



State dependence of aerosol forcing and its implications for climate response

Haozhe He¹, Brian J. Soden², Gabriel A. Vecchi^{1,3}, Ryan J. Kramer⁴, Jianhao Zhang^{5,6}, Chanyoung Park^{2,7}

5 ¹High Meadows Environmental Institute, Princeton University, Princeton, NJ, USA

²Rosenstiel School of Marine, Atmospheric and Earth Science, University of Miami, Miami, FL, USA

³Department of Geosciences, Princeton University, Princeton, NJ, USA

⁴NOAA Geophysical Fluid Dynamics Laboratory, Princeton, NJ, USA

⁵Cooperative Institute for Research in Environmental Sciences (CIRES), University of Colorado Boulder, Boulder, CO, USA

10 ⁶NOAA Chemical Sciences Laboratory, Boulder, CO, USA

⁷Department of Environmental Sciences, University of Virginia, Charlottesville, VA, USA

Correspondence to: Haozhe He (haozhe.he@princeton.edu)

Abstract. Aerosols offset greenhouse-gas warming, but their radiative forcing is often treated as a property of the aerosol perturbation, without accounting for the climate state in which it occurs. Here we show that this assumption is incomplete
15 using atmosphere-only simulations with two GFDL atmospheric model configurations, GFDL-AM4.0 and GFDL-AM4-MG2, across climates from 8 K colder to 8 K warmer than present day. In both models, the same present-day-minus-preindustrial aerosol-emission perturbation produces weaker cooling as climate warms. This weakening arises from changes in both aerosol–radiation and aerosol–cloud interactions. Direct aerosol forcing weakens as aerosol optical depth decreases, reflecting changes in aerosol burden and the optical effectiveness of aerosol mass. Aerosol–cloud forcing weakens through
20 model-dependent mechanisms: shortwave liquid-cloud adjustments dominate in GFDL-AM4.0, consistent with reduced cloud susceptibility in warmer climates, whereas GFDL-AM4-MG2 shows an additional longwave pathway involving aerosol–cloud–ice interactions and high-cloud responses. Aerosol forcing also weakens as climate feedbacks become less negative, revealing consistent within-model co-variation of forcing and feedback. These results highlight the importance of accounting for background climate state when interpreting the role of aerosol forcing in historical attribution, climate model
25 tuning, and future climate projections.

1 Introduction

Anthropogenic aerosols have masked a substantial fraction of greenhouse-gas warming over the industrial era, but the magnitude of this masking remains one of the largest uncertainties in estimates of historical radiative forcing and climate-change attribution (Bellouin et al., 2020; Boucher et al., 2013; Forster et al., 2021; Myhre et al., 2013; Stevens, 2015; Szopa
30 et al., 2021). This uncertainty also contributes substantially to the spread in future warming projections (Watson-Parris and Smith, 2022). Aerosol effective radiative forcing (ERF) is commonly diagnosed operationally as the top-of-atmosphere



radiative response to a specified change in aerosol emissions or concentrations, including rapid atmospheric adjustments, in atmosphere-only simulations with fixed climatological sea surface temperatures (SSTs) and sea-ice distributions (Forster et al., 2016; Hansen et al., 2005). Aerosols perturb Earth's radiation budget through aerosol–radiation interactions, including direct scattering and absorption of radiation, and through aerosol–cloud interactions, in which aerosols acting as cloud condensation nuclei modify cloud droplet number, droplet size, liquid water path, precipitation, cloud amount, and other rapid adjustments (Albrecht, 1989; Bellouin et al., 2020; Bretherton et al., 2007; Pincus and Baker, 1994; Twomey, 1974; Xue and Feingold, 2006). We refer to these diagnostic components as instantaneous radiative forcing from aerosol–radiation interactions (IRF_{ari}) and effective radiative forcing from aerosol–cloud interactions (ERF_{aci}), respectively. Aerosol–radiation interactions can also induce rapid atmospheric adjustments that contribute to ERF_{ari} , although these adjustments are expected to be relatively small and are not a focus of this study. Both IRF_{ari} and ERF_{aci} are difficult to constrain because anthropogenic aerosols are short lived, spatially heterogeneous, and coupled to clouds and precipitation through parameterized microphysical processes in global climate models (Boucher et al., 2013; Ghan et al., 2016; Gryspeerd et al., 2019; Bellouin et al., 2020). In this conventional framing, aerosol forcing is treated primarily as a property of the aerosol perturbation itself. Yet the same aerosol perturbation can produce different radiative effects depending on the climate state into which it is introduced.

This distinction matters because the atmospheric state in which aerosol forcing is diagnosed can itself regulate the radiative response to a given aerosol perturbation. Previous studies have shown that aerosol–cloud interactions depend strongly on background aerosol loading, because the radiative response to additional aerosol particles depends on the preexisting aerosol and cloud droplet population (Boucher et al., 2013; Grandey and Wang, 2019; Twomey, 1977; Wilcox et al., 2015). More broadly, aerosol forcing can depend on meteorological and climate-state variables that affect both aerosol–radiation and aerosol–cloud pathways (Bellouin et al., 2020; Boucher et al., 2013; Forster et al., 2021). For aerosol–radiation interactions, relative humidity regulates aerosol water uptake and particle size, thereby affecting aerosol extinction and the optical effectiveness of aerosol mass (Nemesure et al., 1995; Pilinis et al., 1995; Tang and Munkelwitz, 1994), while surface albedo modifies the radiative contrast against which aerosol scattering and absorption occur (Haywood and Shine, 1995; Haywood et al., 1997). Precipitation and cloud processing influence aerosol lifetime, removal, and vertical distribution (Rasch et al., 2000; Textor et al., 2006). For aerosol–cloud interactions, cloud regime and the associated optical and microphysical state determine cloud susceptibility, shaping how strongly clouds respond radiatively to a given aerosol perturbation (Albrecht, 1989; Platnick and Twomey, 1994; Sorooshian et al., 2009; Storelvmo, 2017; Twomey, 1977; Zhang et al., 2022). Aerosol forcing is therefore not determined solely by the imposed emission perturbation, but also by the climate state in which that perturbation occurs.

The climate-state dependence of ERF_{aci} is especially important because cloud susceptibility varies across cloud regimes and with the background climate. In warm liquid-cloud regimes, aerosol-induced increases in cloud droplet number may produce progressively smaller albedo changes as clouds become optically thicker or less susceptible to additional aerosol



65 perturbations. In colder and mixed-phase regimes, warming can alter the availability of liquid and supercooled liquid water,
potentially modifying aerosol effects on cloud optical thickness. Ice-phase pathways add further complexity: aerosols that
serve as ice-nucleating particles can alter heterogeneous freezing and glaciation, with associated changes in ice-crystal
properties and sedimentation, as well as consequences for high-cloud amount, cloud altitude, and cloud emission temperature
(Boucher et al., 2013; Storelvmo, 2017; Forster et al., 2021). These mechanisms suggest that ERF_{aci} may respond differently
70 across cloud regimes and across models with different cloud microphysics.

Despite this physical basis, the climate-state dependence of aerosol forcing remains insufficiently constrained. Recent studies
have begun to test whether aerosol forcing varies with background state. Using idealized cloud-resolving simulations in
radiative–convective equilibrium, Lorian and Dagan (2024) found that the aerosol-induced cloud radiative response,
diagnosed using changes in cloud radiative effect, weakens with increasing SST, particularly in the shortwave. White et al.
75 (2025), however, found limited dependence of a cloud-radiative-effect-based estimate of the aerosol indirect effect on
background climate state in CESM2, while Zhang et al. (2026) showed that emissions-based anthropogenic aerosol ERF in
GFDL-ESM4.1 depends on background meteorological and chemical conditions. However, these studies either focus on
aerosol–cloud responses in idealized configurations or examine relatively narrow base-state contrasts in global climate
models, including preindustrial-to-present-day differences and moderate warming perturbations. Because aerosol forcing is
80 relatively small and atmospheric variability can be substantial, modest base-state contrasts can make state-dependent
changes difficult to distinguish from sampling variability. A broader assessment therefore calls for simulations that span a
wider range of climate states and diagnostics that more explicitly separate IRF_{ari} , ERF_{aci} , and non-cloud atmospheric
adjustments.

The climate-state dependence of aerosol forcing may also be linked to the way climate feedbacks are inferred and
85 interpreted. Cloud feedback is a dominant source of uncertainty in equilibrium climate sensitivity across climate models
(Forster et al., 2021; Meehl et al., 2020; Sherwood et al., 2020; Zelinka et al., 2020). Across CMIP6 models, Wang et al.
(2021) identified a negative intermodel correlation between ERF_{aci} and cloud feedback, indicating that models with stronger
negative ERF_{aci} tend to exhibit more positive cloud feedback. Subsequent perturbed-parameter ensemble studies have
reproduced a similar anti-correlation, suggesting that this relationship can arise from shared cloud processes that influence
90 both aerosol–cloud interactions and cloud feedback (Gettelman et al., 2024; Feng et al., 2026). For example, Feng et al.
(2026) linked stronger negative ERF_{aci} and more positive cloud feedback to higher autoconversion rates, which enhance
precipitation efficiency and suppress the increase in cloud liquid water path and cloud optical depth with warming. This
forcing–feedback compensation has also been discussed in the context of historical temperature constraints: models with
more positive cloud feedback and higher climate sensitivity may require stronger negative aerosol forcing to reproduce the
95 observed historical temperature evolution. Thus, aerosol forcing, climate feedback, and historical temperature constraints
may not be independent across model ensembles. Here, we ask a complementary question: whether aerosol forcing and
climate feedback vary coherently across climate states within the same model framework.



Here we investigate the state dependence of aerosol forcing using fixed-SST simulations with two GFDL atmospheric model configurations, GFDL-AM4.0 and GFDL-AM4-MG2, across five prescribed base-state climates spanning 8 K colder to 8 K warmer than present day. The two configurations differ in their representation of cloud microphysics and aerosol–cloud interactions, with GFDL-AM4-MG2 including more explicit aerosol–cloud–ice pathways. This contrast provides a framework for comparing liquid-cloud-dominated aerosol responses with responses in which high-cloud and ice-phase adjustments play a relatively larger role. Present-day aerosol perturbations relative to preindustrial aerosol conditions are imposed in each base-state climate to diagnose total aerosol ERF, IRF_{ari} , and ERF_{aci} .

We show that aerosol forcing weakens systematically as the base-state climate warms in both model configurations, but through different mechanisms. In GFDL-AM4.0, the weakening is driven primarily by shortwave ERF_{aci} , alongside a substantial contribution from direct aerosol radiative effects, whereas in GFDL-AM4-MG2 the longwave component plays a larger role, with its state dependence associated with aerosol–cloud–ice interactions and state-dependent high-cloud responses. Direct aerosol forcing also weakens as aerosol optical depth decreases, reflecting changes in aerosol burden and the optical effectiveness of aerosol mass through aerosol removal and humidity-dependent water uptake. Finally, aerosol forcing weakens as climate feedbacks become less negative across base-state transitions, indicating that forcing and feedback can vary coherently with background climate state. These results suggest that aerosol forcing is not independent of the evolving climate system, with implications for historical climate attribution, fixed-aerosol warming experiments, model tuning, and future projections.

2. Materials and Methods

2.1 Model configurations

We use two GFDL atmospheric model configurations: GFDL-AM4.0 (Zhao et al., 2018a,b) and GFDL-AM4-MG2 (Guo et al., 2021). GFDL-AM4.0 is the standard atmospheric model configuration. GFDL-AM4-MG2 implements the two-moment Morrison–Gettelman bulk cloud microphysics scheme with prognostic precipitation (Gettelman and Morrison, 2015), together with a mineral-dust- and temperature-dependent ice-nucleation scheme (Fan et al., 2019). This configuration provides a more detailed representation of cloud microphysics, cloud phase, and aerosol–cloud–ice interactions.

This distinction is central to our interpretation of the aerosol-forcing response. In GFDL-AM4.0, aerosol–cloud interactions are represented primarily through their effects on liquid-cloud properties, including cloud droplet number, cloud optical depth, and precipitation formation. In GFDL-AM4-MG2, aerosols can additionally influence ice nucleation and high-cloud properties through the mineral-dust-dependent ice-nucleation pathway. This contrast allows us to compare aerosol-forcing responses in a configuration emphasizing liquid-cloud aerosol–cloud interactions with responses in a configuration that includes more explicit aerosol–cloud–ice pathways.



2.2 Experimental design

We conduct 30-year fixed-SST atmosphere-only simulations under five prescribed base-state climates: -8 K, -4 K, 0 K, $+4$ K, and $+8$ K. These experiments are referred to as m8K, m4K, 0K, p4K, and p8K, respectively, and were originally developed to investigate the state dependence of cloud feedback (He et al., 2024). The present-day reference simulation (0K) uses prescribed sea surface temperatures (SST) and sea-ice concentrations from the monthly climatology over 1986–2005, with well-mixed greenhouse gases and other non-aerosol forcing agents fixed at corresponding present-day values representative of 1990s conditions. The atmosphere is free-running, without wind nudging. The colder and warmer base states are generated by applying spatially uniform SST perturbations of -8 K, -4 K, $+4$ K, and $+8$ K relative to this 0K climatology, with the SST perturbation serving as the only imposed change used to define the base-state climates.

For each base-state climate, aerosol forcing is diagnosed from the radiative response to present-day anthropogenic aerosol emissions, representative of 1990s conditions, relative to preindustrial aerosol conditions. This present-day-minus-preindustrial aerosol-emission perturbation is applied separately within each prescribed base-state climate. Thus, the imposed aerosol-emission perturbation is defined consistently across m8K, m4K, 0K, p4K, and p8K, while the background atmospheric state responds to the prescribed SST perturbation. This design allows us to isolate the dependence of aerosol forcing on background climate state while keeping the imposed aerosol-emission perturbation fixed.

We also perform fixed-SST aerosol-forcing simulations under preindustrial base-state boundary conditions, following the piClim-control/piClim-aer design of the Radiative Forcing Model Intercomparison Project (RFMIP; Pincus et al., 2016; Kramer et al., 2026). These simulations provide an estimate of present-day anthropogenic aerosol forcing relative to a preindustrial background and are directly relevant to the model-based aerosol forcing estimates used in climate assessments (e.g., Forster et al., 2021). The RFMIP protocols define present-day anthropogenic aerosol emissions as representative of 2014 in CMIP6 (Pincus et al., 2016) and 2021 in CMIP7 (Kramer et al., 2026). In our RFMIP-style experiments, the reference simulation uses preindustrial forcing boundary conditions representative of 1850, whereas the aerosol-perturbed simulation imposes present-day anthropogenic aerosol emissions representative of 1990s conditions, consistent with the five base-state experiments. All other forcing boundary conditions remain at preindustrial levels. The resulting forcing components provide the preindustrial-base-state aerosol-forcing estimates shown as horizontal dashed lines in Figs. 1, 2 and A2. These RFMIP-style estimates are similar to, but not identical with, the 0K aerosol-forcing estimates because the reference climate and non-aerosol forcing conditions differ.

2.3 Decomposition methods

Total aerosol effective radiative forcing (ERF) is diagnosed as the top-of-atmosphere radiative response to the aerosol perturbation under fixed-SST conditions. For diagnostic purposes, we decompose total aerosol ERF into instantaneous radiative forcing from aerosol–radiation interactions (IRF_{ari}), effective radiative forcing from aerosol–cloud interactions (ERF_{aci}), and a residual associated with non-cloud atmospheric adjustments. ERF_{aci} is interpreted as the total cloud-mediated



radiative response to the aerosol perturbation, including cloud adjustments associated with aerosol-induced atmospheric heating; we do not attempt to separate the semi-direct effect from the total cloud response. Shortwave and longwave components are analyzed separately throughout.

The radiative kernel method (Soden et al., 2008) is used to estimate longwave forcing components and non-cloud shortwave atmospheric adjustments. The radiative kernels used here are based on the atmospheric component of HadGEM3, a recent-generation climate model developed by the UK Met Office (Smith et al., 2020). Radiative kernels relate changes in atmospheric and surface variables to their associated top-of-atmosphere radiative responses and are used here to diagnose contributions from temperature, water vapor, and surface albedo. These diagnostics are used to estimate non-cloud rapid adjustments under both clear-sky and all-sky conditions, helping to isolate direct aerosol effects and separate cloud-mediated forcing from non-cloud atmospheric responses.

We use the approximate partial radiative perturbation method (APRP; Taylor et al., 2007) to decompose the shortwave forcing response. APRP uses all-sky and clear-sky shortwave fluxes at the top-of-atmosphere and surface, together with cloud fraction, to attribute shortwave radiative changes to surface albedo, clouds, and non-cloud atmospheric constituents. The cloud component is used to diagnose shortwave ERF_{aci} and is further decomposed into cloud amount, cloud scattering, and cloud absorption contributions. The non-cloud APRP component includes both shortwave direct aerosol radiative effects and shortwave water-vapor adjustments. We therefore estimate shortwave IRF_{ari} by subtracting the radiative-kernel-derived water-vapor response from the APRP non-cloud term.

ISCCP simulator output is used to identify aerosol-induced cloud-regime changes. The ISCCP joint histogram separates clouds by cloud-top pressure and visible optical depth, allowing cloud-fraction changes to be mapped onto standard low-, middle-, and high-cloud categories and optically thin, intermediate, and thick regimes (Rossow and Schiffer, 1999). These diagnostics provide a regime-based interpretation of the cloud changes underlying the ERF_{aci} response.

The cloud-radiative-kernel method is applied to the ISCCP simulator output to estimate the radiative impact of cloud-regime and cloud-layer changes, including low-cloud and non-low-cloud contributions (Zelinka et al., 2012a, 2012b, 2013). The method maps changes in cloud fraction onto radiative sensitivities as a function of cloud-top pressure and optical depth, allowing cloud radiative responses to be decomposed by layer and into cloud amount, cloud scattering, and cloud altitude terms. We use cloud radiative kernels primarily as a structural diagnostic rather than as the primary estimate of ERF_{aci} . This distinction is important because the cloud radiative kernels used here are derived under aerosol-free conditions. When aerosols are located above clouds, overlying aerosol attenuates the shortwave radiation reaching cloud top, so aerosol-free kernels can overestimate the radiative impact of a given cloud perturbation (He et al., 2026). Nevertheless, cloud radiative kernels remain useful for diagnosing the vertical and optical-depth structure of cloud responses.



190 3. Results

3.1 Aerosol forcing weakens as the base-state climate warms

Both GFDL-AM4.0 and GFDL-AM4-MG2 show a systematic weakening of aerosol forcing as the base-state climate warms (Fig. 1). Total ERF becomes less negative from colder to warmer base states, indicating that the same present-day aerosol perturbation exerts weaker cooling in warmer climates. This response demonstrates that aerosol ERF depends not only on the
 195 imposed aerosol perturbation, but also on the background climate state into which that perturbation is introduced.

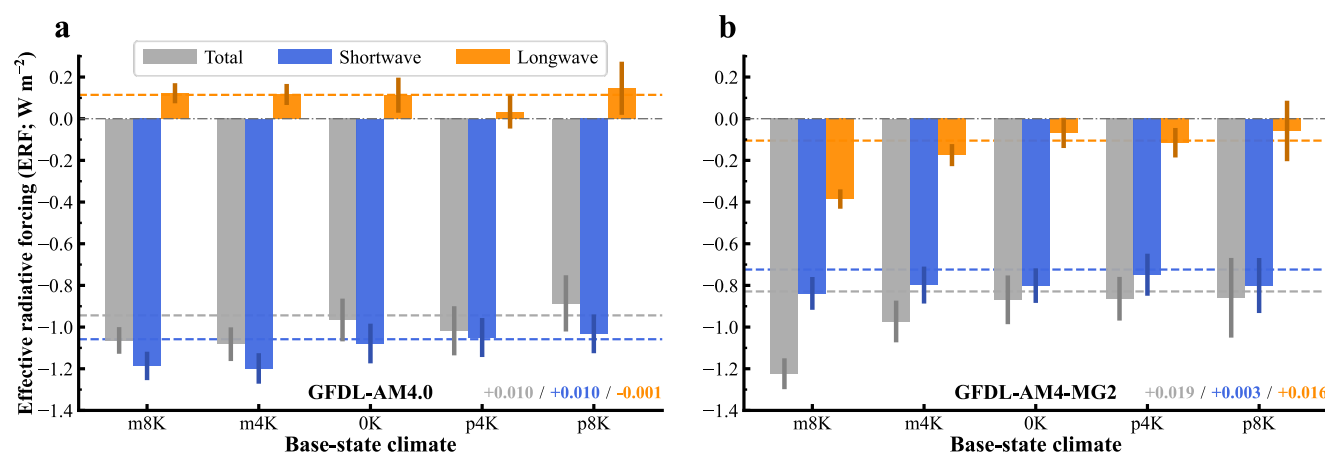


Figure 1. State dependence of aerosol effective radiative forcing in two GFDL models. Total, shortwave, and longwave effective radiative forcing (ERF; W m^{-2}) diagnosed from present-day aerosol perturbations relative to preindustrial aerosol conditions under five prescribed base-state climates: -8 K, -4 K, 0 K, +4 K, and +8 K. Results are shown for (a) GFDL-AM4.0 and (b) GFDL-AM4-MG2. Gray, blue, and orange bars indicate total, shortwave, and longwave ERF, respectively. Error bars represent 95% confidence intervals of the 30-year mean ERF estimates. Horizontal dashed lines indicate the corresponding ERF components diagnosed using preindustrial base-state boundary conditions, consistent with the RFMIP protocol. Colored numbers at the bottom right of each panel indicate the linear slopes of the corresponding ERF components ($\text{W m}^{-2} \text{K}^{-1}$), calculated against global-mean surface air temperature averaged between the preindustrial- and present-day-aerosol simulations across the five base states. These slopes summarize the overall trend but do not capture non-monotonic variations.

Although both models show weaker aerosol cooling with warming, the radiative components responsible for this weakening differ between configurations. In GFDL-AM4.0, variations in shortwave ERF primarily determine the state dependence of total ERF. In contrast, GFDL-AM4-MG2 exhibits a larger longwave contribution: longwave ERF becomes substantially less negative with warming and plays an important role in the total ERF response. Thus, longwave processes contribute more
 210 strongly to the state dependence of aerosol forcing in GFDL-AM4-MG2.

The corresponding zonal-mean ERF distributions show that the global-mean weakening is not spatially uniform (Fig. A1). The state dependence is particularly evident in the Northern Hemisphere, especially in aerosol-influenced midlatitude and high-latitude regions. In these regions, shortwave ERF generally becomes less negative as the base-state climate warms. GFDL-AM4-MG2 also exhibits a stronger longwave response than GFDL-AM4.0, with the largest differences occurring
 215 mainly in the Northern Hemisphere.



3.2 ERF_{aci} dominates the state dependence of total aerosol forcing

We decompose total aerosol ERF into IRF_{ari}, ERF_{aci}, and a residual associated with non-cloud atmospheric adjustments. The residual is generally small compared with ERF_{aci} and IRF_{ari}, with absolute magnitudes typically below 0.1 W m^{-2} (Fig. A2). A notable exception occurs in the colder base states of GFDL-AM4-MG2, where an unexpectedly large stratospheric adjustment contributes substantially to the residual and its state dependence, as discussed later. While acknowledging this additional contribution, we interpret the state dependence of total ERF primarily through ERF_{aci} and IRF_{ari} (Fig. 2). ERF_{aci} is the dominant forcing component and exhibits substantial weakening with warming, although IRF_{ari} also contributes to the overall state dependence. In both models, net ERF_{aci} becomes less negative as the base-state climate warms, but the principal radiative component differs between configurations.

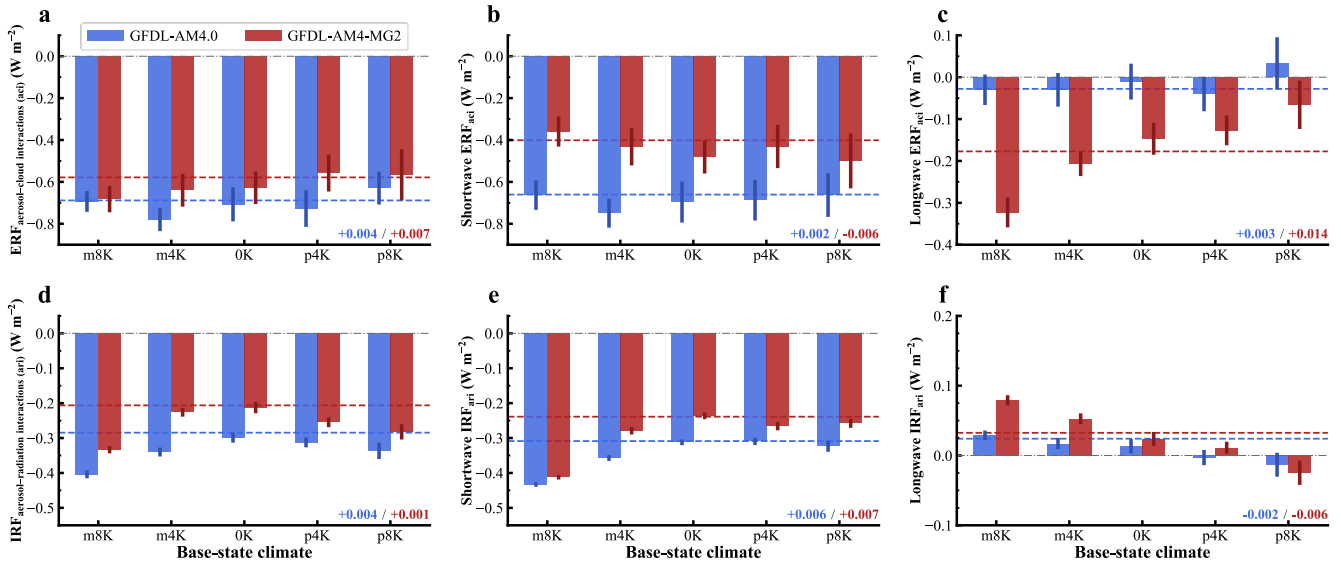


Figure 2. State dependence of aerosol ERF_{aci} and IRF_{ari} in two GFDL models. Total, shortwave, and longwave aerosol–cloud interaction effective radiative forcing (ERF_{aci}; W m⁻²) are shown in the top row, while the corresponding instantaneous radiative forcing from aerosol–radiation interactions (IRF_{ari}; W m⁻²) is shown in the bottom row. Results are diagnosed under five prescribed base-state climates: –8 K, –4 K, 0 K, +4 K, and +8 K. Blue bars indicate GFDL-AM4.0, and red bars indicate GFDL-AM4-MG2. Error bars represent 95% confidence intervals of the 30-year mean forcing estimates. Horizontal dashed lines indicate the corresponding forcing components diagnosed using preindustrial base-state boundary conditions, consistent with the RFMIP protocol. Colored numbers at the bottom right of each panel indicate the linear slopes of the corresponding forcing component for each model (W m⁻² K⁻¹), calculated against the respective model's global-mean surface air temperature, averaged between the preindustrial- and present-day-aerosol simulations for each base state. These slopes summarize the overall trend but do not capture non-monotonic variations.

In GFDL-AM4.0, ERF_{aci} is dominated by the shortwave component, with a comparatively small longwave contribution across all base states. Shortwave ERF_{aci} becomes progressively less negative from the –4 K to the +8 K base state, indicating a weakening of shortwave aerosol–cloud cooling over this range. The main exception is the –8 K base state, where shortwave ERF_{aci} is less negative than at –4 K and comparable to that at +8 K. This non-monotonic behavior may reflect



240 limited liquid or supercooled liquid water availability in the coldest base state, which could constrain aerosol activation and reduce the sensitivity of cloud optical properties to aerosol perturbations.

In GFDL-AM4-MG2, the weakening of net ERF_{aci} is driven more strongly by the longwave component. Shortwave ERF_{aci} is comparatively less state dependent, whereas longwave ERF_{aci} becomes substantially less negative as the base-state climate warms. This contrast reveals a longwave cloud-mediated pathway for aerosol-forcing state dependence in GFDL-AM4-
245 MG2. Thus, the two configurations agree that aerosol forcing weakens with warming, but they express this weakening through different radiative pathways: a shortwave-dominated pathway in GFDL-AM4.0 and a stronger longwave pathway in GFDL-AM4-MG2.

Net IRF_{ari} exhibits a less systematic weakening with warming than ERF_{aci} , although its changes across individual base-state intervals can be substantial. In both models, net IRF_{ari} initially becomes less negative from the -8 K to the 0 K base state, but
250 then becomes more negative again in warmer states. This non-monotonic behavior partly reflects compensation between its shortwave and longwave components. Shortwave IRF_{ari} generally becomes less negative in warmer climates, indicating a weakening of the shortwave direct aerosol radiative effect. Longwave IRF_{ari} is smaller in magnitude and tends to become less positive as the base-state climate warms, partly offsetting the shortwave changes and even turning negative in some warmer states. These negative estimates may partly reflect biases introduced by applying radiative kernels derived for a
255 reference climate to substantially warmer climate states, as discussed in He et al. (2026). Such biases could slightly exaggerate the weakening of longwave IRF_{ari} and consequently lead to a modest underestimation of the weakening of net IRF_{ari} . Thus, the direct aerosol radiative effect is also state dependent, although methodological biases may slightly mute its diagnosed weakening with warming.

3.2.1 Shortwave ERF_{aci} reflects competing cloud-regime responses

260 The state dependence of shortwave ERF_{aci} reflects competing cloud-regime responses rather than a spatially uniform weakening of aerosol–cloud interactions. We use the APRP method to decompose shortwave ERF_{aci} into cloud amount, cloud scattering, and cloud absorption contributions (Fig. 3). The decomposition shows that shortwave ERF_{aci} is controlled mainly by cloud scattering, with an additional model-dependent contribution from cloud amount in GFDL-AM4-MG2. The cloud-scattering component exhibits a clear meridional contrast: over low-to-mid latitudes (0 – 40° N), the scattering
265 contribution generally becomes less negative as the base-state climate warms, whereas over mid-to-high latitudes (40 – 90° N), it becomes more negative. This contrast indicates that the state dependence of aerosol–cloud interactions differs between predominantly warm-cloud and colder or mixed-phase regimes.

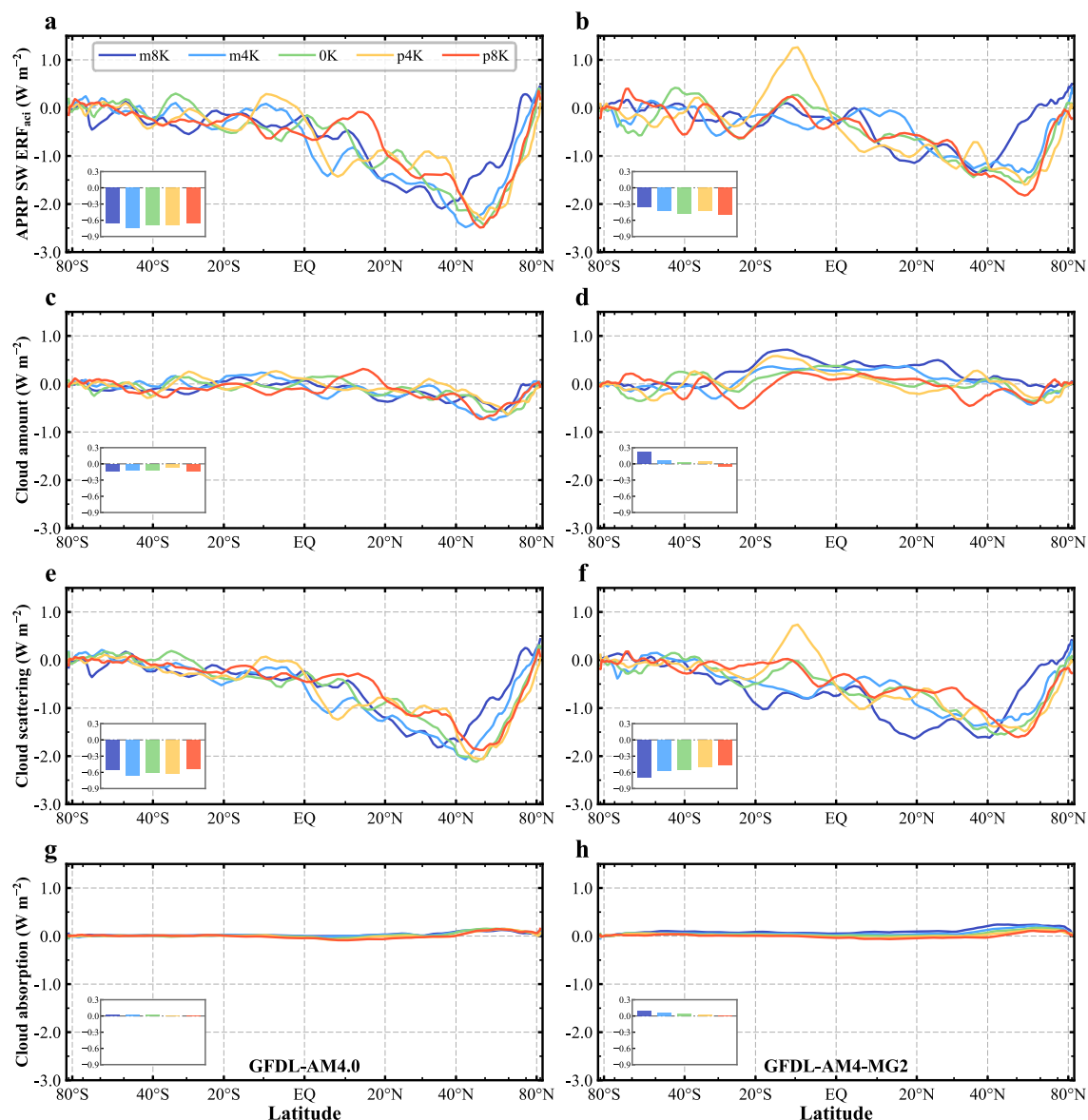


Figure 3. APRP decomposition of shortwave ERF_{aci} under different base-state climates. Zonal-mean distributions of shortwave aerosol–cloud interaction effective radiative forcing (ERF_{aci} ; $W m^{-2}$) and its component contributions, both diagnosed using the APRP method, under five prescribed base-state climates: $-8 K$, $-4 K$, $0 K$, $+4 K$, and $+8 K$. Results are shown for GFDL-AM4.0 in the left column and GFDL-AM4-MG2 in the right column. From top to bottom, rows show the total shortwave ERF_{aci} and the contributions from cloud amount, cloud scattering, and cloud absorption changes, respectively. Insets in each panel show the corresponding global-mean values for the five base-state climates, using the same color scheme as the zonal-mean curves.

Over $0-40^{\circ}N$, where liquid clouds are more prevalent, the weakening of the cloud-scattering contribution with warming is consistent with reduced cloud susceptibility. This may reflect the increasing optical thickness of liquid clouds with warming, which places the clouds in a more saturated regime of the Twomey response: cloud albedo becomes less sensitive to aerosol-



induced increases in cloud optical depth, so comparable optical-depth changes produce smaller albedo changes and a weaker shortwave cooling response (Feingold et al., 2022; Platnick and Oreopoulos, 2008). The Northern Hemisphere mean ISCCP simulator histograms provide a cloud-regime perspective on this behavior, showing that aerosol perturbations generally shift low-cloud fraction from optically thinner cumulus/stratocumulus-like regimes toward optically thicker stratocumulus/stratus-like regimes, with a weaker analogous shift in middle clouds (Fig. 4). This pattern is consistent with aerosol-induced increases in cloud optical thickness and negative shortwave ERF_{aci} .

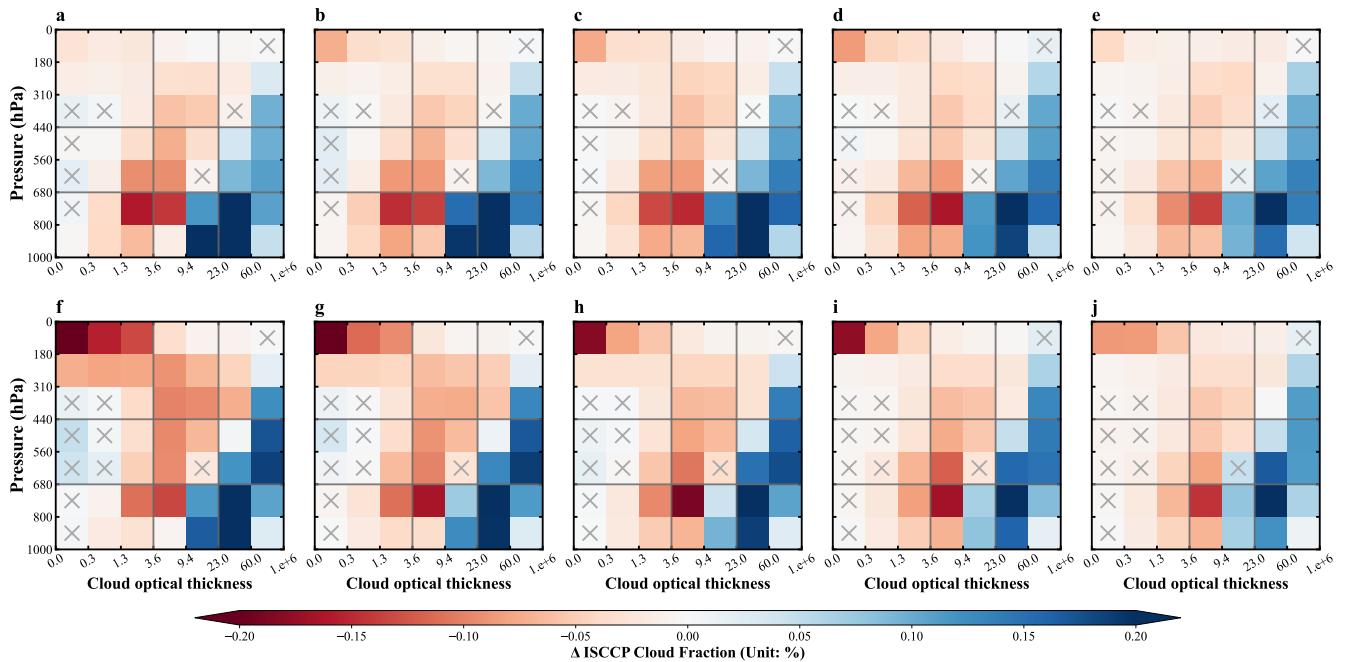


Figure 4. State dependence of aerosol-induced ISCCP cloud-fraction changes in the Northern Hemisphere. Northern Hemisphere mean changes in ISCCP-simulator cloud fraction (%) in response to present-day aerosol perturbations relative to preindustrial aerosol conditions. Results are shown for GFDL-AM4.0 in the top row and GFDL-AM4-MG2 in the bottom row. From left to right, panels show responses diagnosed under five prescribed base-state climates: -8 K , -4 K , 0 K , $+4\text{ K}$, and $+8\text{ K}$. Each panel shows the joint ISCCP 7×7 histogram of cloud visible optical depth and cloud-top pressure, with colors indicating aerosol-induced cloud-fraction changes in each regime. Horizontal and vertical lines denote the standard ISCCP cloud-regime classification; horizontal lines separate low-, middle-, and high-cloud categories, while vertical lines separate cloud optical-depth classes. Crosses indicate bins where the sign of the response is not consistent across the five prescribed base-state climates.

The hemispheric-mean response, however, masks distinct latitudinal behavior. Over $0-40^\circ\text{N}$, the redistribution toward optically thicker low and middle clouds weakens as the base-state climate warms (Fig. A3), whereas over $40-90^\circ\text{N}$, the redistribution remains comparable across base states or strengthens in some regimes (Fig. A4). This latitude contrast supports the interpretation that shortwave ERF_{aci} weakens in warmer liquid-cloud regimes as cloud susceptibility decreases, but can strengthen in colder or mixed-phase regimes as warming increases liquid or supercooled liquid water availability, allowing aerosol perturbations to more effectively modify cloud optical thickness.



The low-cloud and non-low-cloud cloud-radiative-kernel decompositions show how these regime-dependent scattering responses are distributed across cloud layers (Figs. A5 and A6). The low-cloud contribution is dominated by scattering changes and generally becomes less negative with warming, particularly over low-to-mid latitudes, consistent with the reduced warm-cloud susceptibility inferred from the ISCCP histograms. In contrast, the non-low-cloud scattering contribution is relatively weak in the coldest base state, especially over mid-to-high latitudes, and strengthens in some warmer adjacent states. This cold-state minimum is consistent with the nonmonotonic shortwave ERF_{aci} behavior between the -8 K and -4 K base states noted above, suggesting that limited liquid or supercooled liquid water availability in very cold cloud regimes can constrain aerosol-induced changes in cloud optical properties.

The cloud-radiative-kernel decomposition further helps interpret the model-dependent cloud-amount term identified by APRP in GFDL-AM4-MG2. Aerosol-induced increases in low-cloud amount produce a negative shortwave contribution, while reductions in non-low-cloud amount, consistent with the high-cloud decreases shown in Fig. A7, produce a positive shortwave contribution. These two cloud-amount terms partially offset each other, but the positive non-low-cloud term dominates in colder base states, yielding a net positive cloud-amount contribution to shortwave ERF_{aci} (Fig. 3d). As the base state warms and the high-cloud reduction weakens, this positive non-low-cloud term also weakens, causing the net cloud-amount contribution to shift from positive in colder states toward near-neutral or negative values in warmer states. Thus, the shortwave ERF_{aci} response reflects two coupled effects: regime-dependent scattering changes and compensating low- versus non-low-cloud amount adjustments, with the latter controlled mainly by the state-dependent high-cloud response in GFDL-AM4-MG2.

3.2.2 Aerosol–cloud–ice interactions introduce a longwave aerosol-forcing pathway

The strongest model contrast occurs in the longwave component of ERF_{aci} , where GFDL-AM4-MG2 exhibits a response that is largely absent in GFDL-AM4.0 (Fig. 2c). The shortwave cloud-amount decomposition above indicates that aerosol perturbations modify cloud structure more strongly in colder GFDL-AM4-MG2 climates, with reductions in non-low-cloud amount partly offset by increases in low-cloud amount. As the base state warms, the non-low-cloud reduction weakens, reducing the positive shortwave cloud-amount contribution. The longwave cloud-radiative-kernel decomposition provides the corresponding longwave signature of this cloud-structural response (Fig. A8).

The GFDL-AM4-MG2 longwave response is driven primarily by the cloud-altitude component. This contribution is strongly negative in colder base-state climates and weakens as the climate warms, whereas the corresponding response is much smaller in GFDL-AM4.0. Together with the shortwave cloud-amount terms, this behavior indicates stronger aerosol-induced changes in cloud vertical structure, particularly in non-low/high-cloud amount and emission level, in colder GFDL-AM4-MG2 climates. As the base state warms, this cloud-structural response weakens, making longwave ERF_{aci} less negative.

Layer-resolved cloud responses and ISCCP histograms support this high-cloud interpretation. In GFDL-AM4-MG2, aerosol perturbations reduce high-cloud amount most strongly in colder base states, and this reduction weakens substantially as the



base state warms (Fig. A7). The ISCCP histograms further show that the high-cloud reduction occurs primarily in optically thin-to-intermediate high-cloud regimes, rather than in optically thick deep-convective regimes (Fig. 4). In contrast, GFDL-AM4.0 shows much weaker high-cloud changes across the base-state climates. These diagnostics indicate that the longwave state dependence in GFDL-AM4-MG2 is linked to aerosol-induced changes in high-cloud amount and cloud emission level, consistent with aerosol–cloud–ice interactions. In colder GFDL-AM4-MG2 climates, reduced high-cloud amount and a lower effective emission level allow more longwave radiation to escape to space, producing a stronger negative longwave cloud adjustment. As the base state warms, this high-cloud response weakens, reducing the magnitude of the longwave cloud-altitude contribution.

One physically plausible mechanism is that aerosol–cloud–ice interactions are more effective in colder upper-tropospheric conditions. Increased ice-nucleating particles can promote heterogeneous freezing at lower supersaturation, deplete water vapor before homogeneous freezing is reached, reduce homogeneous ice nucleation, and generate fewer but larger ice crystals. These larger ice crystals sediment faster, shortening cirrus lifetime and reducing high-cloud coverage, which allows more longwave radiation to escape to space and produces a more negative longwave cloud adjustment (e.g., Storelvmo, 2017). In warmer base states, this mechanism may weaken as the cloud-phase distribution shifts and the high-cloud response becomes less controlled by cold upper-tropospheric ice nucleation. Although these microphysical steps are not directly isolated by our diagnostics, they are physically consistent with the diagnosed high-cloud and longwave responses.

The aerosol-induced high-cloud reductions may also contribute to the substantial stratospheric adjustment in GFDL-AM4-MG2. Aerosol-induced warming near the tropopause and in the lower stratosphere is most pronounced in the Northern Hemisphere and weakens with increasing background temperature, broadly consistent with the spatial distribution and state dependence of high-cloud reductions (Fig. A9). Previous cirrus cloud thinning simulations have shown that reductions in high-cloud coverage can produce stratospheric warming (Muri et al., 2014; Storelvmo et al., 2014). By reducing the absorption and re-emission of terrestrial radiation by cold high clouds, more longwave radiation from warmer atmospheric layers below can reach the stratosphere, where its absorption by radiatively active gases contributes to warming. A similar mechanism may operate here, while associated changes in upper-tropospheric static stability could further reinforce high-cloud reductions, consistent with the stability-iris mechanism (Bony et al., 2016).

3.3 Direct aerosol forcing weakens through aerosol optical depth and optical effectiveness

The zonal-mean diagnostics confirm the weakening of shortwave IRF_{ari} with warming, particularly poleward of about 30°N (Fig. A10). The separation among base states is larger in colder climates and smaller in warmer climates, suggesting that the incremental weakening of the shortwave direct aerosol radiative effect diminishes with warming. To diagnose the physical basis of this response, we examine zonal-mean changes in column aerosol mass and aerosol optical depth (AOD; Fig. 5). Column aerosol mass represents the aerosol burden within the atmospheric column after transport, transformation, and removal, including both interstitial aerosol and aerosol associated with clouds. AOD is more directly related to the optically

active aerosol that contributes to shortwave scattering and absorption. These quantities need not vary proportionally because AOD depends on the optical effectiveness of aerosol mass, not simply on the amount of aerosol present.

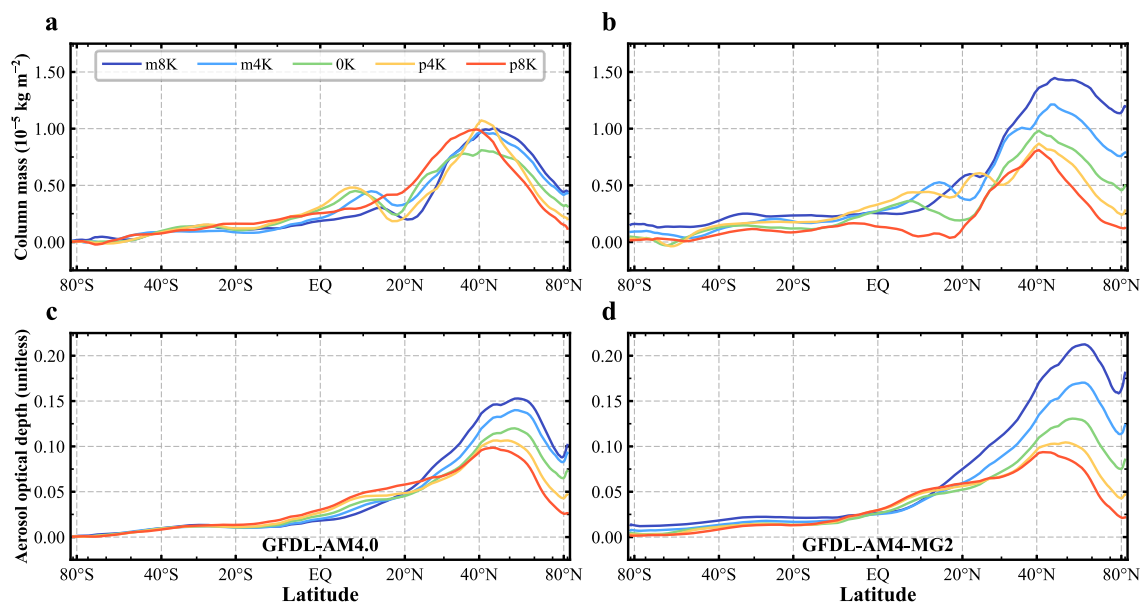


Figure 5. State dependence of aerosol burden and aerosol optical depth responses. Zonal-mean distributions of aerosol-induced changes in column aerosol mass and aerosol optical depth under five prescribed base-state climates: -8 K , -4 K , 0 K , $+4\text{ K}$, and $+8\text{ K}$. Results are shown for GFDL-AM4.0 in the left column and GFDL-AM4-MG2 in the right column. The top row shows changes in column aerosol mass, including both free-tropospheric/interstitial aerosols and aerosols associated with clouds. The bottom row shows changes in aerosol optical depth, which is more directly related to the instantaneous radiative forcing from aerosol–radiation interactions (IRF_{ari}). Different colored lines indicate the five base-state climates.

In both models, AOD generally decreases as the base-state climate warms, consistent with the weakening of shortwave IRF_{ari} . Column aerosol mass also decreases with warming in many regions, indicating that reduced aerosol burden contributes to the weaker direct radiative effect. However, the burden–AOD relationship is not uniform. Over the Northern Hemisphere subtropics and midlatitudes, especially around $20\text{--}40^\circ\text{N}$ in GFDL-AM4.0, AOD decreases with warming even though column aerosol mass remains nearly unchanged or increases. This decoupling indicates that the state dependence of IRF_{ari} is not controlled by aerosol burden alone, but also by changes in the optical effectiveness of aerosol mass.

Precipitation diagnostics help assess the potential role of wet removal in the AOD response (Figs. A11a and A11b). Precipitation climatology increases substantially as the base-state climate warms in both models, indicating a stronger hydrological cycle and greater potential for aerosol scavenging. This background increase in precipitation can contribute to reduced aerosol burden and weaker direct aerosol forcing in some regions. However, precipitation alone cannot explain the full AOD response, because wet removal depends not only on precipitation amount but also on where aerosols reside relative to clouds and precipitation and on how efficiently cloud condensate is converted into precipitation.



The aerosol-induced precipitation-efficiency response provides a complementary diagnostic of how aerosol perturbations affect cloud-to-precipitation conversion within each base state (Figs. A11c and A11d). Here, precipitation efficiency is defined as the ratio of column precipitation rate to condensed water path, following Li et al. (2022). Aerosol perturbations generally reduce precipitation efficiency over broad Northern Hemisphere regions, especially poleward of about 20–30°N, and the magnitude of this reduction weakens as the base state warms, particularly in GFDL-AM4.0. As the background precipitation climatology increases with warming, this weaker aerosol-induced suppression is consistent with greater potential for precipitation scavenging in warmer base states. In GFDL-AM4-MG2, the precipitation-efficiency response is weaker and less clearly separated across base states, consistent with aerosol–cloud–ice processes partly damping or offsetting the liquid-cloud precipitation-suppression response.

The limitation of a wet-removal-only interpretation is most evident over 20–40°N, where AOD decreases with warming more clearly than column aerosol mass (Fig. 5). This latitude band lies outside the tropical precipitation maximum and the main midlatitude storm-track precipitation regions, and the warming-driven increase in precipitation is comparatively modest. Therefore, the pronounced decrease in AOD over 20–40°N, especially where column aerosol mass does not decrease proportionally, is unlikely to be explained primarily by stronger precipitation scavenging alone.

Relative-humidity changes provide a plausible explanation for this decoupling between column aerosol mass and AOD (Fig. A12). Under uniform 4 K warming, relative humidity decreases over the subtropics, especially in the lower-to-middle troposphere over 20–40°N, while increasing in parts of the tropical free troposphere. These relative-humidity changes are consistent with the contrasting AOD responses in Fig. 5. For hygroscopic aerosols, subtropical drying reduces aerosol water uptake, decreases wet particle size, and lowers the ambient extinction per unit dry aerosol mass, even when column aerosol mass does not decrease proportionally. Reduced hygroscopic growth in drier subtropical environments can therefore help explain why AOD and IRF_{ari} weaken with warming despite relatively modest changes in column aerosol mass. Conversely, increased relative humidity in parts of the northern tropics can enhance hygroscopic growth and contribute to local AOD increases with warming.

Changes in aerosol abundance and lifetime may also contribute to the weakening of ERF_{aci} with warming, although their implications for aerosol–cloud interactions are less straightforward. The vertical distribution of aerosol extinction in Fig. A12 further suggests that the weakening of the aerosol optical perturbation with warming is particularly evident at higher altitudes. This pattern may partly reflect changes in aerosol transport, cloud processing, and wet removal in warmer base states. Because aerosol emissions are continuously supplied and aerosols can interact with clouds before being removed, a smaller atmospheric aerosol burden does not necessarily imply a proportionally smaller aerosol perturbation experienced by clouds. This interpretation is further complicated for high clouds, where aerosol availability depends on vertical transport and removal before particles reach the upper troposphere, potentially influencing aerosol–cloud–ice interactions. Without three-dimensional aerosol mass and process-level diagnostics of aerosol evolution, the contributions of changes in aerosol abundance and lifetime to the weakening of ERF_{aci} cannot be readily separated from changes in cloud susceptibility.

3.4 Aerosol forcing and climate feedback covary with climate state

Finally, we examine the relationship between aerosol forcing and climate feedback across adjacent base-state transitions (Fig. 6). Net and cloud feedback values are taken from He et al. (2024), where feedbacks were quantified for the same base-state transitions. That study showed that cloud feedback becomes less negative as the base-state climate warms, primarily because of a weakening of the negative shortwave cloud optical-depth feedback over the middle and high latitudes. This weakening reflects the temperature dependence of cloud liquid water responses, governed by changes in the moist adiabat and cloud adiabaticity, together with the nonlinear dependence of cloud albedo on optical depth. As the background climate warms and clouds become optically thicker, further increases in cloud optical depth produce progressively smaller changes in cloud albedo. Consistent with this state dependence, total aerosol forcing becomes less negative as the corresponding net climate feedback becomes less negative, indicating that aerosol forcing and feedback vary coherently with the background climate state.

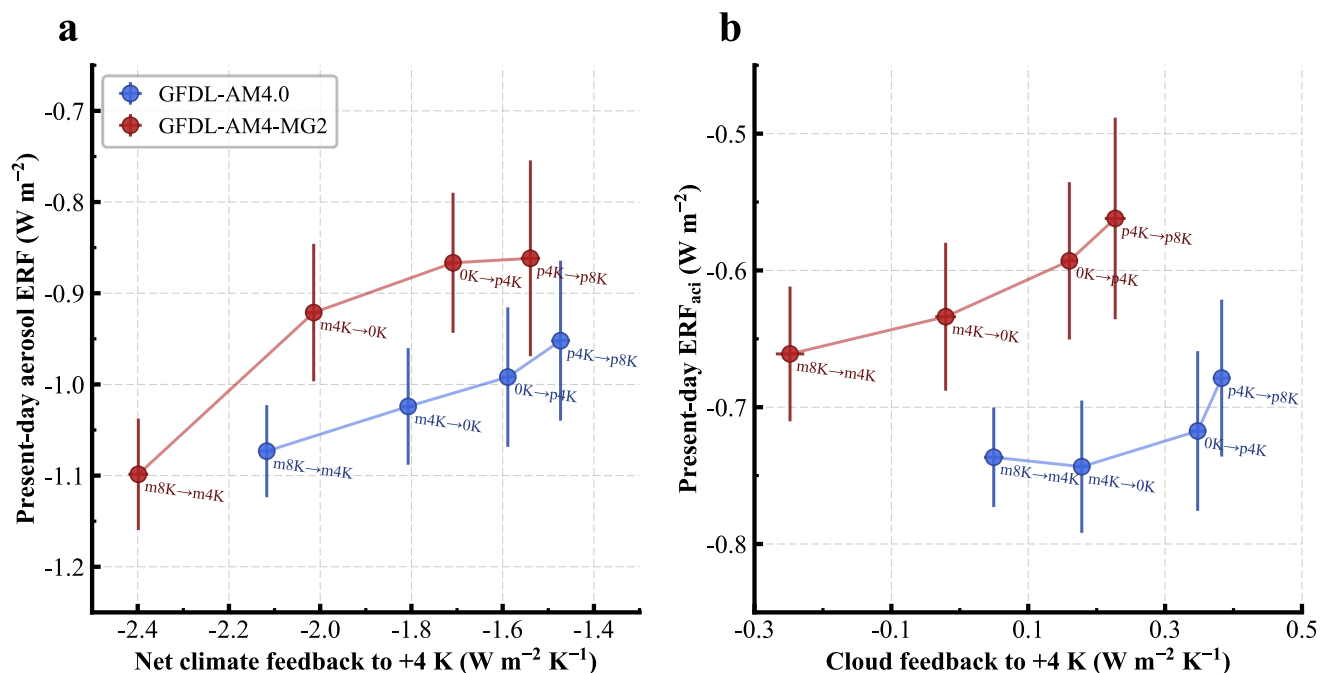


Figure 6. Relationship between aerosol forcing and climate feedback across base-state transitions. Scatter plots showing the relationship between aerosol forcing and climate feedback in GFDL-AM4.0 and GFDL-AM4-MG2. (a) Present-day aerosol effective radiative forcing (ERF; $W m^{-2}$) plotted against net climate feedback to uniform 4 K warming ($W m^{-2} K^{-1}$). (b) Aerosol-cloud interaction effective radiative forcing (ERF_{aci}; $W m^{-2}$) plotted against cloud feedback to uniform 4 K warming ($W m^{-2} K^{-1}$). Blue markers indicate GFDL-AM4.0, and red markers indicate GFDL-AM4-MG2. Each point represents one transition between two adjacent base-state climates: the aerosol forcing is the mean of its values in the two endpoint base states, while the climate feedback is calculated from the change in radiative flux between each pair of adjacent base states, divided by the corresponding change in global-mean surface air temperature. Specifically, the four points correspond to the transitions from -8 K to -4 K, -4 K to 0 K, 0 K to +4 K, and +4 K to +8 K. Horizontal and vertical error bars represent 95% confidence intervals of the corresponding feedback and forcing estimates, respectively. Lines connect adjacent transition points to illustrate the progression across base states.



440 A similar positive association appears between ERF_{aci} and cloud feedback. This relationship may partly reflect the influence
of background cloud properties on both radiative responses. In particular, changes in cloud liquid water content and optical
depth can affect both the sensitivity of cloud albedo to aerosol-induced perturbations and the radiative response of clouds to
surface warming. These mechanisms provide a possible physical basis for their coherent state dependence, although the
processes governing aerosol-induced cloud changes and cloud feedback are not necessarily identical. Moreover, this global-
445 mean covariation does not require the state dependence of forcing and feedback to originate from the same geographical
regions. As shown in He et al. (2024), the state dependence of cloud feedback arises primarily over the extratropics and high
latitudes, particularly the Southern Ocean, whereas the state dependence of aerosol forcing in the present study is
concentrated more strongly over the Northern Hemisphere subtropics and midlatitudes. Thus, aerosol–cloud interactions and
cloud feedbacks may be jointly modulated by changes in the background climate state, even though their state dependence
450 arises primarily from different regions rather than a single shared regional mechanism.

The relationship between ERF_{aci} and cloud feedback is clearer in GFDL-AM4-MG2, where ERF_{aci} weakens as cloud
feedback becomes less negative across base-state transitions. In GFDL-AM4.0, the relationship is weaker, partly because the
−8 K to −4 K transition departs from the smoother progression across warmer base states. This exception is consistent with
the non-monotonic shortwave ERF_{aci} behavior discussed above, whereas the additional cloud-ice and high-cloud pathways in
455 GFDL-AM4-MG2 may contribute to its more coherent relationship. Because the uncertainty ranges are substantial, the
scatter plots should be interpreted as evidence for a systematic tendency rather than as a precise quantitative constraint.
Nevertheless, the result suggests that forcing and feedback are not fully separable properties of a model when the
background climate state changes.

This behavior contrasts with the anti-correlation identified across CMIP6 models and reproduced in several perturbed-
460 parameter ensembles, in which models with stronger negative ERF_{aci} tend to exhibit less negative, or more positive, cloud
feedback (Wang et al., 2021; Gettelman et al., 2024; Feng et al., 2026). The difference likely reflects the nature of the
comparison. Here, the relationship is diagnosed from controlled base-state changes within each model configuration, so the
model formulation and imposed aerosol perturbation are held fixed while the background climate state changes. In contrast,
intermodel and perturbed-parameter relationships combine differences in model formulation, cloud and aerosol
465 parameterizations, mean-state climatology, and historical tuning or calibration. A positive relationship is not unique to the
present analysis: Qin et al. (2025) found a moderate positive correlation between ERF_{aci} and cloud feedback in the global-
ocean mean of an E3SMv2 perturbed-parameter ensemble, while also showing substantial variation in the sign and strength
of the relationship across cloud regimes. Thus, Fig. 6 should be interpreted as a state-dependent relationship within model
configurations rather than as a direct emergent constraint across model or perturbed-parameter ensembles. Nevertheless, it
470 may help interpret broader forcing–feedback relationships by identifying physical pathways through which background
climate state can modulate both aerosol forcing and climate feedback.



4. Conclusions and Implications

Using fixed-SST simulations with GFDL-AM4.0 and GFDL-AM4-MG2 across five prescribed base-state climates, we show that aerosol forcing systematically weakens as the climate warms. The same present-day aerosol perturbation produces weaker cooling in warmer climates, demonstrating that aerosol forcing is not solely a function of emissions or aerosol burden. Instead, it depends on background climate conditions that regulate both aerosol–radiation interactions and aerosol–cloud interactions.

The weakening of aerosol forcing reflects multiple physical mechanisms. For ERF_{aci} , the dominant mechanisms involve cloud optical properties, cloud amount, and cloud vertical structure. The two model configurations express this state dependence through different radiative pathways: GFDL-AM4.0 is dominated by shortwave ERF_{aci} , whereas GFDL-AM4-MG2 includes a stronger longwave contribution associated with high-cloud and cloud-ice processes. Within the shortwave component, ERF_{aci} is governed mainly by cloud scattering, with additional cloud-amount contributions in GFDL-AM4-MG2. In warm-cloud regimes, low-cloud scattering responses weaken with warming, consistent with reduced susceptibility of optically thicker liquid clouds to aerosol perturbations and a weaker incremental cloud-brightening response. In colder and mixed-phase regimes, warming can increase liquid or supercooled liquid water availability, strengthening aerosol-induced cloud brightening in non-low clouds. These competing responses explain why shortwave ERF_{aci} does not weaken uniformly across latitude or cloud regime.

GFDL-AM4-MG2 adds a distinct cloud-ice and high-cloud pathway. In colder base states, aerosol perturbations reduce high-cloud amount and modify cloud-altitude radiative effects more strongly than in warmer base states. These high-cloud changes produce a more negative longwave cloud adjustment by reducing optically thin-to-intermediate high-cloud coverage and lowering the effective cloud emission level, allowing more longwave radiation to escape to space. As the base state warms, this high-cloud pathway weakens, consistent with the reduced high-cloud response, smaller longwave cloud-altitude contribution, and weaker associated shortwave cloud-amount compensation in GFDL-AM4-MG2. This pathway helps explain why the longwave component is much more important for aerosol-forcing state dependence in GFDL-AM4-MG2 than in GFDL-AM4.0, indicating that the representation of aerosol–cloud–ice interactions can alter not only the magnitude of ERF_{aci} , but also the radiative pathway through which ERF_{aci} depends on climate state.

For IRF_{ari} , warming weakens aerosol optical depth and the direct radiative effect. This weakening is largely related to reduced aerosol burden in many regions, with increased wet removal likely contributing, but aerosol optical depth does not always scale with column aerosol mass. In particular, AOD decreases over parts of 20–40°N even where column aerosol mass remains nearly unchanged or increases. This decoupling points to changes in the optical effectiveness of aerosol mass, especially through humidity-dependent water uptake. Thus, direct aerosol forcing also depends on the background climate state, not only through aerosol burden and removal, but also through the optical properties of aerosol mass.



Finally, aerosol forcing and climate feedback vary coherently with the background climate state: aerosol forcing becomes less negative as feedbacks become less negative. This result indicates that forcing and feedback are not fully separable when the background climate changes. Because this relationship is diagnosed from controlled base-state changes within individual model configurations, it should be interpreted as a state-dependent relationship within model configurations rather than as a direct emergent constraint across model or perturbed-parameter ensembles.

4.1 Implications for fixed-aerosol warming experiments

In many idealized warming experiments, background aerosol emissions or concentrations are held fixed while the climate warms. If the radiative effect of those aerosols weakens in a warmer base state, then part of the diagnosed climate response reflects a loss of aerosol cooling. This hidden reduction in aerosol cooling can be incorporated into the apparent feedback response, making the diagnosed effective feedback less negative than it would be if aerosol forcing were invariant to climate state. The reverse can occur in cooling experiments, where fixed aerosols may exert stronger cooling as the climate cools. The resulting forcing change is not externally imposed in the usual sense, but emerges from the interaction between fixed aerosol perturbations and the evolving climate state. This complicates the separation of forcing and feedback in widely used climate-change experiments with fixed background aerosols, potentially affecting the interpretation of diagnosed climate feedback and climate sensitivity. It also highlights the importance of diagnosing aerosol forcing consistently across climate states.

4.2 Implications for historical attribution, model tuning, and future projections

Historical climate simulations often rely on aerosol cooling to offset part of greenhouse-gas warming. If aerosol cooling is stronger in cooler climates and weaker in warmer climates, then time-invariant assumptions about aerosol efficacy may misrepresent the temporal evolution of aerosol masking. For a given aerosol perturbation, aerosol cooling may have been more effective under earlier, cooler climate states, while continued warming may reduce the cooling influence of aerosols even if emissions remain unchanged.

This state dependence also affects model tuning. Models with high climate sensitivity may rely on strong negative aerosol forcing to match observed historical temperature evolution. However, if aerosol forcing weakens with warming, aerosols become a less effective tuning lever in warmer climates. Achieving the same historical offset may therefore involve stronger aerosol forcing in cooler states or compensating differences in atmospheric adjustments, ocean heat uptake, or feedback strength.

For future projections, the results imply that aerosol forcing uncertainty depends not only on future emissions but also on the evolving background climate. The same aerosol emissions may exert weaker cooling in a warmer atmosphere because aerosol–cloud interactions weaken, aerosol optical depth decreases, and high-cloud responses become less sensitive to



aerosol perturbations. This does not remove the importance of future aerosol-emission uncertainty, but it indicates that aerosol forcing should be interpreted as a climate-state-dependent quantity.

535 Several limitations should be considered when interpreting these results. The analysis is based on two atmospheric configurations within the same GFDL model family, so the generality of the diagnosed state dependence across climate models remains to be established. The RFMIP2.0 protocol for CMIP7 offers an opportunity to assess this state dependence in a multi-model framework through paired aerosol-forcing experiments under preindustrial and uniformly warmed (+4 K) SST conditions (Kramer et al., 2026). Subtropical stratocumulus clouds also remain challenging to represent at typical climate-
540 model vertical resolution (e.g., Bogenschutz et al., 2021; Medeiros et al., 2012), and associated cloud-top entrainment and drying processes are only approximately represented (Mellado, 2016). For example, observational evidence suggests that warming can strengthen cloud-top drying through enhanced entrainment, providing an additional pathway by which aerosol-induced cloud cooling may weaken, or potentially reverse, at high SSTs beyond the reduced cloud susceptibility emphasized here (Zhang et al., 2022). If such processes are underrepresented in the present simulations, the magnitude of aerosol-forcing
545 state dependence diagnosed here may be underestimated. Modeling studies using a broader range of models and improved representations of cloud microphysical and boundary-layer processes will therefore be important for assessing the robustness and magnitude of aerosol-forcing state dependence.

5 Appendices

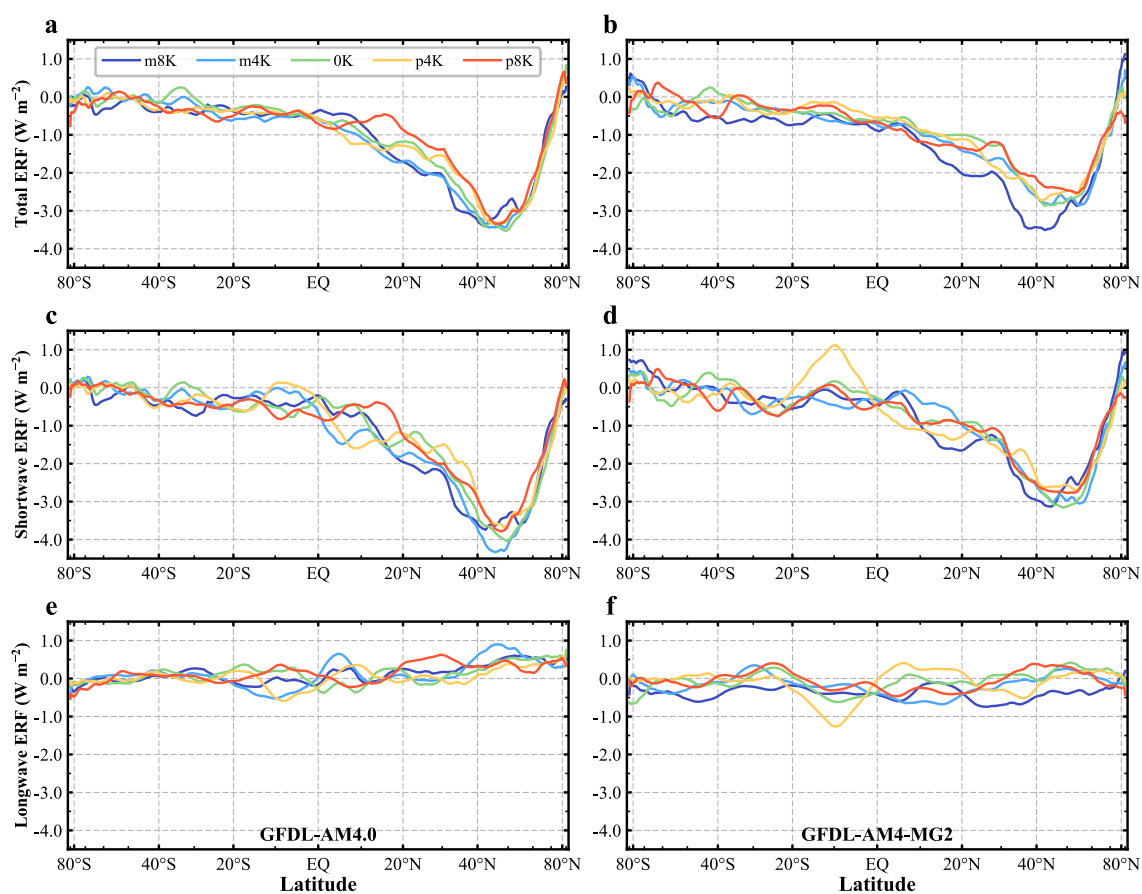


Figure A1. Zonal-mean distributions of effective radiative forcing from aerosol perturbations. Zonal-mean distributions of aerosol effective radiative forcing (ERF; W m^{-2}) in response to present-day aerosol perturbations relative to preindustrial aerosol conditions under five prescribed base-state climates: -8 K , -4 K , 0 K , $+4\text{ K}$, and $+8\text{ K}$. Results are shown for GFDL-AM4.0 in the left column and GFDL-AM4-MG2 in the right column. From top to bottom, rows show the total, shortwave, and longwave ERF, respectively. The corresponding global-mean values are shown in Fig. 1.

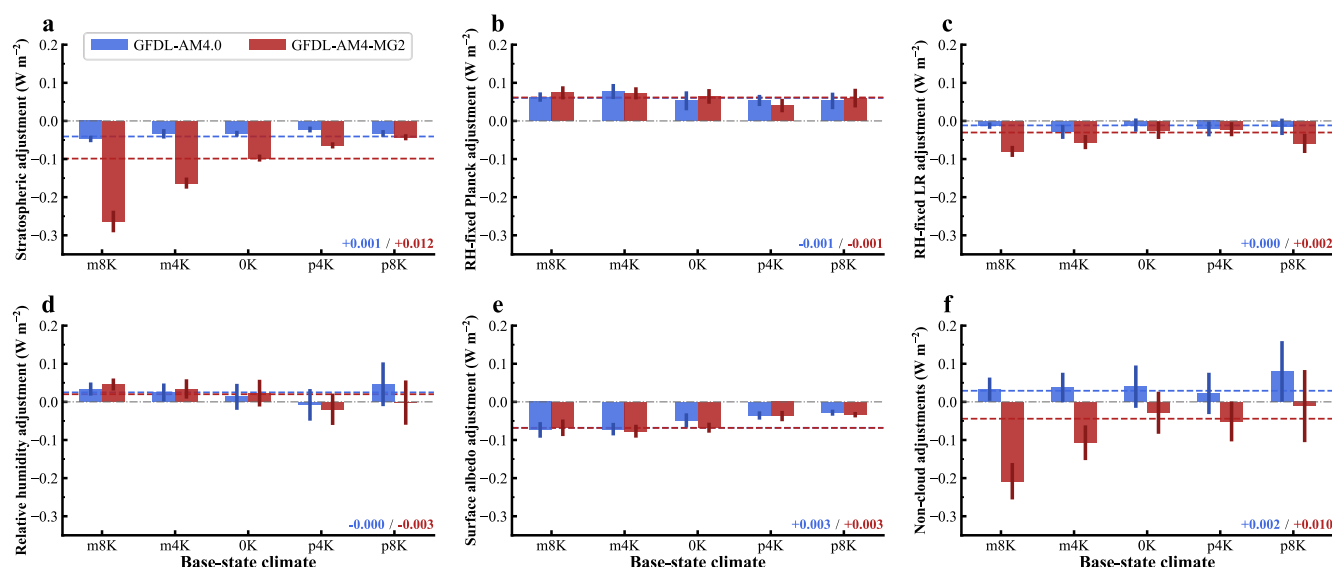
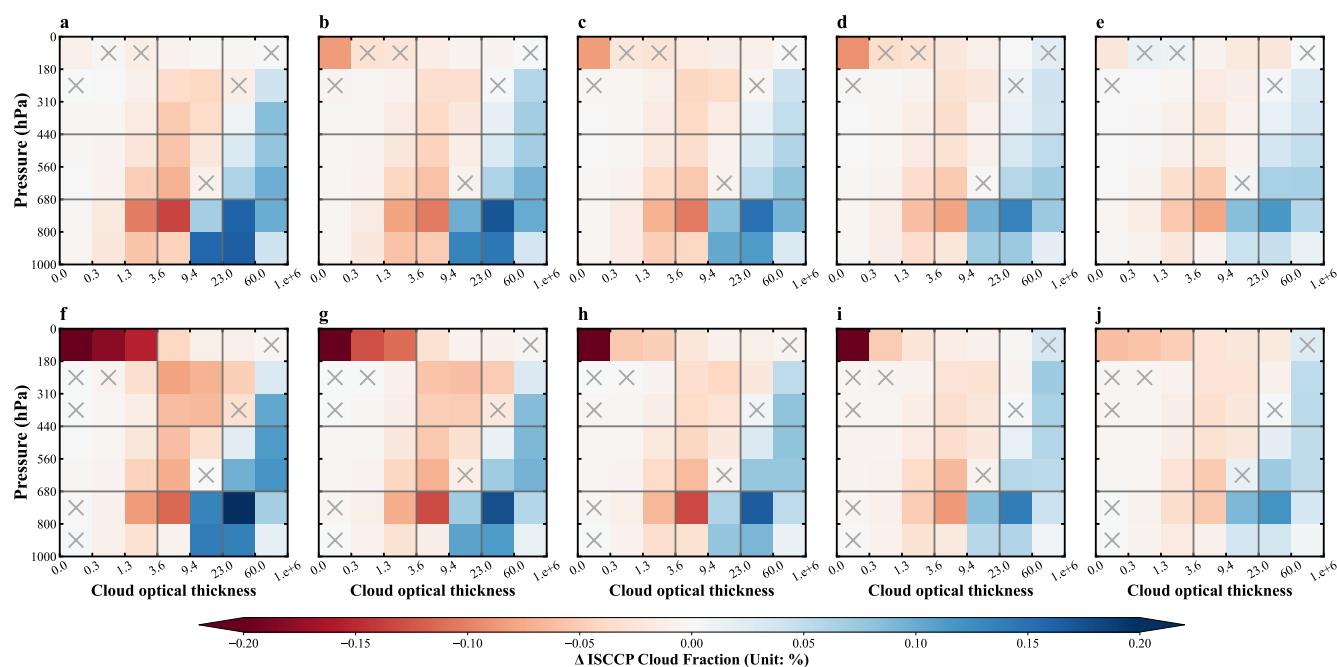
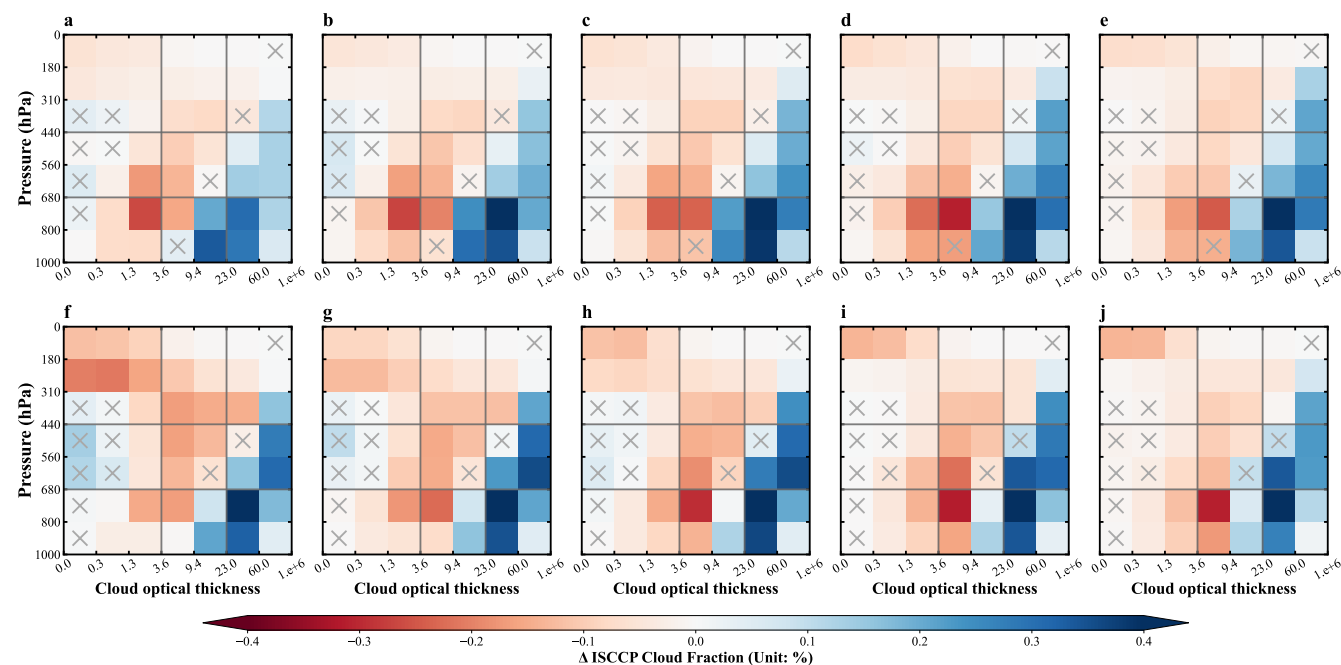


Figure A2. State dependence of non-cloud adjustments in two GFDL models. Stratospheric adjustment, relative-humidity-fixed (RH-fixed) Planck adjustment, RH-fixed lapse-rate (LR) adjustment, relative humidity adjustment, surface albedo adjustment, and the sum of these non-cloud adjustments are shown in panels (a–f), respectively. Results are diagnosed under five prescribed base-state climates: –8 K, –4 K, 0 K, +4 K, and +8 K. Blue bars indicate GFDL-AM4.0, and red bars indicate GFDL-AM4-MG2. Error bars represent 95% confidence intervals of the 30-year mean adjustment estimates. Horizontal dashed lines indicate the corresponding adjustment components diagnosed using preindustrial base-state boundary conditions, consistent with the RFMIP protocol. Colored numbers at the bottom right of each panel indicate the linear slopes of the corresponding adjustment component for each model ($\text{W m}^{-2} \text{ K}^{-1}$), calculated against the respective model's global-mean surface air temperature, averaged between the preindustrial- and present-day-aerosol simulations for each base state. These slopes summarize the overall trend but do not capture non-monotonic variations.



570 **Figure A3. State dependence of aerosol-induced ISCCP cloud-fraction changes in Northern low-to-mid latitudes. Mean changes in**
ISCCP-simulator cloud fraction (%) over 0–40°N in response to present-day aerosol perturbations relative to preindustrial aerosol
conditions. Results are shown for GFDL-AM4.0 in the top row and GFDL-AM4-MG2 in the bottom row. From left to right, panels
show responses diagnosed under five prescribed base-state climates: –8 K, –4 K, 0 K, +4 K, and +8 K. Each panel shows the joint
575 **ISCCP 7 × 7 histogram of cloud visible optical depth and cloud-top pressure, with colors indicating aerosol-induced cloud-fraction**
changes in each regime. Horizontal and vertical lines denote the standard ISCCP cloud classification; horizontal lines separate
low-, middle-, and high-cloud categories, while vertical lines separate cloud optical-depth classes. Crosses indicate bins where the
sign of the response is not consistent across the five prescribed base-state climates.



580 **Figure A4. State dependence of aerosol-induced ISCCP cloud-fraction changes in Northern mid-to-high latitudes. Mean changes in**
ISCCP-simulator cloud fraction (%) over 40–90°N in response to present-day aerosol perturbations relative to preindustrial
aerosol conditions. Results are shown for GFDL-AM4.0 in the top row and GFDL-AM4-MG2 in the bottom row. From left to
right, panels show responses diagnosed under five prescribed base-state climates: –8 K, –4 K, 0 K, +4 K, and +8 K. Each panel
585 **shows the joint ISCCP 7 × 7 histogram of cloud visible optical depth and cloud-top pressure, with colors indicating aerosol-**
induced cloud-fraction changes in each regime. Note that the color scale spans twice the range used in Figs. 4 and A3. Horizontal
and vertical lines denote the standard ISCCP cloud classification; horizontal lines separate low-, middle-, and high-cloud
categories, while vertical lines separate cloud optical-depth classes. Crosses indicate bins where the sign of the response is not
consistent across the five prescribed base-state climates.

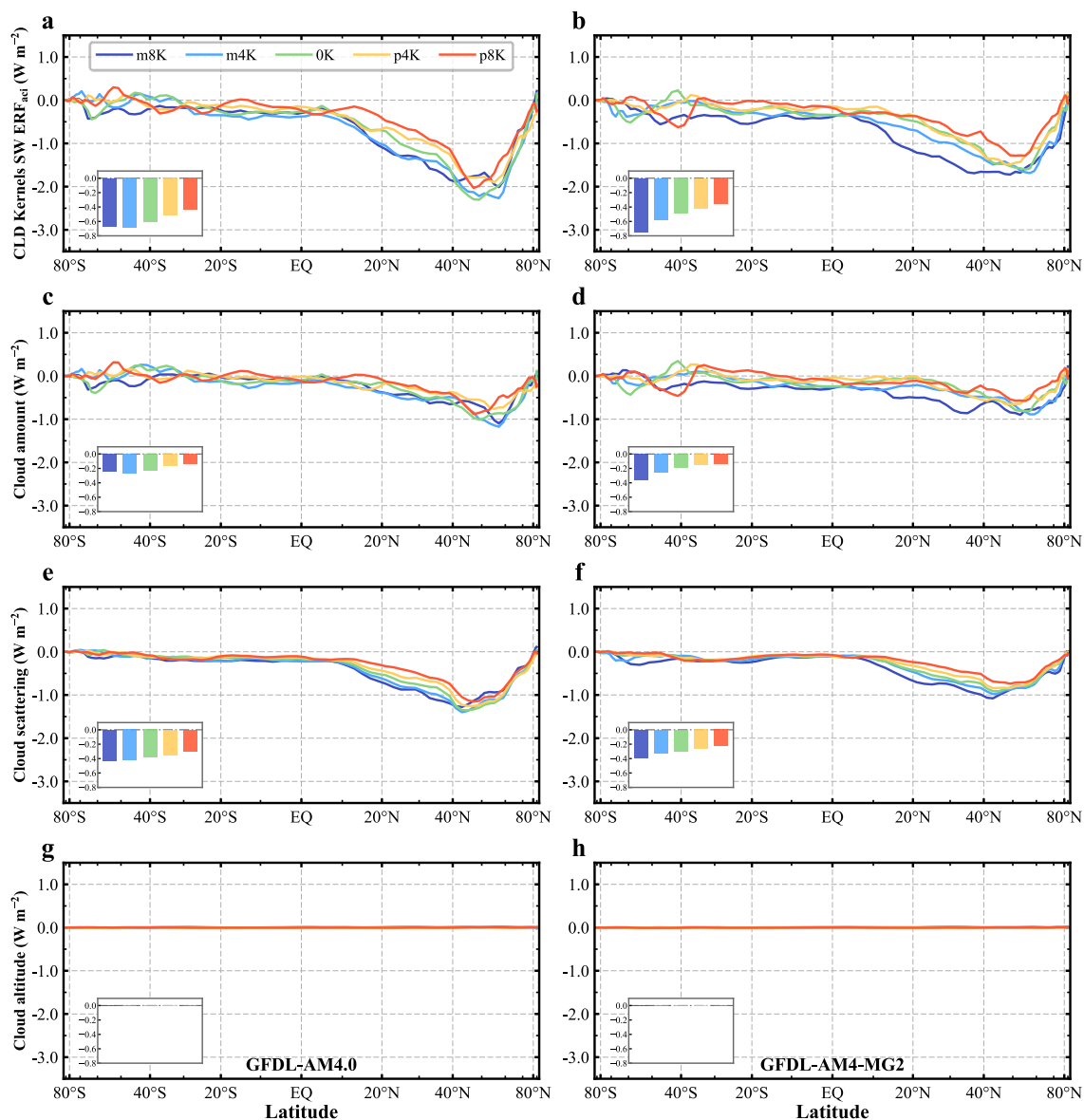
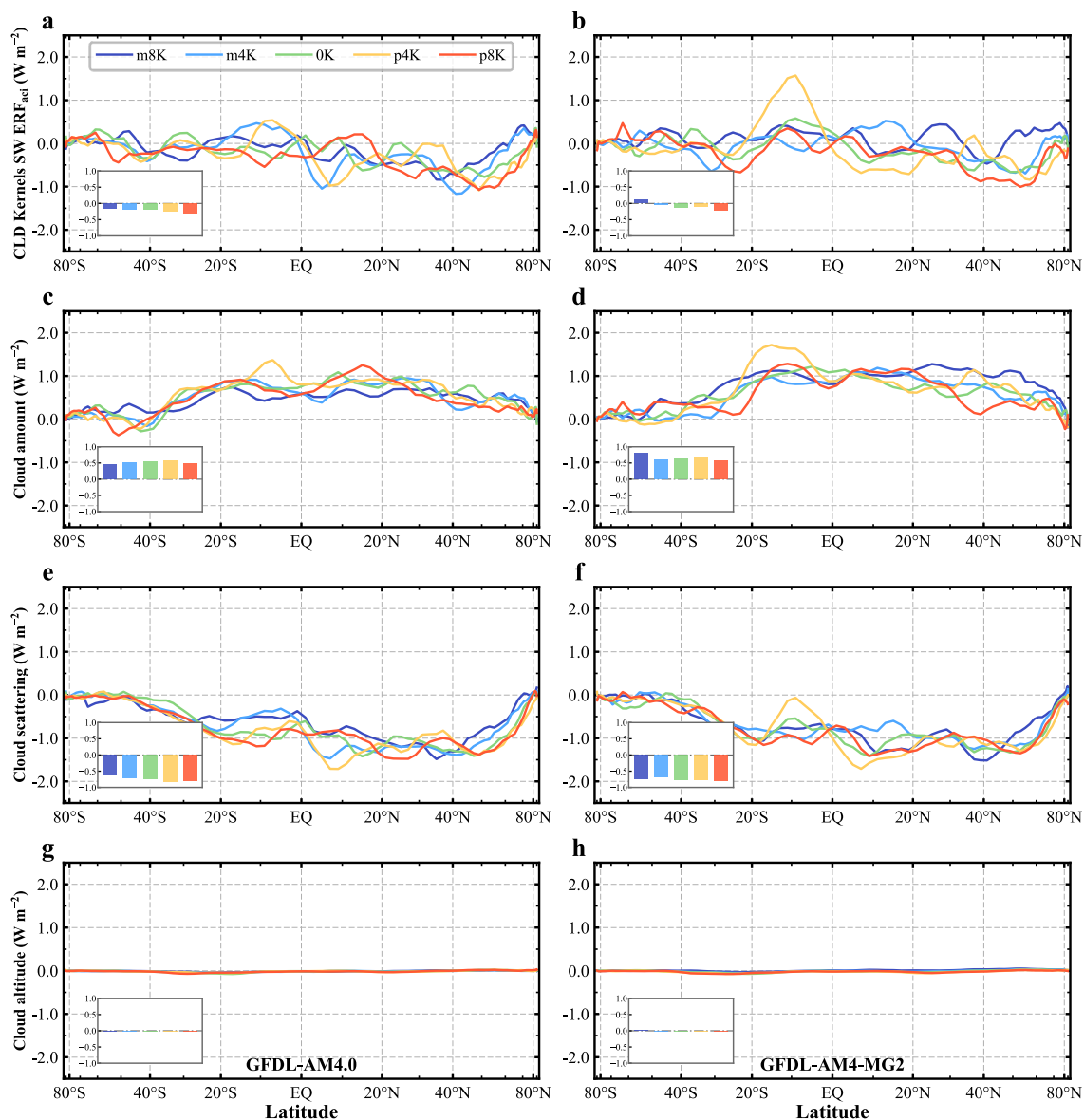


Figure A5. Cloud-radiative-kernel decomposition of low-cloud contributions to shortwave ERF_{aci} . Zonal-mean distributions of the low-cloud contribution to shortwave aerosol-cloud interaction effective radiative forcing (ERF_{aci} ; $W m^{-2}$) and its component contributions diagnosed using the cloud radiative kernel method. Results are shown under five prescribed base-state climates: $-8 K$, $-4 K$, $0 K$, $+4 K$, and $+8 K$. The left column shows GFDL-AM4.0, and the right column shows GFDL-AM4-MG2. From top to bottom, rows show the total low-cloud contribution to shortwave ERF_{aci} and the contributions from low-cloud amount, low-cloud scattering, and low-cloud absorption changes, respectively. Insets show the corresponding global-mean values for each base-state climate, using the same color scheme as the zonal-mean curves.



600 **Figure A6. Cloud-radiative-kernel decomposition of non-low-cloud contributions to shortwave ERF_{aci} .** Zonal-mean distributions of
the non-low-cloud contribution to shortwave aerosol–cloud interaction effective radiative forcing (ERF_{aci} ; W m^{-2}) and its
component contributions diagnosed using the cloud radiative kernel method. Results are shown under five prescribed base-state
climates: -8 K , -4 K , 0 K , $+4\text{ K}$, and $+8\text{ K}$. The left column shows GFDL-AM4.0, and the right column shows GFDL-AM4-MG2.
605 From top to bottom, rows show the total non-low-cloud contribution to shortwave ERF_{aci} and the contributions from non-low-
cloud amount, non-low-cloud scattering, and non-low-cloud absorption changes, respectively. Insets show the corresponding
global-mean values for each base-state climate, using the same color scheme as the zonal-mean curves.

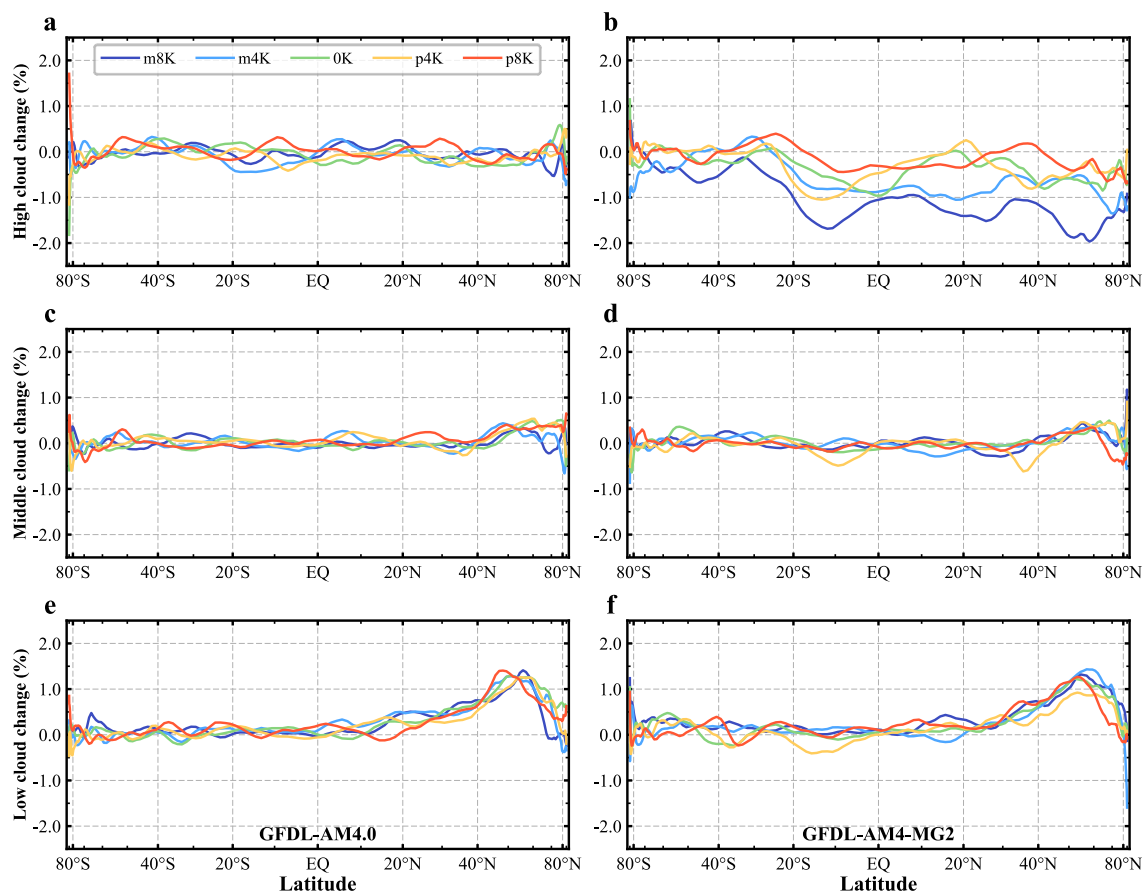


Figure A7. Zonal-mean distributions of aerosol-induced cloud changes by cloud layer under different base-state climates. Zonal-mean distributions of cloud changes in response to present-day aerosol perturbations relative to preindustrial aerosol conditions under five prescribed base-state climates: -8 K, -4 K, 0 K, $+4$ K, and $+8$ K. Results are shown for GFDL-AM4.0 in the left column and GFDL-AM4-MG2 in the right column. From top to bottom, rows show changes in high, middle, and low cloud, respectively. Different colored lines indicate the five base-state climates.

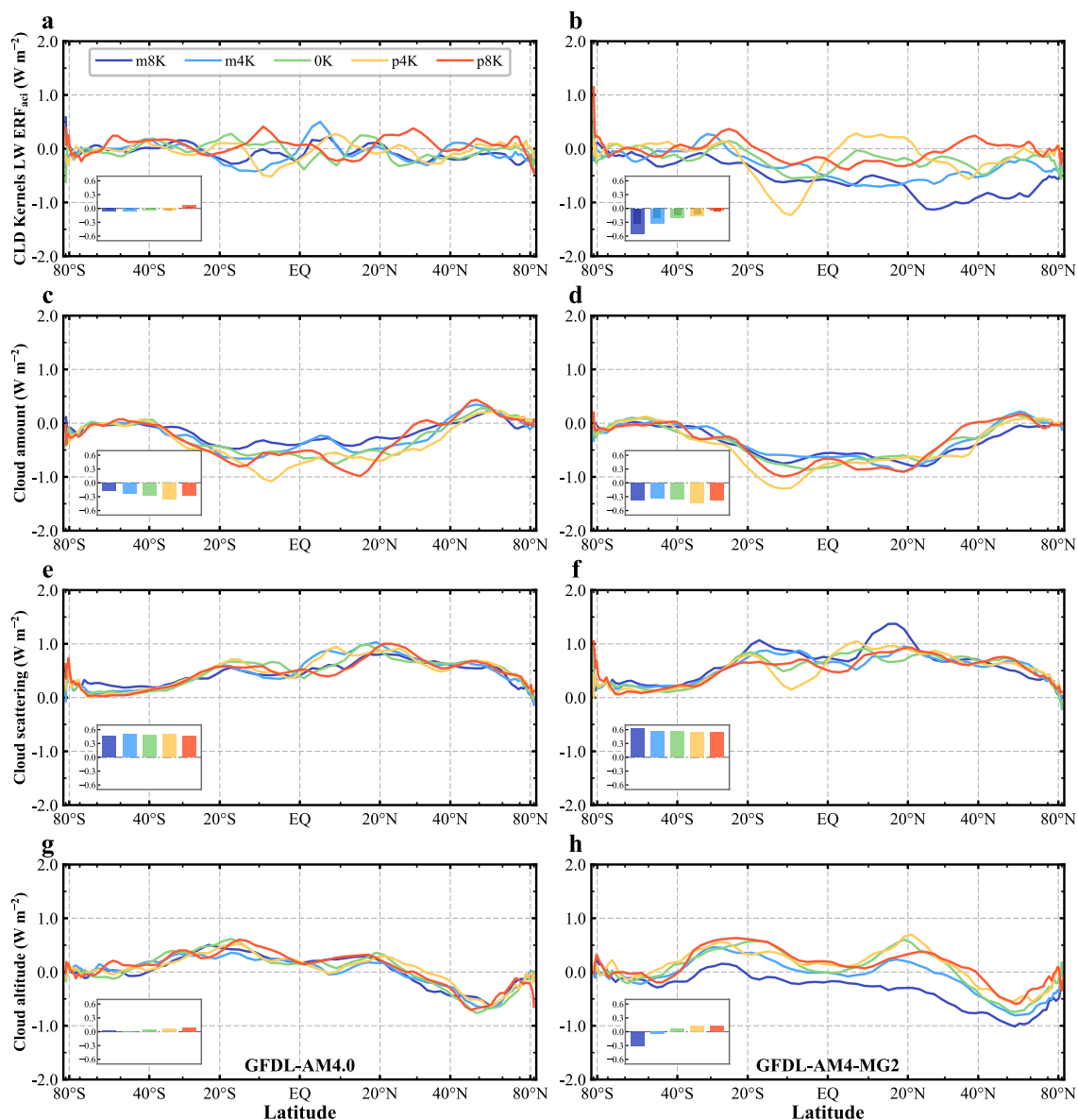


Figure A8. Cloud-radiative-kernel decomposition of longwave ERF_{aci} under different base-state climates. Zonal-mean distributions of longwave aerosol–cloud interaction effective radiative forcing (ERF_{aci} ; W m^{-2}) and its component contributions diagnosed using the cloud radiative kernel method. Results are shown under five prescribed base-state climates: -8 K , -4 K , 0 K , $+4\text{ K}$, and $+8\text{ K}$. The left column shows GFDL-AM4.0, and the right column shows GFDL-AM4-MG2. From top to bottom, rows show total longwave ERF_{aci} and the contributions from cloud amount, cloud scattering, cloud altitude, and the residual term, respectively. Insets show the corresponding global-mean values for each base-state climate, using the same color scheme as the zonal-mean curves. In the total longwave ERF_{aci} panels, the darker and narrower inset bars are overlaid on the corresponding total bars and indicate values diagnosed directly using the radiative kernel method; these correspond to the values shown in Fig. 2.

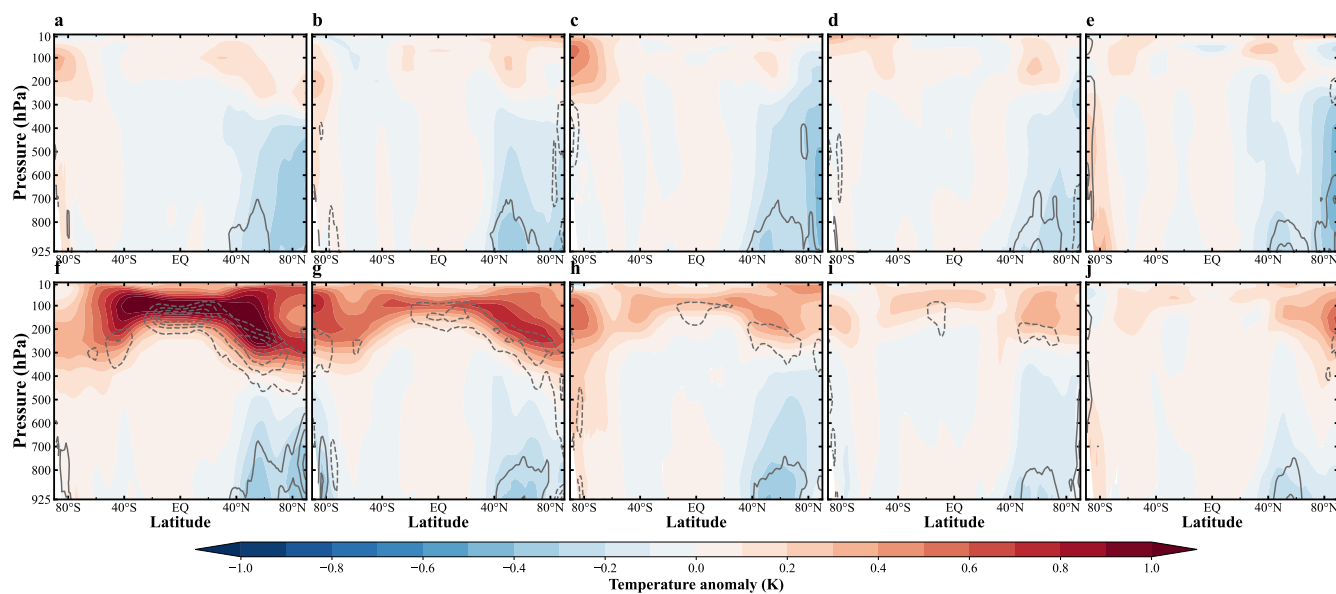


Figure A9. Vertical structure of aerosol-induced temperature and cloud-fraction changes across base-state climates. Zonal-mean atmospheric temperature changes (shading; K) and cloud-fraction changes (contours) in response to present-day aerosol perturbations relative to preindustrial aerosol conditions. Results are shown for GFDL-AM4.0 in the top row and GFDL-AM4-MG2 in the bottom row. From left to right, panels show responses diagnosed under five prescribed base-state climates: -8 K, -4 K, 0 K, +4 K, and +8 K. Solid and dashed contours indicate increases and decreases in cloud fraction, respectively, with a contour interval of 0.005 (0.5 percentage points).

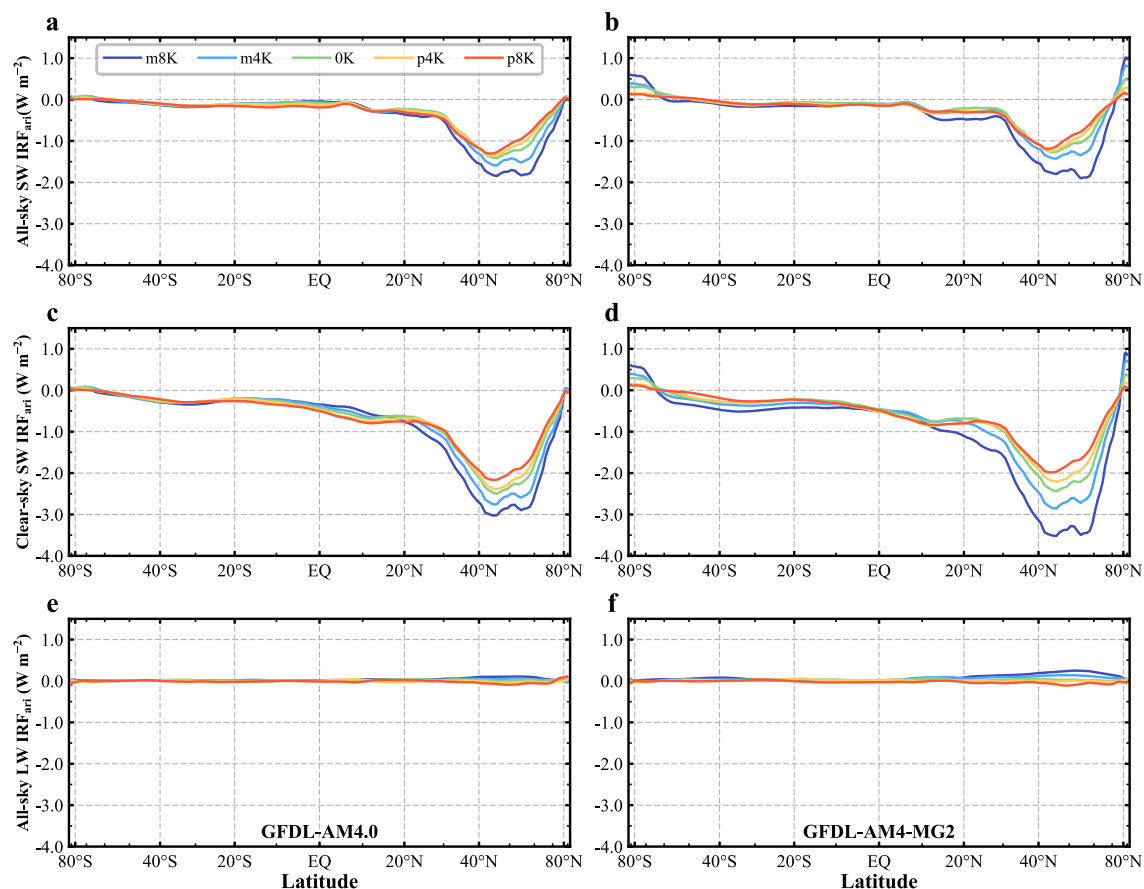


Figure A10. Zonal-mean distributions of IRF_{ari} under different base-state climates. Zonal-mean distributions of instantaneous radiative forcing from aerosol–radiation interactions (IRF_{ari} ; $W m^{-2}$) in response to present-day aerosol perturbations relative to preindustrial aerosol conditions under five prescribed base-state climates: $-8 K$, $-4 K$, $0 K$, $+4 K$, and $+8 K$. Results are shown for GFDL-AM4.0 in the left column and GFDL-AM4-MG2 in the right column. From top to bottom, rows show all-sky shortwave IRF_{ari} , clear-sky shortwave IRF_{ari} , and all-sky longwave IRF_{ari} , respectively. Clear-sky longwave IRF_{ari} is not shown because it differs only slightly from the all-sky longwave IRF_{ari} when a constant cloud-masking value is used. Different colored lines indicate the five base-state climates.

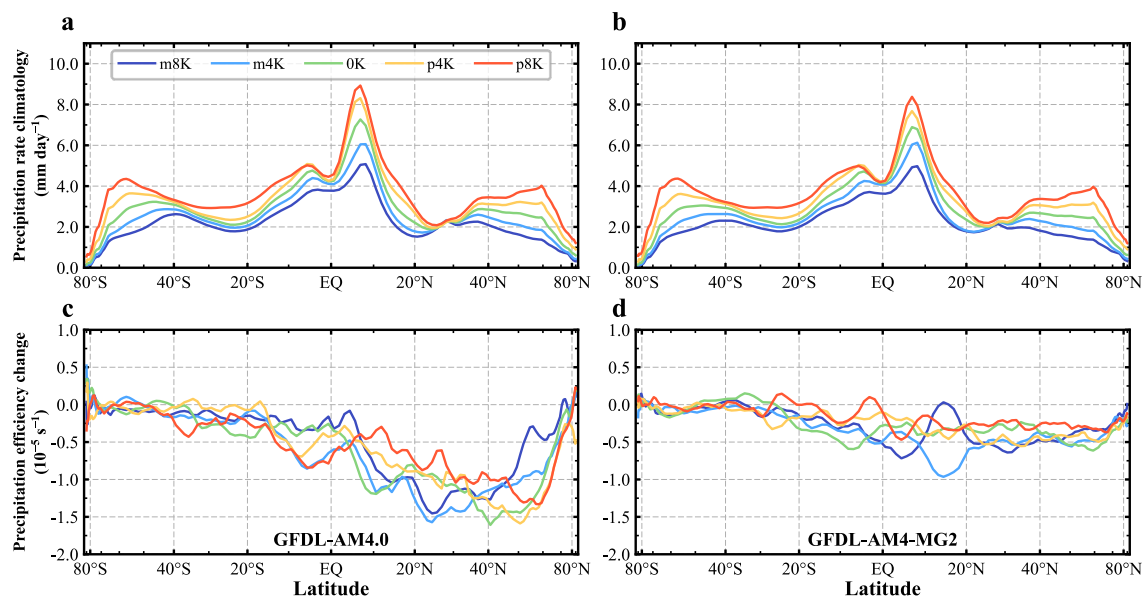


Figure A11. Zonal-mean precipitation climatology and aerosol-induced precipitation-efficiency changes under different base-state climates. Zonal-mean distributions of precipitation climatology without aerosol perturbations and precipitation-efficiency changes in response to present-day aerosol perturbations relative to preindustrial aerosol conditions under five prescribed base-state climates: –8 K, –4 K, 0 K, +4 K, and +8 K. Results are shown for GFDL-AM4.0 in the left column and GFDL-AM4-MG2 in the right column. The top row shows precipitation climatology in the absence of aerosol perturbations, while the bottom row shows aerosol-induced changes in precipitation efficiency. Different colored lines indicate the five base-state climates.

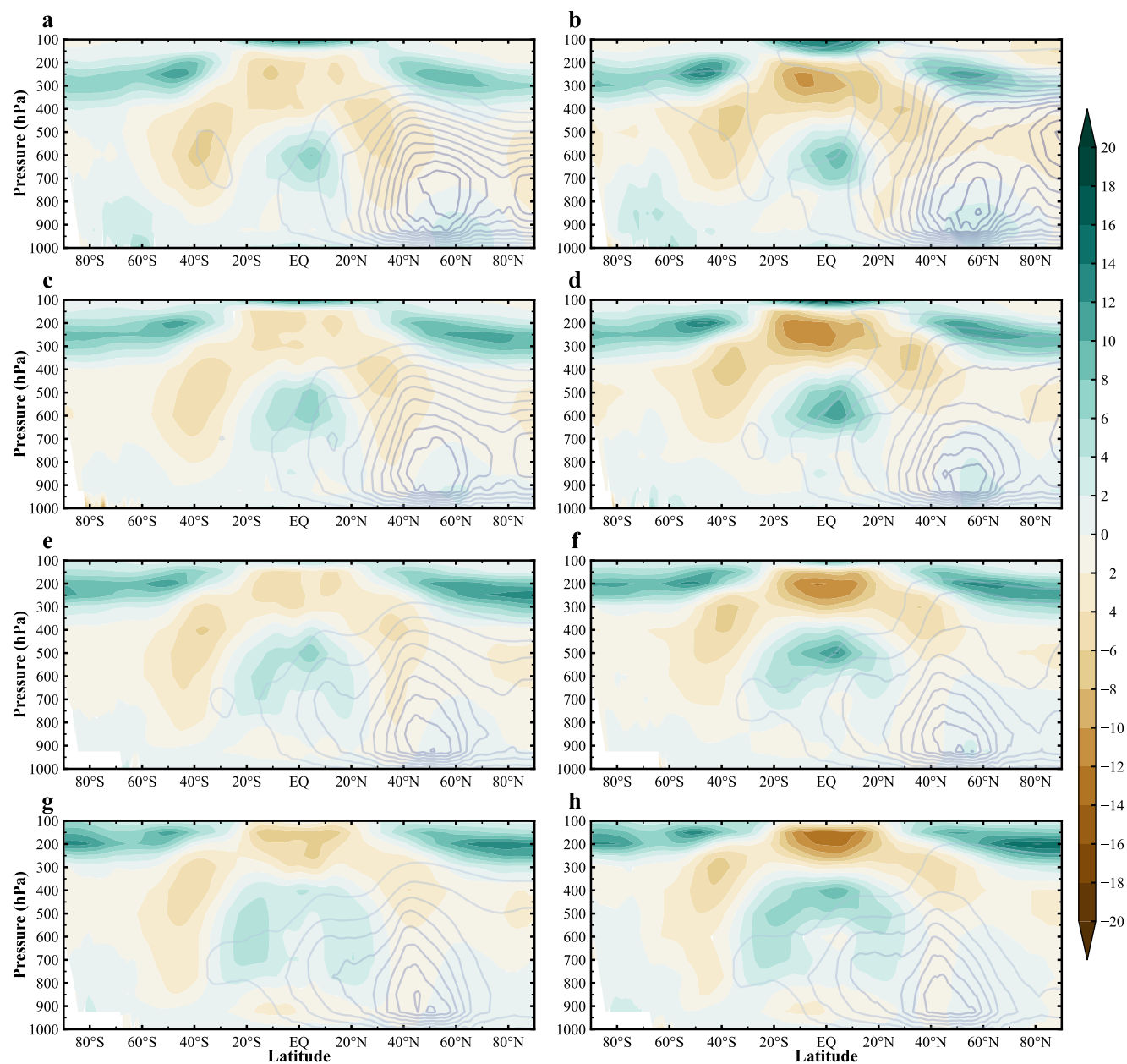


Figure A12. Zonal-mean relative humidity changes associated with uniform 4 K warming across adjacent base-state climates. Zonal-mean changes in relative humidity (%) in response to uniform 4 K warming are evaluated across pairs of adjacent base-state climates. From top to bottom, rows show the transitions from -8 K to -4 K, -4 K to 0 K, 0 K to +4 K, and +4 K to +8 K. Results are shown for GFDL-AM4.0 in the left column and GFDL-AM4-MG2 in the right column. Shading indicates the relative humidity change as a function of latitude and pressure. Semi-transparent contours show the aerosol extinction coefficient anomaly associated with the present-day relative to preindustrial aerosol-emission perturbation, averaged over the two base states defining each 4 K transition.



Code and data availability

660 The processed data and Python scripts used to produce the figures in this study are publicly available on Zenodo (He, 2026; <https://doi.org/10.5281/zenodo.23705667>).

Author contributions

HH performed all simulations and analyses and prepared the initial draft of the manuscript. All authors contributed to reviewing and revising the manuscript and read and approved the final version.

665 Competing interests

Authors have no competing interests.

Acknowledgements

We thank Fei Yang for helpful comments. We also thank Dr. Huan Guo for providing the necessary files and clarifying the namelist modifications required to use GFDL-AM4-MG2.

670 Financial support

H.H., B.J.S., and G.A.V. were supported by Award MPS-SRM-00006584 from the Simons Foundation. R.J.K. acknowledges support from the National Oceanic and Atmospheric Administration, U.S. Department of Commerce. The statements, findings, conclusions, and recommendations are those of the author(s) and do not necessarily reflect the views of the National Oceanic and Atmospheric Administration, or the U.S. Department of Commerce. J.Z. acknowledges the support
675 by the U.S. Department of Commerce, Earth's Radiation Budget grant, NOAA CPO Climate & CI #03-01-07-001 and the NOAA Cooperative Agreement with CIRES, NA22OAR4320151. C.P. was supported by the NASA grant 80NSSC23K0115.

References

Albrecht, B. A.: Aerosols, Cloud Microphysics, and Fractional Cloudiness, *Science*, 245, 1227–1230,
680 <https://doi.org/10.1126/science.245.4923.1227>, 1989.

Bellouin, N., Quaas, J., Gryspeerdt, E., Kinne, S., Stier, P., Watson-Parris, D., Boucher, O., Carslaw, K. S., Christensen, M., Daniau, A. -L., Dufresne, J. -L., Feingold, G., Fiedler, S., Forster, P., Gettelman, A., Haywood, J. M., Lohmann, U.,



- Malavelle, F., Mauritsen, T., McCoy, D. T., Myhre, G., Mülmenstädt, J., Neubauer, D., Possner, A., Rugenstein, M., Sato, Y., Schulz, M., Schwartz, S. E., Sourdeval, O., Storelvmo, T., Toll, V., Winker, D., and Stevens, B.: Bounding Global
685 Aerosol Radiative Forcing of Climate Change, *Rev. Geophys.*, 58, e2019RG000660, <https://doi.org/10.1029/2019rg000660>, 2020.
- Bogenschutz, P. A., Yamaguchi, T., and Lee, H.: The Energy Exascale Earth System Model Simulations With High Vertical Resolution in the Lower Troposphere, *J. Adv. Model. Earth Syst.*, 13, <https://doi.org/10.1029/2020ms002239>, 2021.
- Bony, S., Stevens, B., Coppin, D., Becker, T., Reed, K. A., Voigt, A., and Medeiros, B.: Thermodynamic control of anvil
690 cloud amount, *Proc. Natl. Acad. Sci.*, 113, 8927–8932, <https://doi.org/10.1073/pnas.1601472113>, 2016.
- Boucher, O., Randall, D., Artaxo, P., Bretherton, C., Feingold, G., Forster, P., Kerminen, V.-M., Kondo, Y., Liao, H., Lohmann, U., Rasch, P., Satheesh, S. K., Sherwood, S., Stevens, B., and Zhang, X. Y.: Clouds and Aerosols, in: *Climate Change 2013: The Physical Science Basis. Contribution of Working Group I to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change*, edited by: Stocker, T. F., Qin, D., Plattner, G.-K., Tignor, M., Allen, S. K.,
695 Boschung, J., Nauels, A., Xia, Y., Bex, V., and Midgley, P. M., Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA, 571–659, <https://doi.org/10.1017/cbo9781107415324.016>, 2013.
- Bretherton, C. S., Blossey, P. N., and Uchida, J.: Cloud droplet sedimentation, entrainment efficiency, and subtropical stratocumulus albedo, *Geophys. Res. Lett.*, 34, <https://doi.org/10.1029/2006gl027648>, 2007.
- Fan, S., Ginoux, P., Seman, C. J., Silvers, L. G., and Zhao, M.: Towards Improved Cloud-Phase Simulation with a Mineral
700 Dust and Temperature-Dependent Parameterization for Ice Nucleation in Mixed-Phase Clouds Towards Improved Cloud-Phase Simulation with a Mineral Dust and Temperature-Dependent Parameterization for Ice Nucleation in Mixed-Phase Clouds, *J. Atmos. Sci.*, 76, 3655–3667, <https://doi.org/10.1175/jas-d-18-0287.1>, 2019.
- Feingold, G., Goren, T., and Yamaguchi, T.: Quantifying albedo susceptibility biases in shallow clouds, *Atmos. Chem. Phys.*, 22, 3303–3319, <https://doi.org/10.5194/acp-22-3303-2022>, 2022.
- 705 Feng, C., Liu, X., Zhao, X., Lu, Z., Lin, L., Zelinka, M. D., Qin, Y., Shan, Y., Zheng, Y., Wu, X., and Saravanan, R.: Interconnection of Aerosol-Cloud Interactions and Cloud Feedback Through Warm Rain Process, *J. Geophys. Res.: Atmos.*, 131, <https://doi.org/10.1029/2025jd044854>, 2026.
- Forster, P. M., Richardson, T., Maycock, A. C., Smith, C. J., Samset, B. H., Myhre, G., Andrews, T., Pincus, R., and Schulz, M.: Recommendations for diagnosing effective radiative forcing from climate models for CMIP6, *J. Geophys. Res.: Atmos.*,
710 121, 12,460–12,475, <https://doi.org/10.1002/2016jd025320>, 2016.
- Forster, P., Storelvmo, T., Armour, K., Collins, W., Dufresne, J.-L., Frame, D., Lunt, D. J., Mauritsen, T., Palmer, M. D., Watanabe, M., Wild, M., and Zhang, H.: The Earth’s Energy Budget, Climate Feedbacks and Climate Sensitivity, in: *Climate Change 2021: The Physical Science Basis. Contribution of Working Group I to the Sixth Assessment Report of the*



- Intergovernmental Panel on Climate Change, edited by: Masson-Delmotte, V., Zhai, P., Pirani, A., Connors, S. L., Péan, C.,
 715 Berger, S., Caud, N., Chen, Y., Goldfarb, L., Gomis, M. I., Huang, M., Leitzell, K., Lonnoy, E., Matthews, J. B. R.,
 Maycock, T. K., Waterfield, T., Yelekçi, O., Yu, R., and Zhou, B., Cambridge University Press, Cambridge, United
 Kingdom and New York, NY, USA, 923–1054, <https://doi.org/10.1017/9781009157896.009>, 2021.
- Gettelman, A., Eidhammer, T., Duffy, M. L., McCoy, D. T., Song, C., and Watson-Parris, D.: The Interaction Between
 Climate Forcing and Feedbacks, *J. Geophys. Res.: Atmos.*, 129, <https://doi.org/10.1029/2024jd040857>, 2024.
- 720 Gettelman, A. and Morrison, H.: Advanced Two-Moment Bulk Microphysics for Global Models. Part I: Off-Line Tests and
 Comparison with Other Schemes, *J. Clim.*, 28, 1268–1287, <https://doi.org/10.1175/jcli-d-14-00102.1>, 2015.
- Ghan, S., Wang, M., Zhang, S., Ferrachat, S., Gettelman, A., Griesfeller, J., Kipling, Z., Lohmann, U., Morrison, H.,
 Neubauer, D., Partridge, D. G., Stier, P., Takemura, T., Wang, H., and Zhang, K.: Challenges in constraining anthropogenic
 aerosol effects on cloud radiative forcing using present-day spatiotemporal variability, *Proc. Natl. Acad. Sci.*, 113, 5804–
 725 5811, <https://doi.org/10.1073/pnas.1514036113>, 2016.
- Grandey, B. S. and Wang, C.: Background Conditions Influence the Estimated Cloud Radiative Effects of Anthropogenic
 Aerosol Emissions From Different Source Regions, *J. Geophys. Res.: Atmos.*, 124, 2276–2295,
<https://doi.org/10.1029/2018jd029644>, 2019.
- Gryspeerd, E., Goren, T., Sourdeval, O., Quaas, J., Mülmenstädt, J., Dipu, S., Unglaub, C., Gettelman, A., and Christensen,
 730 M.: Constraining the aerosol influence on cloud liquid water path, *Atmos. Chem. Phys.*, 19, 5331–5347,
<https://doi.org/10.5194/acp-19-5331-2019>, 2019.
- Guo, H., Ming, Y., Fan, S., Zhou, L., Harris, L., and Zhao, M.: Two-Moment Bulk Cloud Microphysics With Prognostic
 Precipitation in GFDL’s Atmosphere Model AM4.0: Configuration and Performance, *J. Adv. Model. Earth Syst.*, 13,
<https://doi.org/10.1029/2020ms002453>, 2021.
- 735 Hansen, J., Sato, M., Ruedy, R., Nazarenko, L., Lacis, A., Schmidt, G. A., Russell, G., Aleinov, I., Bauer, M., Bauer, S.,
 Bell, N., Cairns, B., Canuto, V., Chandler, M., Cheng, Y., Genio, A. D., Faluvegi, G., Fleming, E., Friend, A., Hall, T.,
 Jackman, C., Kelley, M., Kiang, N., Koch, D., Lean, J., Lerner, J., Lo, K., Menon, S., Miller, R., Minnis, P., Novakov, T.,
 Oinas, V., Perlwitz, Ja., Perlwitz, Ju., Rind, D., Romanou, A., Shindell, D., Stone, P., Sun, S., Tausnev, N., Thresher, D.,
 Wielicki, B., Wong, T., Yao, M., and Zhang, S.: Efficacy of climate forcings, *J. Geophys. Res.: Atmos.*, 110,
 740 <https://doi.org/10.1029/2005jd005776>, 2005.
- Haywood, J. M. and Shine, K. P.: The effect of anthropogenic sulfate and soot aerosol on the clear sky planetary radiation
 budget, *Geophys. Res. Lett.*, 22, 603–606, <https://doi.org/10.1029/95gl00075>, 1995.



- Haywood, J. M., Roberts, D. L., Slingo, A., Edwards, J. M., and Shine, K. P.: General Circulation Model Calculations of the Direct Radiative Forcing by Anthropogenic Sulfate and Fossil-Fuel Soot Aerosol, *J. Clim.*, 10, 1562–1577, 745 [https://doi.org/10.1175/1520-0442\(1997\)010<1562:gcmcot>2.0.co;2](https://doi.org/10.1175/1520-0442(1997)010<1562:gcmcot>2.0.co;2), 1997.
- He, H.: Data and scripts used to produce the figures for “State dependence of aerosol forcing and its implications for climate response”, Zenodo [data set], <https://doi.org/10.5281/zenodo.23705667>, 2026.
- He, H., Soden, B. J., Kramer, R. J., Vecchi, G. A., Zhang, J., and Park, C.: Technical note: Aerosol forcing decomposition methods and recommendations, 2026.
- 750 He, H., Soden, B., Zhang, B., Vecchi, G., and Yang, W.: State dependence of cloud feedback and its implications for climate sensitivity, <https://doi.org/10.21203/rs.3.rs-5220475/v1>, 2024.
- Medeiros, B., Williamson, D. L., Hannay, C., and Olson, J. G.: Southeast Pacific Stratocumulus in the Community Atmosphere Model, *J. Clim.*, 25, 6175–6192, <https://doi.org/10.1175/jcli-d-11-00503.1>, 2012.
- Meehl, G. A., Senior, C. A., Eyring, V., Flato, G., Lamarque, J.-F., Stouffer, R. J., Taylor, K. E., and Schlund, M.: Context 755 for interpreting equilibrium climate sensitivity and transient climate response from the CMIP6 Earth system models, *Sci. Adv.*, 6, eaba1981, <https://doi.org/10.1126/sciadv.aba1981>, 2020.
- Mellado, J. P.: Cloud-Top Entrainment in Stratocumulus Clouds, *Annu. Rev. Fluid Mech.*, 49, 1–25, <https://doi.org/10.1146/annurev-fluid-010816-060231>, 2016.
- Muri, H., Kristjánsson, J. E., Storelvmo, T., and Pfeffer, M. A.: The climatic effects of modifying cirrus clouds in a climate 760 engineering framework, *J. Geophys. Res.: Atmos.*, 119, 4174–4191, <https://doi.org/10.1002/2013jd021063>, 2014.
- Kramer, R. J., Smith, C., and Andrews, T.: The Radiative Forcing Model Intercomparison Project (RFMIP2.0) for CMIP7, *Geosci. Model Dev.*, 19, 4447–4466, <https://doi.org/10.5194/gmd-19-4447-2026>, 2026.
- Li, R. L., Studholme, J. H. P., Fedorov, A. V., and Storelvmo, T.: Precipitation efficiency constraint on climate change, *Nat. Clim. Chang.*, 12, 642–648, <https://doi.org/10.1038/s41558-022-01400-x>, 2022.
- 765 Lorian, S. and Dagan, G.: On the sensitivity of aerosol–cloud interactions to changes in sea surface temperature in radiative–convective equilibrium, *Atmos. Chem. Phys.*, 24, 9323–9338, <https://doi.org/10.5194/acp-24-9323-2024>, 2024.
- Myhre, G., Shindell, D., Bréon, F.-M., Collins, W., Fuglestedt, J., Huang, J., Koch, D., Lamarque, J.-F., Lee, D., Mendoza, B., Nakajima, T., Robock, A., Stephens, G., Takemura, T., and Zhang, H.: Anthropogenic and Natural Radiative Forcing, in: Climate Change 2013: The Physical Science Basis. Contribution of Working Group I to the Fifth Assessment Report of the 770 Intergovernmental Panel on Climate Change, edited by: Stocker, T. F., Qin, D., Plattner, G.-K., Tignor, M., Allen, S. K., Boschung, J., Nauels, A., Xia, Y., Bex, V., and Midgley, P. M., Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA, 659–740, <https://doi.org/10.1017/cbo9781107415324.018>, 2013.



- Nemesure, S., Wagener, R., and Schwartz, S. E.: Direct shortwave forcing of climate by the anthropogenic sulfate aerosol: Sensitivity to particle size, composition, and relative humidity, *J. Geophys. Res.: Atmos.*, 100, 26105–26116,
 775 <https://doi.org/10.1029/95jd02897>, 1995.
- Pilinis, C., Pandis, S. N., and Seinfeld, J. H.: Sensitivity of direct climate forcing by atmospheric aerosols to aerosol size and composition, *J. Geophys. Res.: Atmos.*, 100, 18739–18754, <https://doi.org/10.1029/95jd02119>, 1995.
- Platnick, S. and Oreopoulos, L.: Radiative susceptibility of cloudy atmospheres to droplet number perturbations: 1. Theoretical analysis and examples from MODIS, *J. Geophys. Res.: Atmos.*, 113, <https://doi.org/10.1029/2007jd009654>,
 780 2008.
- Platnick, S. and Twomey, S.: Determining the Susceptibility of Cloud Albedo to Changes in Droplet Concentration with the Advanced Very High Resolution Radiometer, *J. Appl. Meteorol.*, 33, 334–347, [https://doi.org/10.1175/1520-0450\(1994\)033<0334:dtsoa>2.0.co;2](https://doi.org/10.1175/1520-0450(1994)033<0334:dtsoa>2.0.co;2), 1994.
- Pincus, R. and Baker, M. B.: Effect of precipitation on the albedo susceptibility of clouds in the marine boundary layer,
 785 *Nature*, 372, 250–252, <https://doi.org/10.1038/372250a0>, 1994.
- Pincus, R., Forster, P. M., and Stevens, B.: The Radiative Forcing Model Intercomparison Project (RFMIP): experimental protocol for CMIP6, *Geosci. Model Dev.*, 9, 3447–3460, <https://doi.org/10.5194/gmd-9-3447-2016>, 2016.
- Qin, Y., Ma, P., Zelinka, M. D., Klein, S. A., Zhang, T., Zheng, X., Larson, V. E., and Huang, M.: Impact of Turbulence Representation on the Relationship Between Cloud Feedback and Aerosol-Cloud Interaction in an E3SMv2 Perturbed
 790 Parameter Ensemble, *J. Adv. Model. Earth Syst.*, 17, <https://doi.org/10.1029/2024ms004756>, 2025.
- Rasch, P. J., Feichter, J., Law, K., Mahowald, N., Penner, J., Benkovitz, C., Genthon, C., Giannakopoulos, C., Kasibhatla, P., Koch, D., Levy, H., Maki, T., Prather, M., Roberts, D. L., Roelofs, G.-J., Stevenson, D., Stockwell, Z., Taguchi, S., Kritz, M., Chipperfield, M., Baldocchi, D., McMurry, P., Barrie, L., Balkanski, Y., Chatfield, R., Kjellstrom, E., Lawrence, M., Lee, H. N., Lelieveld, J., Noone, K. J., Seinfeld, J., Stenchikov, G., Schwartz, S., Walcek, C., and Williamson, D.: A
 795 comparison of scavenging and deposition processes in global models: results from the WCRP Cambridge Workshop of 1995, *Tellus B*, 52, 1025–1056, <https://doi.org/10.3402/tellusb.v52i4.17091>, 2000.
- Sherwood, S. C., Webb, M. J., Annan, J. D., Armour, K. C., Forster, P. M., Hargreaves, J. C., Hegerl, G., Klein, S. A., Marvel, K. D., Rohling, E. J., Watanabe, M., Andrews, T., Braconnot, P., Bretherton, C. S., Foster, G. L., Hausfather, Z., Heydt, A. S. von der, Knutti, R., Mauritsen, T., Norris, J. R., Proistosescu, C., Rugenstein, M., Schmidt, G. A., Tokarska, K.
 800 B., and Zelinka, M. D.: An Assessment of Earth’s Climate Sensitivity Using Multiple Lines of Evidence, *Rev. Geophys.*, 58, e2019RG000678, <https://doi.org/10.1029/2019rg000678>, 2020.
- Smith, C. J., Kramer, R. J., and Sima, A.: The HadGEM3-GA7.1 radiative kernel: the importance of a well-resolved stratosphere, *Earth Syst. Sci. Data*, 12, 2157–2168, <https://doi.org/10.5194/essd-12-2157-2020>, 2020.



- Soden, B. J., Held, I. M., Colman, R., Shell, K. M., Kiehl, J. T., and Shields, C. A.: Quantifying Climate Feedbacks Using Radiative Kernels, *J. Clim.*, 21, 3504–3520, <https://doi.org/10.1175/2007jcli2110.1>, 2008.
- Rossow, W. B. and Schiffer, R. A.: Advances in Understanding Clouds from ISCCP, *Bull. Am. Meteorol. Soc.*, 80, 2261–2287, [https://doi.org/10.1175/1520-0477\(1999\)080<2261:aiucfi>2.0.co;2](https://doi.org/10.1175/1520-0477(1999)080<2261:aiucfi>2.0.co;2), 1999.
- Sorooshian, A., Feingold, G., Lebsock, M. D., Jiang, H., and Stephens, G. L.: On the precipitation susceptibility of clouds to aerosol perturbations, *Geophys. Res. Lett.*, 36, <https://doi.org/10.1029/2009gl038993>, 2009.
- 810 Stevens, B.: Rethinking the Lower Bound on Aerosol Radiative Forcing, *J. Clim.*, 28, 4794–4819, <https://doi.org/10.1175/jcli-d-14-00656.1>, 2015.
- Storelvmo, T., Boos, W. R., and Herger, N.: Cirrus cloud seeding: a climate engineering mechanism with reduced side effects?, *Philos. Trans. R. Soc. A: Math., Phys. Eng. Sci.*, 372, 20140116, <https://doi.org/10.1098/rsta.2014.0116>, 2014.
- Storelvmo, T.: Aerosol Effects on Climate via Mixed-Phase and Ice Clouds, *Annu. Rev. Earth Planet. Sci.*, 45, 1–24, <https://doi.org/10.1146/annurev-earth-060115-012240>, 2017.
- 815 Szopa, S., Naik, V., Adhikary, B., Artaxo, P., Berntsen, T., Collins, W. D., Fuzzi, S., Gallardo, L., Kiendler-Scharr, A., Klimont, Z., Liao, H., Unger, N., and Zanis, P.: Short-Lived Climate Forcers, in: *Climate Change 2021: The Physical Science Basis. Contribution of Working Group I to the Sixth Assessment Report of the Intergovernmental Panel on Climate Change*, edited by: Masson-Delmotte, V., Zhai, P., Pirani, A., Connors, S. L., Péan, C., Berger, S., Caud, N., Chen, Y., Goldfarb, L., Gomis, M. I., Huang, M., Leitzell, K., Lonnoy, E., Matthews, J. B. R., Maycock, T. K., Waterfield, T., Yelekçi, O., Yu, R., and Zhou, B., Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA, 817–922, <https://doi.org/10.1017/9781009157896.008>, 2021.
- Tang, I. N. and Munkelwitz, H. R.: Water activities, densities, and refractive indices of aqueous sulfates and sodium nitrate droplets of atmospheric importance, *J. Geophys. Res.: Atmos.*, 99, 18801–18808, <https://doi.org/10.1029/94jd01345>, 1994.
- 825 Taylor, K. E., Crucifix, M., Braconnot, P., Hewitt, C. D., Doutriaux, C., Broccoli, A. J., Mitchell, J. F. B., and Webb, M. J.: Estimating Shortwave Radiative Forcing and Response in Climate Models, *J. Clim.*, 20, 2530–2543, <https://doi.org/10.1175/jcli4143.1>, 2007.
- Textor, C., Schulz, M., Guibert, S., Kinne, S., Balkanski, Y., Bauer, S., Berntsen, T., Berglen, T., Boucher, O., Chin, M., Dentener, F., Diehl, T., Easter, R., Feichter, H., Fillmore, D., Ghan, S., Ginoux, P., Gong, S., Grini, A., Hendricks, J., Horowitz, L., Huang, P., Isaksen, I., Iversen, I., Kloster, S., Koch, D., Kirkevåg, A., Kristjansson, J. E., Krol, M., Lauer, A., Lamarque, J. F., Liu, X., Montanaro, V., Myhre, G., Penner, J., Pitari, G., Reddy, S., Seland, Ø., Stier, P., Takemura, T., and Tie, X.: Analysis and quantification of the diversities of aerosol life cycles within AeroCom, *Atmos. Chem. Phys.*, 6, 1777–1813, <https://doi.org/10.5194/acp-6-1777-2006>, 2006.



- Twomey, S.: Pollution and the planetary albedo, *Atmos. Environ.*, 8, 1251–1256, [https://doi.org/10.1016/0004-6981\(74\)90004-3](https://doi.org/10.1016/0004-6981(74)90004-3), 1974.
- Twomey, S.: The Influence of Pollution on the Shortwave Albedo of Clouds, *J. Atmos. Sci.*, 34, 1149–1152, [https://doi.org/10.1175/1520-0469\(1977\)034<1149:tiopot>2.0.co;2](https://doi.org/10.1175/1520-0469(1977)034<1149:tiopot>2.0.co;2), 1977.
- Wang, C., Soden, B. J., Yang, W., and Vecchi, G. A.: Compensation Between Cloud Feedback and Aerosol-Cloud Interaction in CMIP6 Models, *Geophys. Res. Lett.*, 48, <https://doi.org/10.1029/2020gl091024>, 2021.
- Watson-Parris, D. and Smith, C. J.: Large uncertainty in future warming due to aerosol forcing, *Nat. Clim. Chang.*, 12, 1111–1113, <https://doi.org/10.1038/s41558-022-01516-0>, 2022.
- White, K., Liu, D., and Persad, G.: Absence of Aerosol Indirect Effect Dependence on Background Climate State in NCAR CESM2, *J. Clim.*, 38, 147–163, <https://doi.org/10.1175/jcli-d-23-0755.1>, 2025.
- Wilcox, L. J., Highwood, E. J., Booth, B. B. B., and Carslaw, K. S.: Quantifying sources of inter-model diversity in the cloud albedo effect, *Geophys. Res. Lett.*, 42, 1568–1575, <https://doi.org/10.1002/2015gl063301>, 2015.
- Xue, H. and Feingold, G.: Large-Eddy Simulations of Trade Wind Cumuli: Investigation of Aerosol Indirect Effects, *J. Atmos. Sci.*, 63, 1605–1622, <https://doi.org/10.1175/jas3706.1>, 2006.
- Zelinka, M. D., Myers, T. A., McCoy, D. T., Po-Chedley, S., Caldwell, P. M., Ceppi, P., Klein, S. A., and Taylor, K. E.: Causes of Higher Climate Sensitivity in CMIP6 Models, *Geophys. Res. Lett.*, 47, <https://doi.org/10.1029/2019gl085782>, 2020.
- Zelinka, M. D., Klein, S. A., and Hartmann, D. L.: Computing and Partitioning Cloud Feedbacks Using Cloud Property Histograms. Part I: Cloud Radiative Kernels, *J. Clim.*, 25, 3715–3735, <https://doi.org/10.1175/jcli-d-11-00248.1>, 2012a.
- Zelinka, M. D., Klein, S. A., and Hartmann, D. L.: Computing and Partitioning Cloud Feedbacks Using Cloud Property Histograms. Part II: Attribution to Changes in Cloud Amount, Altitude, and Optical Depth, *J. Clim.*, 25, 3736–3754, <https://doi.org/10.1175/jcli-d-11-00249.1>, 2012b.
- Zelinka, M. D., Klein, S. A., Taylor, K. E., Andrews, T., Webb, M. J., Gregory, J. M., and Forster, P. M.: Contributions of Different Cloud Types to Feedbacks and Rapid Adjustments in CMIP5*, *J. Clim.*, 26, 5007–5027, <https://doi.org/10.1175/jcli-d-12-00555.1>, 2013.
- Zhang, J., Zhou, X., Goren, T., and Feingold, G.: Albedo susceptibility of northeastern Pacific stratocumulus: the role of covarying meteorological conditions, *Atmos. Chem. Phys.*, 22, 861–880, <https://doi.org/10.5194/acp-22-861-2022>, 2022.
- Zhang, S., Naik, V., Horowitz, L., and Paynter, D.: The Dependence of Emissions-Based Anthropogenic Aerosol ERFs on Background Composition and Meteorological States in the GFDL-ESM4.1 Model, *J. Geophys. Res.: Atmos.*, 131, <https://doi.org/10.1029/2026jd046386>, 2026.



865 Zhao, M., Golaz, J. -C., Held, I. M., Guo, H., Balaji, V., Benson, R., Chen, J. -H., Chen, X., Donner, L. J., Dunne, J. P.,
Dunne, K., Durachta, J., Fan, S. -M., Freidenreich, S. M., Garner, S. T., Ginoux, P., Harris, L. M., Horowitz, L. W.,
Krasting, J. P., Langenhorst, A. R., Liang, Z., Lin, P., Lin, S. -J., Malyshev, S. L., Mason, E., Milly, P. C. D., Ming, Y.,
Naik, V., Paulot, F., Paynter, D., Phillipps, P., Radhakrishnan, A., Ramaswamy, V., Robinson, T., Schwarzkopf, D., Seman,
C. J., Shevliakova, E., Shen, Z., Shin, H., Silvers, L. G., Wilson, J. R., Winton, M., Wittenberg, A. T., Wyman, B., and
870 Xiang, B.: The GFDL Global Atmosphere and Land Model AM4.0/LM4.0: 1. Simulation Characteristics With Prescribed
SSTs, *J. Adv. Model. Earth Syst.*, 10, 691–734, <https://doi.org/10.1002/2017ms001208>, 2018a.

Zhao, M., Golaz, J. -C., Held, I. M., Guo, H., Balaji, V., Benson, R., Chen, J. -H., Chen, X., Donner, L. J., Dunne, J. P.,
Dunne, K., Durachta, J., Fan, S. -M., Freidenreich, S. M., Garner, S. T., Ginoux, P., Harris, L. M., Horowitz, L. W.,
Krasting, J. P., Langenhorst, A. R., Liang, Z., Lin, P., Lin, S. -J., Malyshev, S. L., Mason, E., Milly, P. C. D., Ming, Y.,
Naik, V., Paulot, F., Paynter, D., Phillipps, P., Radhakrishnan, A., Ramaswamy, V., Robinson, T., Schwarzkopf, D., Seman,
875 C. J., Shevliakova, E., Shen, Z., Shin, H., Silvers, L. G., Wilson, J. R., Winton, M., Wittenberg, A. T., Wyman, B., and
Xiang, B.: The GFDL Global Atmosphere and Land Model AM4.0/LM4.0: 2. Model Description, Sensitivity Studies, and
Tuning Strategies, *J. Adv. Model. Earth Syst.*, 10, 735–769, <https://doi.org/10.1002/2017ms001209>, 2018b.