



1 **Sensitivity of the Summertime Diurnal Precipitation Cycle over**
2 **China to Convective Trigger and Closure Treatments in a**
3 **Single-Column Model**

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24 **Abstract:**

25 Using a single-column model, we investigate how trigger and closure treatments in ZM,
26 TDK, and UNICON convection schemes affect the diurnal cycle of precipitation over
27 China. For late-afternoon rainfall, TDK and ZM schemes with standard CAPE closure
28 simulate precipitation peaks around noon, substantially earlier than observed. The
29 UNICON scheme, with cold-pool feedback deactivated, also produces a noontime peak.
30 For nocturnal rainfall, all schemes spuriously produce a noontime peak, in stark contrast
31 to the observed nocturnal maximum. These behaviors are closely related to closure
32 assumptions and the resulting cloud-base mass flux, which is primarily governed by
33 CAPE in ZM and TDK, and responds more directly to surface fluxes in UNICON. By
34 accounting for the imbalance between deep convection and boundary-layer production
35 through specific modifications: adopting the dCAPE trigger in ZM, activating cold-pool
36 feedback in UNICON, and applying the non-equilibrium closure in TDK, the simulated
37 afternoon precipitation is delayed and in better agreement with observations. In addition,
38 allowing elevated convection in ZM and UNICON improves nocturnal rainfall. The
39 improved diurnal cycle is also evident in the diabatic heating fields. All modifications
40 lead to a delayed and more pronounced upward-tilting heating structure as a result of
41 the suppression of early-afternoon convection, with ZM and TDK extending to higher
42 altitudes, while both intensified and extended in UNICON. These changes alleviate the
43 upper-level cold bias in TDK and ZM. In UNICON, however, the overly strong
44 convection reverses the upper-tropospheric bias from cold to warm with stronger
45 magnitude, thereby worsening the bias.



46 **1. Introduction**

47 The diurnal cycle of precipitation (DCP), which plays a critical role in modulating the
48 surface energy budget and hydrology, has long been recognized as a fundamental mode
49 of variability in the climate system (Trenberth et al., 2003). In addition, DCP also acts
50 as a building block for the large-scale organization of precipitation systems, including
51 convectively coupled equatorial waves and the Madden–Julian Oscillation (Mapes et
52 al., 2006). Therefore, accurate simulation of DCP is not only essential for advancing
53 the precision of “timing–location–intensity” in fine-scale forecasting (Dai and
54 Trenberth, 2004; Chen et al., 2017), but also important for improving processes across
55 a range of temporal and spatial scales—from convective systems to intraseasonal
56 disturbances (Benedict et al., 2013). Nevertheless, DCP modeling in general circulation
57 models (GCMs) remains a persistent challenge, as evidenced by a premature afternoon
58 peak over land and a frequent inability to capture nocturnal rainfall maxima downwind
59 of major orography (Dai, 2006; Lee et al., 2007; Tang et al., 2022).

60 Over land, the observed DCP typically peaks in late afternoon during summer,
61 partly due to the strong diurnal cycle of surface heat flux that modulates convection
62 (Dai et al., 1999; Li et al., 2008). However, on the leeward side of mountainous regions such
63 as the Tibetan Plateau and the Rockies, the maximum precipitation is often observed in
64 late night, owing to the propagation of mesoscale convective systems induced by
65 terrain–forced circulations (Carbone and Tuttle, 2008; Zhang et al., 2014). Because
66 DCP is closely related to changes in surface temperature and boundary-layer structure
67 on much shorter timescales, it poses a fundamental challenge to the equilibrium



68 assumption commonly used in convection parameterizations (Davies et al., 2013). In
69 this sense, DCP provides a measure of how well a GCM represents the response of
70 cumulus convection to large-scale and surface forcing. Many schemes initiate
71 convection based on convective available potential energy (CAPE), which responds
72 rapidly to solar heating, resulting in overly frequent and premature triggering of
73 afternoon convection (e.g., Xie and Zhang, 2000; Lee et al., 2007; Yang et al., 2018).
74 On the other hand, the failure to capture nocturnal precipitation is often attributed to the
75 inadequate representation of elevated convection originating above the boundary layer.
76 (Lee et al., 2008; Wang et al., 2015; Xie et al., 2019). To overcome these deficiencies,
77 increasing efforts have been devoted to developing improved trigger and closure
78 treatments. For example, Xie and Zhang (2000) proposed dynamic CAPE (dCAPE)
79 triggering, which initializes convection only when the large-scale CAPE production
80 exceeds a critical value. Bechtold et al. (2014) accounted for CAPE consumption by
81 shallow convection and modified the closure to subtract its contribution from the total
82 mass flux. Park (2014a) incorporated subgrid mesoscale organized flow and its
83 feedback into the convection parameterization, enabling the successful simulation of
84 various convection-related transitional phenomena including DCP. To better simulate
85 nocturnal rainfall, Wang et al. (2015) allows parcels to be initiated from above the
86 boundary layer, thereby improving the simulation of elevated nocturnal convection over
87 the central Great Plains. This triggering concept was later used in Xie et al. (2019) and
88 Park et al. (2024). While these modifications have mitigated certain DCP biases at the
89 testing sites, it remains unclear how well they generalize to regions with markedly



90 different climatic and topographic conditions, such as China. The present study is
91 therefore motivated to assess their effectiveness for DCP simulation over China.
92 Additionally, despite the documented improvements in DCP simulation achieved by
93 various approaches, a systematic intercomparison of their performance and underlying
94 mechanisms has yet to be conducted. Liu et al. (2026) performed a process-oriented
95 analysis that traces DCP errors over China in three reanalyses to specific parameterized
96 physical processes, highlighting the critical role of convective trigger in shaping DCP.
97 This study builds on Liu et al. (2026) by using a single-column model (SCM) to further
98 explore the sensitivity of DCP to modifications in convective trigger and closure
99 treatments. Specifically, we assess how precipitation varies diurnally under different
100 trigger and closure treatments, aiming to elucidating the physical mechanisms behind.
101 The remainder of this paper is structured as follows. Section 2 describes the case setup,
102 model configurations, and experimental design. Section 3 presents modifications to the
103 convection schemes and examines the DCP sensitivity to different parameterization
104 settings, along with the underlying mechanisms. Section 4 analyzes the diurnal
105 variations of diabatic heating and convective eddy transport of moist static energy
106 (MSE). Section 5 discusses the implications for temperature and moisture simulations,
107 and Section 6 summarizes the main findings.

108

109 **2. Case, Model and Experiments**

110 **2.1 Case Description**

111 The three representative regions used in this study are $4^{\circ} \times 4^{\circ}$ subregions of those in Liu



112 et al. (2026): the Tibetan Plateau (TP, 31°–35°N, 90°–94°E), southeastern China (SECN,
113 24°–28°N, 112°–116°E), and the Sichuan Basin (SCB, 28°–32°N, 103°–107°E).
114 Among these, TP and SECN are characterized by late-afternoon rainfall regimes, while
115 SCB features a nocturnal regime. For each region, DCP is simulated and analyzed over
116 the summer from 1 June to 31 August 2021. The initial and large-scale forcing fields
117 used to drive SCMs are derived from ERA5 reanalysis (Hersbach et al., 2020),
118 facilitated by the open-source Python package named the “Large-eddy simulation and
119 Single-column model—Large-Scale Dynamics” (van Stratum et al., 2023).

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121 **2.2 Model and Experimental Design**

122 We use the single column version of the Community Atmospheric Model (CAM)
123 for SCM simulations, which features a suite of parameterization schemes for cumulus
124 convection (Gettelman et al., 2019). In this study, three distinct configurations are
125 compared: CAM_ZM, CAM_UNICON, and CAM_TDK. In the CAM_ZM setup, deep
126 convection is parameterized using the Zhang & McFarlane (1995) scheme, whereas
127 shallow convection are treated with the Park & Bretherton (2009) scheme.
128 CAM_UNICON employs a unified framework for both shallow and deep convection
129 (Park, 2014a). In addition to ZM and UNICON, the bulk mass-flux scheme originally
130 proposed by Tiedtke (TDK; 1989) and later updated by Bechtold et al. (2014) was also
131 implemented for comparison, referred to as CAM_TDK. Other parameterizations are
132 set the same across all configurations. The parameterization of planet boundary layer
133 (PBL) turbulence is according to Bretherton & Park (2009). Cloud microphysics is



134 represented by the double moment scheme of Morrison & Gettelman (2008). Radiative
135 transfer is calculated with the Rapid Radiative Transfer Model for GCMs (Clough et al.,
136 2005). For each convection scheme, two simulations are conducted: a control
137 experiment (CTRL) and a sensitivity experiment (SENS). For CAM_ZM, the
138 sensitivity experiment adopts the dCAPE trigger and permits elevated convection,
139 following Xie et al. (2019). In CAM_UNICON, the control experiment disables
140 cold-pool feedback, whereas the sensitivity experiment activates cold-pool feedback
141 and allows elevated convection as described in Park et al. (2024). For CAM_TDK, the
142 control experiment employs the equilibrium closure, while the sensitivity experiment
143 applies the non-equilibrium closure as proposed by Bechtold et al. (2014). Table 1
144 summarizes the key differences in parameterization components across various
145 configurations. To reduce model error accumulation, all simulations are reinitialized
146 daily at 00 UTC from June 1 to August 31 and integrated for 36 hours. While
147 momentum fields are relaxed toward the reanalysis with a 3-hour timescale, no
148 relaxation is applied to other variables. The last 24 hours of each 36-hour simulation
149 are retained for analysis.

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156 **Table 1** Parameterization configurations

Parameterization		CAM_ZM	CAM_TDK	CAM_UNICON
Deep convection	CTRL	CAPE-based trigger and closure	Equilibrium closure	Cold-pool feedback disabled
		<i>Zhang & McFarlane [1995]</i>	<i>Bechtold et al. [2014]</i>	<i>Park [2014a]</i>
	SENS	dCAPE trigger + elevated convection allowed	Non-equilibrium closure	Cold-pool feedback enabled + elevated convection allowed
		<i>Xie et al. [2019]</i>	<i>Bechtold et al. [2014]</i>	<i>Park et al. [2024]</i>
Shallow convection		<i>Bretherton & Park [2008]</i>		<i>Park [2014a]</i>
PBL turbulence		<i>Bretherton & Park [2009]</i>		
Cloud macrophysics		<i>Park et al. [2014]</i>		
Cloud microphysics		<i>Morrison & Gettelman [2008]</i>		
Radiation		<i>Clough et al. [2005]</i>		

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159 **2.3 Composite Analysis**

160 To isolate the typical DCP behavior, we performed a composite analysis by selecting
161 precipitation events that represent the characteristic DCP category for each region.
162 Following Liu et al. (2026), we selected afternoon-peaking (12:00–20:00 LST, local
163 solar time) days for TP and SECN, and nocturnal-peaking (20:00–08:00 LST) days for
164 SCB. We also required the daily maximum precipitation to exceed thresholds of 5 mm
165 d⁻¹ for TP and 10 mm d⁻¹ for SECN and SCB. By doing so, the influence of dry days,
166 light precipitation, and days with diurnal phases inconsistent with the target regime is
167 excluded, thereby enhancing the DCP signal. Unlike Liu et al. (2026) that used different
168 event dates across the datasets, we used the same fixed event dates derived from
169 precipitation observations for all model simulations.

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172 **3. Sensitivity of Diurnal Precipitation to Convection Schemes**

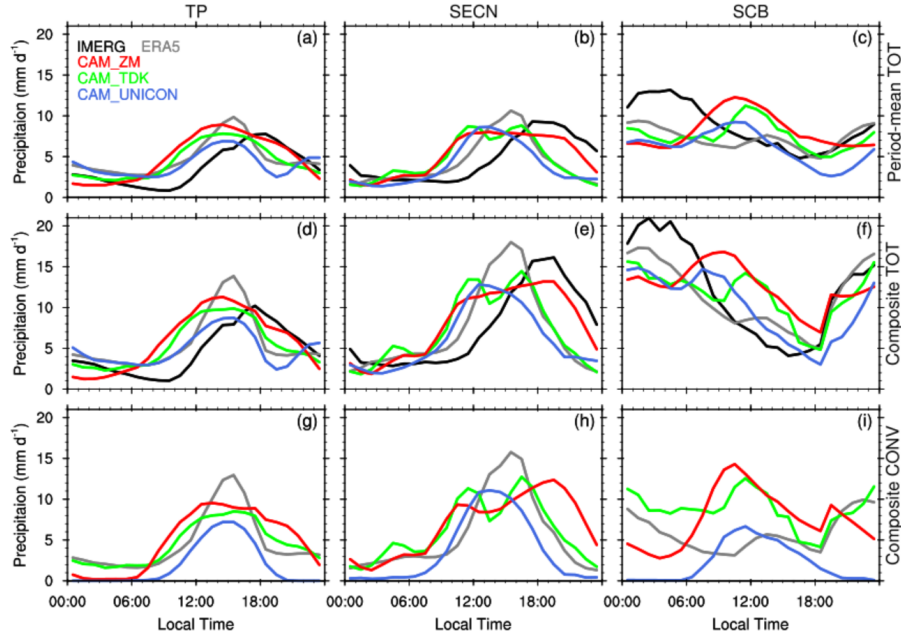


173 **3.1 Diurnal Precipitation in the Control Experiments**

174 Figure 1 presents the period-averaged and case-composite diurnal variations of
175 precipitation over the three regions from the control experiment. The observation is the
176 Integrated Multi-satellitE Retrievals for Global Precipitation Measurement (IMERG)
177 Final Run Version 07B (Huffman et al., 2023), with ERA5 precipitation also
178 superimposed for reference. A comparison between the upper and middle rows reveals
179 that the composite cycles align closely with the period-mean in phase, and the diurnal
180 signals are amplified due to the removal of light-precipitation days. Over TP and SECN,
181 the IMERG precipitation peaks around 18:00 LST, whereas all simulations produce
182 rainfall peaks notably earlier than observed, with the maximum occurring near noon.
183 In comparison, the ZM and TDK schemes initiate rainfall even earlier than UNICON
184 and tend to produce a secondary peak over SECN—at 19:00 LST for ZM and at 17:00
185 LST for TDK. Note that ERA5 also exhibits an early phase bias over TP and SECN
186 relative to IMERG, though less pronounced than the SCM simulations. The convective
187 precipitation shown in the bottom closely resembles the total precipitation, suggesting
188 that deficiencies in convection parameterization are largely responsible for the poor
189 simulation of DCP. Over SCB, the observed precipitation exhibits a distinct nocturnal
190 maximum around 03:00 LST, in stark contrast to the late-morning peak simulated by
191 CAM_ZM (Figure 1f). While CAM_TDK reasonably captures the nocturnal peak, it
192 also produces an additional noontime maximum. Similar to ZM and TDK, the UNICON
193 scheme also produces a noontime peak in convective precipitation (Figure 1i), with
194 nocturnal rainfall being primarily contributed by large-scale precipitation.



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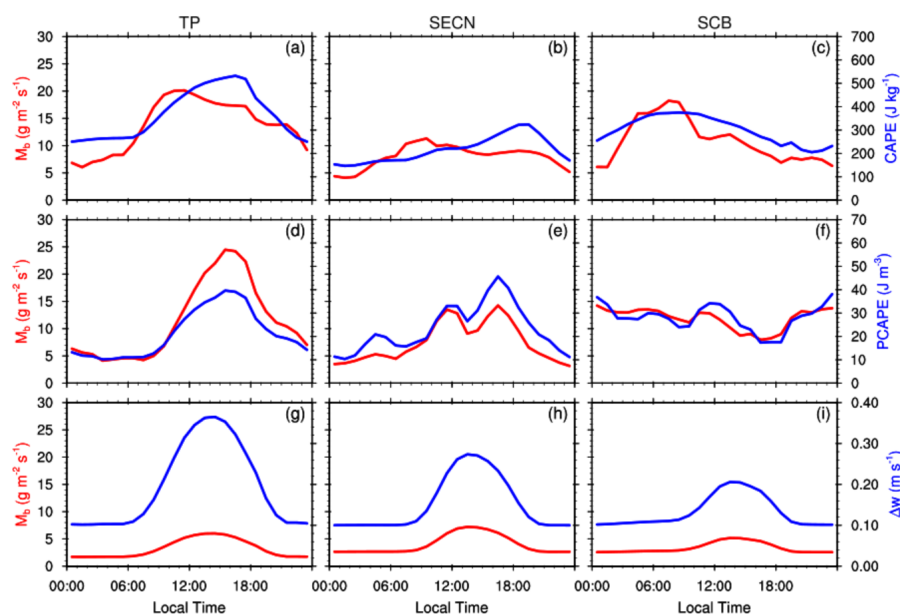
197 **Figure 1** Period-averaged (a-c) and case-composite (d-i) diurnal precipitation cycles over TP
198 (left), SECN (middle) and SCB (right) from IMERG, ERA5 and the SCM control simulation.
199 Panels (g–i) show convective precipitation. In each panel, the red, green, and blue lines represent
200 precipitation simulated by CAM_ZM, CAM_TDK, and CAM_UNICON, respectively.

201

202 To understand why SCM simulations produce noontime peaks in convective
203 precipitation, we show in Figure 2 the cloud-base cumulus mass flux M_b and closure-
204 related quantities in each convection scheme. Note that convective precipitation is
205 proportional to the product of M_b and the updraft rain/snow content, with M_b itself
206 determined by the standard quasi-equilibrium closure as $\frac{CAPE}{\tau * F}$, where τ is the
207 convective adjustment time scale and F is the CAPE consumption rate per unit mass
208 flux. As noticed, M_b is generally in phase with PCAPE (density-weighted CAPE) in
209 TDK, but shows little phase coherence with CAPE in ZM, both being computed within
210 their respective convection schemes during active convection. Over TP, the ZM M_b



211 peaks near 11:00 LST, well ahead of the CAPE maximum at 18:00 LST, due to the
212 lower F at that time (Figure S1; see Supplement for additional figures). The phase
213 advance of M_b relative to CAPE is even more pronounced over SECN (Figure 2b).
214 Higher CAPE is usually associated with deeper convection and greater vertically
215 accumulated condensate, which, together with the M_b peak, tends to produce dual
216 precipitation peaks in ZM (Figure 1h). The UNICON scheme does not rely on any
217 equilibrium assumptions; instead, it initializes the thermodynamic properties of updraft
218 plumes using the input surface flux. In all three regions, the UNICON M_b peaks around
219 14:00 LST, closely tracking the evolution of the launching vertical velocity derived
220 from the surface turbulent kinetic energy (TKE).



221
222 **Figure 2** Cloud-base cumulus mass flux M_b and closure-related quantities in the control
223 experiment of (a–c) CAM_ZM, (d–f) CAM_TDK and (g–i) CAM_UNICON over TP (left),
224 SECN (middle) and SCB (right). The closure-related quantities are CAPE for ZM, PCAPE for



225 TDK, and perturbations of updraft vertical velocity driven by symmetric turbulent eddies in the
226 UNICON scheme.

227

228 **3.2 Modification of Convection Schemes and Their impact on Diurnal** 229 **Precipitation**

230 The cloud-base mass flux in ZM and TDK is based on the standard equilibrium closure,
231 which assumes that CAPE is consumed at an exponential rate by cumulus convection
232 with characteristic time scale. Donner and Phillips (2003) and Zhang (2002) pointed
233 out that, although a CAPE quasi-equilibrium closure is well-suited for the free
234 troposphere, it does not apply well to the entire troposphere including the boundary
235 layer, particularly under rapidly varying boundary-layer forcing. Indeed, under weak
236 large-scale forcing, boundary layer generation of CAPE dominates surface-driven
237 convection (Bechtold et al., 2014). Hence, the strategy for improvement centers on how
238 to properly account for the imbalance between deep convection and PBL production.
239 Xie and Zhang (2000) addressed this by reducing M_b when dCAPE is below a
240 threshold value of $100 \text{ J kg}^{-1} \text{ hr}^{-1}$, with the reduction factor linearly related to dCAPE.
241 The underlying reasoning is that when large-scale dynamic forcing is weak, *i.e.*, small
242 positive dCAPE, the convective adjustment time scale is typically larger. Bechtold et
243 al. (2014) addressed this by subtracting from the total mass flux a fraction of shallow
244 convective contribution that is supposed to approximately balance the boundary layer
245 forcing. Park (2014a) attributed the inability to accurately simulate the DCP, in part, to
246 the absence of plume memory under the diagnostic quasi-equilibrium approximation.
247 He then developed an approach that incorporates additional plume memory through the



248 prognostic cold pool and employs this memory to reconstruct convective plumes at
249 subsequent time steps. Figure 3 presents M_b for each scheme in the sensitivity
250 experiment, with the aforementioned modifications applied. A notable change is that
251 for both TP and SECN, the M_b peak shifts to approximately 18:00 LST across all
252 schemes, matching the observed precipitation peak much more closely. Each scheme
253 accomplishes this improvement through a different mechanism. In CAM_ZM, dCAPE
254 (dashed lines in Figure 3a, b) peaks around 18:00 LST and is much smaller near noon,
255 which suppresses the early M_b peak in CTRL and shifts the main maximum toward
256 18:00 LST. In CAM_TDK, the PBL contribution to PCAPE peaks around 16:00 LST
257 (dashed lines in Figure 3d, e), accounting for a large portion of the total PCAPE at that
258 time. After subtracting this contribution, the resulting PCAPE and M_b peaks are
259 delayed to 17:00 LST over TP and 19:00 LST over SECN, approximately 1–2 hours
260 later than in CTRL. It is worth noting that the dual M_b peak in the control experiment
261 of both CAM_ZM and CAM_TDK over SECN now merges into a single peak. In
262 CAM_UNICON, while the updraft perturbations from symmetric turbulent eddies still
263 peak around noon, the enabled cold-pool-driven mesoscale flow generates pronounced
264 dynamic uplift forcing around 18:00 LST, ultimately shifting the M_b peak toward
265 18:00 LST.
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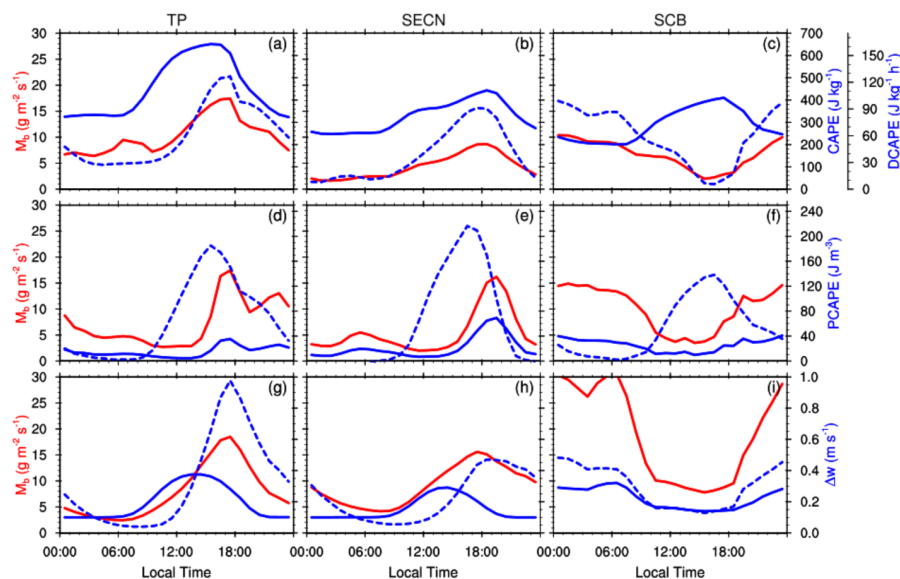
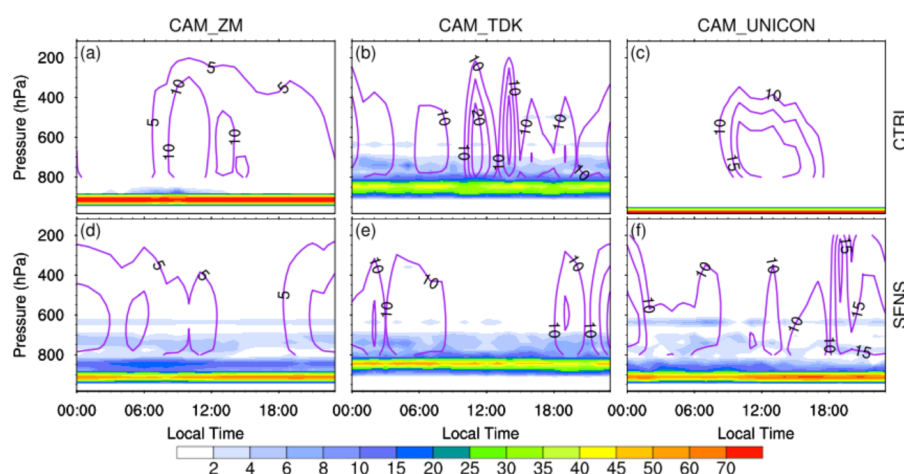


Figure 3 Same as Figure 2, but for the sensitivity experiments. The dashed blue lines in each row denote, respectively: dCAPE in ZM, the PBL contribution to PCAPE in TDK, and perturbations induced by subgrid cold pools and external mesoscale upflow in UNICON.

For elevated convection over SCB, Figure 4 compares the time-pressure cross-sections of the launching level frequency and updraft mass flux between the control and sensitivity runs. In all control simulations except CAM_TDK, afternoon convection prevails, whereas nocturnal convection remains very weak. Both CAM_ZM and CAM_UNICON restrict the launch level to near the surface, inhibiting convection initialization during nighttime because of the strong nocturnal inversion that develops above the PBL. CAM_TDK, however, is not subject to this restriction. It produces significant updraft mass during nighttime and early morning, frequently triggering elevated convection below 700 hPa (Figure 4b). Removing the boundary-layer origin constraint in ZM and UNICON leads to stronger nocturnal convection, weaker



282 afternoon convection, and a markedly raised launch level relative to the control runs
283 (Figure 4d, f). This improvement is also manifested in the diurnal variation of M_b
284 (rightmost column of Figure 3), with a maximum at night and a minimum around
285 midday. Most strikingly, CAM_TDK also yields stronger nocturnal convection in the
286 sensitivity run (Figure 4e), likely because the suppression of daytime convection favors
287 subsequent nocturnal development.



288
289 **Figure 4** Time-pressure cross-sections of launching-level frequency (shaded, units: %) and
290 updraft mass flux (contour, units: $\text{g m}^{-2} \text{s}^{-1}$) over SCB for the control (top) and sensitivity (bottom)
291 experiments. From left to right, the panels show results from CAM_ZM, CAM_TDK, and
292 CAM_UNICON, respectively.

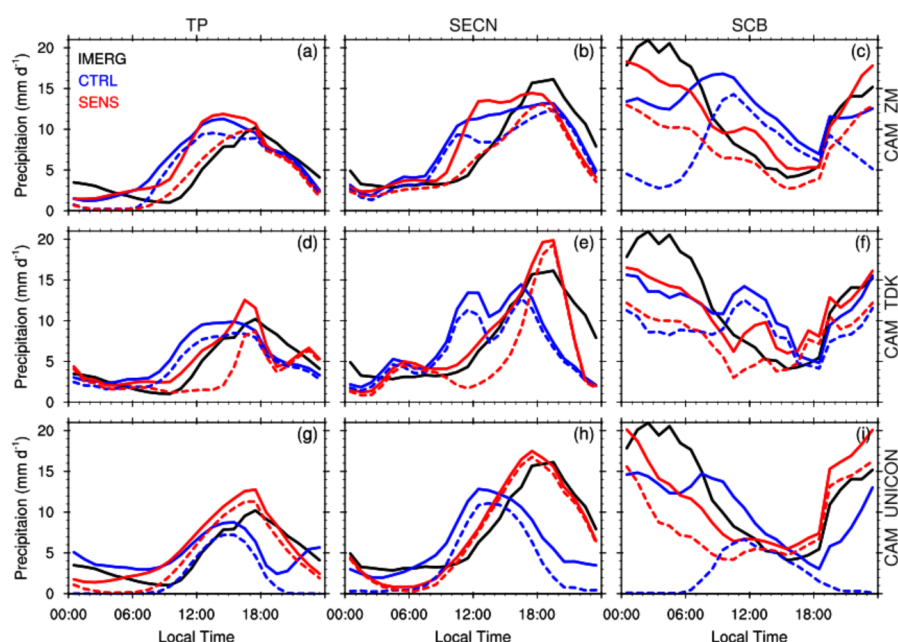
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294 Figure 5 compares the DCP simulation between CTRL and SENS experiments. As
295 anticipated, all modifications result in improved diurnal precipitation cycles when
296 compared with the control runs. For all three schemes, the premature afternoon
297 precipitation over TP and SECN is substantially delayed, aligning better with
298 observations. This delayed response is largely attributable to the delay in simulated
299 convective precipitation (dashed lines). Note the stronger precipitation in the



300 CAM_UNICON sensitivity experiment is due to positive perturbations in updraft
301 vertical velocity, temperature, and moisture, driven by mesoscale organized flow (Park,
302 2014b; Wang et al., 2026). Over SCB, the excessive afternoon rainfall and insufficient
303 nocturnal precipitation in the control experiments of CAM_ZM and CAM_UNICON
304 are substantially improved in the sensitivity experiments. Although no explicit
305 modification is made for elevated convection in CAM_TDK, nocturnal precipitation is
306 also enhanced in the sensitivity experiment (Figure 5f), presumably due to the reduced
307 afternoon convective stabilization resulting from the removal of shallow convective
308 contribution in the closure. A similar daytime-to-nocturnal convective interaction has
309 also been documented by Xie et al. (2019) and Park et al. (2024), who found that
310 suppressing daytime convection can strengthen nocturnal precipitation.

311



312



313 **Figure 5** Diurnal cycles of total (solid) and convective (dashed) precipitation over TP (left),
314 SECN (middle), and SCB for the control (blue) and sensitivity (red) runs of each convection
315 scheme. From top to bottom, the panels show DCP simulated by CAM_ZM, CAM_TDK, and
316 CAM_UNICON, respectively.

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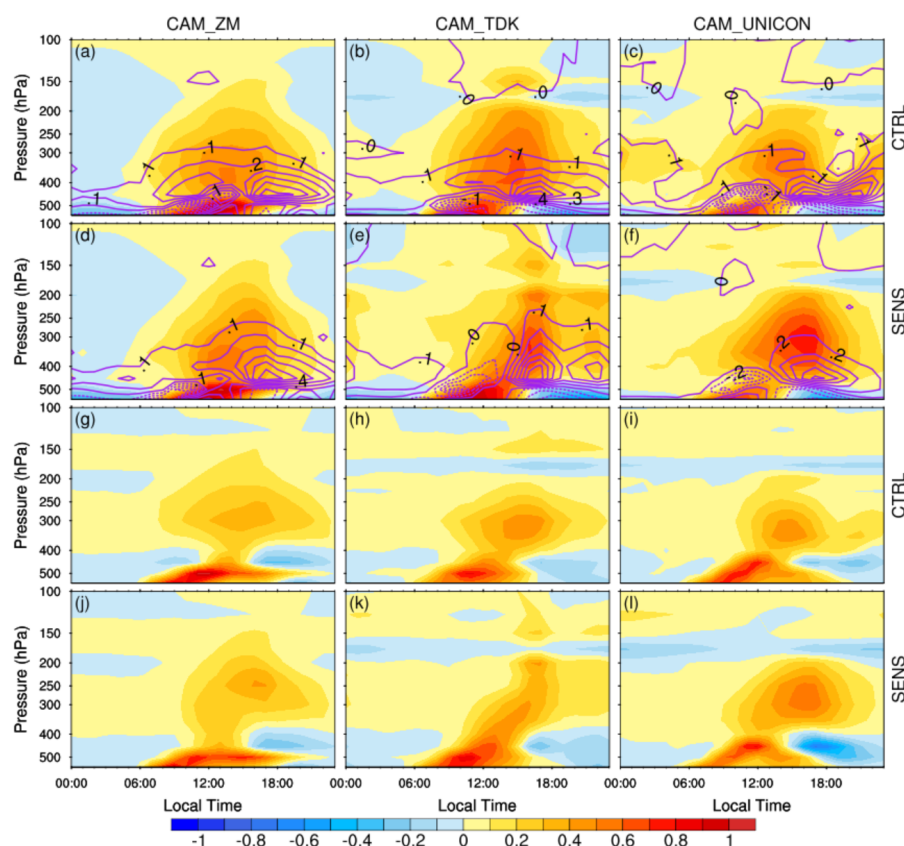
318 **4. Diurnal variation of Diabatic Heating**

319 To examine the impact of modified convection schemes on diabatic heating, Figure 6
320 presents the diurnal cycles of apparent heating Q_1 and moisture sink Q_2 over TP for
321 both control and sensitivity experiments. These quantities are computed as the sum of
322 tendencies from individual parameterizations. A common feature is the upward-tilting
323 Q_1 structure, which extends from PBL in the morning to mid-tropospheric levels later
324 in the day. This evolution features an early stage with positive Q_1 and negative Q_2 in
325 the lower troposphere, followed by a transition to positive values for both Q_1 and Q_2
326 aloft once deep convection develops. Although all SCMs reasonably reproduce this
327 upward-tilting structure, they differ notably in the timing and height of the peak heating.
328 Relative to their control experiments, CAM_TDK and CAM_UNICON delay and
329 elevate the heating peak, resulting in a more pronounced upward-tilt structure (Figure
330 6e, f). Moreover, CAM_UNICON exhibits a stronger heating amplitude, consistent
331 with its intensified convection. While the ZM heating center shows no clear phase shift
332 between the control and sensitivity experiments, it tends to extend upward to higher
333 altitudes over time in the sensitivity run (Figure 6d). Notably, the noontime Q_2 peak
334 near 400 hPa is strongly suppressed in all sensitivity experiments, thereby alleviating
335 the overly strong early-afternoon precipitation in the control. The lower positioning of
336 the Q_2 peak relative to Q_1 is indicative of pronounced subgrid-scale MSE transport,



337 which is presented in the bottom two rows of Figure 6 as the term $Q_1 - Q_2 - Q_{rad}$,
 338 where Q_{rad} denotes radiative heating. After sunrise, $Q_1 - Q_2 - Q_{rad}$ is initially
 339 positive and continues to rise due to subgrid-scale transport from turbulence and
 340 shallow convection. As the cycle progresses, a characteristic dipole structure emerges
 341 in its vertical profile—negative anomalies dominate below 400 hPa, while positive
 342 anomalies prevail aloft—a typical signature of well-organized deep convection. All
 343 three schemes show a more pronounced upward-tilting structure in the sensitivity
 344 experiment than in the control, owing to the delayed onset of deep convection.

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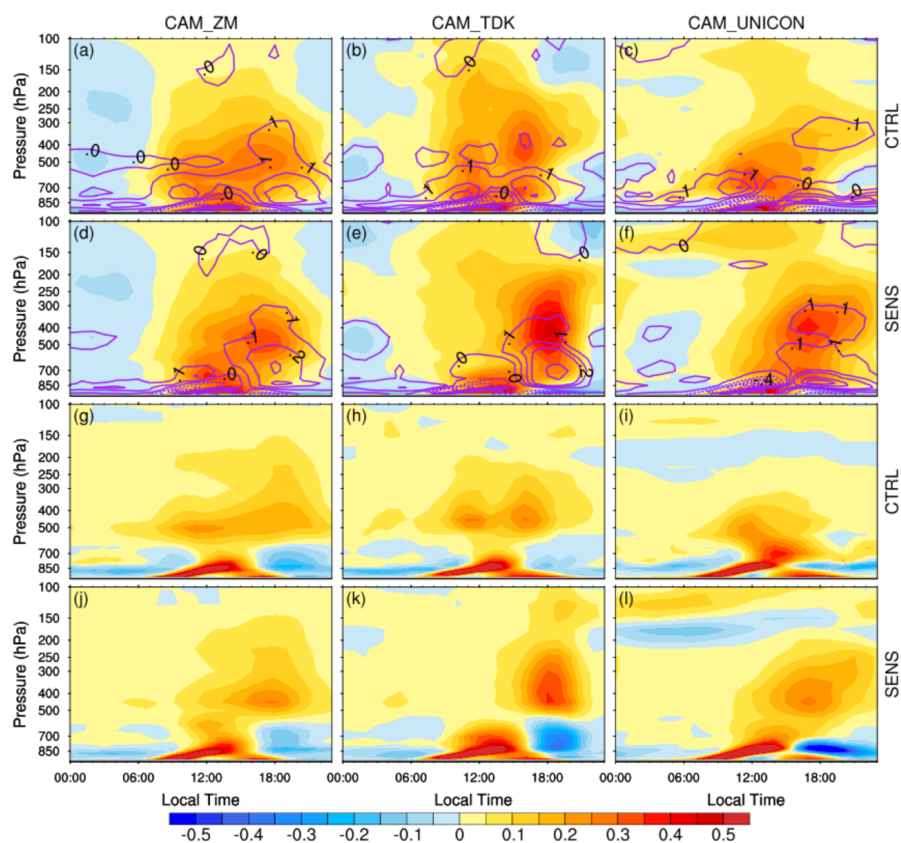
347 **Figure 6** The diurnal cycles of (a–f) apparent heating Q_1 (shaded), moisture sink Q_2 (contour)
348 and (g–l) subgrid-scale MSE transport $Q_1 - Q_2 - Q_{rad}$ over TP for both control and sensitivity
349 experiments. From left to right, the panels show results from CAM_ZM, CAM_TDK, and
350 CAM_UNICON, respectively. In the upper two panels, solid lines are for contours greater than
351 or equal to zero and dashed lines for contours less than zero. Units: K h^{-1} .

352

353 Similar discrepancies between control and sensitivity runs are also evident over SECN
354 (Figure 7). All sensitivity runs exhibit a suppression of the early Q_2 peak,
355 accompanied by a delayed and elevated Q_1 peak and a more pronounced
356 upward-tilting structure in subgrid-scale MSE transport. It is worth noting that
357 CAM_UNICON exhibits more coherent vertical organization in subgrid-scale MSE
358 transport than CAM_ZM and CAM_TDK, particularly during the transition from
359 shallow to deep convection, underscoring the benefit of UNICON's unified convection
360 framework. Figure 8 presents the diurnal cycles of diabatic heating over SCB, where
361 a notable difference from TP and SECN is the dominance of nocturnal heating. This
362 nocturnal heating in the control runs of CAM_ZM and CAM_UNICON is primarily
363 due to large-scale condensation, as indicated by the relatively weak subgrid-scale MSE
364 transport at night (Figure 8g, 8i). In contrast, the stronger MSE transport in CAM_TDK
365 (Figure 8h) suggests that convection contributes substantially to its heating. All
366 sensitivity experiments exhibit enhanced nocturnal convection, with both Q_1 and
367 subgrid-scale MSE transport strengthened during nighttime. Conversely, the diabatic
368 heating at noontime is substantially suppressed, in line with the reduced precipitation
369 at that time.



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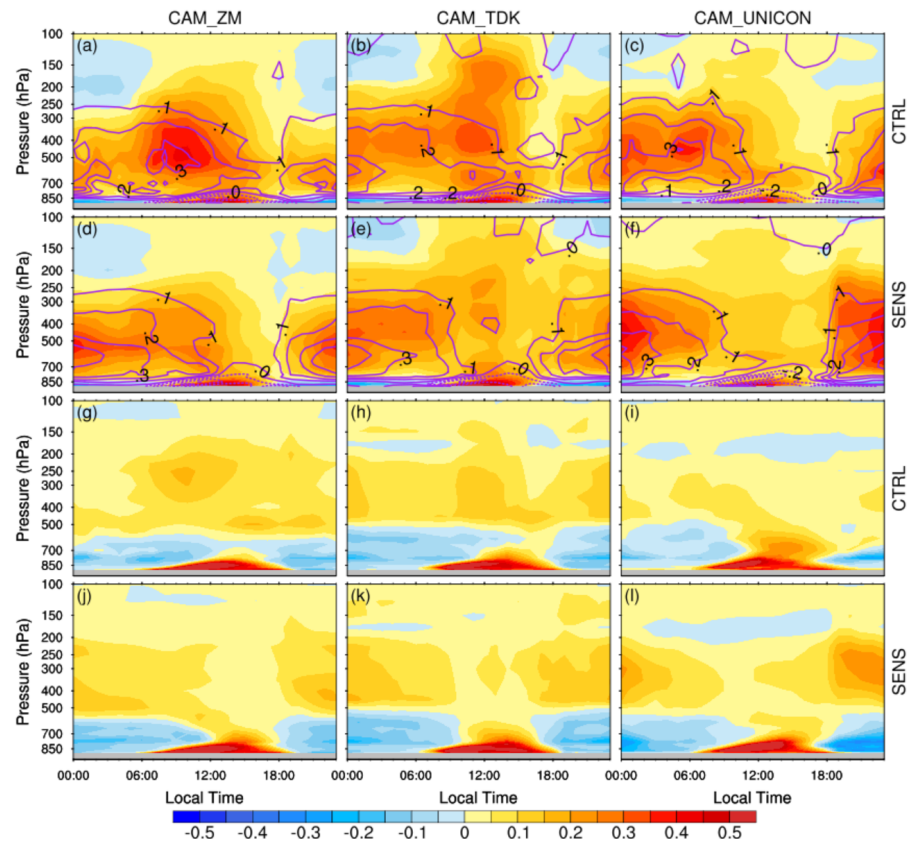
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372 **Figure 7** Same as Figure 6, but for SECN.

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376 **Figure 8** Same as Figure 6, but for SCB.

377

378 **5. Impact of Convection Scheme Modifications on Thermodynamic** 379 **Fields**

380 While the modified convection schemes improve the diurnal phase of precipitation and
381 associated diabatic heating, they also tend to produce deep and/or stronger convection,
382 particularly in CAM_UNICON. To assess whether these changes have an adverse
383 impact on temperature and moisture biases, we examine the model's thermodynamic
384 performance in this section.

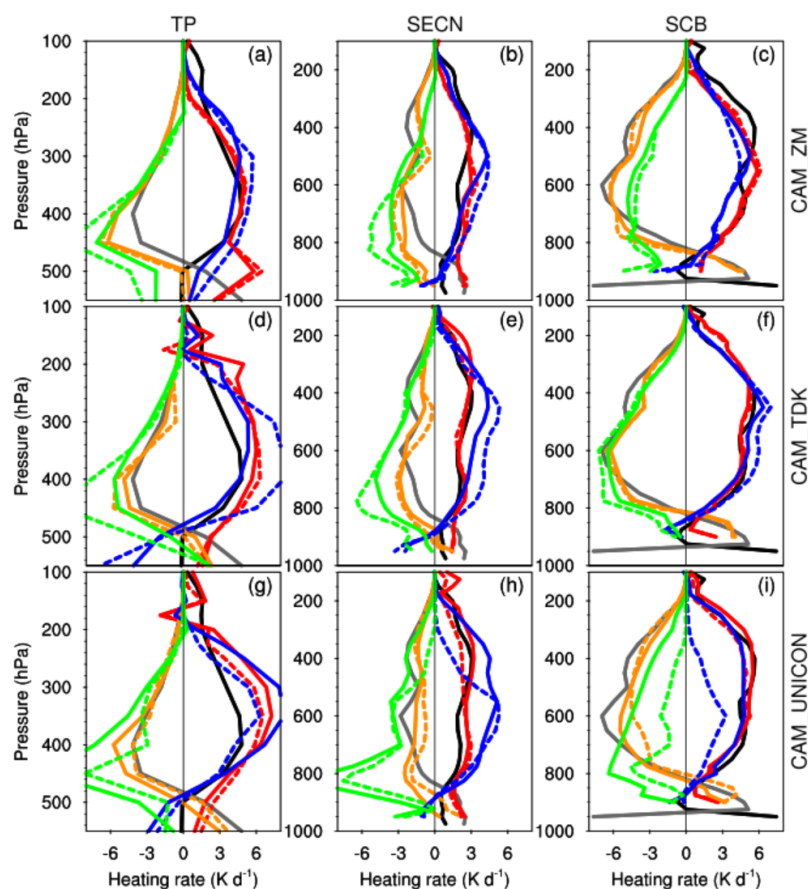


385 **5.1 Diurnally averaged Q_1 and Q_2 : Observation vs. Simulation**

386 Figure 9 compares the simulated diurnally averaged Q_1 and Q_2 profiles against
387 observations, along with their deep convective contributions in each SCM. The
388 observed Q_1 and Q_2 are derived from ERA5 reanalysis as residuals of the
389 thermodynamic equation and moisture continuity equation, following Yanai et al.
390 (1973). Relative to observations, the CAM_TDK control run over TP exhibits a positive
391 Q_1 bias of approximately 2 K d^{-1} throughout the troposphere and a negative Q_2 bias
392 of approximately 2 K d^{-1} between 400 and 450 hPa (Figure 9d). While these biases are
393 considerably alleviated in the sensitivity experiment below 450 hPa, a positive Q_1 bias
394 is exaggerated aloft, largely a consequence of convection penetrating to higher altitudes.
395 The CAM_UNICON control run exhibits a bias pattern similar to that of CAM_TDK
396 (Figure 9g), yet the biases are further amplified in the sensitivity run, as a result of the
397 strengthened deep-convective component. The ZM scheme shows a notably smaller Q_1
398 bias than TDK and UNICON above 450 hPa, yet exhibits a pronounced positive bias of
399 up to 6 K d^{-1} near 500 hPa (Figure 9a). While the Q_1 and Q_2 profiles remain broadly
400 comparable between the control and sensitivity experiments, the convective component
401 is notably weaker in the latter, implying that other physical processes (e.g., large-scale
402 condensation) compensate for the reduced convective heating (Wang and Zhang, 2013).
403 Over SECN, the Q_1 simulations from all SCMs show better overall performance than
404 over TP, with biases remaining mostly within $\pm 1 \text{ K d}^{-1}$. Compared with observations,
405 all simulations systematically overestimate Q_2 in the lower troposphere and
406 underestimate it in the upper troposphere. As over the TP, the convective component in



407 the sensitivity runs of CAM_TDK and CAM_ZM extends to higher altitudes, while in
408 CAM_UNICON it is both intensified and extends further upward. For nocturnal
409 convection over SCB, the ZM scheme performs satisfactorily for Q_2 , yet exhibits a
410 positive Q_1 bias below 500 hPa and a negative bias above. In contrast,
411 CAM_UNICON simulates Q_1 well, but substantially underestimates Q_2 across the
412 entire troposphere. In comparison, CAM_TDK exhibits overall good agreement with
413 observations for both Q_1 and Q_2 . While the modifications substantially improve Q_2
414 in UNICON, they yield no marked improvement in Q_1 for ZM.



415



Figure 9 Simulated profiles of Q_1 (red) and Q_2 (orange) compared with observations (black for Q_1 , gray for Q_2), along with the deep convective component of Q_1 (blue) and Q_2 (green) in each SCM. Dashed and solid lines denote the control and sensitivity experiments, respectively. From top to bottom, the panels show results from CAM_ZM, CAM_TDK, and CAM_UNICON, respectively. Units: K d^{-1} .

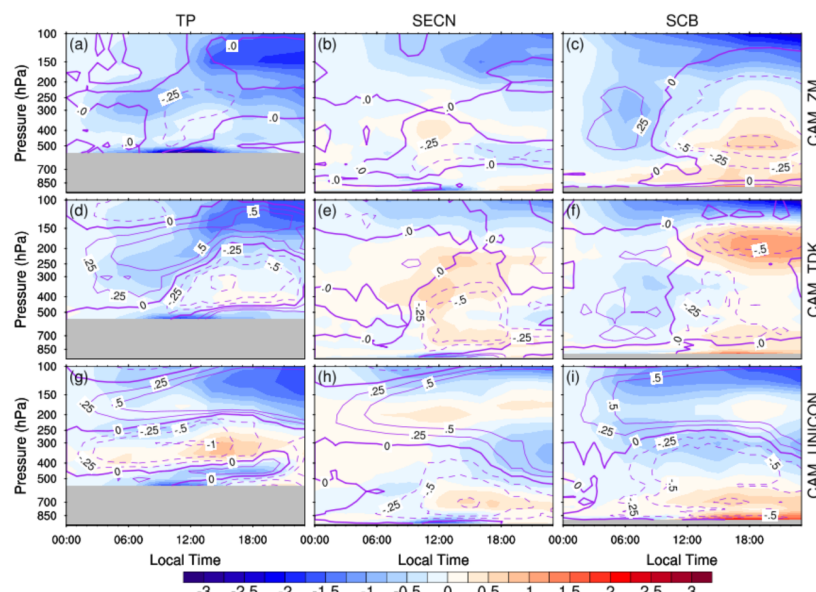
421

422 5.2 Temperature and moisture biases

423 Figures 10 and 11 show the temperature and moisture differences, respectively, between
424 the control run and ERA5 reanalysis (shaded), overlaid with the sensitivity-control
425 differences (contours). Over TP, all three schemes exhibit a band of cold bias of up to
426 2 K in the upper troposphere during the late afternoon, as well as cold biases of similar
427 magnitude near the surface at noon. Note that the surface cold bias is accompanied by
428 a dry bias of -1 g kg^{-1} , reflecting excessive subgrid-scale MSE transport at noon. A dry
429 bias is also found around 400 hPa after 18:00 LST in CAM_ZM and CAM_TDK, which
430 is consistent with their excessive precipitation relative to ERA5 (Figure 1d). In the
431 sensitivity experiments, the upper-level cold bias in CAM_TDK and CAM_UNICON
432 is substantially reduced (Figures 10d, g), consistent with the enhanced Q_1 above
433 250 hPa (Figures 9d, g). The ZM scheme exhibits no significant differences, owing to
434 the close resemblance in Q_1 between the two runs (Figure 9a). The late-afternoon dry
435 bias is also alleviated in the modified TDK and ZM schemes (Figures 11a, d). In
436 CAM_UNICON, the delayed onset and enhanced intensity of convection in the
437 sensitivity run cause the original mid-to-lower tropospheric wet bias into a dry bias
438 (Figure 11g, Figure S2g). The bias patterns over SECN are broadly analogous to those
439 over TP, characterized by a cold bias above 200 hPa, a cold and dry bias near the surface

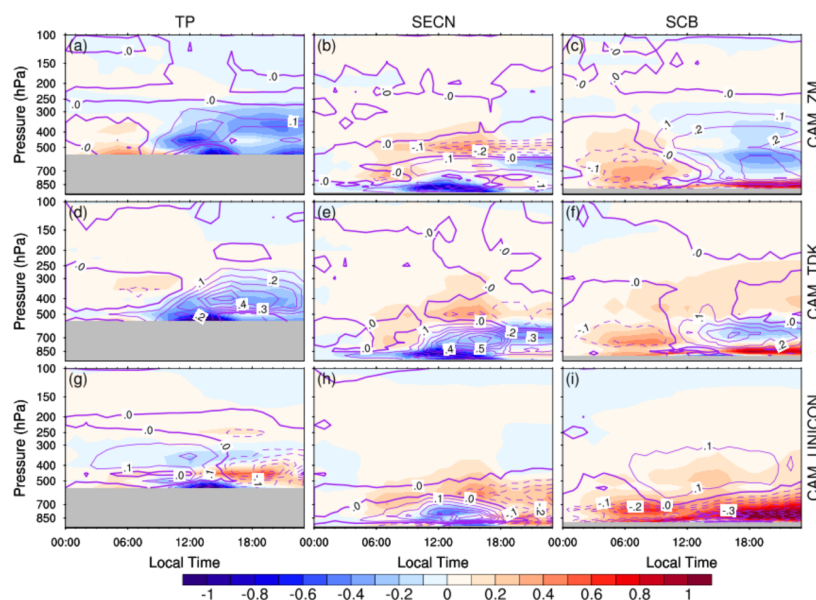


440 at noontime, and a negative moisture bias in the lower troposphere during the late
441 afternoon. Moreover, a warming bias between 200 and 700 hPa around noon is present
442 in all three schemes, which is linked to excessive early-afternoon convection. This bias
443 is largely alleviated in the modified runs, as reflected by the negative SENS-CTRL
444 differences (dashed lines in the middle column of Figure 10), owing to the suppression
445 of early convective activity. As over the TP, the lower-tropospheric dry bias after 18:00
446 LST is also reduced in the modified TDK and ZM schemes (Figure 11 b, e). In
447 CAM_UNICON, the temperature differences between sensitivity and control runs are
448 generally opposite in sign to the control bias, largely offsetting the bias (Figure 10h).
449 Nevertheless, the magnitude of these differences is so large that the sign is reversed
450 completely, leaving a residual warm bias aloft and a cold bias in the lower troposphere
451 (Figure S3h). Over SCB, the mid-tropospheric warm and dry bias in CAM_ZM is
452 substantially alleviated in the sensitivity run (Figure 10c; Figure 11c), owing to the
453 suppression of overly active early-afternoon convection. However, in the UNICON
454 sensitivity run, the overly strengthened convection warms the upper troposphere,
455 yielding a positive bias of up to 2.5 K (Figure S3i), similar to that over TP and SECN.
456 This raises a caveat for the UNICON scheme: while enabling cold-pool feedback may
457 improve the diurnal cycle, the overly strong intensity, stemming from positive
458 perturbations in plume initialization, may adversely affect thermodynamic fields.



459

460 **Figure 10** Temperature differences between the control run and ERA5 reanalysis (CTRL – ERA5,
461 shading) and between the sensitivity and control runs (SENS – CTRL, contours) over TP (left),
462 SECN (middle), and SCB (right). Panels from top to bottom show results from CAM_ZM,
463 CAM_TDK, and CAM_UNICON. In these figures, solid lines are for contours greater than or
464 equal to zero and dashed lines for contours less than zero. Units: K.



465

466 **Figure 11** Same as Figure 10, but for specific humidity (units: g kg^{-1}).



467

468 **6. Conclusions**

469 Using a single-column model, this study investigates the impact of convective
470 trigger and closure treatments on the summertime diurnal cycle of precipitation over
471 China, with a focus on three convection schemes: ZM, TDK, and UNICON. The
472 analysis targets two regions characterized by late-afternoon rainfall (TP and SECN) and
473 one region with a nocturnal regime (SCB). Over TP and SECN, the TDK and ZM
474 schemes with the standard CAPE closure simulate precipitation peaks that are
475 substantially earlier than the observed. The UNICON scheme, with cold-pool feedback
476 deactivated, also exhibits a noontime peak, albeit with a somewhat delayed initiation
477 relative to ZM and TDK. Over SCB, both ZM and UNICON produce a pronounced
478 noontime precipitation peak, in stark contrast to the observed nocturnal maximum.
479 TDK reasonably reproduces the nocturnal peak, yet it also produces an additional
480 noontime maximum. These SCM precipitation behaviors are closely related to closure
481 assumptions and the resulting cloud-base mass flux. While convective precipitation in
482 ZM and TDK is more closely tied to CAPE, in UNICON it aligns more directly with
483 the surface fluxes.

484 Modifications were then applied to each scheme: adopting the dCAPE trigger and
485 permitting elevated convection in ZM; activating cold-pool feedback and allowing
486 elevated convection in UNICON; and applying the non-equilibrium closure in TDK. A
487 notable improvement is evident for both late-afternoon and nocturnal rainfall. Over TP
488 and SECN, the precipitation peak shifts to approximately 18:00 LST across all schemes,



489 aligning much more closely with the observation. Each scheme achieves this
490 improvement through a distinct mechanism. In ZM, the dCAPE constraint suppresses
491 the early peak at 10:00 LST and redirects the main peak toward 18:00 LST. In TDK,
492 subtracting the boundary-layer contribution to PCAPE delays the precipitation peak to
493 19:00 LST. In UNICON, the cold-pool-driven mesoscale flow generates pronounced
494 dynamic uplift forcing around 18:00 LST, ultimately shifting the peak toward that time.
495 Over SCB, permitting convection to be launched above PBL in ZM and UNICON
496 enhances nocturnal convection while suppressing afternoon convection. Although no
497 explicit modification is made for elevated convection in TDK, applying the
498 non-equilibrium closure also enhances nocturnal precipitation, due to the reduced
499 convective stabilization in the afternoon.

500 The improved diurnal cycle is manifested in the diabatic heating fields as well. All
501 modifications lead to a delayed and more pronounced upward-tilting Q_1 heating
502 structure. Subgrid-scale MSE transport is likewise delayed and more upward tilted, with
503 ZM and TDK extending to higher altitudes, while in UNICON it is both intensified and
504 extends further upward. These changes alleviate the upper-level cold bias in TDK and
505 ZM and largely reduce the late-afternoon dry bias in the lower troposphere. However,
506 the overly strengthened convection in UNICON yields a warm bias of up to 2.5 K in
507 the upper troposphere, as the positive perturbations in plume properties induced by
508 mesoscale organized flow are too strong and deteriorates the thermodynamic fields.



509 **Data availability**

510 The simulation data used in this study are available at <https://doi.org/10.5281/zenodo.21207758> (Wang
511 and Liu, 2026).

512 **Author contribution**

513 XW: conceptualization, funding acquisition, methodology, validation, writing (original draft, review
514 and editing). YL: data curation, visualization, writing (review and editing). YL: resources, writing
515 (review and editing).

516 **Competing interests**

517 The authors declare that they have no conflict of interest.

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529

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