

Differences between daytime and nighttime in the impacts of anthropogenic aerosol-radiative-cloud interaction on the summer precipitation over the mid-lower Yangtze River Basin

Zhuohui Wang^a, Shu Li^{a,*}, Anning Huang^a, Xinsheng Zhu^{b,**}, Tijian Wang^a, Bingliang Zhuang^a, Mengmeng Li^a, Rongchang Wu^a

^a School of Atmospheric Sciences, Nanjing University, Nanjing, 210023, China

^b Nanjing Institute of Environmental Sciences, Ministry of Ecology and Environment of the People's Republic of China, Nanjing, 210033, China

HIGHLIGHTS

- RegCM5 demonstrates stronger ACI impacts than ARI on summer rainfall in MLYRB.
- RegCM5 shows stronger daytime than nighttime impacts of ARI and ACI on MLYRB rainfall.
- ARI cools surface, weakening daytime convection and local moisture supply.
- ACI increases high clouds and droplet number, suppressing daytime light rain.

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ABSTRACT

Aerosol-Radiation Interaction (ARI) and Aerosol-Cloud Interaction (ACI) significantly affect radiation budgets and hydrological cycles. However, quantifying the contribution of anthropogenic aerosols to regional precipitation remains a major challenge. In this study, utilizing the Regional Climate Model version 5 (RegCM5), we unraveled the mechanisms related to the disparities between daytime and nighttime in the impact of the anthropogenic aerosols on the summer precipitation over the middle and lower Yangtze River Basin (MLYRB). The results indicate that the ARI decreases the daily total precipitation amount (PA) by 0.05 mm (12 h)⁻¹ and precipitation frequency (PF) by 0.74 %. This suppression occurs primarily through a reduction in daytime convective precipitation, driven by radiative cooling (up to 14.02 W m⁻² reduction in surface net shortwave radiation) that enhances lower-atmospheric stability, manifested as a significant suppression of the planetary boundary layer height (PBLH), leading to moisture divergence and widespread subsidence. Compared to ARI, ACI exerts a stronger suppression, reducing daily total PA (PF) by 0.16 mm (12 h)⁻¹ (1.91 %) mainly by reducing daytime light to moderate convective precipitation. This is attributed to increased high cloud cover (3.89 %), which reduces surface net radiation and profoundly enhances atmospheric stability, quantified by a significant reduction in Convective Available Potential Energy (CAPE), thereby weakening ascending motion. This study quantifies that ACI has a stronger impact than ARI, and both effects are significantly more pronounced during the daytime, revealing the critical role of diurnal processes in modulating aerosol-precipitation interactions.

1. Introduction

Anthropogenic aerosols, characterized as critical atmospheric components, possess complex chemical compositions and short atmospheric

residence times. Their spatiotemporal distributions exhibit significant heterogeneity due to emission source variability, contributing to substantial uncertainties in climate change projections while exerting a profound impact on both the environment and public health

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* Corresponding author.

** Corresponding author.

E-mail addresses: lishu@nju.edu.cn (S. Li), zhuxinsheng@nies.org (X. Zhu).

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(Hodnebrog et al., 2024; Lelieveld et al., 2019; Li et al., 2022; Rosenfeld et al., 2014; Tan et al., 2023).

Aerosols significantly modulate the earth climate system by altering radiative budgets and modulating the hydrological cycle through two dominant mechanisms (Li et al., 2019; Mahowald et al., 2011; Menon, 2004): Aerosol-radiation interaction (ARI) alters atmospheric albedo through scattering and absorption of solar radiation, thereby reducing surface solar irradiance. This modification of the surface-atmosphere radiative balance subsequently modulates precipitation patterns (Dagan et al., 2021; Hong et al., 2020; Zhao et al., 2024). Aerosol-cloud interaction (ACI) arises from aerosols acting as cloud condensation nuclei (CCN), which influence cloud optical properties and lifetime, ultimately affecting precipitation efficiency (Boucher et al., 2013; Penner et al., 2004). These distinct effects collectively drive regional heterogeneity in aerosol-induced climate responses (Fan et al., 2012; Huang et al., 2014; Tao et al., 2012).

Aerosol effects on precipitation encompass various aspects, including precipitation amount (PA), frequency (PF), and intensity (PI), with their impacts varying based on aerosol properties and regional meteorological conditions (Alizadeh-Chooari and Gharaylou, 2017; Liu et al., 2020a; Matho Lontio et al., 2024; Stjern et al., 2011; Wang, 2013). Several studies have shown that ARI and ACI lead to a reduction in precipitation in East Asia, which is attributed to increased thermodynamic stability (Guo et al., 2017; Huang et al., 2007, 2016; Jiang et al., 2013; Zhao et al., 2006). Specifically, ARI reduces the surface solar radiation, thereby suppressing convection and consequently decreasing the precipitation rate and PF (Guo et al., 2016; Lee et al., 2016). Similarly, increased aerosols have been considered as the primary cause of reduced light precipitation in Eastern China, with ACI playing a more significant role than ARI (Wang et al., 2016; Xu et al., 2017), while the PF of heavy precipitation tends to increase (Alizadeh-Chooari, 2018; Han et al., 2022; Shao et al., 2022). In terms of precipitation type, convective precipitation is more sensitive to aerosol effects than stratiform precipitation (Sun et al., 2022a; Sun and Zhao, 2021). Moreover, aerosol effects on precipitation are also closely related to local meteorological conditions, such as relative humidity, the stability in the low-troposphere, wind shear, and Convective Available Potential Energy (CAPE) (Tang et al., 2025; Da Silva et al., 2020; Bai et al., 2020; Lee et al., 2008).

Most existing studies have focused on the aerosol effects on daily-scale precipitation, yet their impacts on the diurnal variations of precipitation remain underexplored and warrant further investigation. Previous researches indicate that the aerosol effects show distinct regional and temporal differences (Liu et al., 2022; Wei et al., 2022). During the daytime, absorbing aerosols advance the precipitation peak time by 1–3 h in Northern China through low-level heating and convective invigoration. Conversely, scattering aerosols delay the precipitation from afternoon to evening in Southern China by enhancing atmospheric stability (Guo et al., 2016; Sun and Zhao, 2021; Zhang et al., 2025; Zheng et al., 2020; Zhou et al., 2020). Gao and Zhang (2018) further reveal that anthropogenic aerosols cause a notable decrease in precipitation across both Southern and Northern China. At nighttime, dust aerosols promote PI through ACI, counteracting daytime suppression (Wang et al., 2024). Furthermore, the thermodynamic feedbacks further intensify night precipitation in Northern China, as a reduced diurnal temperature range strengthens nocturnal moisture convergence (Lin and Chen, 2023). An increase in nocturnal precipitation over Central China has also been reported (Gao and Zhang, 2018). Mechanistically, a reduction in scattering aerosols during the day weakens ARI-induced cooling, which decreases atmospheric stability and enhances PF and PI. In contrast, an increase in absorbing aerosols strengthens ARI-induced heating, stabilizing the lower atmosphere and thus suppressing daytime precipitation. At nighttime, absorbing aerosols continue to suppress precipitation through similar ARI mechanisms, whereas the influence of scattering aerosols is considerably weaker (Li et al., 2024). Yang et al. (2012) also demonstrated that increased

anthropogenic aerosols reduce precipitation by enhancing cloud albedo and suppressing drizzle formation, which ultimately lowers PF and PI. However, researches on the day-night differences in aerosol impacts on precipitation and associated mechanisms remain limited.

The middle and lower Yangtze River Basin (MLYRB), as one of China's most economically advanced regions, has seen a marked rise in aerosol concentrations due to rapid urbanization, especially within the Yangtze River Delta (He et al., 2018; Peng et al., 2025; Zhang et al., 2018; Zhong et al., 2017). In this region, anthropogenic emissions overwhelmingly dominate regional aerosol concentrations, far exceeding natural aerosol levels (An et al., 2021; Chen et al., 2020; Kang et al., 2019; Xie et al., 2019). Governed by the East Asian monsoon, the summer precipitation in MLYRB exhibits unique features of diurnal variation, with daytime precipitation exceeding that observed at night (Wu et al., 2025; Yao et al., 2022; Chen et al., 2010; Yin et al., 2009; Ding, 1992). Recent studies indicate that the anthropogenic aerosols reduce light precipitation frequency but intensify extreme precipitation occurrences in the MLYRB (Han et al., 2023; Luo et al., 2024). However, previous studies mostly focused on the impact of total aerosol effect on precipitation, with few distinguishing the roles of the two aerosol effects and their differences (Risser et al., 2024; Koren et al., 2012; Stevens and Feingold, 2009). Moreover, most researches remain case-based, with a notable lack of studies analyzing the competition mechanism between the impact of ARI and ACI on summer precipitation in the MLYRB from a long-term climate perspective (Bai et al., 2020; Wang et al., 2011; Zhang et al., 2021; Zhou et al., 2024).

This study employs the Regional Climate Model version 5 (RegCM5) to systematically investigate the impacts of ARI and ACI induced by the anthropogenic aerosols on summer precipitation over the MLYRB. It aims to address three key scientific questions: (1) What are the respective contributions of ARI and ACI to changes in PA and PF in this region? (2) How do the impacts of ARI and ACI differ between daytime and nighttime? (3) What are the associated mechanisms? By tackling these questions, this study provides new quantitative insights into the competing roles of ARI and ACI in modulating regional precipitation characteristics from a long-term climate perspective.

2. Data, model, experimental design, and methods

2.1. Data

The data used in this study are listed as follows.

1. The hourly precipitation data observed at 300 gauge stations, which are uniformly distributed across the MLYRB during the summers (June to August), were quality-controlled and provided by the China Meteorological Administration (CMA). These station observations offer reliable representations of the spatial and temporal characteristics of precipitation over the MLYRB for the study period.
2. The ERA5 dataset with a temporal resolution of 1 h and spatial resolution of $0.25^\circ \times 0.25^\circ$ from 00:00 UTC 1 June to 23:30 UTC 31 August in each year of 2009–2018 (Hersbach et al., 2020) produced by the European Center for Medium-Range Weather Forecasts (ECMWF), includes wind components (u and v), air temperature, and specific humidity at different levels.
3. The daily aerosol optical depth (AOD) data at 550 nm are obtained from MODIS aboard the Terra platform (MOD08_D3 v6.1), with a spatial resolution of $1^\circ \times 1^\circ$ from 1 June to 31 August in each year of 2009–2018, derived from the combined Dark Target and Deep Blue algorithms (Levy et al., 2013; Sayer et al., 2013; Wei et al., 2019).
4. The cloud data uses the CLARA-A3 version of the cloud product from the Ultra High Resolution Radiometer (AVHRR) with a daily temporal resolution and a spatial resolution of $0.25^\circ \times 0.25^\circ$ from 1 June to 31 August in each year of 2009–2018 (Karlsson et al., 2023).

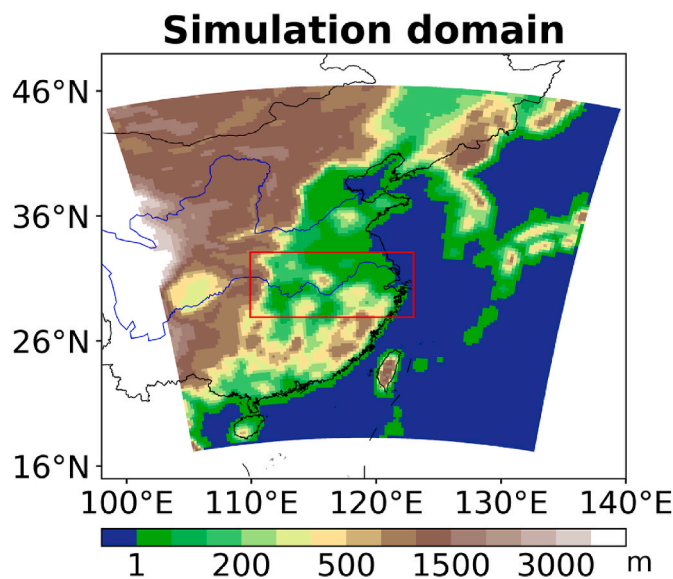


Fig. 1. Study domain and its topography (shaded), the region of MLYRB is indicated by the red box. (For interpretation of the references to colour in this figure legend, the reader is referred to the Web version of this article.)

5. The 6-hourly ERA-Interim reanalysis data (Dee et al., 2011) with the horizontal resolution of $1.5^\circ \times 1.5^\circ$ from June to August in each year of 2009–2018 provided by ECMWF.
6. The weekly Optimum Interpolation Sea Surface Temperature (SST) data from the National Oceanic and Atmospheric Administration (NOAA), with a spatial resolution of $1^\circ \times 1^\circ$, covering June to August for each year from 2009 to 2018 (Reynolds et al., 2002)
7. The monthly anthropogenic emission data from June to August in each year of 2009–2018 are sourced from the International Institute for Applied Systems Analysis (IIASA) emission inventory at a horizontal resolution of $0.5^\circ \times 0.5^\circ$ (Klimont et al., 2017).

2.2. RegCM5 model

The model used in this study is the latest version 5.0 of the regional climate model (RegCM5), which has been developed through a collaboration between the Abdus Salam International Center for Theoretical Physics (ICTP) and the Institute of Atmospheric Sciences and Climate (ISAC) of the National Research Council (CNR) in Italy (Giorgi et al., 1993, 2012, 2023; Pal et al., 2007). One major improvement of RegCM5 over RegCM4 is the integration of a non-hydrostatic dynamical core derived from the MOLOCH weather prediction model, which enhances both accuracy and computational efficiency, particularly at convection-permitting resolutions of a few kilometers. This new dynamical core, coupled with updated physics schemes, allows for improved performance in high-resolution simulations. While different versions of RegCM have been extensively used in East Asia and China in past studies (Fan et al., 2023; Gao et al., 2017; Pu et al., 2017; Wang et al., 2003), the application of RegCM5 in these regions remains limited. Compared to earlier versions, RegCM5 shows considerable improvement, but further optimization and validation are needed to fully exploit its capabilities in various climate settings (Kankeu et al., 2024; Ngo-Duc et al., 2024; Ren et al., 2024).

2.3. Experimental design

The model domain uses a Lambert conformal projection centered at 32°N , 119°E , with 125 grid points in both the zonal and meridional directions, a horizontal resolution of 25 km, and 18 vertical layers. The red rectangle in Fig. 1 shows the MLYRB (27°N – 33°N , 110°E – 123°E). In

Table 1
Experimental design.

Experiments	ARI	ACI
CTRL	✓	✓
SEXP_D	✓	×
SEXP_DI	×	×

the present study, the BATS1e land surface parameterization scheme is employed (Dickinson et al., 1993), along with the exponential relaxation lateral boundary condition scheme and the Zeng ocean flux scheme (Zeng et al., 1998). The physics configurations comprise the NCAR CCM3 atmospheric radiative transfer scheme (Kiehl et al., 1996), the UW PBL boundary layer scheme (Bretherton et al., 2004), and the SUBEX explicit moisture scheme (Pal et al., 2000). Cumulus convection schemes are parameterized using the Tiedtke cumulus parameterization scheme over land (Tiedtke, 1996) and the Kain-Fritsch cumulus parameterization scheme over the ocean (Kain and Fritsch, 1990). In this study we focus on anthropogenic aerosol effects and explicitly include sulfate, black carbon (BC), and organic carbon (OC) emissions from the IIASA inventory. Natural aerosol sources (e.g., mineral dust and sea salt) were not activated in the present configuration in order to isolate the anthropogenic pathways of aerosol–radiation interaction (ARI) and aerosol–cloud interaction (ACI) over the MLYRB, where human emissions dominate aerosol loading. As a result, our quantitative estimates of radiative forcing and precipitation responses should be viewed as conservative with respect to potential contributions from natural aerosols, while the qualitative day–night contrasts and the relative role of ACI versus ARI are expected to remain robust in this anthropogenic-aerosol-dominated region.

To explore the impacts of ARI and ACI on precipitation, as shown in Table 1, a control experiment (CTRL) and two sensitivity experiments (SEXP_D and SEXP_DI) are conducted to distinguish the effects of ARI and ACI separately. In this design, ARI includes direct and semi-direct effects through radiative–thermodynamic adjustments with fixed cloud microphysics, whereas ACI isolates microphysical pathways associated with cloud droplet number and lifetime by differencing CTRL and SEXP_D. Specifically, the CTRL experiment includes both ARI and ACI, the SEXP_D experiment only includes the ARI, while the SEXP_DI experiment excludes both ARI and ACI. Differences between the CTRL and SEXP_D experiments reveal the impacts of ACI on precipitation, whereas differences between the SEXP_D and SEXP_DI experiments highlight the effects of ARI on precipitation.

For all three experiments, the RegCM5 model is driven by the same atmospheric initial and lateral conditions derived from the ERA-Interim reanalysis data and the oceanic surface conditions derived from the NOAA SST data. In each experiment, the model integration is carried from 00:00 UTC on May 1st to 23:00 UTC on August 31st in each year of 2009–2018, and the model output interval is set to 1 h. Following Gu et al. (2022), we take the first month of simulation in each year as the model spin-up time. So the model outputs during June to August in each year of 2009–2018 are used for subsequent analysis to reveal the impact of ARI and ACI on the summer precipitation over the MLYRB.

2.4. Methods

In this study, precipitation is analyzed from multiple perspectives, including PA, PF, and PI. Following the approach of Sun et al. (2022b) we divide a day into daytime and nighttime. The daytime is defined as 07:00–19:00 Beijing Time (BJT), and the nighttime is defined as 19:00 on the preceding day to 07:00 BJT on the following day. The hourly simulated precipitation or the hourly rainfall gauge data is then aggregated into 12-h rainfall. All subsequent analyses are conducted based on this daytime–nighttime classification. A rainy day or night is identified when the accumulated precipitation during the daytime or nighttime is greater than or equal to $0.1 \text{ mm} (12 \text{ h})^{-1}$. The PA (PF) for daytime or

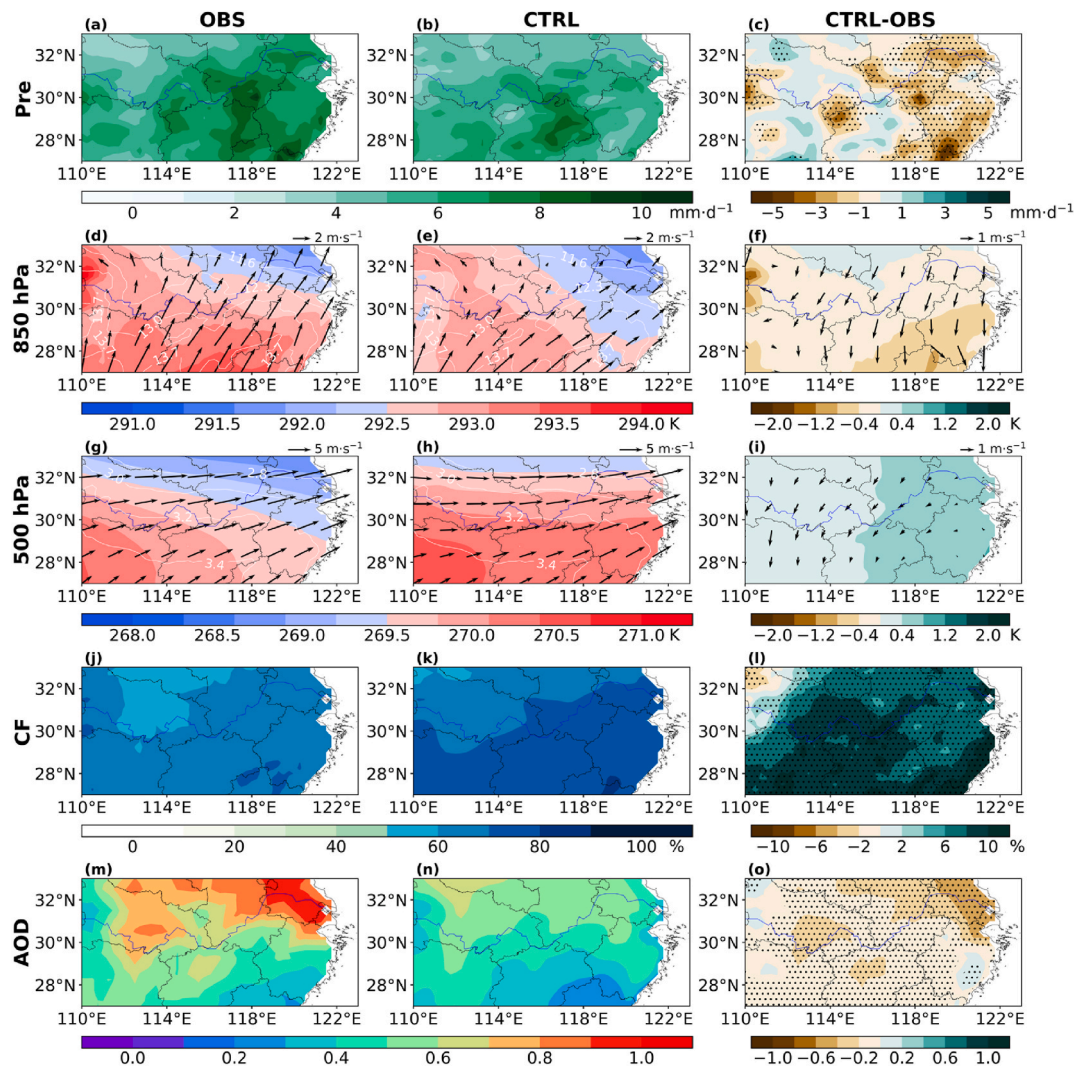


Fig. 2. Spatial distribution of summer climatic characteristics over the MLYRB during 2009–2018. (a–c) Precipitation (mm d^{-1}); (d–f) air temperature (shaded, K), wind field (vector, m s^{-1}), and specific humidity (contour, g kg^{-1}) at 850 hPa; (g–i) same as (d–f) but at 500 hPa; (j–l) cloud fraction (%); (m–o) aerosol optical depth. Panels show observations (left), CTRL experiment (middle), and the difference (CTRL-OBS, right).

nighttime is calculated as the total rainfall (the number of rainy days or nights multiplied by 100 %) during the ten summers divided by 920 (92 days or nights per year $\times$ 10 years). Then we can get PI, which equals PA/PF in daytime or nighttime.

According to Shao et al. (2022) and Guo et al. (2016), the 12 hourly rainfall is categorized into light precipitation ($0.10 \text{ mm (12 h)}^{-1} \leq R < 5.0 \text{ mm (12 h)}^{-1}$), moderate precipitation ($5 \text{ mm (12 h)}^{-1} \leq R < 12.5 \text{ mm (12 h)}^{-1}$), heavy precipitation ($12.5 \text{ mm (12 h)}^{-1} \leq R < 25 \text{ mm (12 h)}^{-1}$), and torrential precipitation ($R \geq 25 \text{ mm (12 h)}^{-1}$). In addition, the Student's t-test is used to test the significance of the differences and to eliminate the effects of internal variability introduced by the climate model. (Lehmann, 1992; Student, 1908). All observed data, reanalysis data, and model output data are interpolated uniformly using linear interpolation with a horizontal resolution of $0.25^\circ \times 0.25^\circ$. Model simulation skill is quantified using the spatial correlation coefficient (SCC) to measure pattern similarity, root mean square error (RMSE) to gauge magnitude error, and Taylor scores (TS; Taylor, 2001) for a combined assessment. Stronger model agreement with observations is indicated by higher SCC and TS, alongside lower RMSE.

3. Results

3.1. Model evaluation

The applicability of the RegCM model in simulating climate as well as aerosols in East Asia has been well demonstrated by previous studies (Hu et al., 2024; Gao et al., 2017; Ge et al., 2021; Lu et al., 2019; Zhou et al., 2013). To further validate the performance of RegCM5, this study compares the results from the CTRL experiment with the observations or reanalysis data. The evaluation focuses on key meteorological variables, including wind components, air temperature, specific humidity, precipitation, cloud cover, and AOD. Metrics such as SCC, RMSE, and TS are used to evaluate the model performance.

As shown in Fig. 2a and b, the CTRL experiment well reproduces the spatial distribution of precipitation in the MLYRB, despite of underestimation in eastern MLYRB with particularly large deviations in mountainous areas (Fig. 2c), similar to results of previous studies (Gao et al., 2017; Huang et al., 2020; Oh and Suh, 2018). This spatial bias pattern is partially inherited from the ERA-Interim forcing and has also been found in most GCM simulations (Gao et al., 2017; Xu et al., 2006).

From Fig. 2d–i, the distribution of the summer mean air temperature, specific humidity, and wind fields at 850 hPa and 500 hPa over the MLYRB from the CTRL experiment simulation is overall similar to the

Table 2

Statistical metrics (SCC, RMSE, TS) for model evaluation of selected variables over the MLYRB in the summers of 2009–2018.

Variables	Altitude	SCC	RMSE	TS
Temperature (K)	850 hPa	0.81	0.25	0.89
	500 hPa	0.92	0.47	0.95
Wind-v component (m s^{-1})	850 hPa	0.62	0.88	0.90
	500 hPa	0.97	0.39	0.89
Wind-u component (m s^{-1})	850 hPa	0.91	0.42	0.90
	500 hPa	0.99	0.31	0.98
Specific humidity (g kg^{-1})	850 hPa	0.98	0.18	0.98
	500 hPa	0.99	0.35	0.99
Precipitation (mm d^{-1})	–	0.44	1.36	0.67
Cloud fraction (%)	–	0.80	7.96	0.74
AOD	–	0.81	0.29	0.42

ERA5. The CTRL experiment successfully reproduces the overall south-to-north decreasing gradients of air temperature and specific humidity, with prevailing southwesterly winds at 850 hPa and westerly winds

at 500 hPa. However, Fig. 2f and i shows systematic model biases: there is a cold (-0.27°C) and dry (-0.01 g kg^{-1}) bias at 850 hPa, while a slight warm (0.4°C) and moist (0.04 g kg^{-1}) bias at 500 hPa. The wind difference fields further show that the model slightly underestimates the southwesterly flow at 850 hPa and westerly flow at 500 hPa, indicating weaker simulated monsoon strength compared with ERA5.

From Fig. 2j and k, the CTRL experiment reasonably reproduces the distribution of observed cloud cover over the MLYRB, although the magnitude is slightly overestimated across most parts of the MLYRB (Fig. 2l). AOD plays a crucial role in determining aerosol radiative forcing, which tends to increase as AOD rises (Bellouin et al., 2020; Stier et al., 2007; Xu et al., 2014). The MODIS observations indicate that AOD decreases from north to south over the MLYRB (Fig. 2m), and this gradient is generally captured by the CTRL experiment (Fig. 2n), despite that the model slightly underestimates AOD across the region (Fig. 2o), this may be due to the limited representation of anthropogenic aerosol sources in the experiment configuration. Given the slight regional underestimation of AOD and the omission of natural aerosols (dust, sea

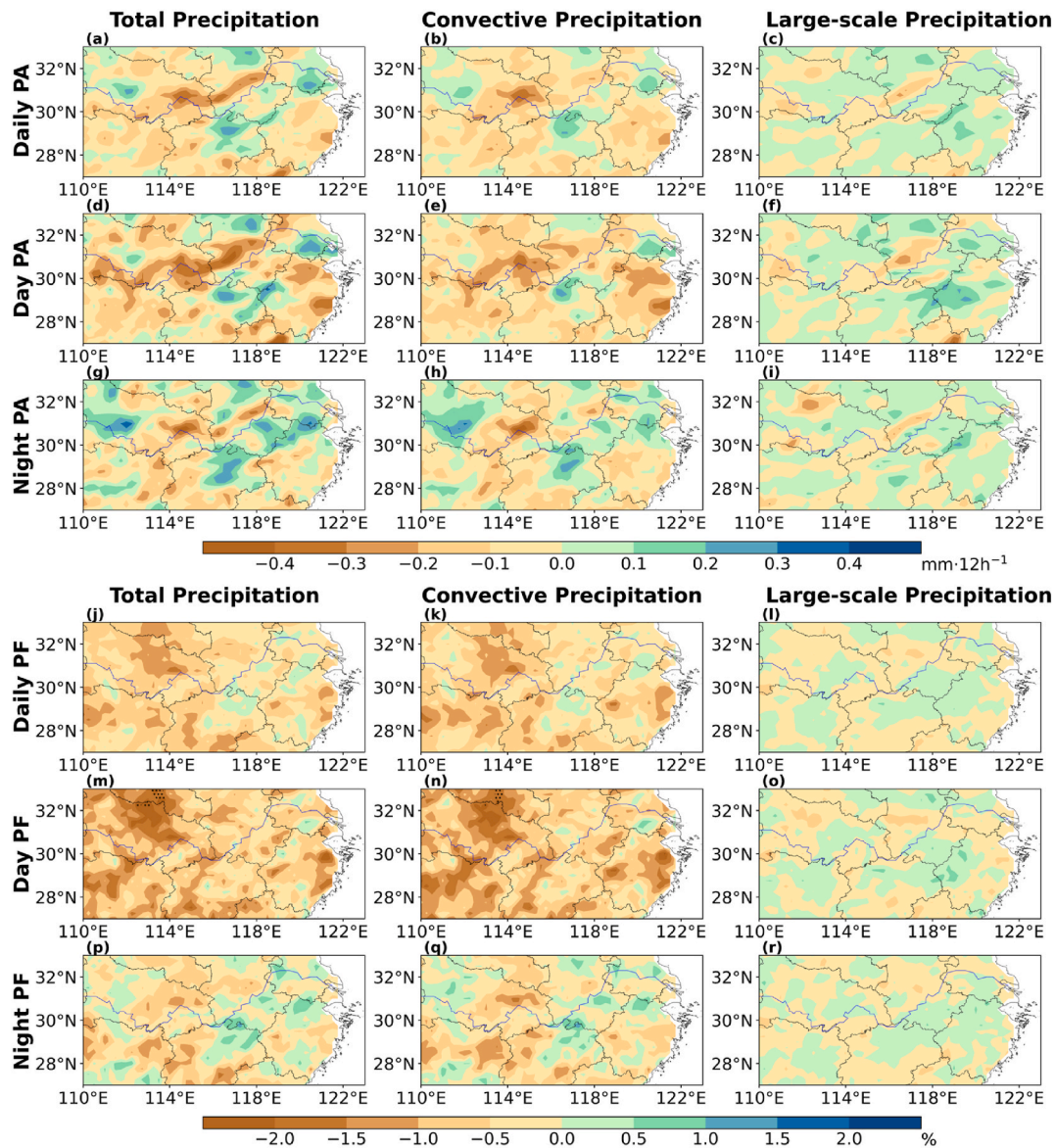


Fig. 3. The spatial distribution of mean differences of PA (mm (12 h)^{-1} , a-i) and PF (% , j-r) over the MLYRB during summer (June–August) from 2009 to 2018 induced by the ARI. Each row shows the differences during the daily mean (a-c, j-l), daytime (d-f, m-o), and nighttime (g-i, p-r), respectively. Columns represent total precipitation (left), convective precipitation (middle), and large-scale precipitation (right). Stippling indicates regions where the differences are statistically significant at the 90 % confidence level.

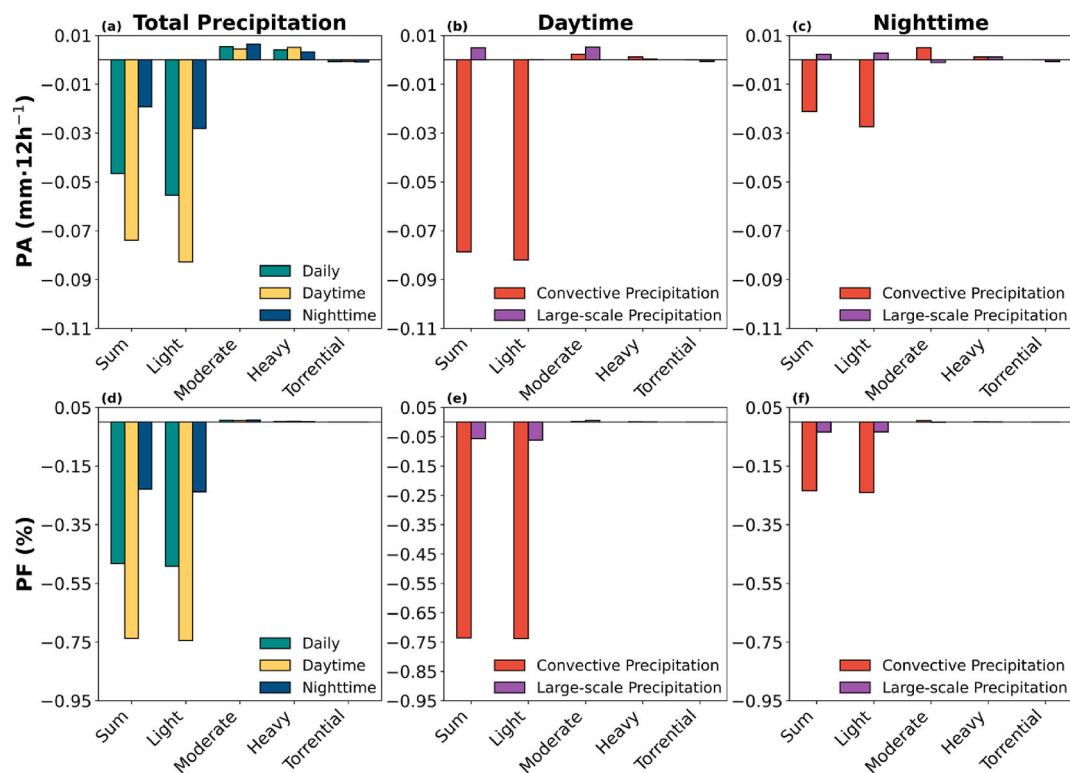


Fig. 4. Differences induced by ARI in PA (mm (12 h)⁻¹) and PF (%) for total, convective, and large-scale precipitation across five precipitation intensities (Sum, Light, Moderate, Heavy, Torrential) over the MLYRB. (a–c) show mean PA differences for daily, daytime, and nighttime; (d–f) show the corresponding PF differences. Convective and large-scale components are shown separately for daytime (b, e) and nighttime (c, f).

salt), the absolute magnitudes of surface radiative perturbations and precipitation responses reported here likely represent lower bounds.

Table 2 shows the statistical metrics of the simulation compared to the observation or reanalysis data for selected variables across different levels. The SCC values of modeled air temperature, specific humidity, and wind components are mostly above 0.90 at both 850 hPa and 500 hPa levels and low RMSEs, indicating general consistency with the reanalysis data. For the simulated cloud fraction and precipitation, the SCC values range from 0.44 to 0.80. The SCC of the simulated AOD is 0.81, indicating that the model can capture the overall spatial pattern of aerosol loading, despite some underestimation in magnitude. All metrics pass the 99.9 % confidence level of the significance test, confirming the robustness of the statistical relationships between the simulation and reference data. Overall, these results demonstrate that RegCM5 is capable of reasonably reproducing the key climatic factors and aerosol characteristics over the MLYRB, making it a reliable tool for further carrying out sensitivity experiments and regional climate research.

3.2. Changes of precipitation over the MLYRB in summer induced by ARI

Fig. 3 presents the spatial distribution of the mean changes in the PA and PF over the MLYRB during the summers from 2009 to 2018 induced by the ARI.

The changes in PA caused by ARI are relatively minor and not statistically significant. Daily total PA is suppressed by ARI by approximately 0.05 mm (12 h)⁻¹, with a larger reduction during the daytime than at night (Fig. 3a–d and g), particularly over central and northern MLYRB, indicating that the overall daily reduction in precipitation is predominantly driven by daytime changes. The convective precipitation is suppressed over most of the MLYRB, with a regional mean reduction of 0.08 mm (12 h)⁻¹ during the daytime (Fig. 3e). However, the suppression is relatively weaker at night, with a regional mean reduction of 0.02 mm (12 h)⁻¹ and a more scattered spatial distribution compared to

daytime (Fig. 3h), indicating that the ARI primarily weakens convective processes during the daytime. In contrast, the large-scale PA is slightly enhanced during the daytime, with a regional mean increase of 0.005 mm (12 h)⁻¹ (Fig. 3f), and an even smaller increase is observed at night (Fig. 3i).

The differences in PF are more pronounced than those in PA, particularly during the daytime. ARI predominantly reduces the daily total PF across most of the MLYRB by an average decline of 0.74 % (Fig. 3j), and exhibits pronounced suppression of up to 1.1 % over the 112°–114°E and 32°–33°N region during the daytime (Fig. 3m). At nighttime, the suppression weakens, with a mean reduction of 0.23 % (Fig. 3p), although slight enhancements in PF are observed in mountainous regions (e.g., the northern Wushan Mountains, Huangshan Mountain) and lake zones (e.g., Poyang Lake). The spatial distribution of convective PF differences (Fig. 3k–n and q) mirrors the patterns of total PF (Fig. 3j–m and p), with daytime PF decreased by an average of 0.74 % (Fig. 3n), while nighttime decrease remains marginal at 0.23 % (Fig. 3q). The large-scale PF shows minimal changes, with average decreases of 0.06 % during the daytime and 0.03 % at night (Fig. 3o and r).

Overall, the summer PA and PF over most MLYRB are suppressed by the ARI. The impact of ARI on large-scale precipitation is much weaker than on convective precipitation, and the reduction of daily total PA and PF induced by the ARI is mainly contributed by the reductions of the nighttime convective precipitation.

ARI exhibits distinct diurnal variations in its impacts on summer precipitation, regionally averaged over the MLYRB, across intensity categories (sum, light, moderate, heavy, torrential) (Fig. 4). ARI imposes stronger suppression on PF than on PA, with a particularly pronounced effect on convective precipitation. For the total precipitation, Fig. 4a indicates a reduction in PA induced by the ARI, with daily regionally averaged PA over the MLYRB decreased by ~0.05 mm (12 h)⁻¹, and a larger daytime reduction of ~0.08 mm (12 h)⁻¹ compared to ~0.02 mm (12 h)⁻¹ at night. The most substantial impact is observed in light

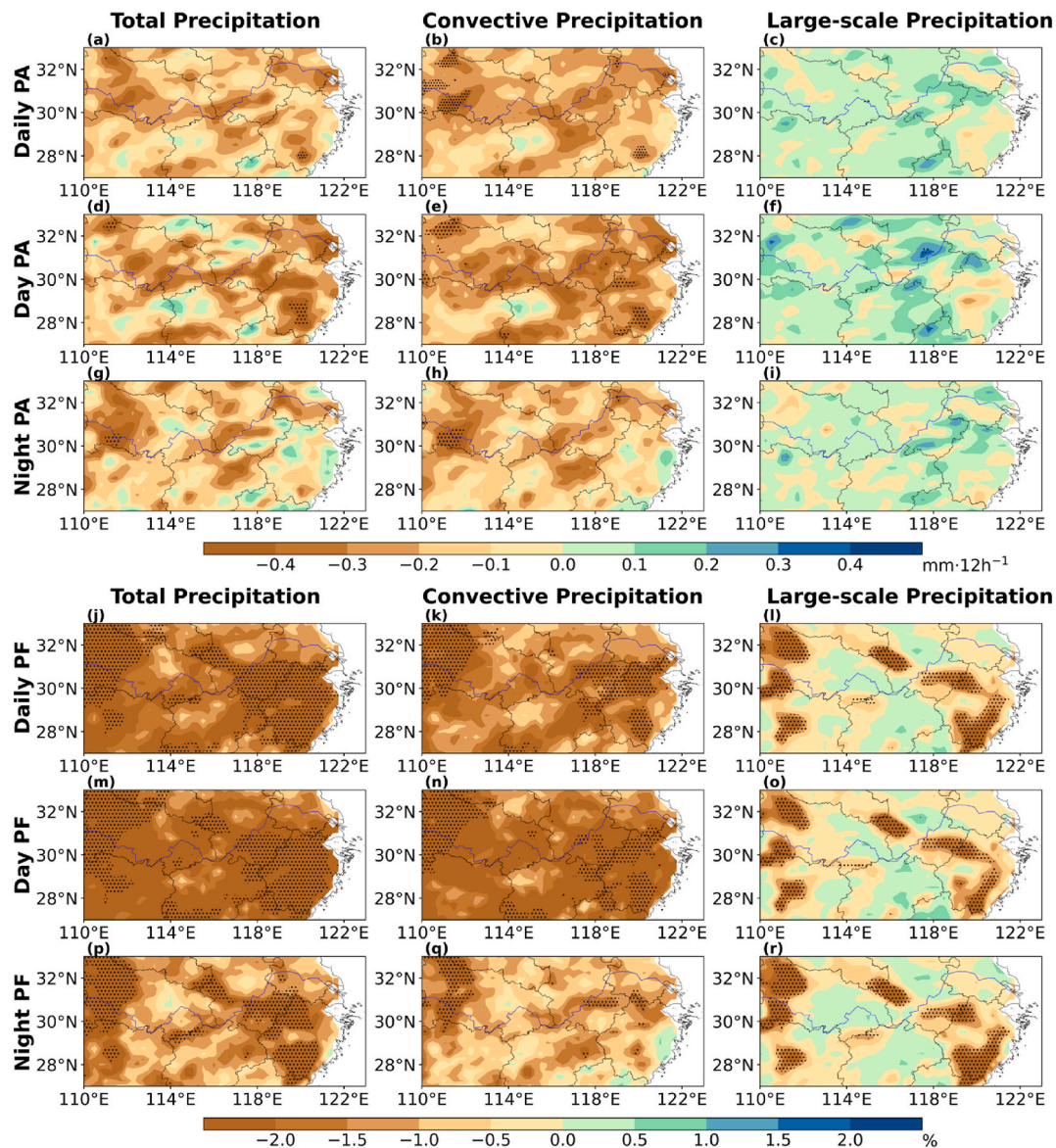


Fig. 5. The spatial distribution of mean differences of PA (mm (12 h)^{-1} , a–i) and PF (% , j–r) over the MLYRB during summer (June–August) from 2009 to 2018 induced by the ACI. Each row shows the differences during the daily mean (a–c, j–l), daytime (d–f, m–o), and nighttime (g–i, p–r), respectively. Columns represent total precipitation (left), convective precipitation (middle), and large-scale precipitation (right). Stippling indicates regions where the differences are statistically significant at the 90 % confidence level.

precipitation, with a reduction of $\sim 0.06 \text{ mm (12 h)}^{-1}$, accounting for most of the total PA decrease. In contrast, slight increases in PA are observed for heavier precipitation categories, all of which are below $0.01 \text{ mm (12 h)}^{-1}$. Fig. 4d reveals a corresponding decline in PF, with a total reduction of 0.48 % on the daily scale, primarily driven by the decrease in daytime light precipitation. This suggests that ARI predominantly suppresses light precipitation, with its influence diminishing as precipitation intensity increases. Comparison of convective and large-scale precipitation reveals significant diurnal differences. During the daytime (Fig. 4b), convective precipitation is suppressed, with light rain PA decreased by $0.08 \text{ mm (12 h)}^{-1}$, while large-scale PA shows a slight increase of less than $0.01 \text{ mm (12 h)}^{-1}$. A slight increase in convective moderate PA induced by ARI is observed, though its magnitude is lower than that of large-scale moderate precipitation (Fig. 4b). Meanwhile, ARI leads to a 0.74 % decline in the sum of convective PF, primarily due to the reduction in light precipitation (Fig. 4e). Differences in other intensity categories remain negligible. Fig. 4c and f shows that convective precipitation suppression persists at night but with reduced magnitude

compared to daytime. Correspondingly, the sum of PA decreases by $0.02 \text{ mm (12 h)}^{-1}$ at nighttime (Fig. 4c), while the sum of PF declines by 0.23 % (Fig. 4f), accompanied by a slight increase of moderate PA and PF. The Large-scale precipitation remains nearly unchanged in both PA and PF. Notably, ARI leads to a decrease in large-scale precipitation PF and an increase in PA, suggesting an enhancement in precipitation intensity, primarily driven by changes in light and moderate precipitation. These findings suggest that ARI predominantly suppresses light-to-moderate convective precipitation, especially during the daytime, while exerting minimal influence on large-scale or high-intensity precipitation events.

3.3. Changes of precipitation over the MLYRB in summer induced by ACI

The precipitation differences illustrated in Fig. 5 are the same as those in Fig. 3 but are attributed to ACI. Overall, ACI leads to a general reduction in both PA and PF across most of the MLYRB.

ACI exerts a stronger suppressive effect on PA than ARI, with significant impacts observed in certain regions (Fig. 5). The daily total PA

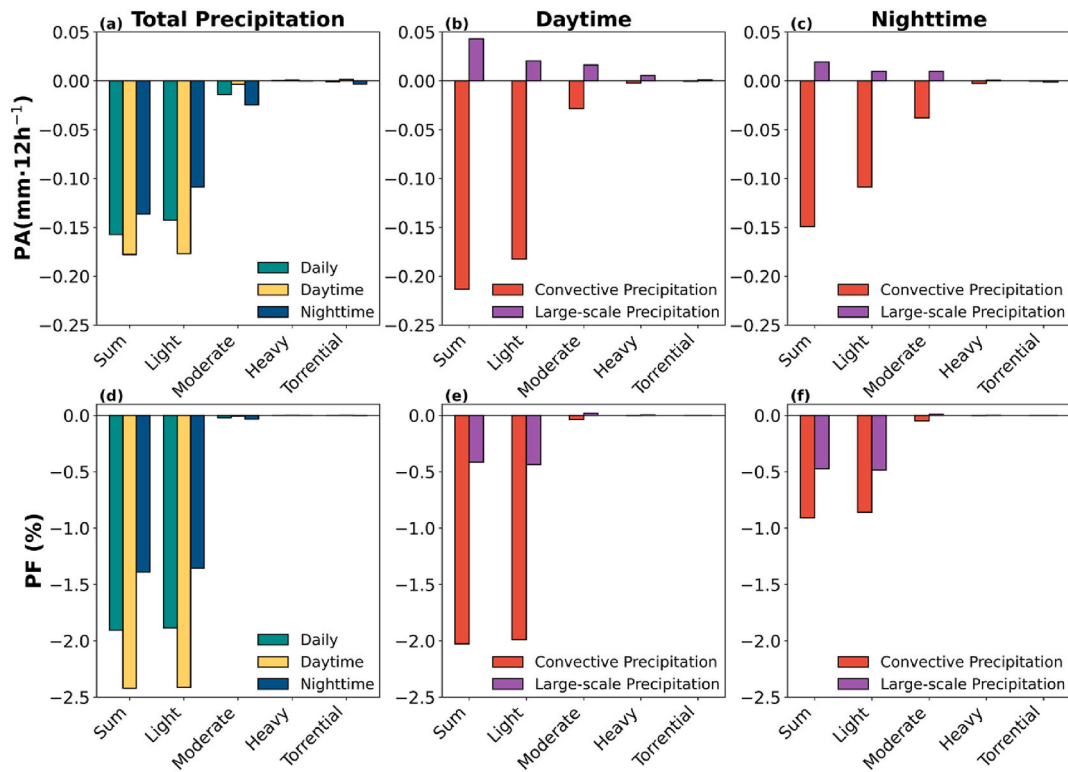


Fig. 6. Differences induced by ACI in PA (mm (12 h)^{-1}) and PF (%) for total, convective, and large-scale precipitation across five precipitation intensities (Sum, Light, Moderate, Heavy, Torrential) over the MLYRB. (a–c) show mean PA differences for daily, daytime, and nighttime; (d–f) show the corresponding PF differences. Convective and large-scale components are shown separately for daytime (b, e) and nighttime (c, f).

regionally averaged over the MLYRB decreases by $0.16 \text{ mm (12 h)}^{-1}$, with statistically significant decreases in the southeastern MLYRB (Fig. 5a). During the daytime, the regional mean PA decreases by $0.18 \text{ mm (12 h)}^{-1}$, with notable reductions in the eastern and northwestern regions (Fig. 5d). At nighttime, the regional mean decrease of PA induced by ACI is slightly smaller at $0.14 \text{ mm (12 h)}^{-1}$ compared to daytime, with significant reductions confined to the central and western regions (Fig. 5g). ACI causes convective PA to be reduced by an average of $0.21 \text{ mm (12 h)}^{-1}$ during the daytime and $0.15 \text{ mm (12 h)}^{-1}$ at night (Fig. 5e and h). The spatial distribution of convective PA mirrors that of total precipitation, though the extent of statistically significant areas is slightly broader than in total precipitation. The effect of ACI on the large-scale PA is weakly positive, with a regional mean increase of $0.04 \text{ mm (12 h)}^{-1}$ during the daytime and $0.02 \text{ mm (12 h)}^{-1}$ at night; however, these changes are not statistically significant.

Similarly, ACI exerts a more significant suppressive effect on PF compared to ARI across most MLYRB (Fig. 5j). The daily total PF exhibits a regional mean decrease of 1.91% (Fig. 5j) with daytime reductions reaching 2.42% , statistically significant in the northwestern and southeastern areas (Fig. 5m). In mountainous regions such as the Wushan Mountains, Huangshan Mountains, and Wuyi Mountains, the suppression is more pronounced, with reductions ranging from 3.29% to 3.38% during the daytime. Although nighttime suppression is weaker than daytime, it remains prevalent, with a regional average decrease of 1.39% ; mountainous regions still exhibit reductions of approximately 1.86% – 2.18% (Fig. 5p). The spatial distribution of convective precipitation PF mirrors that of total precipitation, contributing most significantly to the overall variation. ACI can lead to the daytime (nighttime) regional mean convective PF decreased by 2.03% (0.91%) with slight enhancements in the southeastern coastal regions of the MLYRB (Fig. 5n and q). Large-scale PF exhibits minimal changes but demonstrates terrain-dependent patterns: daytime regional mean reductions of 0.41% , with suppression in mountainous regions (e.g., Wushan, Dabie, and

Wuyi Mountains) and slight enhancement in lake regions. At nighttime, the suppression intensifies, with an average reduction of 0.47% , maintaining a spatial pattern similar to daytime (Fig. 5o and r).

Fig. 6 illustrates the distinct diurnal variations of ACI impact on PA and PF across different intensity categories over the MLYRB. As shown in Fig. 6a and d, ACI consistently suppresses both PA and PF for all intensity levels, with light precipitation experiencing the strongest reduction. Specifically, the daily PA for light precipitation decreases by approximately $0.15 \text{ mm (12 h)}^{-1}$ (Fig. 6a), while PF drops by about 1.9% (Fig. 6d), largely driving the total reduction. Compared to daytime (Fig. 6b), nighttime suppression (Fig. 6c) is weaker but still evident. Convective PA and large-scale PA respond differently to ACI. Fig. 6b and c shows that convective PA is consistently negative across all intensities, indicating strong suppression, especially for light precipitation during the daytime ($-0.18 \text{ mm (12 h)}^{-1}$). In contrast, large-scale PA shows slight positive anomalies, particularly for light and moderate precipitation, suggesting that ACI inhibits convective development while slightly enhancing stratiform processes. However, this enhancement weakens as intensity increases, becoming negligible for heavy and torrential rainfall. The PF response (Fig. 6e and f) does not mirror this clear dichotomy in PA. Both convective and large-scale precipitation generally exhibit reduced PF across most intensities. The largest decrease appears in light convective precipitation during the daytime (-1.99%), while large-scale PF also declines slightly (-0.44% for light rain) (Fig. 6e). For moderate to torrential large-scale precipitation, PF remains nearly unchanged or shows minor increases, possibly due to limited compensatory effects under stable nocturnal conditions. In summary, ACI strongly suppresses both the frequency and amount of light and moderate convective precipitation, especially during the daytime. While large-scale PA tends to increase, its PF mostly declines or remains near neutral, indicating that enhanced stratiform precipitation occurs less frequently but may be more intense when it does occur.

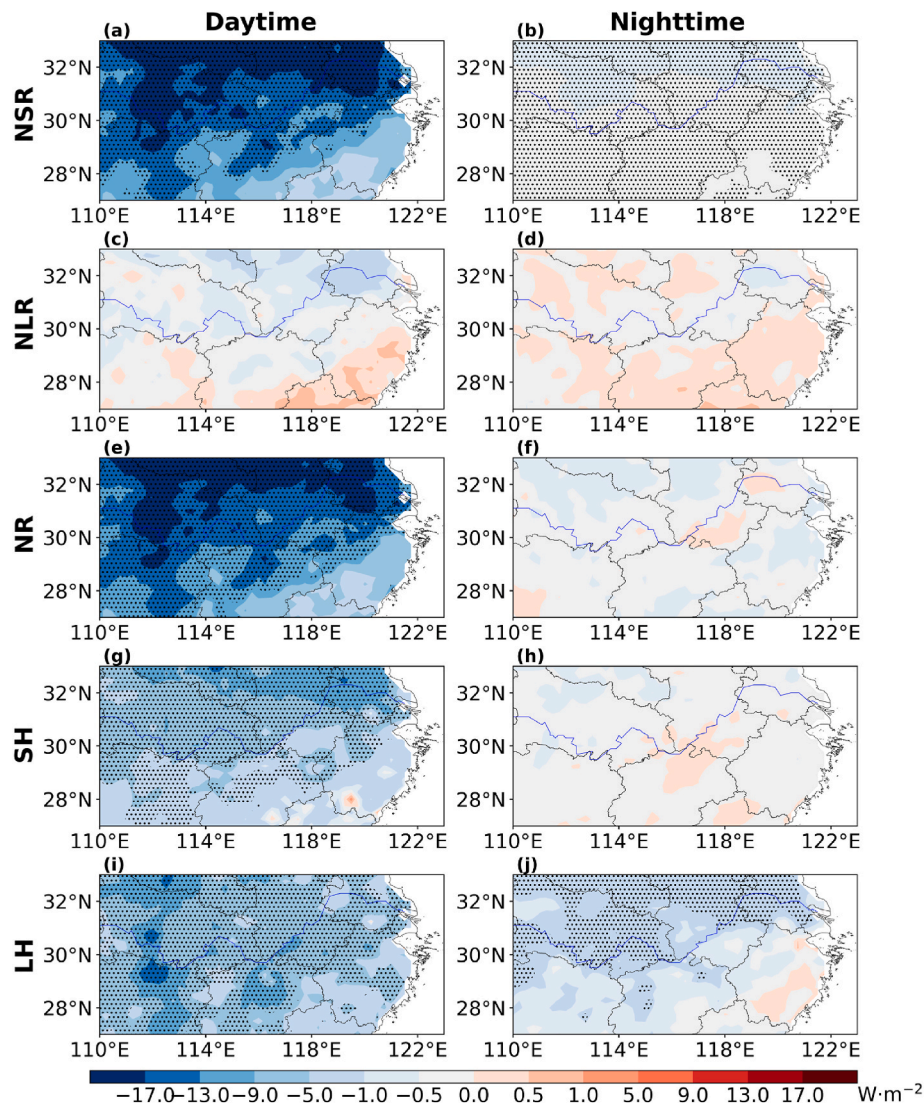


Fig. 7. The spatial distribution of mean differences of aerosol effective radiative forcing (W m^{-2}) over MLYRB during summer (June–August) from 2009 to 2018 induced by ARI. Panels (a–b) show net downward shortwave radiation (NSR), (c–d) Net upward longwave radiation (NLR), (e–f) Net downward radiation (NR), (g–h) Sensible heat flux (SH), and (i–j) Latent heat flux (LH) during the daytime (left) and nighttime (right). Stippling indicates regions where the differences are statistically significant at the 90 % confidence level.

3.4. Mechanisms of ARI impacts on precipitation

The precipitation differences can be traced back to changes induced by ARI in surface energy balance. Fig. 7 presents the spatial distribution of aerosol effective radiative forcing (ERF) caused by ARI, including the daytime and nighttime (Ming et al., 2010). During the daytime, aerosols substantially reduce incoming shortwave radiation (NSR) by up to 14.02 W m^{-2} (Fig. 7a) over the northern MLYRB, where aerosol loading is the highest (Fig. 2j). This strong radiative cooling results from aerosol scattering and absorption, leading to a marked decline in the surface net radiation (NR) (Fig. 7e). Consequently, both sensible heat flux (SH) and latent heat flux (LH) at the surface decrease substantially (Fig. 7g and i), indicating weakened surface heating and evapotranspiration and, thereafter, weakened thermal convection and reduced moisture availability, both of which are essential for convective precipitation (Tao et al., 2012). These thermodynamic changes correspond spatially with the strongest reductions of PA and PF in daytime (Fig. 3d–f, 3m–o), indicating that precipitation suppression is closely tied to aerosol-induced energy deficits at the surface. At nighttime, the impact of ARI on radiation are minimal due to the absence of solar input, with NSR changes below 0.4 W m^{-2} (Fig. 7b), and associated perturbations in

NR, SH, and LH all less than 1 W m^{-2} (Fig. 7f–h and j). Correspondingly, nighttime precipitation responses are much weaker and spatially fragmented (Fig. 3g–i, 3p–r), suggesting a limited role of ARI at nighttime.

These thermodynamic changes, driven by the surface energy deficit, alter the lower atmospheric structure (Fig. 8). While the impacts on convective potential indices such as CAPE and CIN are minor and not statistically significant (Fig. 8a–c), a robust and significant average reduction of 15.3 m in the planetary boundary layer height (PBLH) is observed over the northern MLYRB during the daytime (Fig. 8e). This suppression of the boundary layer height is a direct consequence of reduced surface heating, which limits vertical mixing and further stabilizes the lower atmosphere, thus creating conditions less favorable for the development of convection. This suppression of the boundary layer height restricts vertical mixing and confines moisture to lower levels, which, combined with large-scale dynamic adjustments, leads to changes in vertical motion and moisture transport. Fig. 9 further presents the mean differences in vertical velocity and water vapor flux divergence caused by ARI. During the daytime, the latitude–pressure cross-section (Fig. 9a) exhibits positive water vapor flux divergence centered around 30°N from the surface to 850 hPa , accompanied by widespread downward motion throughout the entire tropospheric

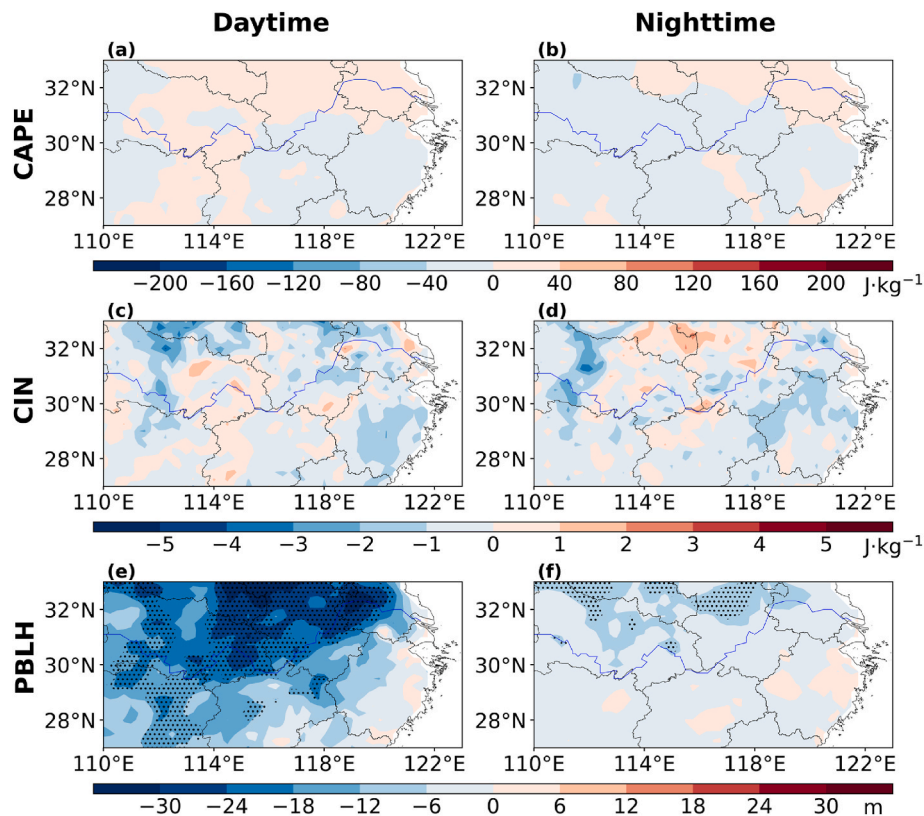


Fig. 8. The spatial distribution of mean differences of key thermodynamic parameters over the MLYRB during summer (June–August) from 2009 to 2018 induced by ARI. The panels show differences in (a, b) Convective Available Potential Energy (CAPE; J kg^{-1}), (c, d) Convective Inhibition (CIN; J kg^{-1}), and (e, f) Planetary Boundary Layer Height (PBLH; m). The left column represents daytime, and the right column represents nighttime. Stippling indicates regions where the differences are statistically significant at the 90 % confidence level.

column (from the surface to 300 hPa). This combination of low-level moisture divergence and subsidence indicates a suppressed convective environment (Li et al., 2016), consistent with the reductions in convective precipitation observed in Fig. 3e and n. A similar pattern appears in the longitude–pressure cross-section (Fig. 9c), where persistent anomalous downward motion and moisture divergence dominate over 114°E – 116°E . These dynamic signals spatially correspond to the reductions in both total and convective precipitation (Fig. 3d, e, m and n), reinforcing the role of ARI in suppressing daytime rainfall through thermodynamic stabilization and reduced moisture availability. At nighttime, the circulation anomalies weaken. In the latitude–pressure section (Fig. 9b), the anomalous moisture divergence diminishes, and anomalous downward motion becomes weaker or even slightly changes to upward, indicating a more neutral atmospheric state. Localized moisture convergence appears near 30°N below 850 hPa, with weakened subsidence compared to daytime. Furthermore, an area of enhanced convergence and upward motion develops near 31°N up to 700 hPa, suggesting the potential for localized nocturnal convection (Fig. 3h and q). This nocturnal enhancement is more evident in the longitude–pressure section (Fig. 9d), which displays intensified moisture divergence near 112°E along with an anomalous upward motion extending to 600 hPa. These features coincide with the areas of slightly increased convective precipitation at nighttime (Fig. 3h and q), implying that the ARI may locally enhance nighttime convection under specific dynamic conditions.

In summary, ARI suppresses daytime precipitation primarily by altering the surface energy balance and weakening the conditions necessary for convection. The strong reduction in incoming shortwave radiation decreases net surface radiation, leading to substantial declines in sensible and latent heat fluxes. This surface energy deficit directly triggers a series of atmospheric responses: it weakens surface heating

and evapotranspiration, thereby reducing moisture availability and convective instability. This stabilization is evidenced by a significant reduction in the planetary boundary layer height, which in turn enhances low-level moisture divergence and promotes widespread subsidence. Collectively, this cascade of an unfavorable thermodynamic and dynamic environment significantly suppresses the development of convective precipitation, particularly light and moderate rain. Consequently, the reduction in convective precipitation largely explains the observed decreases in total PA and PF during the daytime. At nighttime, the absence of solar radiation input limits the impact of ARI on surface energy balance, resulting in minimal changes in heat fluxes and a weaker suppression of convection. In some areas, localized moisture convergence and upward motion may even induce slight nocturnal convective enhancements, but these are generally scattered and weak. The impact on large-scale precipitation remains limited, with slight increases in PA but negligible changes in PF, suggesting that stratiform precipitation may be more intense when it occurs but does not become more frequent. Overall, ARI exerts a stronger influence on light-to-moderate convective precipitation during the daytime, with a much weaker and more localized effect at night.

3.5. Mechanisms of ACI impacts on precipitation

ACI primarily affects summer precipitation over the MLYRB primarily through microphysical and radiation–cloud feedbacks. Fig. 10a–d shows that both daytime and nighttime total cloud cover increase markedly over the MLYRB, with respective enhancements of 3.89 % and 4.19 %. The increase in total cloud cover is mainly attributed to the enhancement of high clouds. Middle clouds and low clouds show slight change below 0.5 % (Fig. 10e–h), indicating their weaker response to aerosol-induced microphysical perturbations. Elevated aerosol

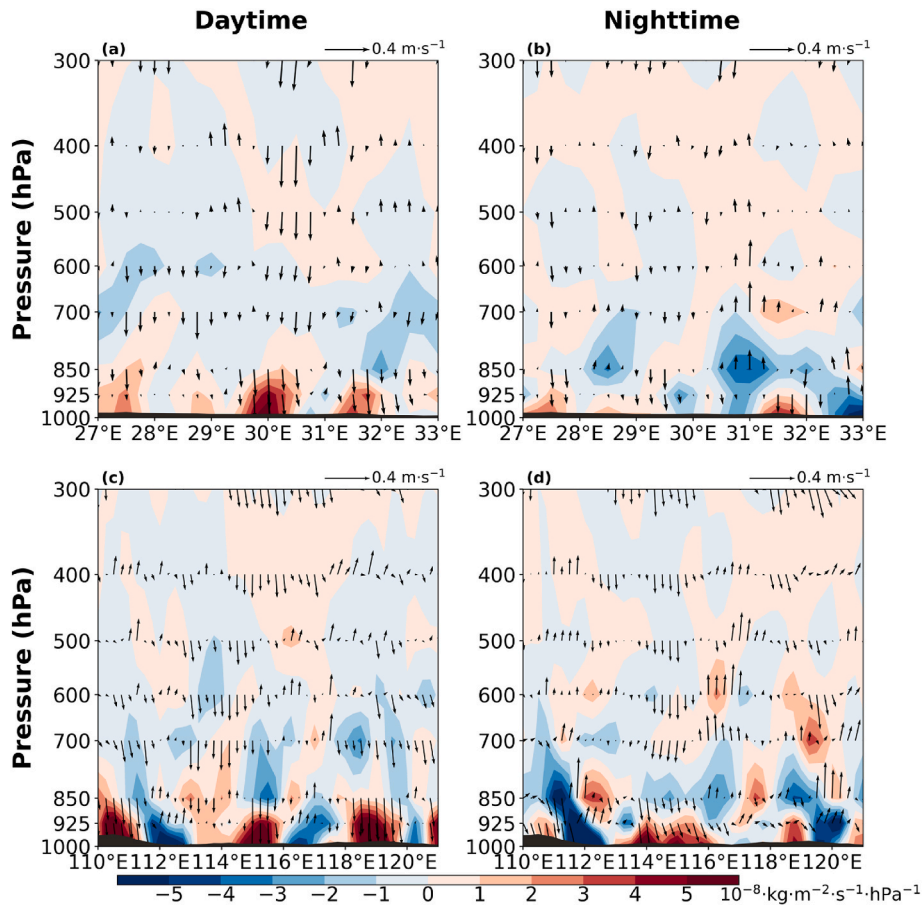


Fig. 9. The cross-section of mean differences of vertical velocity and integrated water vapor flux divergence over MLYRB during summer (June–August) from 2009 to 2018 induced by ARI. Panels (a) and (b) show latitude–pressure cross-sections (27°N–33°N, averaged over 110°E–121°E) for daytime and nighttime, respectively. Panels (c) and (d) show longitude–pressure cross-sections (110°E–121°E, averaged over 27°N–33°N) for daytime and nighttime, respectively. Shading indicates water vapor flux divergence ($10^{-8} \text{ kg m}^{-2} \text{ s}^{-1} \text{ hPa}^{-1}$), and arrows denote vertical wind components. Terrain height is shown by the black shading at the bottom of each panel.

concentrations increase CCN, which in turn reduces cloud droplet size and coalescence efficiency. This directly suppresses the conversion of cloud water into precipitation, meaning more water vapor can be transported to higher altitudes, facilitating the formation of persistent ice-phase clouds (Twomey, 1977; Albrecht, 1989). This inhibits the warm rain process, prolongs cloud lifetime, and enhances total cloud cover, especially high clouds (Xue et al., 2008). The increase in high clouds lowers outgoing longwave radiation because the colder cloud tops emit less infrared energy than the warmer surface, retaining more heat in the lower troposphere (Lohmann and Feichter, 2005; Hartmann et al., 1992), especially at night when longwave radiation dominates the surface energy balance (Fig. 11d and f). The increase in high cloud cover traps additional longwave radiation aloft, leading to warming in the upper troposphere and a steeper lapse rate. Together, these changes enhance atmospheric stability and create conditions that are less favorable for daytime convective development (Fan et al., 2016; Tao et al., 2012).

These cloud-induced radiative changes directly alter the surface energy balance. Fig. 11 shows the aerosol ERF caused by ACI, which is slightly stronger than that caused by ARI (Allen et al., 2016). During the daytime, NSR decreases by up to 13.72 W m^{-2} , mainly over the western and southern MLYRB (Fig. 11a). NLR is reduced by 0.94 W m^{-2} , while weak positive anomalies are observed over eastern MLYRB (Fig. 11c). The reduced NSR further lowers NR by as much as 12.78 W m^{-2} (Fig. 11e), resulting in a daytime SH decrease of 2.30 W m^{-2} across most areas except for slight local increases in the Yangtze River Delta (Fig. 11g). LH also drops notably by 9.68 W m^{-2} on average, with larger

reductions over the western MLYRB (Fig. 11i). These changes correspond to the enhanced high cloud cover and reduced low clouds (Fig. 10a and c), which together weaken shortwave radiation reaching the surface and increase atmospheric stability. The decrease in SH indicates weaker turbulent mixing and boundary-layer convection, while lower LH reflects reduced surface evaporation under enhanced relative humidity and decreased available energy, together limiting moisture supply and convective development, and ultimately resulting in suppressed precipitation over the MLYRB (Figs. 5 and 6). At nighttime, increased high clouds reduce outgoing longwave radiation, leading to slightly positive NLR anomalies that expand compared to daytime, although the regional average still shows a mild suppression of 0.70 W m^{-2} (Fig. 11d). This retention of longwave energy slightly raises NR by about 0.56 W m^{-2} and SH by 0.16 W m^{-2} in some areas (Fig. 11f and h), but this warming effect remains limited under stable nocturnal conditions (Fig. 11j). Overall, the enhanced high clouds trap more outgoing longwave radiation due to their cold tops, partly offset by any increased low clouds reflecting longwave energy downward, but this slight nocturnal warming is insufficient to sustain strong convection, so nighttime precipitation remains weakly suppressed.

The cloud and radiation changes induced by ACI ultimately adjust the atmospheric thermodynamic structure significantly, thereby suppressing convection (Fig. 12). During the daytime, the reduction in surface net radiation due to increased cloud cover weakens the energy input into the lower atmosphere, leading to a substantial decrease of 105.57 J kg^{-1} in CAPE (Fig. 12a). Although CIN also slightly decreases by 0.93 J kg^{-1} (Fig. 12c), which would theoretically make convection

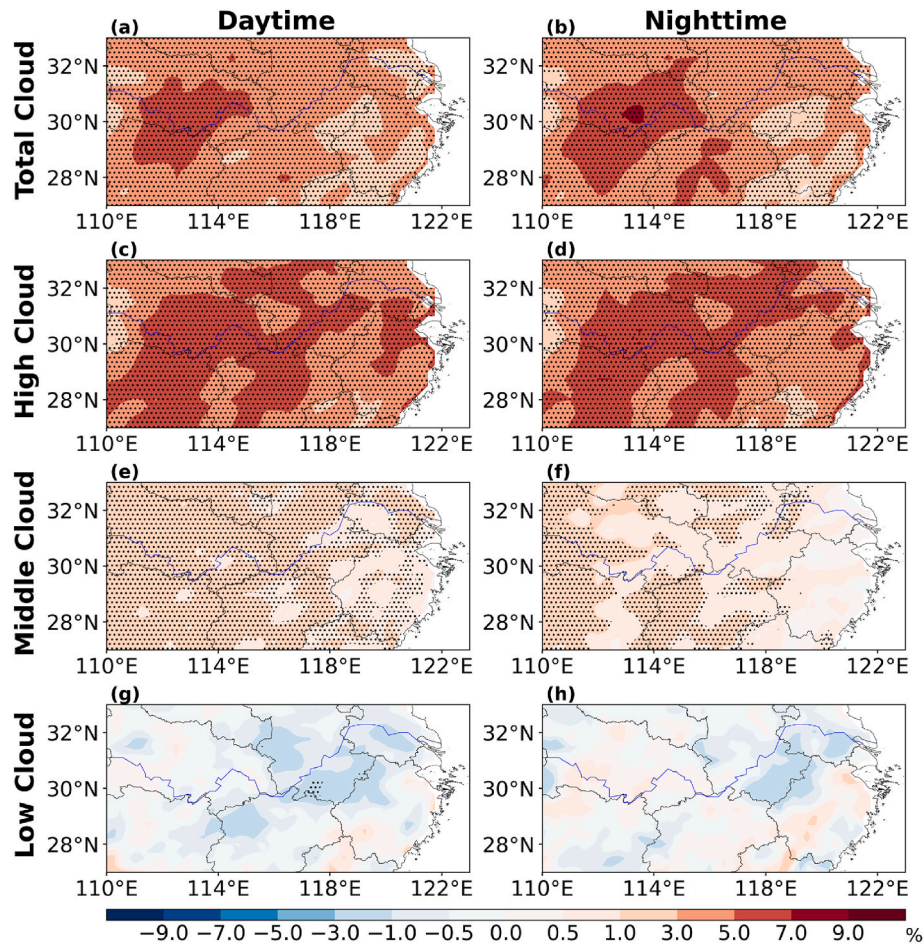


Fig. 10. The spatial distribution of mean differences of total cloud cover (a,b), high cloud cover (c, d), middle cloud cover (e, f), and low cloud cover (g, h) over the MYLRB during the daytime (left) and nighttime (right) in the summer (June–August) from 2009 to 2018 induced by ACI. Shading indicates regions where the differences are statistically significant at the 90 % confidence level.

easier to trigger, this minor effect is completely overwhelmed by the massive reduction in CAPE, thus the net impact remains a strong suppression. In some regions, the PBLH also shows a notable decrease (Fig. 12e), further limiting the vertical transport of heat and moisture. At nighttime, although the increase in high clouds provides a warming effect at the surface through longwave radiation feedback (Fig. 11f), it is insufficient to counteract the microphysical suppression of precipitation. CAPE at night still exhibits a consistent and significant decrease of 115.51 J kg^{-1} across the entire region (Fig. 12b). This large energy deficit far outweighs the minor 0.60 J kg^{-1} decrease in CIN (Fig. 12d), indicating that the atmosphere remains in a suppressed, stable state unfavorable for the maintenance and development of convective precipitation. These changes in thermodynamic parameters provide direct evidence for how ACI suppresses convective precipitation during both day and night.

These thermodynamic adjustments are coupled with large-scale dynamic changes. Fig. 13 shows that ACI generates widespread downward motion and positive moisture flux divergence during both daytime and nighttime (Fig. 5). During the daytime, the latitude–pressure cross-section (Fig. 13a) shows a negative water vapor flux divergence centered and slight upward motion around 29°N from the surface to 850 hPa, corresponding to the slight increase in precipitation in Fig. 5e. The longitude–pressure cross-section (Fig. 13c) further indicates broad downward motion and divergence across 110°E – 121°E , corresponding to the general decrease in convective precipitation observed in Fig. 5e and n. These circulation anomalies suppress vertical development and hinder moisture accumulation, resulting in reduced convective activity.

At nighttime, enhanced moisture divergence and subsidence dominate from 28°N – 33°N and 110°E – 118°E (Fig. 13b and d), further limiting moisture convergence and vertical development. Although orographic uplift and coastal breezes can locally enhance LH in mountainous and coastal areas at night, this local effect is too weak to offset the broader suppression. Thus, while ACI continues to promote atmospheric stabilization and moisture divergence at night, its influence on precipitation remains secondary to daytime effects.

In summary, ACI suppresses summer precipitation over the MYLRB mainly through changes in cloud microphysics and subsequent cloud–radiation feedbacks. First, from a microphysical perspective, elevated aerosols increase CCN, which reduces cloud droplet size and coalescence efficiency, thereby inhibiting warm rain processes and prolonging cloud lifetime. This leads to an increase in total cloud cover, especially high clouds. Second, these changes in cloud properties induce strong cloud–radiation feedbacks. During the daytime, increased high clouds and reduced low clouds limit incoming solar radiation and lower surface NR. This reduces SH and LH, weakens surface heating and evaporation, and stabilizes the boundary layer, which jointly suppresses convection and moisture supply. Combined with large-scale subsidence and moisture divergence, this leads to a notable reduction in daytime convective precipitation. This atmospheric stabilization is confirmed by a significant decrease in CAPE, which creates conditions less favorable for convective development. At nighttime, more high clouds restrict outgoing longwave radiation, which slightly buffers surface cooling and weakens the stable boundary layer. However, this modest warming effect is insufficient to overcome both the persistent microphysical

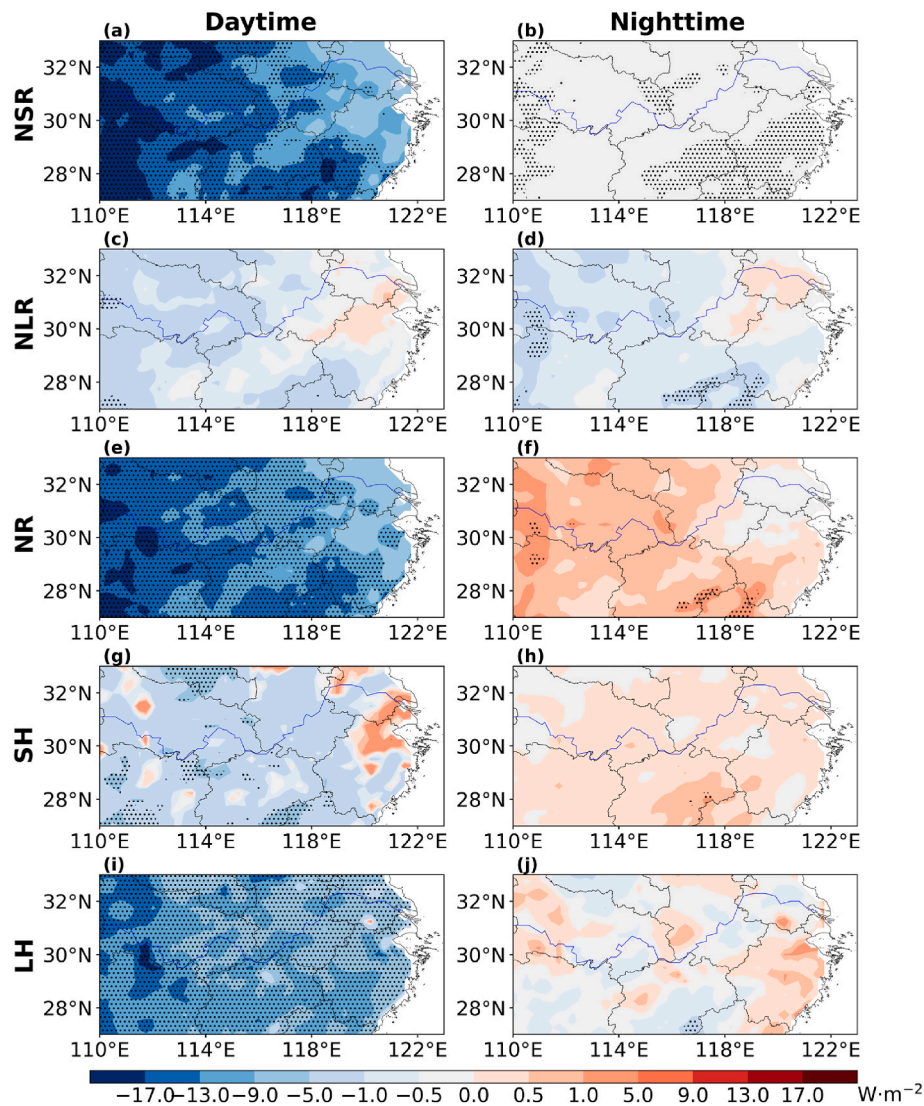


Fig. 11. The spatial distribution of mean differences of aerosol effective radiative forcing (W m^{-2}) over MLYRB during summer (June–August) from 2009 to 2018 induced by ACI. Panels (a–b) show net downward shortwave radiation (NSR), (c–d) Net upward longwave radiation (NLR), (e–f) Net downward radiation (NR), (g–h) Sensible heat flux (SH), and (i–j) Latent heat flux (LH) during the daytime (left) and nighttime (right). Stippling indicates regions where the differences are statistically significant at the 90 % confidence level.

suppression from smaller, less efficient cloud droplets and the large-scale subsidence established during the day, which is reinforced by the inherently stable nocturnal conditions. As a result, nocturnal precipitation remains generally suppressed, with only minor local increases possible in mountainous and coastal areas due to topographic lifting and sea-land breeze effects.

4. Conclusion and discussion

This study systematically analyzes the diurnal differentiation characteristics and physical mechanisms of ARI and ACI on summer precipitation in MLYRB based on the RegCM5 model. The main conclusions are as follows:

ARI exerts a suppressive impact on summer precipitation over the MLYRB from 2009 to 2018, primarily by weakening convective precipitation, while having little effect on large-scale precipitation. ARI leads to a regional mean decrease in daily total PA of approximately $0.05 \text{ mm (12 h)}^{-1}$, with a stronger reduction during the daytime compared to nighttime. The effect is more pronounced in PF, with a total PF reduction of 0.74% , mainly driven by the decline in light precipitation during the daytime. The physical mechanism is governed by

thermodynamic cooling and dynamic stabilization of the lower atmosphere. During the daytime, aerosols reduce NSR by up to 14.02 W m^{-2} , resulting in marked declines in NR, SH, and LH, especially over northern and central MLYRB. This leads to thermodynamic stabilization, characterized by cooler surface temperatures, weakened convection, and reduced evapotranspiration, which is further confirmed by a significant reduction in the PBLH. Correspondingly, vertical circulation patterns reveal enhanced moisture divergence and subsidence around 30°N and 114°E – 116°E , which suppresses convective development. At nighttime, these effects largely diminish, with net radiation perturbations below 1 W m^{-2} and weaker circulation anomalies, resulting in more limited and spatially scattered precipitation responses.

ACI significantly inhibits precipitation over the MLYRB, especially light to moderate convective precipitation during the daytime. ACI leads to a decrease in daily average total PA of $0.16 \text{ mm (12 h)}^{-1}$ and PF of 1.91% , with a stronger reduction during the daytime compared to nighttime. Its physical mechanism is mainly dominated by the coupling effect of cloud microphysics and radiation. ACI increases cloud droplet number concentrations, which reduces droplet size and coalescence efficiency, directly limiting the conversion of cloud water to rainwater and lowering precipitation efficiency. As a result, total cloud coverage

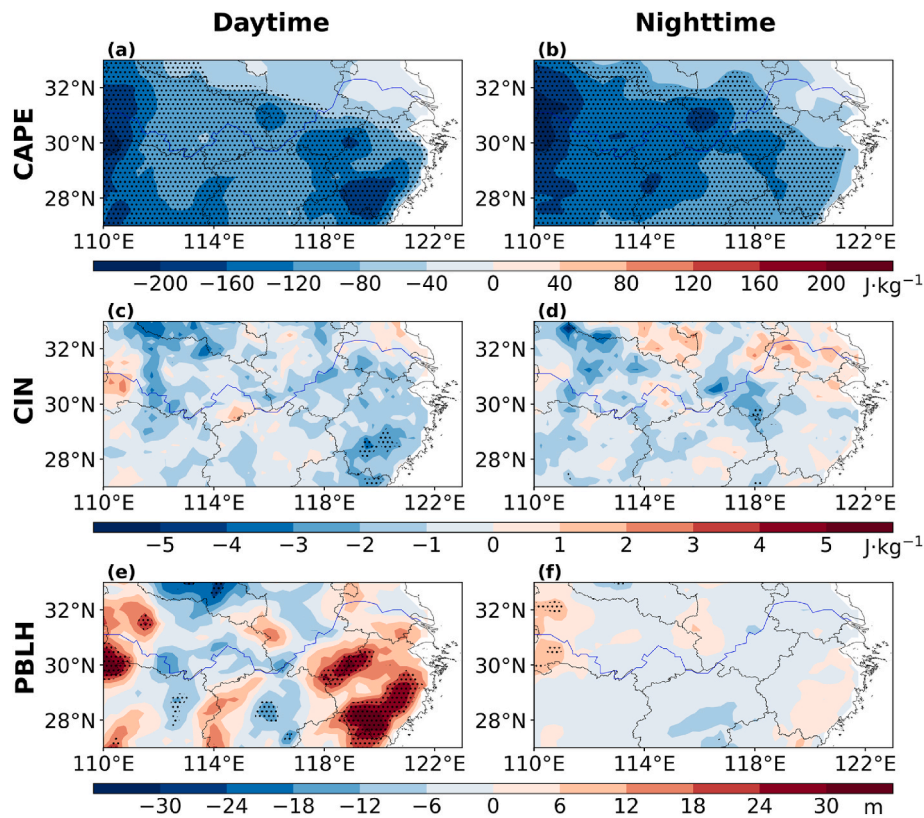


Fig. 12. The spatial distribution of mean differences of key thermodynamic parameters over the MLYRB during summer (June–August) from 2009 to 2018 induced by ACI. The panels show differences in (a, b) Convective Available Potential Energy (CAPE; J kg^{-1}), (c, d) Convective Inhibition (CIN; J kg^{-1}), and (e, f) Planetary Boundary Layer Height (PBLH; m). The left column represents daytime, and the right column represents nighttime. Stippling indicates regions where the differences are statistically significant at the 90 % confidence level.

increasing by 3.89 % during the day and 4.19 % at night, mainly manifested in a significant increase in high clouds. This expansion of high clouds reduces shortwave radiation and longwave radiation, resulting in a decrease of 12.78 W m^{-2} in NR during the day. Radiative cooling weakens the SH and LH during the daytime, thereby suppressing thermal convection and evaporation processes. At nighttime, due to changes in longwave radiation, NR increases slightly, but SH and LH still decrease respectively. ACI further suppresses precipitation by promoting atmospheric stability and water vapor flux divergence, which manifests as extensive sinking movement and water vapor divergence during the day, limiting vertical development and water vapor accumulation, thereby significantly reducing convective activity. At nighttime, the increased high clouds restrict outgoing longwave radiation, which slightly buffers surface cooling and weakens the stable boundary layer. However, this modest warming effect is insufficient to overcome the persistent microphysical suppression and lingering large-scale subsidence. Consequently, the net effect of ACI at night remains a general suppression of precipitation.

Additionally, it should be noted that while this study analyzed the effects of ARI and ACI respectively, the synergistic or offsetting mechanisms of them were not fully explored. Our experimental design was intentionally chosen to elucidate the distinct physical pathways through which each process modulates precipitation. However, our findings suggest their combined effect is likely synergistic, acting through distinct but complementary pathways to enhance atmospheric stability: ARI primarily reduces surface solar radiation directly through scattering and absorption, while ACI achieves a similar cooling effect indirectly through enhanced cloud albedo. Both pathways lead to a more stable lower atmosphere that is less conducive to convection. This conclusion is strongly supported by numerous studies in East Asia. For instance, regional modeling has consistently shown that the coexistence of ARI

and ACI tends to inhibit precipitation by reducing incoming shortwave radiation at the surface, which in turn increases lower-atmospheric stability and weakens convection (e.g., Fan et al., 2015; Jiang et al., 2017; Wu et al., 2013). A systematic review of coupled models in Asia further confirms that these mechanisms are common pathways for aerosol-induced precipitation changes (Gao et al., 2022). While the interaction can be complex, with some case studies showing ARI partially offsetting ACI-induced precipitation enhancement (Liu et al., 2020b), the synergistic suppression of rainfall by anthropogenic aerosols is recognized as a significant factor in East Asian climate (Dong et al., 2019; Ryu and Min, 2022). Therefore, the mechanisms identified in our study are consistent with the prevailing understanding of synergistic aerosol effects. Quantifying the precise degree of this synergy, however, remains a valuable direction for future research. This study reveals the multi-scale regulatory mechanism of anthropogenic aerosol effects on regional precipitation in MLYRB. However, several limitations remain: Our present configuration does not include natural aerosols (dust, sea salt) and nitrate, which may modulate cloud microphysics and radiative forcing, especially during episodic dust intrusions and marine inflow. Because anthropogenic aerosols are the dominant sources over the MLYRB, our estimates likely provide conservative lower bounds of the total aerosol impact, while preserving the qualitative diurnal contrasts and the relative strength of ACI over ARI. Future work will incorporate multi-component aerosol modules (including nitrate, dust, and sea salt) to reassess the magnitude of forcings and to quantify possible nonlinear interactions. Additionally, the insufficient description of the cloud droplet–raindrop conversion process in the current scheme may lead to biases in the estimation of ACI. The SUBEX scheme is a single-moment parameterization, and the choice of a different scheme (e.g., WSM5) would likely alter the quantitative results due to different representations of ice-phase physics and precipitation efficiency. More advanced

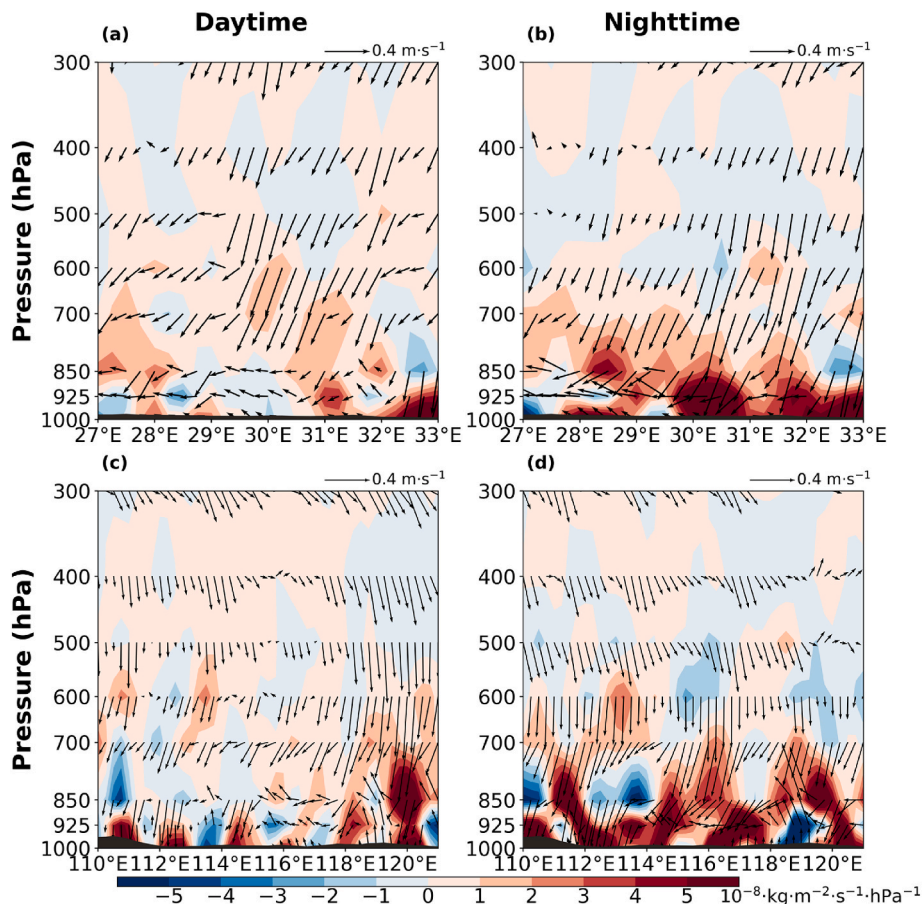


Fig. 13. The cross-section of mean differences of vertical velocity and integrated water vapor flux divergence over MLYRB during summer (June–August) from 2009 to 2018 induced by ACI. Panels (a) and (b) show latitude–pressure cross-sections (27°N–33°N, averaged over 110°E–121°E) for daytime and nighttime, respectively. Panels (c) and (d) show longitude–pressure cross-sections (110°E–121°E, averaged over 27°N–33°N) for daytime and nighttime, respectively. Shading indicates water vapor flux divergence ($10^{-8} \text{ kg m}^{-2} \text{ s}^{-1} \text{ hPa}^{-1}$), and arrows denote vertical wind components. Terrain height is shown by the black shading at the bottom of each panel.

two-moment schemes would better capture the explicit link between aerosols and cloud droplet number, but their high computational cost makes them challenging for long-term climate studies. Therefore, we acknowledge that the magnitude of the simulated ACI effect is scheme-dependent, and constraining this uncertainty with a multi-scheme ensemble remains an important direction for future research. Furthermore, the model does not capture the fine-scale characteristics of regional precipitation accurately. While it can replicate the general trend of precipitation changes, it struggles to precisely reproduce the timing of peak precipitation and the start and end times of precipitation events. These limitations highlight the need for future studies to incorporate multi-component aerosol mixtures and more physically realistic cloud microphysical schemes, which will improve the mechanistic understanding of aerosol effects in regional precipitation modeling.

In conclusion, this study systematically investigates the diurnal differentiation, relative importance, and physical mechanisms of ARI and ACI impacts on summer precipitation over the MLYRB from a long-term climate perspective. Our findings reveal that both ARI and ACI exert a net suppressive effect on summer precipitation, with the impact of ACI being stronger than that of ARI. A key finding is the pronounced diurnal asymmetry of these effects; both interactions are significantly stronger during the daytime, primarily through the suppression of convective precipitation. By elucidating these distinct day-night mechanisms, this study provides a crucial scientific basis for understanding how anthropogenic activities modulate regional weather, climate, and the hydrological cycle through complex aerosol processes tied to the diurnal cycle.

CRediT authorship contribution statement

Zhuohui Wang: Writing – original draft, Methodology, Investigation, Data curation. **Shu Li:** Writing – review & editing, Supervision, Funding acquisition, Conceptualization. **Anning Huang:** Writing – review & editing, Supervision, Project administration. **Xinsheng Zhu:** Project administration, Methodology. **Tijian Wang:** Project administration. **Bingliang Zhuang:** Software. **Mengmeng Li:** Visualization. **Rongchang Wu:** Validation.

Declaration of competing interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

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Data availability

Data will be made available on request.

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