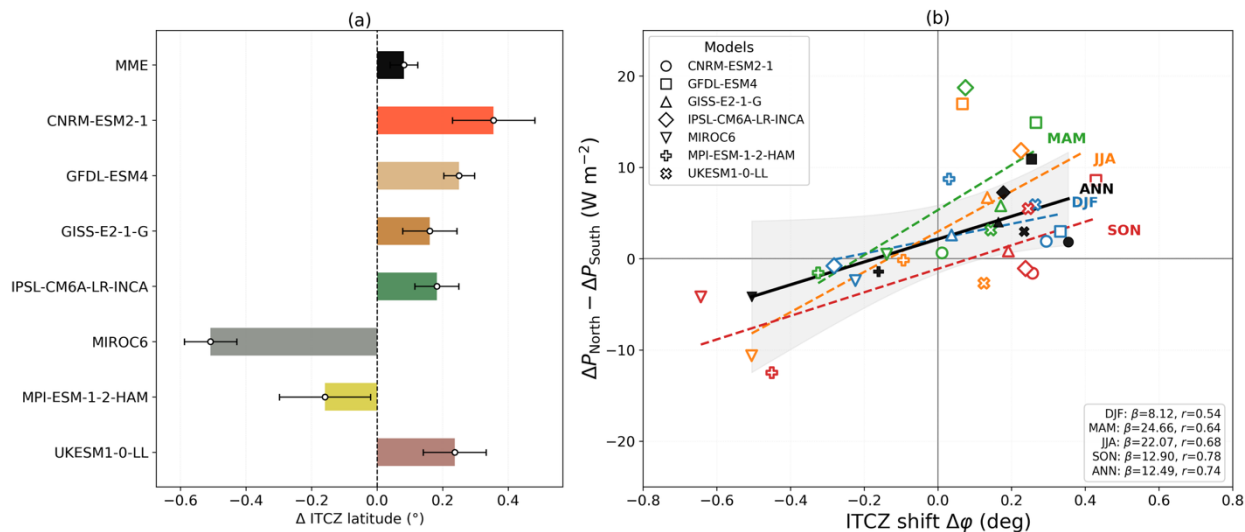
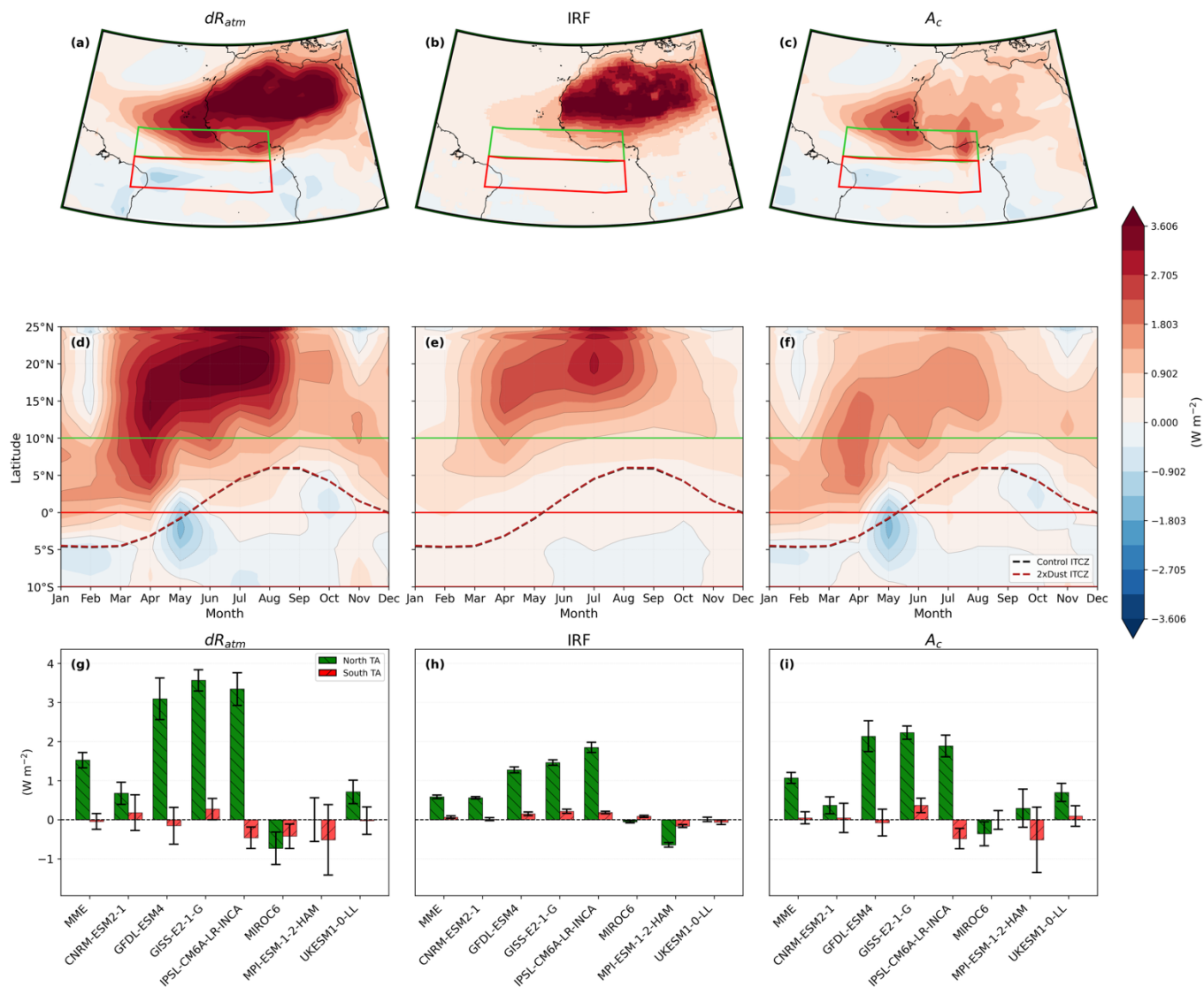


Figure 3. Dust-induced ITCZ shifts and precipitation redistribution over the tropical Atlantic. (a) Annual mean change in ITCZ latitude between piClim-2xDust and piClim-Control for the multi-model ensemble mean (MME) and individual models. Positive values indicate a northward ITCZ shift. Bars show the mean annual shift, and error bars indicate the standard error. (b) Relationship between ITCZ shift and the precipitation-response contrast, defined as $\Delta P_{\text{north}} - \Delta P_{\text{south}}$, where the North tropical Atlantic box spans 40°W–5°E, 0°–10°N, and the South tropical Atlantic box spans 40°W–5°E, 10°S–0°. Regression slopes (β) and Pearson correlation coefficients (r) are shown in each panel. Precipitation changes are expressed in W m^{-2} .

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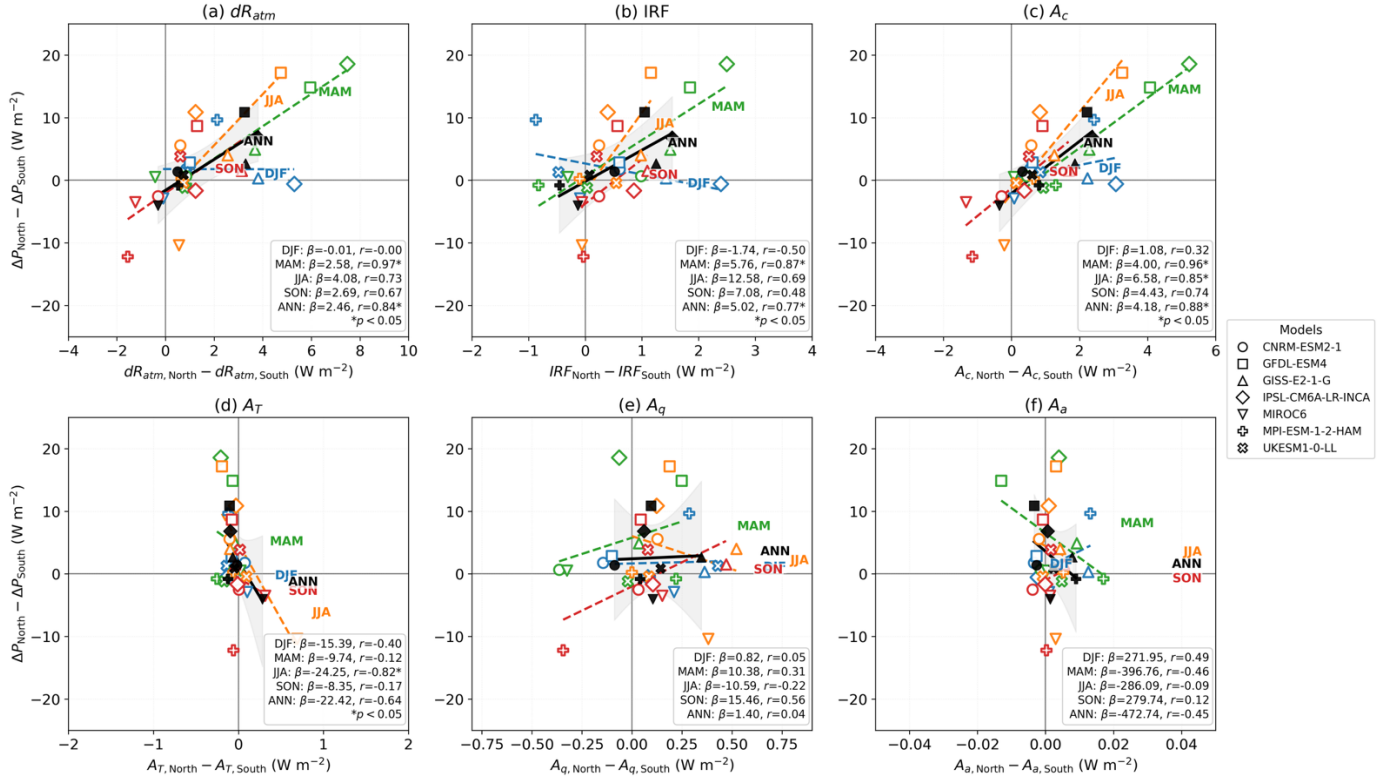
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896 Figure 4. Radiative kernel decomposition of the dust response. Multi-model mean change in atmospheric radiative
897 flux anomaly, (dR_{atm}), instantaneous radiative forcing, IRF, and cloud adjustment, (A_c). Panels (a–c) show the
898 annual mean response. Panels (d–f) show the corresponding latitude–month distribution averaged over $40^{\circ}W -$
899 $5^{\circ}E$, and Panels (g–i) show the regional mean responses for the multi-model ensemble and individual models over
900 the North TA and South TA. Bars show annual means and error bars indicate the standard error. Units are expressed
901 in $W m^{-2}$.

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908 *Figure 5.* Inter-model relationships between the regional precipitation response difference, ($\Delta P_{\text{north}} - \Delta P_{\text{south}}$) and
909 the corresponding North TA–South TA difference in (a) atmospheric radiative flux anomaly, (dR_{atm}), (b)
910 instantaneous radiative forcing, IRF, (c) cloud adjustment, (A_c), (d) temperature adjustment, (A_T), (e) water-vapor
911 adjustment, (A_q), and (f) surface-albedo adjustment, (A_a). Each marker represents one model for a given season;
912 marker shapes identify models and colors identify seasons. Dashed lines show seasonal least-squares regressions.
913 Regression slopes (β) and Pearson correlation coefficients (r) are shown in each panel. Positive relationships
914 indicate that models with larger radiative heating over the North TA relative to the South TA tend to produce a
915 larger North TA relative precipitation response. Units are expressed in Wm^{-2} .

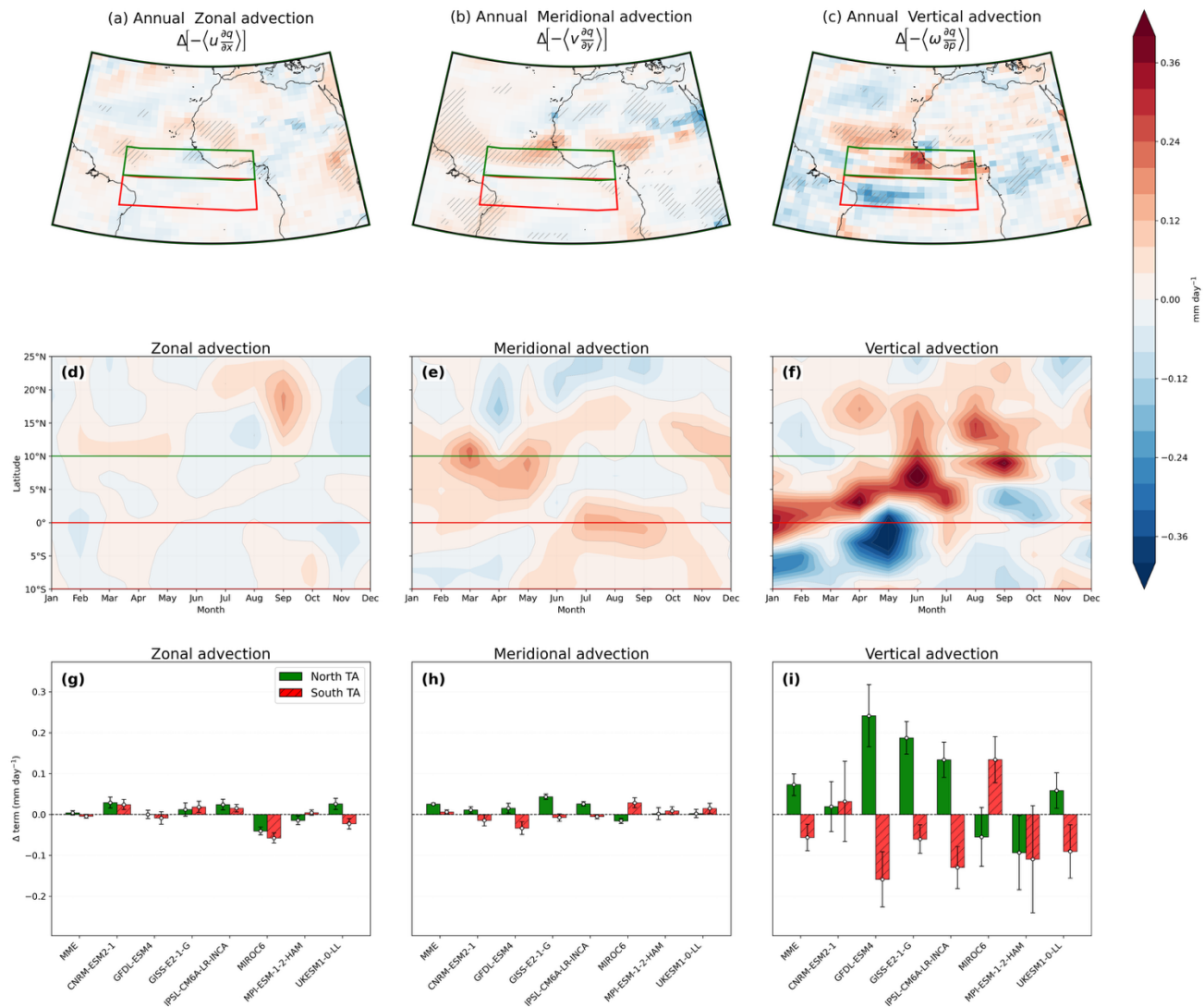
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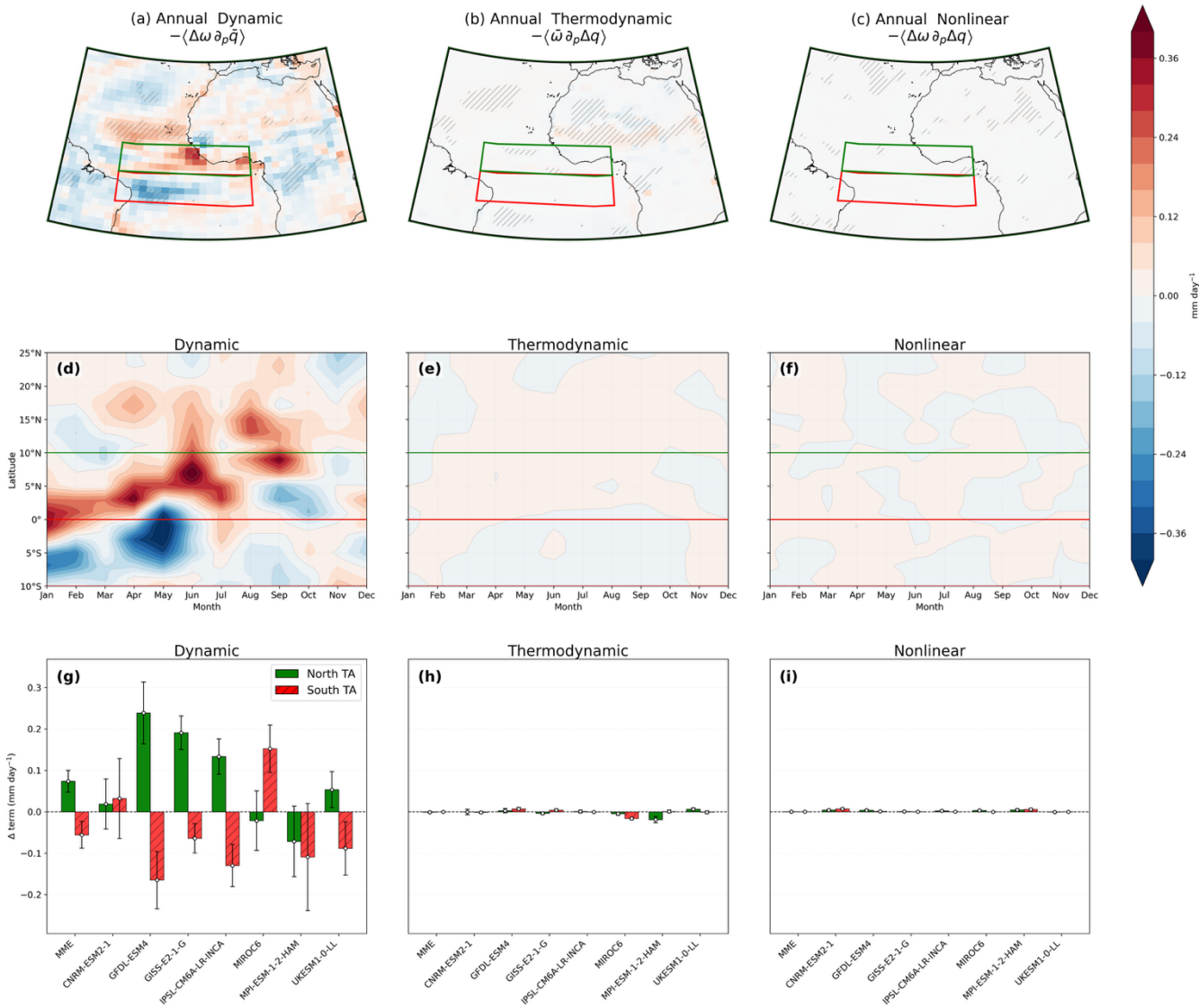
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924 Figure 6. MME Decomposition of the dust-induced moisture advection response over the tropical Atlantic and
925 West Africa. Annual mean changes in vertically integrated moisture advection are shown for the zonal, meridional,
926 and vertical components in panels (a–c), respectively. Stippling indicates grid cells where the response is
927 statistically significant at 5% level. Panels (d–f) show the seasonal zonal-mean response averaged over 40°W–5°E,
928 with the green and red horizontal lines marking the North Tropical Atlantic (North TA; 0°–10°N) and South
929 Tropical Atlantic (South TA; 10°S– 0°), respectively. Panels (g–i) show the area-mean response for the multi-
930 model ensemble and individual models over the North TA and South TA regions. Error bars represent the standard
931 error of the mean. Units are mm/day.

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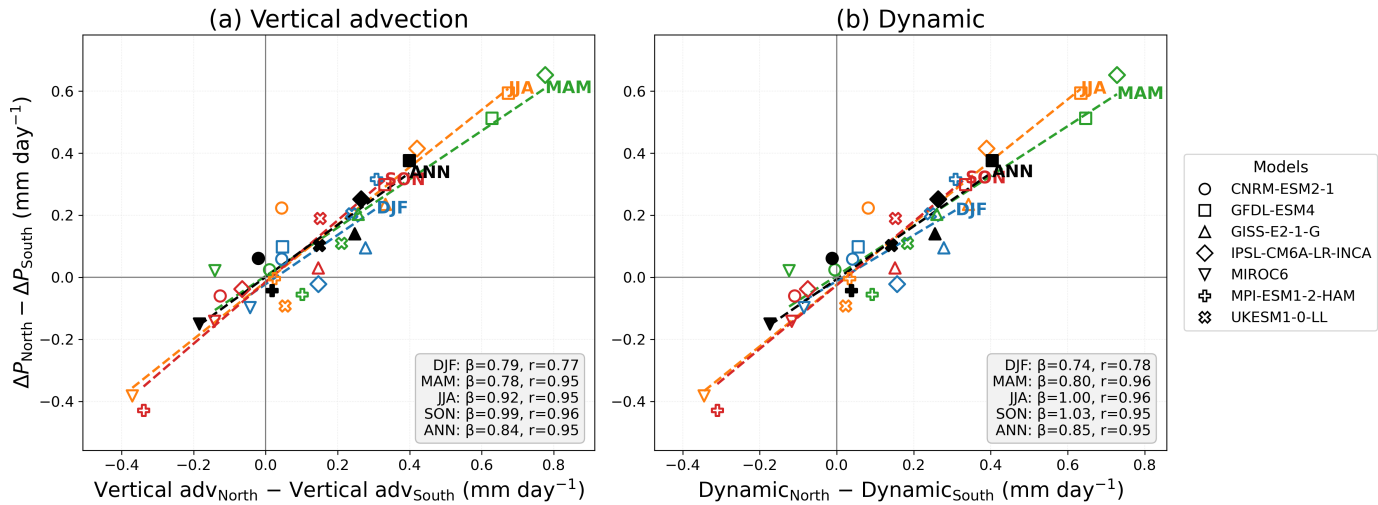
936 Figure 7. MME Decomposition of the Vertical moisture advection ($(\Delta \left[-\langle \omega \frac{\partial q}{\partial p} \rangle \right])$ component under dust
937 perturbation. Annual mean changes are shown for the dynamic, thermodynamic, and nonlinear components in
938 panels (a-c), respectively. Stippling indicates grid cells where the response is statistically significant ($p < 0.05$).
939 Panels (d-f) show the seasonal zonal-mean response of the vertical moisture advection decomposition averaged
940 over 40°W-5°E. Panels (g-i) show the area-mean response for MME and individual CMIP6 models over the North
941 TA (0°-10°N) and South TA (10°S-0°), regions. Error bars represent the standard error of the mean. Units are
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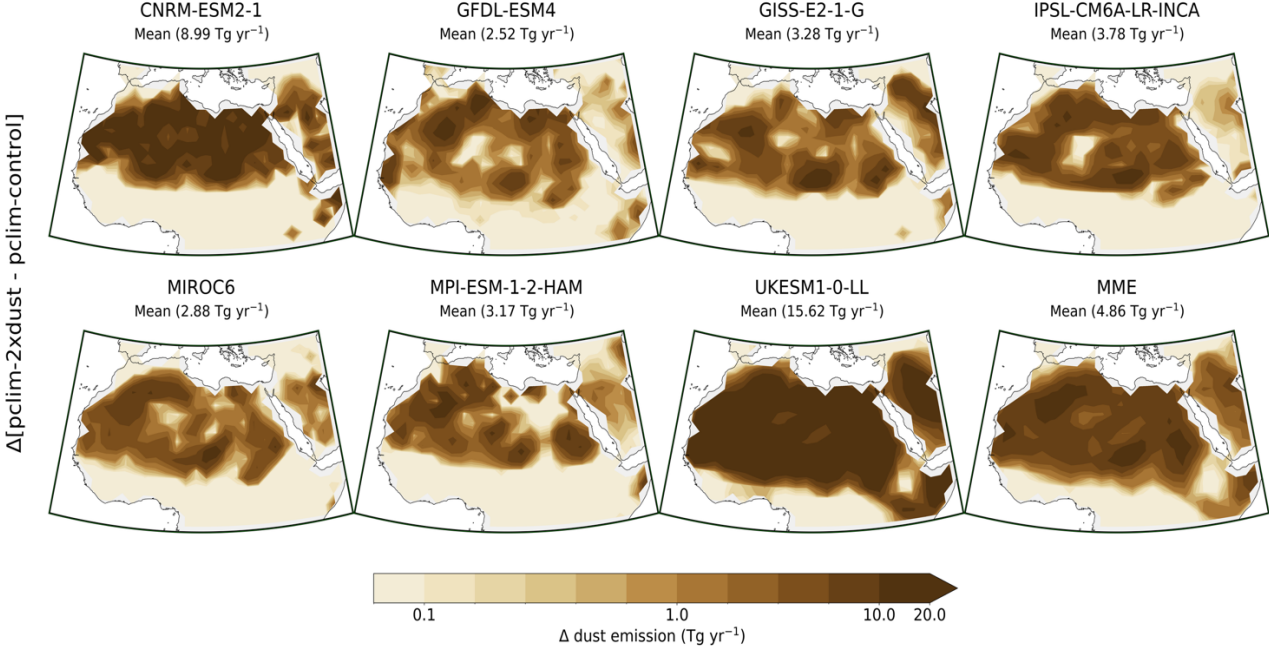
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949 Figure 8. Dust-induced precipitation response difference between the North Tropical Atlantic and South Tropical
950 Atlantic, and the corresponding regional difference in (a) vertical moisture advection and (b) the dynamic
951 component of vertical moisture advection. Each symbol shape represents individual CMIP6 models as a function
952 of annual and different seasons. Dashed lines show seasonal least-squares regressions. Regression slopes (β) and
953 Pearson correlation coefficients (r) are shown in each panel. Positive slopes indicate that models producing larger
954 precipitation increases over the North TA (0° – 10° N) relative to the South TA (10° S– 0°) also produce larger North
955 TA relative increases in vertical moisture advection.

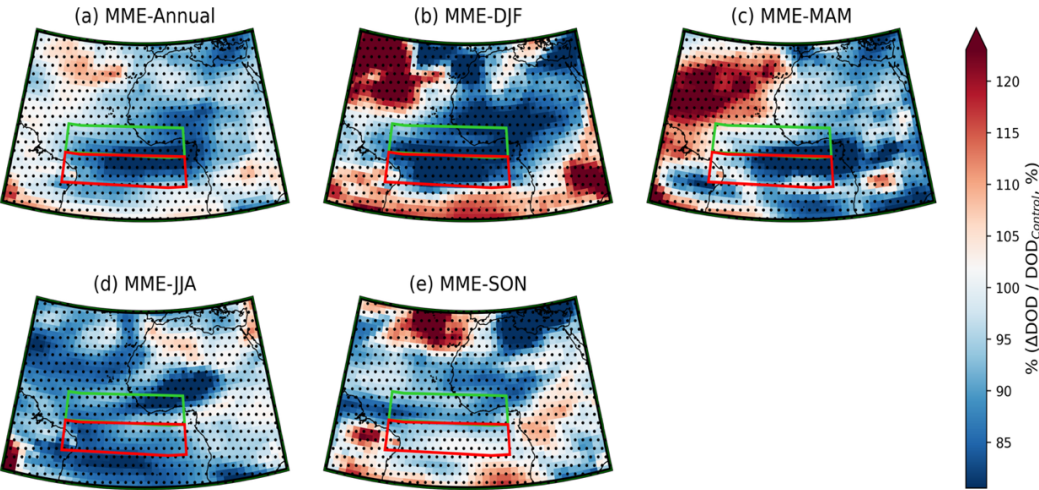
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968 **Supplementary Figures**



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970 *Figure S1: Dust emission (Tg/yr) response to the dust perturbation experiment, shown as the difference between piClim-2xdust*
971 *and piClim-control . Panels show the individual model responses with the multi-model ensemble mean (MME).*
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975 *Figure S2: MME response of DOD to doubled dust perturbation. Annual and seasonal changes in DOD are expressed*
976 *as percentage changes relative to the control simulation. Panels(a-e) show the annual mean and seasonal means for*
977 *DJF, MAM, JJA, and SON. Black stippling indicates grid cells where the difference between piClim-2xdust and piClim-*
978 *control simulations is statistically significant ($p < 0.05$).*

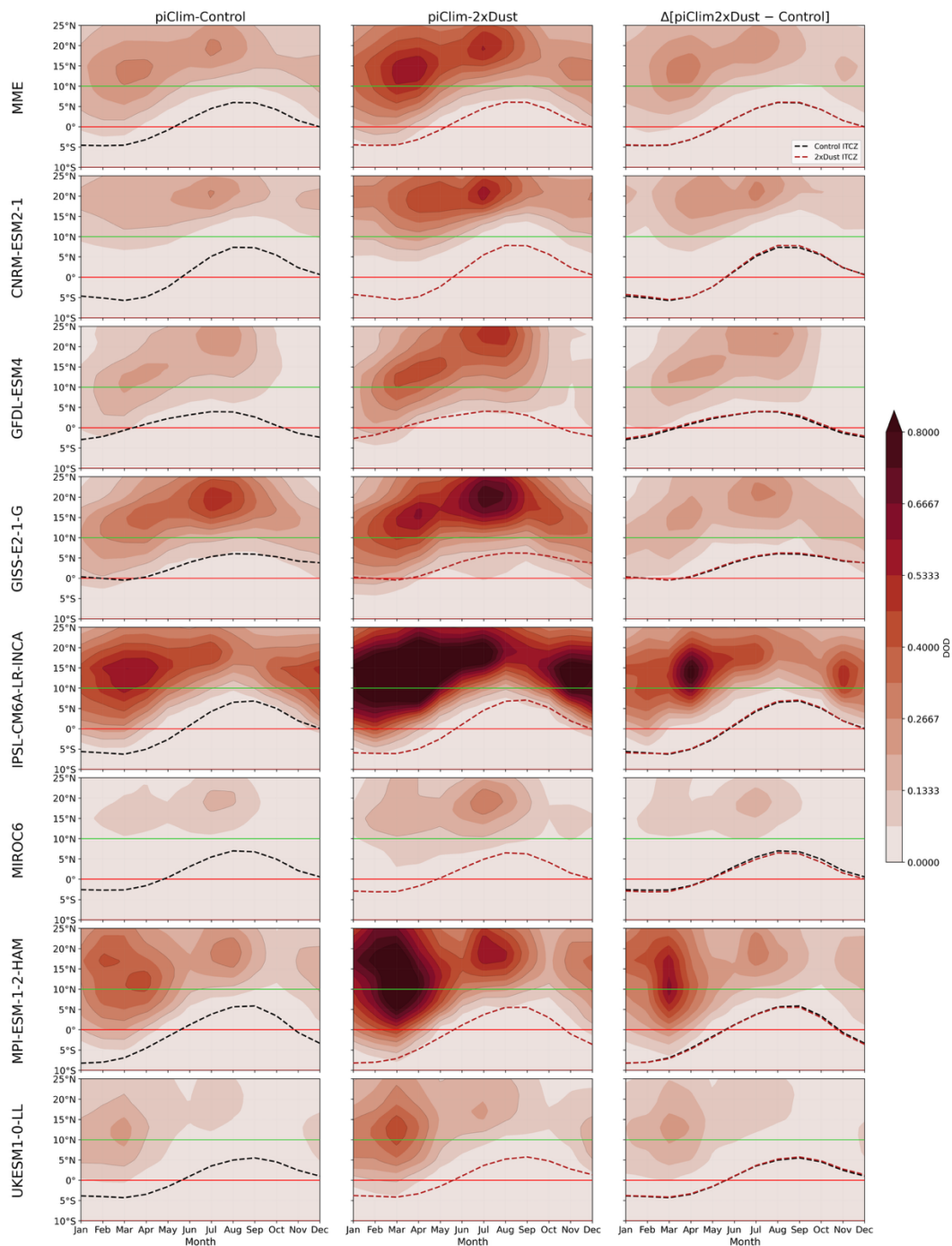


Figure S3: DOD and ITCZ seasonal migration under dust perturbation. Latitude-month DOD averaged over 40°W – 5°E for the multi-model mean and individual CMIP6 models. Columns show piClim-control, piClim-2xDust, and the positive dust response, $\Delta[\text{piClim-2xDust} - \text{control}]$. Dashed lines show the precipitation-centroid ITCZ in the control (black) and 2xDust (red) simulations.

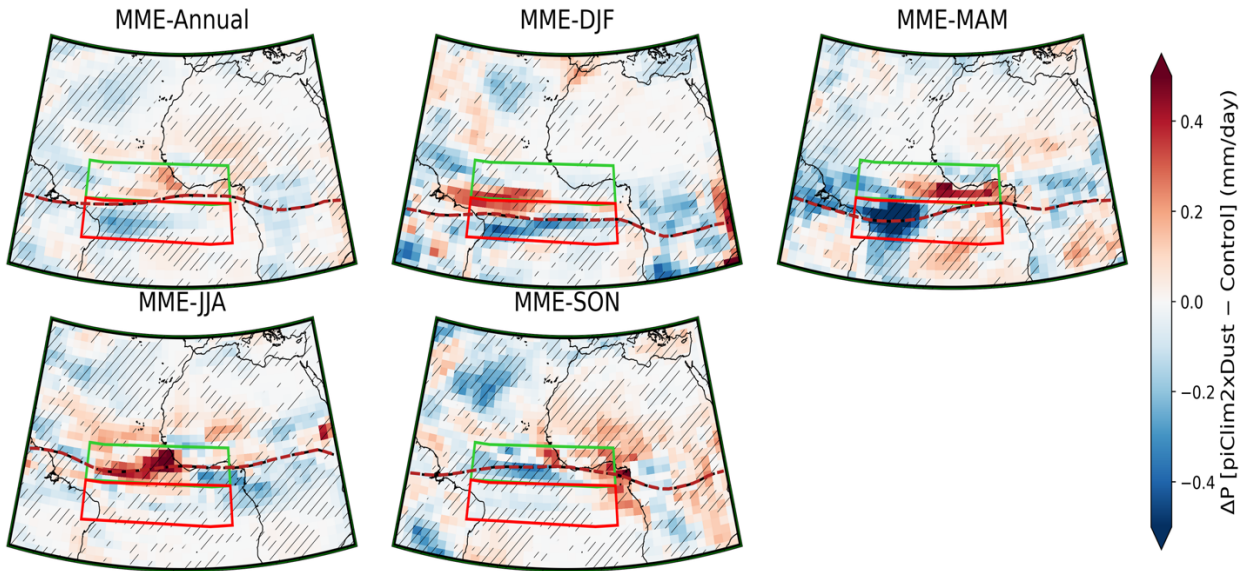


Figure S4: Dust-driven precipitation anomalies and ITCZ displacement. MME annual and seasonal precipitation response to dust perturbation in piClim-2xDust relative to piClim-control. Shading shows $\Delta P(\text{mm/day}) = \text{piClim-2xDust} - \text{control}$ with positive and negative anomalies. Dashed curves show the precipitation-centroid ITCZ in the control simulation (black) and the 2xDust simulation (red). Green and red boxes indicate the northern tropical Atlantic region, $0\text{--}10^\circ\text{N}$, and the southern tropical Atlantic region, $10^\circ\text{S}\text{--}0^\circ$, respectively. Hatching indicates regions where 70% of the models agree on the sign.

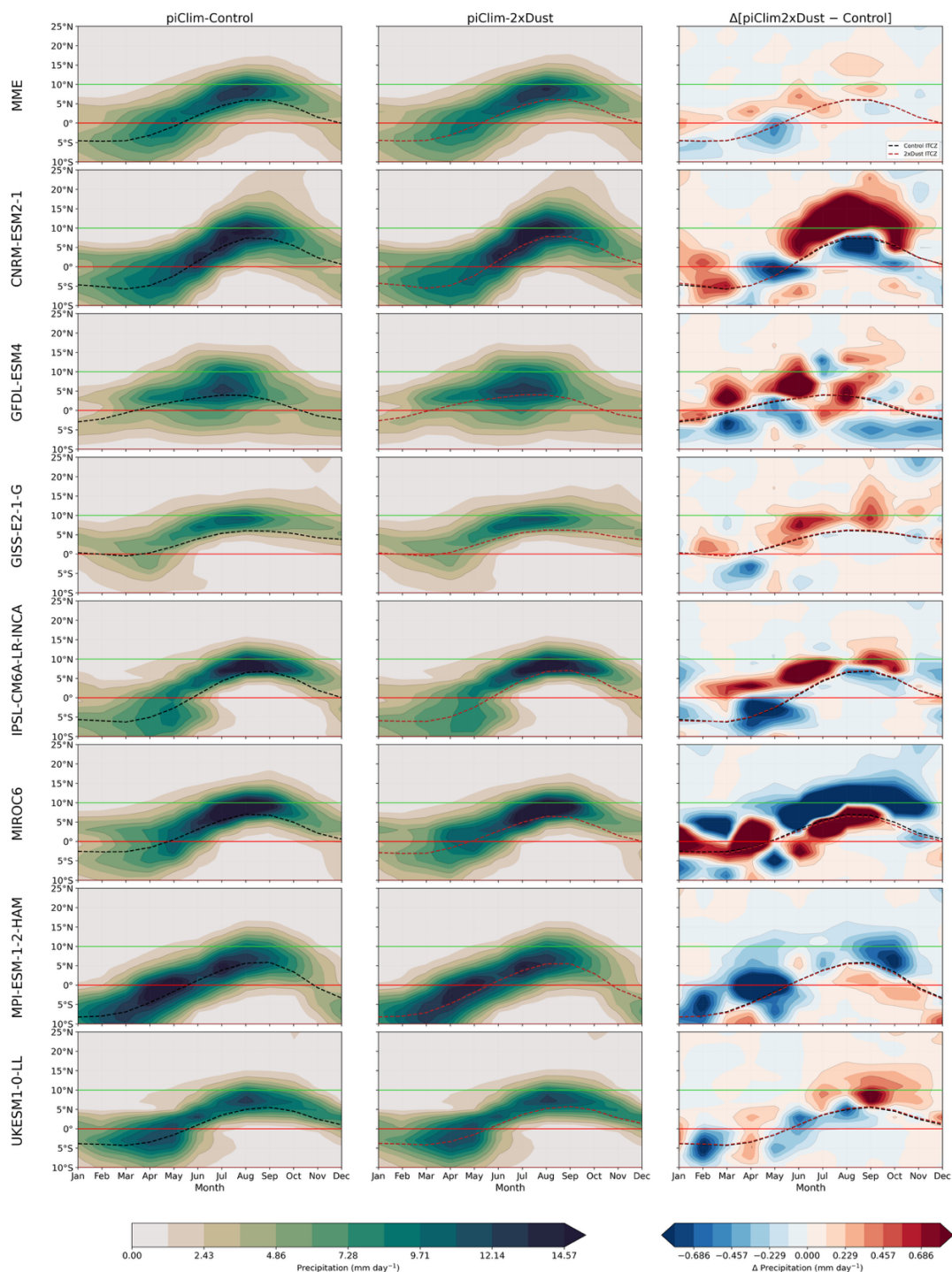


Figure S5: Latitude-month precipitation and ITCZ response to dust perturbation. Monthly precipitation averaged over 40°W–5°E for the multi-model mean and individual CMIP6 models. The columns show piClim-control, piClim-2xDust, and their difference. Dashed lines show the precipitation-centroid ITCZ in the control simulation (black) and 2xDust simulation (red).

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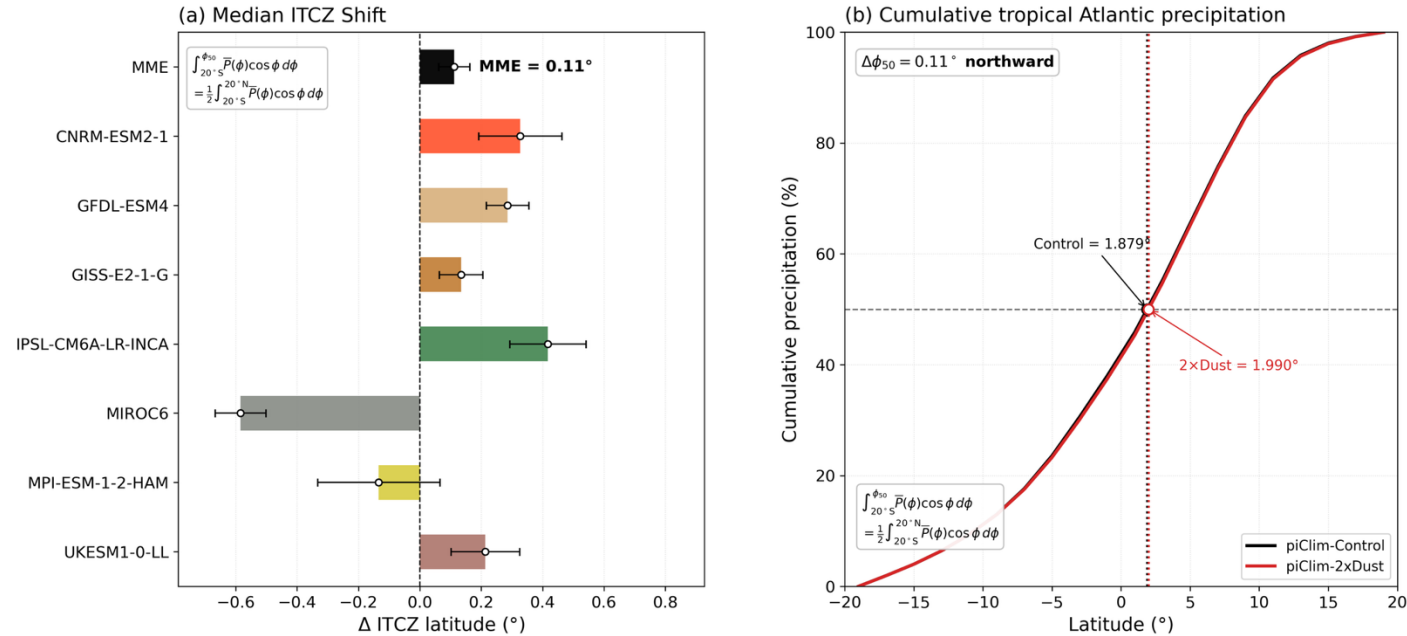


Figure S6: Median-based diagnostics of the tropical Atlantic ITCZ shift (a) Median-based annual ITCZ shift for the multi-model ensemble mean (MME) and individual models. Positive values indicate a northward ITCZ shift. Bars show the mean annual shift, and error bars indicate the standard error of the annual mean shift. (b) Cumulative area-weighted tropical Atlantic precipitation for MME in the piClim-Control (black) and piClim-2xDust (red). The latitude at which the cumulative precipitation reaches 50% defines the median precipitation latitude (ϕ_{50}), yielding an MME northward shift of 0.11° .

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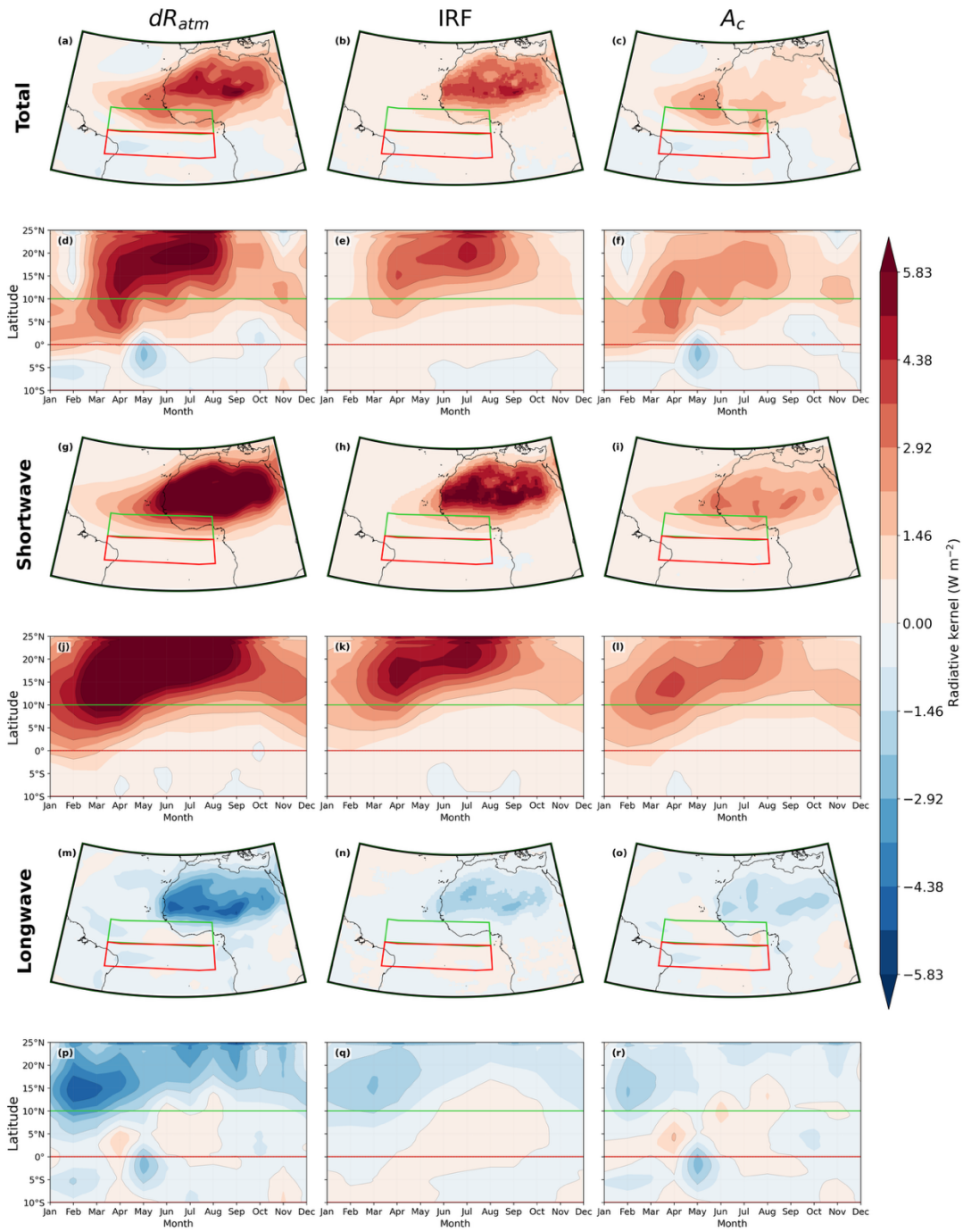


Figure S7: Radiative kernel decomposition of the dust response. Multi-model mean change in atmospheric radiative flux anomaly, (dR_{atm}), instantaneous radiative forcing, IRF, and cloud adjustment, (A_c). Panels (a–c), (g–i), and (m–o) show annual mean maps for the total, shortwave, and longwave components, respectively. Panels (d–f), (j–l), and (p–r) show the corresponding latitude–month distribution averaged over $40^\circ W - 5^\circ E$.

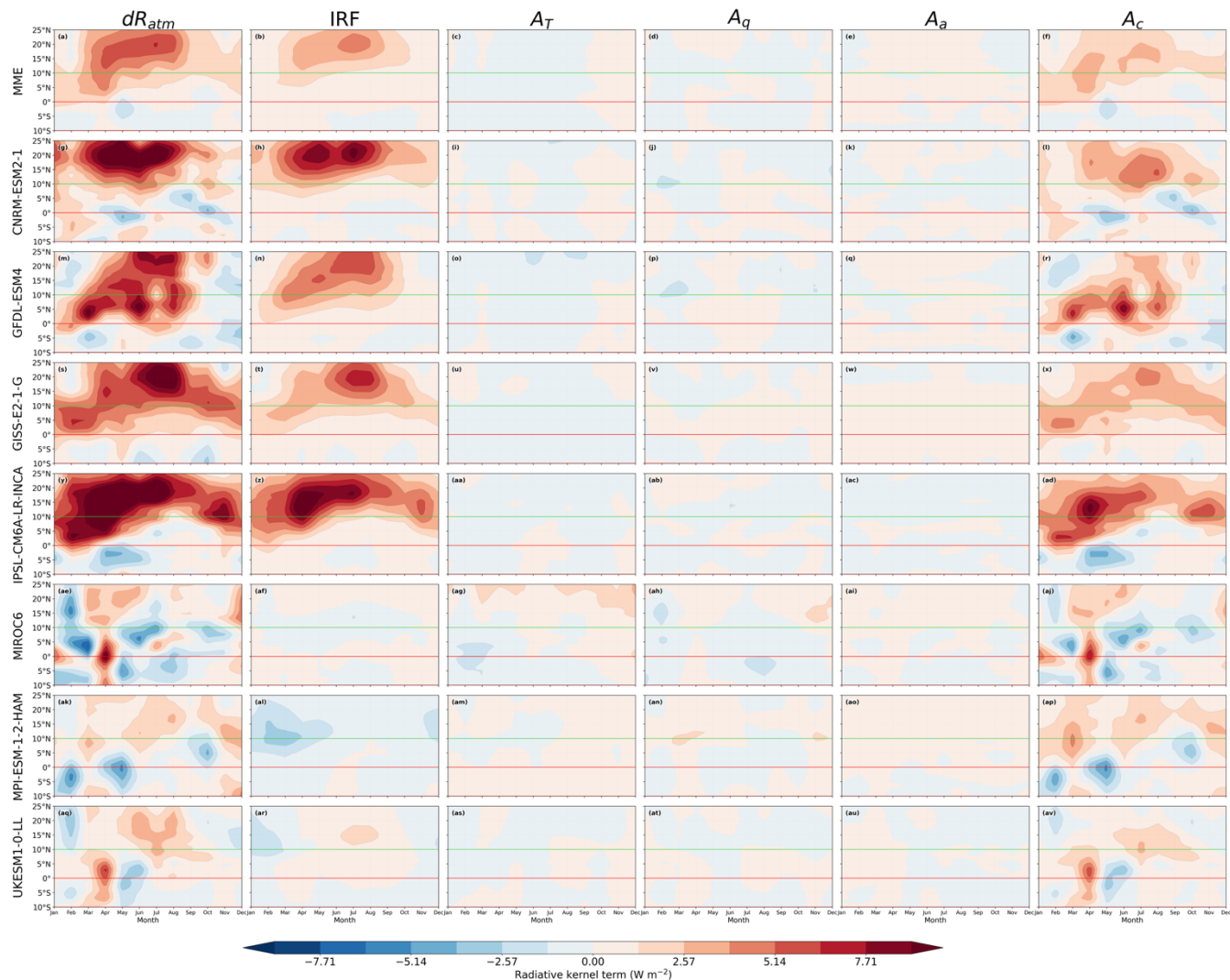
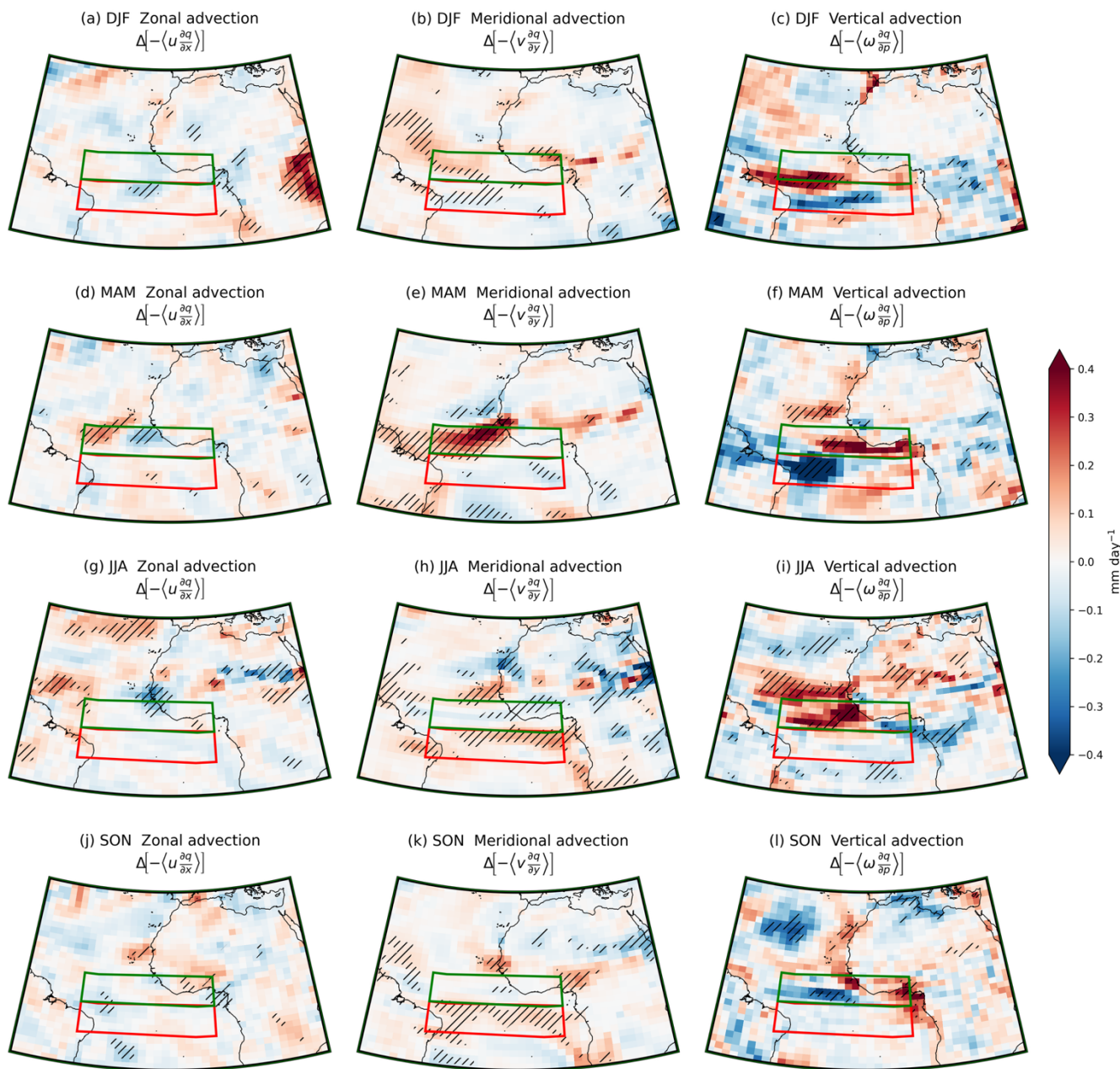


Figure S8: Latitude–month decomposition of radiative adjustments under dust perturbation. Multi-model ensemble mean (MME), and individual CMIP6 models’ radiative responses averaged over 40°W – 5°E. Columns show the atmospheric radiative flux anomaly, (dR_{atm}); instantaneous radiative forcing, IRF ; atmospheric temperature adjustment, (A_T); water-vapor adjustment, (A_q); surface albedo adjustment, (A_a); and cloud adjustment, (A_c).

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1043 *Figure S9: Seasonal moisture-advection response to dust perturbation. Multi-model mean changes in vertically integrated*
 1044 *moisture-advection terms between piClim-2xDust and piClim-control. Panels (a, d, g, j) show zonal advection, panels (b, e, h,*
 1045 *k) show meridional advection, and panels (c, f, i, l) show vertical advection for DJF, MAM, JJA, and SON, respectively. The*
 1046 *terms are integrated over 1000-100 hPa and expressed in mm/day. Hatching indicates regions significant at ($P < 0.05$).*

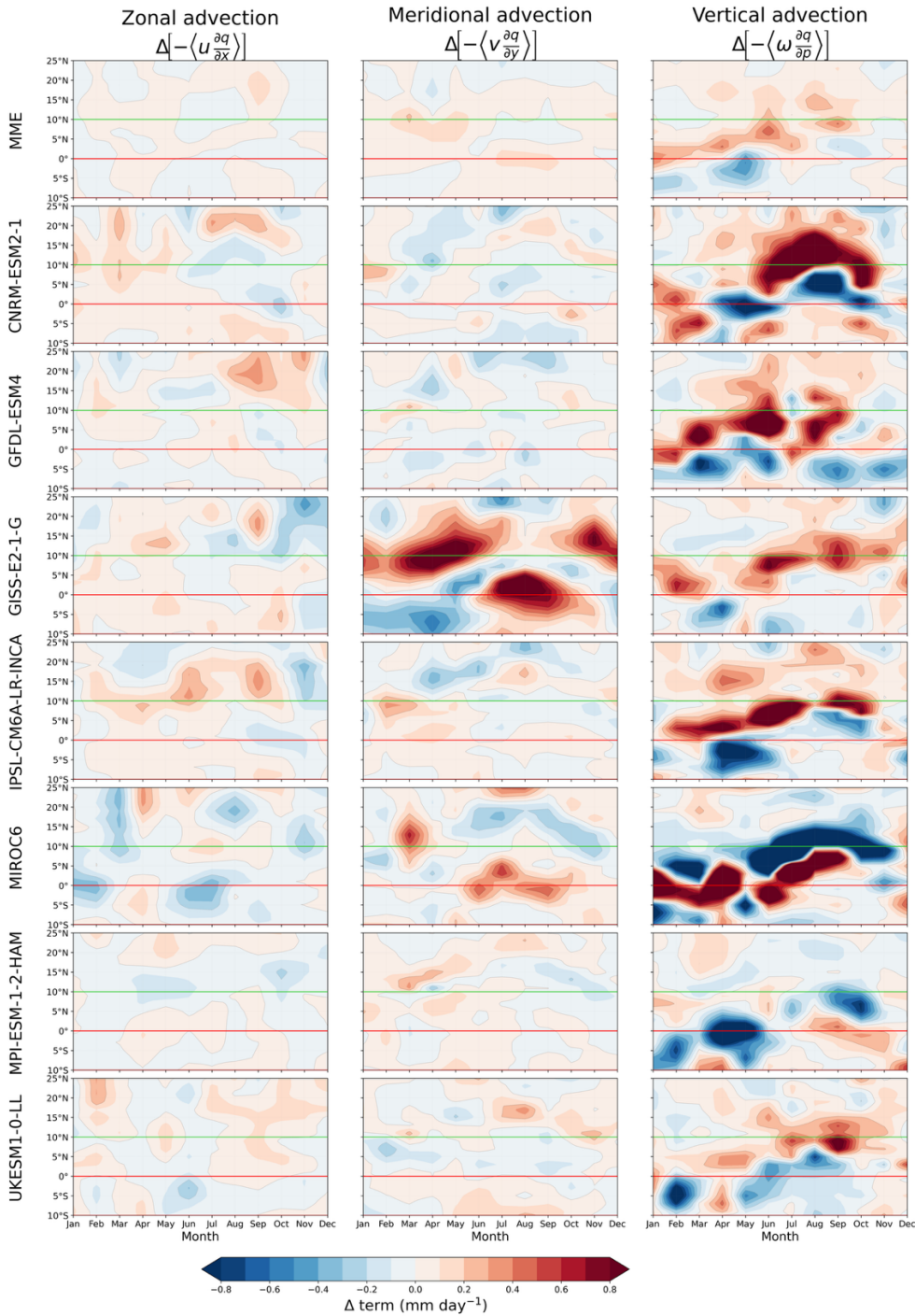


Figure S10: Latitude–month decomposition of moisture budget under dust perturbation averaged over 40°W – 5°E. Columns show the zonal advection term, $(\Delta[-\langle u \frac{\partial q}{\partial x} \rangle])$; meridional advection term $(\Delta[-\langle v \frac{\partial q}{\partial y} \rangle])$; and vertical advection term, $(\Delta[-\langle \omega \frac{\partial q}{\partial p} \rangle])$, for Multi model ensemble mean (MME) and individual CMIP6 models.

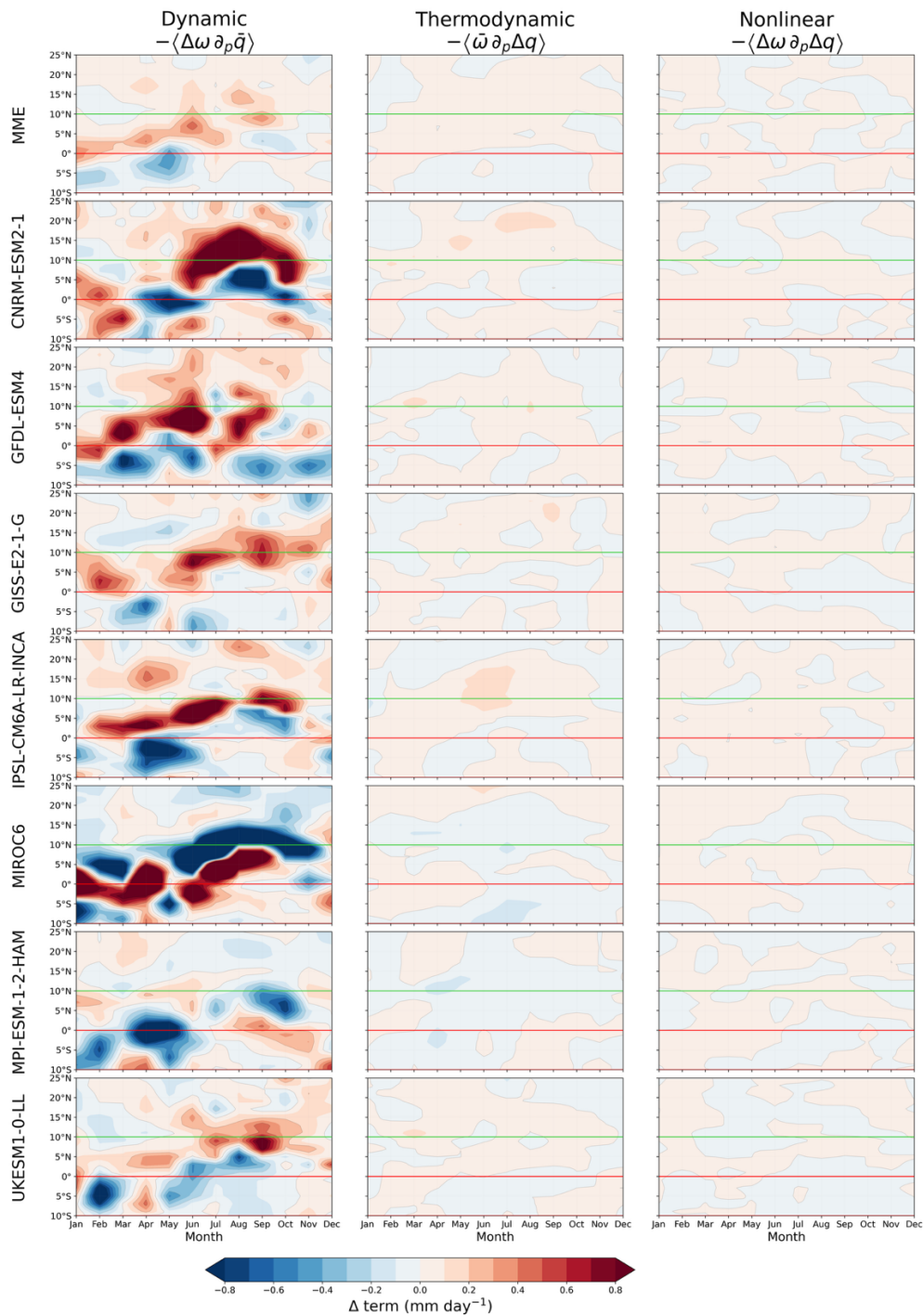


Figure S11: Latitude–month decomposition of vertical advection term under dust perturbation averaged over 40°W – 5°E. Columns show the dynamic component, $(-\langle \Delta \omega \partial_p \bar{q} \rangle)$; thermodynamic component $(-\langle \bar{\omega} \partial_p \Delta q \rangle)$; and nonlinear component, $(-\langle \Delta \omega \partial_p \Delta q \rangle)$ for Multi model ensemble mean (MME) and individual CMIP6 models.

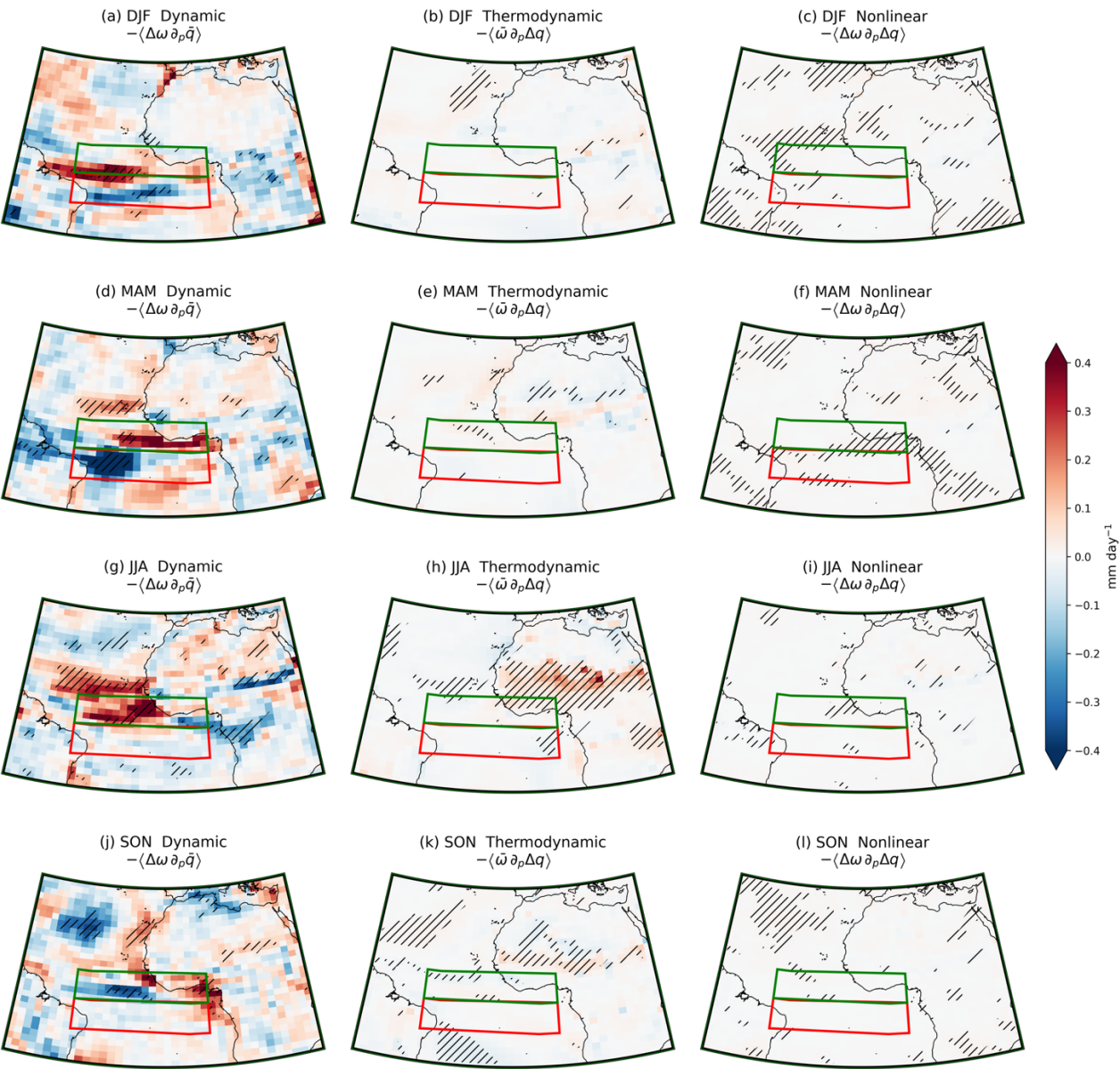


Figure S12: Multi-model ensemble mean (MME) changes to dust perturbation in vertical moisture advection term ($(\Delta [-\langle \omega \partial q / \partial p \rangle])$ between piClim-2xDust and piClim-control. Panels (a, d, g, j) show dynamic term, panels (b, e, h, k) show thermodynamic term, and panels (c, f, i, l) show nonlinear term for DJF, MAM, JJA, and SON, respectively. The terms are integrated over 1000 to 100 hPa and expressed in mm/day. Hatching indicates regions significant at ($P < 0.05$).

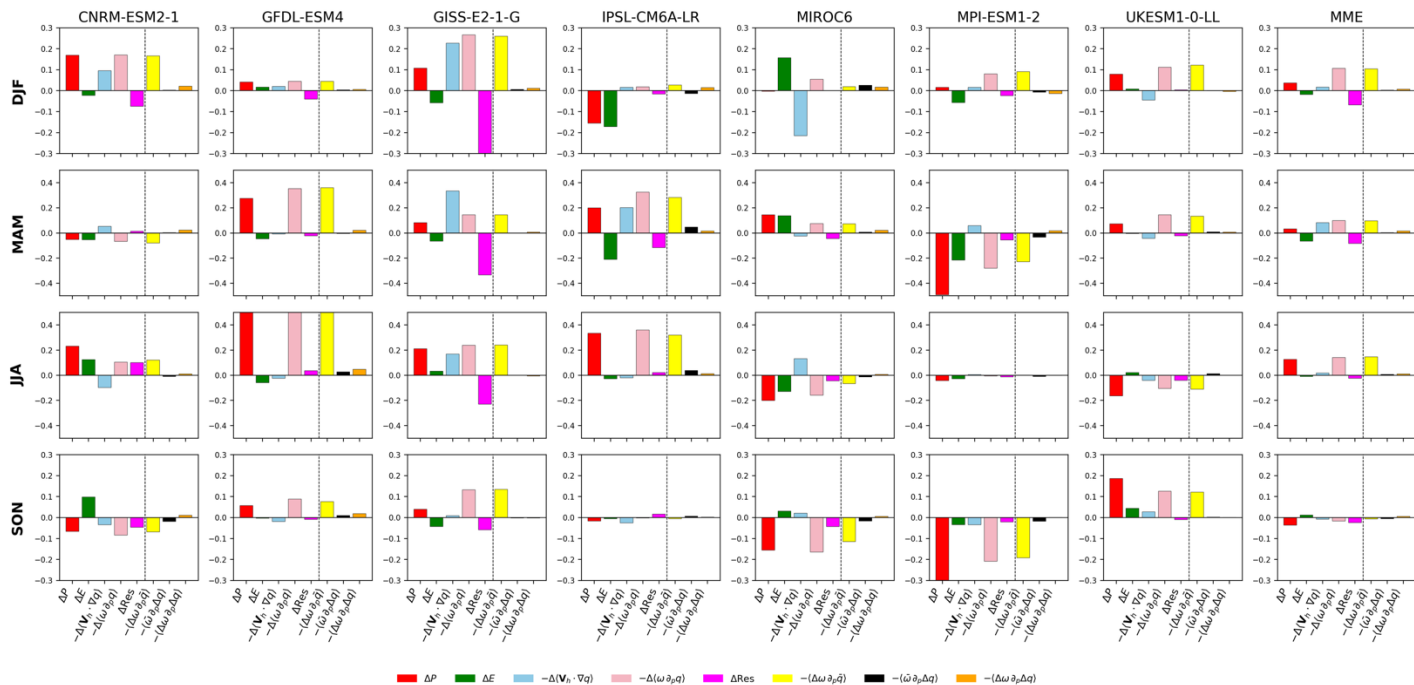


Figure S13: Seasonal decomposition of the moisture budget response in [piclim-2xdust-control] over the northern tropical Atlantic region, 0–10°N, for each model and the multi-model ensemble mean (MME). Columns show individual models and the MME, while rows show DJF, MAM, JJA, and SON. Bars represent anomalous precipitation, evaporation, horizontal moisture advection, total vertical moisture advection, the dynamic and thermodynamic contributions to vertical moisture advection, the nonlinear term, and the residual. The dashed vertical lines separate the full vertical moisture advection term from its component contributions. Units are mm/day.

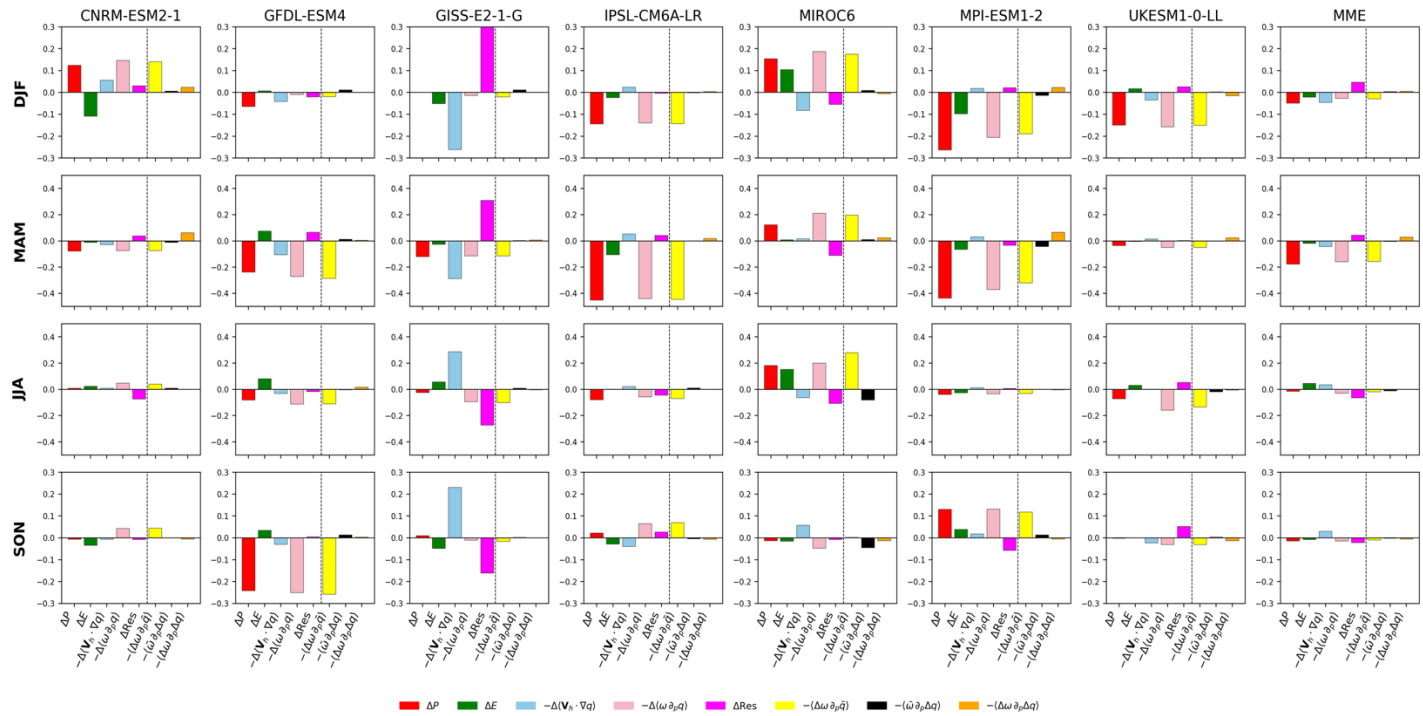


Figure S14: Same as Fig. S13 but over the southern tropical Atlantic region, 10°S–0°.

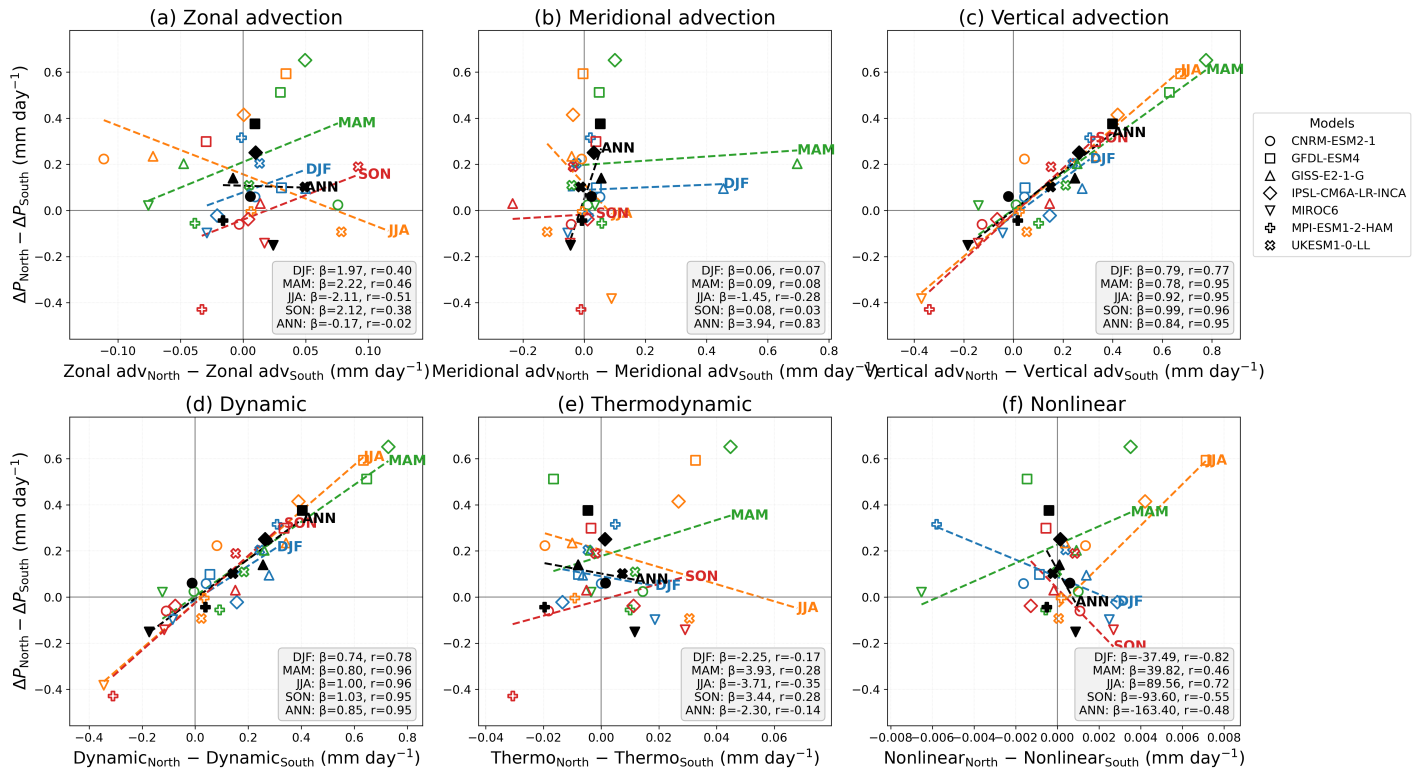


Figure S15: Relationship between the dust-induced north-south precipitation contrast, and the corresponding north-south contrast in moisture-budget terms. Panels show (a) zonal advection, (b) meridional advection, (c) vertical advection, and the vertical-advection decomposition into (d) dynamic, (e) thermodynamic and (f) nonlinear components. Each marker represents an individual model, with colours indicating seasons and black markers indicating the annual mean. Dashed lines show seasonal linear fits; β and r denote the regression slope and correlation coefficient, respectively. All terms are expressed in mm/day.