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Boundary Layer Recovery and Precipitation Symmetrization Preceding Rapid Intensification of Tropical Cyclones under Shear

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Abstract

22

23 This study investigates the precipitation symmetrization preceding rapid intensification (RI)

24 of tropical cyclones (TCs) experiencing vertical wind shear by analyzing numerical simulations

25 of Typhoon Mujigae (2015) with warm (CTL) and relatively cool (S1) sea surface temperatures

26 (SSTs). A novel finding is that precipitation symmetrization is maintained by the continuous

27 development of deep convection along the inward flank of a convective precipitation shield

28 (CPS), especially in the downwind part. Beneath the CPS, downdrafts flush the boundary layer

29 with low-entropy parcels. These low-entropy parcels do not necessarily weaken the TCs; instead,

30 they are "recycled" in the TC circulation, gradually recovered by positive enthalpy fluxes, and

31 develop into convection during their propagation toward a downshear convergence zone. Along-

32 trajectory vertical momentum budget analyses reveal the predominant role of buoyancy

33 acceleration in the convective development in both experiments. The boundary layer recovery is

34 more efficient for warmer SST, and the stronger buoyancy acceleration accounts for the higher

35 probability of these parcels developing into deep convection in the downwind part of the CPS,

36 which helps maintain the precipitation symmetrization in CTL. In contrast, less efficient

37 boundary layer recovery and less upshear deep convection hinder the precipitation

38 symmetrization in S1. These findings highlight the key role of boundary layer recovery in

39 regulating the precipitation symmetrization and upshear deep convection, which further accounts

40 for an earlier RI onset timing of the CTL TC. The inward rebuilding pathway also illuminates

41 why deep convection is preferentially located inside the radius of maximum wind of sheared TCs

42 undergoing RI.

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1. Introduction

Accurate forecasts of rapid intensification (RI) of tropical cyclones (TCs) remain a challenging task (DeMaria et al. 2014), especially under moderate vertical wind shear (VWS) (Bhatia and Nolan 2013). Whether and when a TC is going to undergo RI under moderate VWS depends crucially on the other environmental factors including sea surface temperature and environment humidity (Tao and Zhang 2014). This forecast challenge becomes even more acute for prelandfall RI forecasts. One recent example in point is Hurricane Michael (2018), which underwent an unexpected prelandfall RI[1] under moderate VWS and became a category-5 hurricane near landfall.

In the presence of VWS, dry TC-like vortices are usually vertically tilted (Jones 1995, 2004). When coupled with moist processes, the balanced mesoscale lifting associated with the tilted vortex organizes a convective precipitation shield (CPS) in the downtilt side (e.g., Wang and Holland 1996; Corbosiero and Molinari 2002; Reasor et al. 2013; Gu et al. 2019). Observational and modeling studies of early-stage TCs[2] under moderate VWS indicate a common feature before RI onset: as the midlevel TC vortex precesses from the downshear-left to upshear and the TC vortex becomes nearly vertically aligned, the CPS also propagates into the upshear flank and spirals inward toward the formation of an incipient eyewall (e.g., Rappin and Nolan 2012; Zhang and Tao 2013; Alvey et al. 2015; Rogers et al. 2016; Chen et al. 2017; Leighton et al. 2018; Rios-Berrios et al. 2018; Ryglicki et al. 2018). This process is also termed precipitation symmetrization. The radius of maximum wind (RMW) contracts inward significantly during the vortex alignment and precipitation symmetrization, as documented in

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[1] The prelandfall RI means that the RI occurs within 1-2 days of landfall and the RI can last to the landfall time.
[2] Early-stage TCs include the tropical depressions, tropical storms, and Category-1 hurricanes.

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64 previous studies (e.g., Judt and Chen 2016; Chen et al. 2018a, b; Miyamoto and Nolan 2018;

65 Tang et al. 2019). The coupled inner-core structural and precipitation evolution preceding RI

66 onset for TCs in shear indicates the necessity to examine these processes in an integrated

67 framework before proposing a theoretical explanation for RI onset in shear.

68 The contribution of the deep convection in the CPS to vortex alignment has been examined

69 in previous studies, and two different pathways have been proposed, either through the

70 downshear reformation that involves a newly formed center (Molinari et al. 2004; Molinari et al.

71 2006; Nguyen and Molinari 2015; Chen et al. 2018b; Rogers et al. 2020) or through an inner-

72 core vorticity "restructuring" process (Rios-Berrios et al. 2018; Miyamoto and Nolan 2018;

73 Shimada and Horinouchi 2018). Both pathways involve sustained deep convection in the

74 azimuthally-propagating CPS and continuous merger of convectively-induced vorticity

75 anomalies. In an analytical study using a shallow-water model, Schecter (2020) demonstrated

76 that the two different pathways mentioned above can be explained by the relative strength

77 between the velocity-convergence generated by the mass sink on the downshear side (cf. the

78 downshear convergence zone in Chen et al. 2018b) and a critical value that is determined by the

79 horizontal scale of the mass sink as well as the absolute value of the drift velocity of the mass

80 sink relative to the background cyclonic flow.

81 The linkage between the precipitation symmetrization and RMW contraction was also

82 examined in a modeling study (Chen et al. 2018a). By performing a set of numerical simulations

83 for Typhoon Mujigae (2015) over various sea surface temperatures (SSTs), Chen et al. (2018a)

84 found that TCs over different SSTs all undergo vertical alignment and precipitation

85 symmetrization before RI onset. However, over warmer SST TCs exhibit a higher degree of

86 precipitation symmetry, and the RMW contraction and RI occur much earlier. Diagnoses using

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87 the Sawyer-Eliassen equation indicate that the stronger diabatic heating due to more midlevel

88 and deep convection within the inner core (also within the CPS) contributes to the earlier RMW

89 contraction of the TCs over warmer SST. These results are consistent with earlier analytical

90 analyses invoking balanced dynamics in that diabatic heating near/inside of the RMW benefits

91 the RMW contraction and TC intensification (Schubert and Hack 1982; Pendergrass and

92 Willoughby 2009).

93   Given the pivotal role of deep convection in both vortex alignment and RMW contraction,

94 understanding the mechanisms that maintain the deep convection in the CPS during precipitation

95 symmetrization is key. Convective downdrafts can bring low-entropy air parcels into the

96 boundary layer and cool the inflow layer (Tang and Emanuel 2010; Riemer et al. 2010; Zhang et

97 al. 2013; Gu et al. 2015; Wadler et al. 2018b; Chen et al. 2019), i.e., the low-level ventilation,

98 which is argued as the most detrimental pathway of VWS to weaken a TC (Riemer et al. 2010,

99 2013). The boundary layer recovery of these downdraft-cooled parcels by the surface enthalpy

100 fluxes is argued as the key to compensate for the low-level ventilation and impacts the

101 subsequent TC intensity change (Powell 1990; Tang and Emanuel 2012; Molinari et al. 2013;

102 Zhang et al. 2017b; Zhang and Rogers 2019; Nguyen et al. 2019). However, the linkage between

103 the boundary layer recovery and convective development in sheared TCs remains elusive: recent

104 idealized simulations with the same SST attribute convective initiation in the azimuthally-

105 propagating CPS before RI onset to dynamical forcing, rather than buoyancy forcing (Gu et al.

106 2019).

107   As a follow-up of Chen et al. (2018a), this study will further examine the numerical

108 simulation dataset for Typhoon Mujigae (2015) over various SSTs. It is hypothesized that the

109 boundary layer recovery is more effective under warmer SST conditions, and a comparison of

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110 two representative experiments with warm and relatively cool SSTs provides a unique

111 opportunity to gain insight into the role of boundary layer recovery in governing the distribution

112 of deep convection within the TC inner core, which is the key to further understand the

113 relationship between precipitation symmetrization, RMW contraction, and RI onset for sheared

114 TCs. The specific scientific questions to be addressed in this study include:

115   1) Before RI onset, how is the CPS organized and maintained during precipitation

116   symmetrization under VWS?

117   2) What is the role of boundary layer recovery in the convective development and

118   precipitation symmetrization?

119   3) What is the relative importance between dynamical and buoyancy forcing in the

120   convective development?

121   The remainder of this paper is organized as follows. Section 2 describes the methods and

122 simulation datasets used in this study. Section 3 provides an overview of the vortex intensity and

123 structural change, and precipitation evolution prior to the RI onset in warm and relatively cool

124 SST experiments. Section 4 compares the organization of the CPS during precipitation

125 symmetrization in the two experiments. The role of boundary layer recovery in the convective

126 development and precipitation symmetrization is discussed in section 5. Additional discussion

127 and concluding remarks are given in sections 6 and 7, respectively.

128 **2. Data and Methods**

129   Following Chen et al. (2018a), the same two representative WRF-ARW experiments of

130 Typhoon Mujigae (2015) with warm (CTL) and relatively cool (S1) SSTs are compared in this

131 study. The model setup for these two experiments is the same except for the initial SST. In CTL,

132 the SST is set as the initial condition at 0000 UTC 2 October, while in S1 the SST is set as the

133   climatological monthly-mean state (MMSST) averaged from 1990 to 2013. The MMSST

134   averaged in the South China Sea is 28.6°C, 1°C cooler than that in CTL. For simplicity, the

135   ocean coupling is not included and the SST is not updated during the simulation in these

136   experiments. The horizontal resolution of the triple-nested domains is 12, 4, and 1.33 km,

137   respectively. The outermost domain is static and the inner two domains move with the simulated

138   TC. All three domains contain 51 sigma levels with the top level at 50 hPa. The CTL simulation

139   successfully captures the track, prelandfall RI, and storm structure evolution. In comparison, the

140   RI onset timing lags by 13 h in the S1 experiment and the intensification rate is much weaker.

141   For more details of model setup, verification, and differences in the two experiments, we refer

142   interested readers to Chen et al. (2018a).

143   The objective partitioning method proposed by Rogers (2010) is adopted to separate the

144   convective, stratiform, and other (typically flanking the stratiform) type precipitation. This

145   method uses reflectivity criteria at 0.9 and 3 km heights and a threshold of vertical velocity

146   averaged between 0.9 and 2.1 km (i.e., >0.5 m s^{-1}) to identify convective points. If one grid point

147   is not flagged as a convective point and 3-km reflectivity is >20 dBZ, it is flagged as stratiform

148   precipitation [for more details, see Rogers (2010)]. Then, based on the height of cloud top,

149   indicated by 20-dBZ echo top, the convective region is further divided into shallow (< 4 km),

150   midlevel (4-8 km), and deep (>8 km) convection following Fritz et al. (2016). We also pay

151   attention to one type of extreme deep convection, namely convective bursts (CBs) and adopt the

152   definition proposed by Rogers (2010). A CB is defined as a grid point where the layer-averaged

153   vertical velocity within the 300–700-hPa layer exceeds 5 m s^{-1}.

154   A forward trajectory analysis is performed in section 5 to investigate the boundary layer

155   recovery of the downdraft-cooled parcels. To compute the trajectory of air parcels ending in

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156   rapidly changing convection, output from the innermost domain is saved every 1 min. The

157   parcels to be tracked are selected near the TC center and beneath the vortex-tilt-related CPS, and

158   their trajectories over the analysis period do not cross the boundary of the innermost moving

159   nest. The predictor-corrector technique is used for the trajectory calculation, following

160   Onderlinde and Nolan (2016). The essence of this technique is to use the wind information at the

161   predicted midpoint to advect the parcel at the initial location for a full-time step. The predicted

162   midpoint is determined by advecting the parcel from the initial location by a half time step using

163   the wind information at the initial location. The time step selected is 30s, since a higher temporal

164   resolution produces similar results.

165   The TC center at a given pressure level is defined as the geopotential height centroid (Chen

166   et al. 2018b), which is skillful at locating the TC center for weak storms (Nguyen et al. 2014).

167   Vortex tilt is calculated as the distance between the TC centers at 450 hPa and 850 hPa pressure

168   levels.

169   **3. Overview of vortex intensity and structural change prior to RI onset**

170   Typhoon Mujigae underwent RI over the warm water in the South China Sea under

171   moderate VWS. The magnitude of 200-850 hPa VWS remained 7-8 m s^{-1} before RI onset[3] (i.e.,

172   0000 UTC 3 October). Note that the 200–850-hPa VWS is calculated for the area between 200

173   and 800 km from the surface TC center. The VWS subsequently decreased to 4-5 m s^{-1} during

174   the early RI period and then increased to 7-9 m s^{-1} near the end of RI, which was followed by the

175   landfall near 0600 UTC 4 October. Both simulations have similar VWS evolution to

176   observations. The CTL TC successfully reproduces the intensity evolution, and its RI lasts from

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[3] RI onset in this study is defined as the time when the increase in 10-m maximum wind speed (VMAX) exceeds 15 m s^{-1} in the subsequent 24 h or shorter period, if the RI duration is less than 1 day. An additional requirement is that the VMAX should increase in the first 6 h of the subsequent 24-h or shorter period.

177 0000 UTC 3 October to 0300 UTC 4 October (Fig. 1a). In comparison, the RI onset of the S1 TC

178 is delayed by 13 h. The RI duration of the S1 TC is ~14h, as the maximum surface wind

179 increases from 29 m s^{-1} at 1300 UTC 3 October to 46 m s^{-1} at 0300 UTC 4 October.

180   In addition to the similar evolution of VWS in the two experiments, the CTL and S1 TCs

181 undergo similar vertical alignment prior to RI onset, as the midlevel vortex gradually precesses

182 from the downshear-left quadrant to upshear, and the magnitude of the 400-850 hPa TC vortex

183 tilt decreases with time (Fig. 1b). However, the different RI onset timing between the two TCs

184 suggests vertical alignment, albeit necessary, is not a sufficient RI indicator under moderate

185 VWS. Instead, RI onset in both experiments is effectively indicated when the RMW contraction

186 (Fig. 1d) reaches a certain threshold measured in terms of the local Rossby number at 10-m

187 height (R_o > 12, Fig. 1c). The local Rossby number is defined as $R_o = v_m/(r_m f)$, where v_m is

188 the maximum azimuthal mean tangential wind at 10-m height, r_m represents the RMW, and f

189 represents the Coriolis parameter at the TC center. Of note, the R_o threshold (i.e., R_o > 12)

190 cannot be overgeneralized beyond this case, which is latitude dependent based on its definition.

191 Additionally, the value of R_o at RI onset varies by changing the size and intensity of the initial

192 vortex, translational speed, and VWS magnitude (Miyamoto and Nolan 2018). Nonetheless, this

193 metric shows that the CTL TC contracts much earlier and its RI starts much earlier too.

194   Figure 2 shows three snapshots of the simulated radar reflectivity and background

195 horizontal convergence at 1.5-km height for the CTL and S1 experiments. Concurrent with

196 vertical alignment, both TCs undergo precipitation symmetrization before RI onset. The

197 background horizontal convergence is computed with the coarser data resolution of 0.5°× 0.5°

198 that is interpolated from the outermost model domain with a horizontal resolution of 12 km,

199 following Chen et al. (2018b). A mesoscale convergence zone exists in the downshear quadrants

200 of both TCs, which is consistent with the findings in the simulated Typhoon Vicente (2012)

201 (Chen et al. 2018b). The formation of the downshear convergence zone can be explained by the

202 differential vorticity advection by VWS, which induces mesoscale lifting and low-level

203 convergence in the downshear side (Bender 1997; Bracken and Bosart 2000). The convergence

204 zone remains in the downshear side during precipitation symmetrization.

205 **4. Inward rebuilding of CPS during precipitation symmetrization**

206   In this section, we mainly focus on the comparison between CTL and S1 experiments over

207 the 12-h period preceding the RI onset of the CTL TC (i.e., 1200 UTC 2 October−0000 UTC 3

208 October) following Chen et al. (2018a). This is the period when the RMW evolution differs

209 between the two experiments (Fig. 1d), which further impacts the RI onset timing (Fig. 1a).

210   Figures 3a-d show the location of deep convection and CBs within $r = 100$ km in three

211 consecutive 3-h periods at a 10-min interval after 1200 UTC 2 October. In both TCs, deep

212 convection propagates azimuthally from downshear to upshear and meanwhile radially inward

213 toward the TC center over the three periods. The evolution of CBs location over the same

214 periods exhibits similar features in CTL, while the radially inward shift of CBs location is less

215 notable in S1. In this study, *we define the inward rebuilding of the CPS as deep convection*

216 *continuously develops at the inward flank and downwind part of the CPS during precipitation*

217 *symmetrization*. The inward rebuilding process in these two experiments is a reminiscence of the

218 "inward progression" of cloud-to-ground lightning clusters from large radii downshear to smaller

219 radii upshear in RI TCs under moderate VWS (e.g., Molinari et al. 2004; Molinari and Vollaro

220 2010; Stevenson et al. 2014; Zawislak et al. 2016). The inward rebuilding is more notable in

221 CTL, and the CTL TC has much more (1091) grid points of deep convection within the inner-

222 core region (i.e., $r = 60$ km), particularly in the upshear side (Figs. 3a-b and 3e). The CTL TC

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223 also has 90 more CBs within the inner core (Fig. 3f), and the difference in the number of CBs

224 between the two TCs is most prominent in the downshear-left quadrant. Nevertheless, the CTL

225 TC has slightly more CBs in the upshear-left quadrant (see Figs. 3c-d and 3f). The difference in

226 the deep convection or CBs in the upshear-left quadrant between the two experiments is

227 consistent with previous observational studies that found deep convection in the upshear-left

228 quadrant is key to determining subsequent intensity change (e.g., Wadler et al. 2018a).

229   An examination of the animation of radar reflectivity at the lowest model level indicates

230 that the newly-developed deep convection related to inward rebuilding are mostly initiate in the

231 downshear-left quadrant inside the RMW and mature along their path toward the downshear

232 convergence zone (see the supplemental movie). Figure 4 presents two examples for the CTL

233 and S1 TCs. In the CTL TC, a 40-50-km long spiral rainband (circled by a thick dashed line) is

234 visible in the downshear-right quadrant at 1710 UTC 2 October (Figs. 4a-b). Of note, this spiral

235 rainband is only visible below the lowest 1.5 km at this moment (not shown), suggesting that it

236 remains in the boundary layer. This rainband gradually develops above the boundary layer and

237 matures (>55 dBZ) during the propagation from the downshear-left to upshear-left quadrant

238 (Figs. 4c-f). The 40-50-km long spiral rainband then becomes the leading edge of the

239 azimuthally propagating CPS as preexisting convection weakens due to its own lifecycle,

240 resulting in an inward rebuilding event. Figures 4g-l show a similar phenomenon occurring over

241 a later period for the S1 TC, i.e., from 1920 UTC 2 October to 2030 UTC 2 October. A notable

242 difference is that the newly-developed spiral rainband in the S1 TC is much weaker than that of

243 the CTL TC in terms of the radar reflectivity. Thus, the convective activity at the downwind part

244 of the CPS in S1 is weaker than that in CTL after this inward rebuilding event (Figs. 4k-l).

245   Figure 5 further compares the composite vertical structure of the newly-developed deep

246   convection between the CTL and S1 TCs over the period of inward rebuilding (see Fig. 4).

247   Dashed lines in Figs. 5a-b mark the locations of the vertical slices in Figs. 5c-f. At $r = 50$ km, the

248   maximum microphysics diabatic heating of the discrete convective tower within the 40-50 km

249   long spiral rainband in CTL exceeds 40 K h^{-1} and the top of the strong diabatic heating ($>$40 K h^{-}

250   ^{1}) extends to 11 km height (Fig. 5c), indicating that the newly-developed convection (Fig. 5a) is

251   evolving toward its mature stage during its propagation toward the upshear-left quadrant (see

252   Figs. 4d-f). In comparison, the strong diabatic heating ($>$40 K h^{-1}) of the newly-developed deep

253   convection in S1 is vertically confined in the 5-8 km layer at $r = \sim58$ km. The region of larger

254   value of absolute vorticity ($>0.5\times10^{-3}$ s^{-1}) outside of $r = 40$ km in S1 is ~4 km shallower than that

255   in CTL (Figs. 5c-d).

256   The composite storm-relative streamlines in Figs. 5e-f indicate two sources of convective

257   updrafts for the matured newly-developed deep convection within $r = 60$ km (Figs. 5e-f). The

258   first source comes from the radial inflow jet ($\theta_e < 354$ K) that descends from the freezing level

259   into the boundary layer, pass through the high-entropy ($\theta_e >358$ K) central area within $r = 40$ km,

260   and then becomes outflow above the boundary layer. The inflow to outflow transition is

261   indicative of supergradient wind. The second source is directly traced back to the "eye" region

262   within the lowest 2 km, which is closely related to the inward rebuilding process. Of note, the

263   incipient eyewall with a clear eye appears 2 hours later than the composite period (e.g., Fig. 2c).

264   A comparison of the red streamline in Figs. 5e-f demonstrates that the maximum height related

265   to the second source of convective updrafts differs between the two experiments, as convective

266   updrafts in CTL vertically extend to $\sim$17 km height, 5 km taller than the updrafts in S1. These

267   findings in Fig. 5 again suggest that the discrepancies in the strength of newly-developed deep

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268 convection inside of the RMW affect the convective activity at the downwind part of the
269 azimuthally propagating CPS.

270   To systematically examine the outward propagation of newly-formed deep convection
271 inside the RMW, Figure 6 shows the evolution of radar reflectivity azimuthally averaged within
272 the downshear-right and downshear-left quadrants below 500-m height in CTL and S1. We only
273 select the downshear quadrants, given that the newly-developed convection typically become
274 visible in the boundary layer of the downshear-left quadrant and convection at the smaller radii
275 in upshear-left may mask the signal of outward propagation of newly-formed deep convection
276 toward the RMW (cf. Fig. 4). This averaging method can capture the distinct inward rebuilding
277 events in which the newly-developed deep convection projects significantly onto the azimuthal
278 mean. Figure 6 presents five and four visually trackable inward rebuilding events over the 12-h
279 period before 0000 UTC 3 October for the CTL and S1 TCs, respectively. The S1 TC undergoes
280 a notable inward rebuilding over 1600-1630 UTC (i.e., the event 1), with the maximum radar
281 reflectivity comparable to that of the inward rebuilding events of the CTL TC. Nevertheless, the
282 strength of outward propagating convection in terms of radar reflectivity is weaker in S1 than
283 that in CTL on average. The outward propagation of the circled spiral rainband shown in Figs.
284 4a-f and Figs. 4g-l corresponds to the event 3 in Fig. 6a and Fig. 6b, respectively.

285   Figures 7a-b further assess the impact of inward rebuilding on precipitation symmetrization
286 and show a time-azimuthal plot of the radar reflectivity averaged within the 20-50 km radii and
287 within the 0-500 m layer for the CTL and S1 TCs, respectively. We select the annulus within the
288 initial RMW (i.e., 60 km) to better illustrate the inward rebuilding of the CPS during
289 precipitation symmetrization. The CPS in CTL spans a broader azimuthal coverage and generally
290 exhibits a more vigorous convective activity in terms of radar reflectivity than the CPS in S1

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291   over the 12-h period, and the latter is consistent with Fig. 6. Precipitation symmetrization in CTL

292   is sustained with more vigorous convection over the 12-h period; however, precipitation

293   symmetrization in S1 is only notable when the convective activity is most vigorous over 1600-

294   1830 UTC (see the reflectivity maximum in Fig. 7b). These findings demonstrate that

295   precipitation symmetrization is closely related to the strength of the outward propagating newly-

296   developed deep convection inside the RMW (see also Figs. 4-6). Additionally, the relatively

297   weak echo (10-25 dBZ) in the right-of-shear quadrants of the CTL TC is related to stratiform

298   precipitation (see Figs. 8a-b). The area of stratiform precipitation substantially increases in the

299   right-of-shear quadrants of the CTL TC while those quadrants of the S1 TC are almost devoid of

300   stratiform precipitation, which can be inferred from the comparison between Figs. 7a and 7b and

301   clearly seen from one example in Fig. 8.

302   A large patch of low-θ_e air (<358 K) at the lowest model level appears in the upshear-left

303   quadrant near 1800 UTC 2 October and is superimposed by the downward motion (<−0.2 m s^{-1})

304   at 1-km height at the leading edge of the spiral rainband (Figs. 7c-d). This finding suggests that

305   the low-θ_e air originates from above the boundary layer and is transported downward by

306   convective downdrafts. This low-level ventilation pathway is comparable with the one proposed

307   by Riemer et al. (2010), although they discussed this process in a mature hurricane with a well-

308   defined eyewall. These low-θ_e parcels are then advected downwind and their θ_e gradually

309   recover to higher values during their propagation toward the downshear-right quadrant (i.e.,

310   ~1800-2100 UTC 2 October), indicative of a boundary layer recovery process (Powell 1990;

311   Molinari et al. 2013; Zhang et al. 2013; Nguyen et al. 2019). Multiple low-level ventilation and

312   subsequent boundary layer recovery events can be found in both experiments over the 12-h

313   period. In CTL, the θ_e averaged within the lowest 500 m is generally 2-K warmer than that in S1

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314   (cf. Figs. 7c and 7d). Given the inward rebuilding events in CTL are generally stronger than

315   those in S1, a key question arises as whether the stronger newly-developed deep convection in

316   CTL is attributed to the more effective boundary layer recovery for the warmer SST. In the next

317   section, we will address this issue by performing trajectory analyses and along-trajectory vertical

318   momentum budgets.

319   **5. Boundary layer recovery**

320   *a. Trajectory analyses*

321   To examine the role of boundary layer recovery in the inward rebuilding process, a forward

322   Lagrangian trajectory analysis of the downdraft-related low-θ_e air parcels in the lower boundary

323   layer, beneath the leading edge of the CPS, is carried out. The 4-h trajectory analysis starts from

324   1700 UTC and 1740 UTC 2 October for the CTL and S1 TCs, respectively, when the midlevel

325   vortex of both TCs is located upshear (Fig. 1b), and the pattern and intensity of the CPS (Figs. 9a

326   and 9d) as well as the low-θ_e values beneath the rainband are comparable (Figs. 9b-c and 9e-f).

327   Over the 4-h period, precipitation symmetrization is sustained in CTL while it is hindered after

328   1830 UTC 2 October in S1 (Figs. 7a-b). A total of 320 parcels are tracked from the low-θ_e

329   region, with 64 parcels at 3rd, 5th, 7th, 9th, and 11th lowest model levels, respectively. The mean

330   height of all of these 5 model levels is below 450 m. The initial locations of the 64 parcels at

331   each model level are the same (see black dots in Figs. 9a and 9d), and the horizontal spacing

332   between each parcel at the same model level is 4 km. Figures 9b-c and 9e-f show the initial

333   trajectory points colored by the maximum height of the subsequent 4-h forward trajectories. The

334   parcels with their maximum height below 1.5 km are referred to as "boundary layer parcels"

335   (PBL). The others with their maximum height within 1.5-4 km, 4-8 km, and >8 km are grouped

336   into shallow, midlevel, and deep convection categories, respectively, which is analogous to the

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337 partitioning of convection in section 2. A comparison of the parcel trajectories starting from the

338 3rd (Figs. 9b-c) and 7th (Figs. 9e-f) lowest model levels between CTL and S1 indicates that the

339 parcels in CTL are more likely to develop into deep convection, while most of the parcels in S1

340 fail to escape from the boundary layer (<1.5 km). Similar results are also found for the

341 trajectories starting from other model levels in the lower boundary layer (not shown). Statistics

342 based on the maximum height of these 4-h trajectories confirm this statement (Fig. 10). More

343 than 80% of the tracked parcels in CTL develop into convection while the ratio decreases ~40%

344 in S1. Moreover, ~35% of these parcels in CTL develop into deep convection while the ratio

345 decreases to <3% in S1. These results are consistent with the difference in the strength of inward

346 rebuilding between the event 4 in CTL and event 3 in S1 (Fig. 6).

347   To quantitatively examine the differences in the trajectories between the two experiments,

348 the top 20% of the tracked parcels based on their maximum height of the 4-h trajectory are

349 selected for a comparison. Figure 11 shows the evolution of height and vertical velocity along

350 the trajectories. Over the first 2 hours of the trajectories ($t = 0$-2 h), these parcels in both CTL

351 and S1 generally stay below 1.5 km height, and their vertical velocity is generally <1 m s^{-1}. The

352 parcel height evolution in CTL and S1 diverges afterwards (Fig. 11a), as the upward motion of

353 these parcels in CTL accelerates more rapidly than that in S1 (Fig. 11b). Considering these facts,

354 the boundary layer recovery of these parcels over $t = 0$-2 h is examined in Fig. 12. In CTL, the

355 mean θ_e increase over $t = 0$-2 h for these parcels is ~5 K (Fig. 12a). In contrast, the mean θ_e

356 increase in S1 is 1.5 K over the same period, which is 3.6 K lower than that in CTL. Figure 12b

357 further shows that at $t = 0$ h the mean θ_e of these parcels in CTL is 2.1 K cooler than that in S1,

358 while at $t = 2$ h the mean θ_e of these parcels in CTL is 1.5 K warmer. Additionally, the mean

359 parcel height in both experiments is <600 m over $t = 0$-2 h, and mean parcel height in CTL is 190

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360   m lower than that in S1 (Fig. 12a). Given that the period of trajectory analyses is before sunrise

361   at local time (0100-0500 LST for CTL), there is no incoming shortwave radiation; in right-of-

362   shear quadrants, weak radiative heating is only found in the boundary layer of the downshear-left

363   quadrant and is one order smaller in magnitude than the diabatic heating due to upward enthalpy

364   fluxes (not shown). Thus, the upward enthalpy fluxes from the ocean surface is the dominant

365   energy source in the lower boundary layer; the more efficient boundary layer recovery in CTL is

366   attributed to the warmer SST as well as the ability of these parcels to stay at a lower height

367   where the upward enthalpy fluxes are typically larger (Zhang and Drennan 2012).

368   ### b. Along-trajectory vertical momentum budgets

369   To investigate the mechanisms responsible for the convective initiation and the subsequent

370   convective development during precipitation symmetrization, a vertical momentum budget along

371   the trajectories of the top 20% of the tracked parcels is performed for both experiments

372   (Jeevanjee and Romps 2015):

373    $$\frac{dw}{dt} = a_i + a_b, \tag{1}$$

374   in which w is the vertical velocity and the vertical acceleration ($\frac{dw}{dt}$) is decomposed into dynamic

375   (a_i) and buoyancy (a_b) accelerations. The buoyancy acceleration (or "effective buoyancy") is

376   defined as the Lagrangian acceleration that would result if the wind were instantaneously zeroed

377   out. Similarly, the dynamic acceleration is defined as the Lagrangian vertical acceleration

378   resulting from an instantaneous zeroing out of any horizontal density anomalies. The relative

379   roles of these two acceleration terms could be quantified by solving the Poisson equation:

380    $$-\nabla^2(\bar{\rho} a_b) = g\nabla_h^2 \rho, \tag{2}$$

381    $$-\nabla^2(\bar{\rho} a_i) = -\partial_z \nabla \cdot [\bar{\rho}(\mathbf{u} \cdot \nabla)\mathbf{u}], \tag{3}$$

16

382   where $\bar{\rho}$ is horizontal average of air density in the budget domain, ρ is the air density, g is the

383   gravitational acceleration, and **u** is the three-dimensional wind vector, ∇^2 is the three-

384   dimensional Laplacian operator, and ∇_h^2 is the horizontal Laplacian operator. Dirichlet boundary

385   conditions of $a_i = 0$ and $a_b = 0$ are specified on the top and bottom boundaries following

386   Jeevanjee and Romps (2015). The main advantage of this form of vertical momentum budget

387   over other forms (e.g., Zhang et al. 2000; Braun 2002; Eastin et al. 2005) is that it refrains from

388   the ambiguity in the arbitrary definition of the reference state $\bar{\rho}$ when calculating the

389   Archimedean buoyancy (Davies-Jones 2003; Doswell and Markowski 2004). Note that a_b

390   includes both Archimedean buoyancy and the environment response to vertical acceleration

391   driven by Archimedean buoyancy. To improve the accuracy of the interpolated vertical

392   acceleration along the trajectories, the vertical momentum budget is performed from 50 m to 18

393   km height at a 50-m interval.

394   Figures 13a and 13d show that $\frac{dw}{dt}$ ($=a_i + a_b$) averaged below 1.5 km height and over t = 0-

395   2 h is marginal in the right-of-shear semicircle. This finding is consistent with the fact that the

396   top 20% of the tracked parcels stratified by their maximum height of the 4-h trajectory in both

397   experiments stay in the boundary layer before arriving at the downshear convergence zone. The

398   a_b within r = 40 km is positive in the right-of-shear semicircle below 1.5 km height (Figs. 13c

399   and 13f), which is largely counteracted by a_i (Figs. 13b and 13e). The mean $\frac{dw}{dt}$ along the

400   trajectory over t = 0-2 h is positive in both experiments (Fig. 14a). Of note, the mean w of the

401   tracked parcels over the initial half hour is negative in both experiments (Fig. 11b), and negative

402    $\frac{dw}{dt}$ is not required to retain these parcels in the boundary layer. The larger upward acceleration in

403   S1 over t = 0-2 h is mainly attributed to the much larger a_b in S1 (Fig. 14a), which is further

404   related to the smaller mean orbital radius ($\approx$30 km) in S1 (Fig. 13) such that parcels can tap into

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405   the warm reservoir near the TC center (see Fig. 5e-f). In contrast, the mean orbital radius in CTL

406   is 50 km, which is generally outside the region of relatively large a_b (>0.2 m s^{-2}).

407   The smaller mean $\frac{dw}{dt}$ over t=0-2 h in CTL (Fig. 14a) accounts for the lower mean parcel

408   height in CTL than in S1 (Fig. 12a). These parcels subsequently arrive at the inward flank of the

409   CPS, indicated by the spiral band of upward motion at 1.5 km height (Fig. 15), or downshear

410   convergence zone, where both the dynamic and buoyancy forcing play an important role in

411   lifting the parcels out of the boundary layer (i.e., convective initiation). The upward dynamic

412   forcing is likely attributed to the low-level convergence in the downshear convergence zone, and

413   it is difficult to further decompose the dynamical forcing to quantify the relative contribution of

414   the low-level convergence and other mechanisms. The role of dynamic forcing in the convective

415   initiation in the incipient eyewall is consistent with the findings in previous modeling studies of

416   TCs (Zhang et al. 2000; Braun 2002; Gu et al. 2019). *One new finding in this study is that the*

417   *buoyancy forcing also plays a role in the convective initiation at the inward flank of the CPS*

418   (Figs. 15c and 15f). Figures 15 and 11 also indicate a striking difference between the two

419   experiments: in CTL a large portion of these parcels have already developed or are going to

420   develop into deep convection in the downwind part of the azimuthally-propagating CPS, as seen

421   in Figs. 4-6, while in S1 the parcels reaching the downwind part of the CPS mostly develop into

422   shallow and midlevel convection.

423   Figure 14b shows the vertical acceleration terms averaged over t = 2-3 h for the top 20% of

424   the tracked parcels. Both the a_b and $\frac{dw}{dt}$ are significantly larger than those over t=0-2 h (Fig. 14a).

425   Clearly, the acceleration of the upward motion above the boundary layer is mainly attributed to

426   the a_b in both experiments. Parcels in CTL exhibit larger mean a_b over t = 2-3 h than those in

427   S1, which is mainly attributed to the stronger a_b over t = 2.5-3 h in CTL (see Fig. 16). In S1, a_i

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428   plays a secondary role in accelerating the upward motion; whereas in CTL, the mean a_i for these

429   parcels is negative (Fig. 14b). Figure 16 further shows that in both experiments dynamic forcing

430   accelerates the upward motion of parcels over $t = 2\text{-}2.5$ h, when these parcels still reside at low

431   levels; as convection matures over $t = 2.5\text{-}3$ h, these parcels experience substantial dynamic

432   deceleration. The dynamical deceleration in CTL is more notable over $t = 2.5\text{-}3$ h, leading to a

433   net negative value of a_i over $t = 2\text{-}3$ h in CTL (Fig. 14b). The dynamical deceleration comes

434   from the effect of a downward-pointing perturbation pressure gradient force (cf. Braun 2002). In

435   short, the stronger acceleration of the upward motion during the convective development in CTL

436   is attributed to the stronger a_b. Given the a_b during convective development is closely related to

437   the θ_e values of the parcels at convective initiation, the above findings confirm that the boundary

438   layer recovery of the downdraft-cooled parcels is a key mechanism in maintaining the convective

439   activity during precipitation symmetrization.

440   **6. Discussion**

441   *a. Observational support and additional discussions on precipitation symmetrization*

442   As mentioned in section 4, the successive inward rebuilding of the CPS during precipitation

443   symmetrization under moderate VWS is reminiscent of the observed "inward progression" of the

444   cloud-to-ground lightning clusters in the RI TCs under moderate VWS (e.g., Molinari et al.

445   2004; Molinari and Vollaro 2010; Stevenson et al. 2014; Zawislak et al. 2016). Note that cloud-

446   to-ground lightning can be treated as an indicator of strong/deep convection. These observational

447   case studies clearly show that during precipitation symmetrization the lightning cluster drifted

448   cyclonically from downshear-left at *large* radii to upshear-left at *smaller* radii and mostly inside

449   the RMW. The relationship between boundary layer recovery and convective development in the

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450 sheared TCs was also alluded to in observational studies of Tropical Storm Edouard (2002)

451 (Molinari et al. 2013), Hurricane Cristobal and Bertha (2014) (Nguyen et al. 2017), and early-

452 stage TCs from 1997 to 2017 (see fig. 16 in Nguyen et al. 2019). These observational facts imply

453 that the identified "inward rebuilding" pathway is not limited to one single sheared TC, and the

454 related the dynamical (e.g., downshear convergence zone and the associated low-level

455 convergence) and thermodynamic processes (e.g., boundary layer recovery) are intrinsic to

456 sheared early-stage TCs.

457   Previous observational studies (e.g., Molinari et al. 2013; Nguyen et al. 2017) generally

458 recognized that the high-θ_e parcels in the downshear-right quadrant contribute to the convective

459 development in the left-of-shear quadrants of sheared TCs. However, to our knowledge, no direct

460 evidence has been presented to prove that the downshear-right high-θ_e parcels that contribute to

461 the subsequent left-of-shear convective development are an outcome of the boundary layer

462 recovery of downdraft-cooled parcels beneath the CPS at earlier times. Thus, one unique

463 contribution of this study is that the identified "inward rebuilding" pathway directly relates the

464 boundary layer recovery of downdraft-cooled parcels to the development of deep convection in

465 the left-of-shear quadrants. Additionally, observational and modeling studies of TCs in shear

466 frequently pointed out the deep convection in the upshear quadrant and inside the RMW as a key

467 factor in differentiating the RI and non-RI TCs (e.g., Rogers et al. 2016; Hazelton et al. 2017;

468 Wadler et al. 2018a; Leighton et al. 2018), while mechanisms responsible for the radius and

469 quadrant preference of deep convection for RI TCs remain elusive. The "inward rebuilding"

470 pathway provides a reasonable explanation to this phenomenon. This pathway is considered

471 applicable to the sheared TCs at tropical storm or minimal hurricane stage, when the eyewall has

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472 not yet typically formed, but when RI occurs more frequently (Kaplan et al. 2010; Chen et al.

473 2015).

474   Over warm SST, the CTL TC also has a much larger area of stratiform precipitation in the

475 right-of-shear quadrants than the S1 TC after 1800 UTC 2 October (Figs. 7a-b, 8, and 17a-b).

476 Figure 17 shows azimuthal-height plots of radar reflectivity, θ_e, relative humidity, and vertical

477 velocity averaged within r = 20-60 km and over the period of 1800-1850 UTC 2 October for the

478 CTL and S1 TCs. Given stratiform precipitation is preferred in an environment with high

479 saturation fraction (López and Raymond 2005), the prevailing stratiform precipitation in the

480 right-of-shear quadrants of the CTL TC suggest the inner-core environment therein is very moist,

481 as seen from the nearly saturated 5-9 km layer above the freezing level (cf. Figs. 17c-d). In

482 contrast, the RH of the 5-9 km layer in the right-of-shear quadrants of the S1 TC is <85%. An

483 idealized simulation study (Rios-Berrios et al. 2018) also found that the right-of-shear quadrants

484 of the TC inner core become very moist before RI onset. The humidification of the layer above

485 the freezing level in the right-of-shear quadrants is most likely attributed to the horizontal

486 advection of water vapor and the evaporation/sublimation of condensates coming from the

487 upwind quadrants (Rappin and Nolan 2012; Alvey et al. 2020). All of these processes are closely

488 related to the convection that transports the boundary layer moisture upward into the mid-to-

489 upper troposphere. The left-of-shear convective activity within the inner core is generally more

490 vigorous in CTL than in S1 over the 12-h period preceding the RI onset of the CTL TC (Figs. 7a-

491 b), which helps account for the more moist midlevels in the right-of-shear quadrants of the CTL

492 TC. In the numerical simulation study of Typhoon Vicente (2012), the amount of deep

493 convection in upshear quadrants steadily increases 5 hours before RI onset (Chen et al. 2018b)

494 and a nearly-saturated inner core forms at RI onset (Chen et al. 2019), which lends support to the

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495   hypothesis. A detailed diagnostic analysis of the moistening processes in a shear-relative

496   framework is beyond the scope of this study and will be left for future work.

b. Comparison with a mesoscale convective system with trailing stratiform

498   The analyses in section 5 demonstrate that the precipitation symmetrization of the sheared

499   TCs is essentially a three-dimensional rather than an axisymmetric process (also see Fig. 18).

500   Particularly, the boundary layer recovery of downdraft-cooled parcels and the subsequent inward

501   rebuilding of deep convection in the sheared TCs cannot be described in the simulations using an

502   axisymmetric framework. Before the formation of a complete eyewall, the CPS propagates

503   cyclonically and radially inward during precipitation symmetrization. The CPS in a sheared

504   early-stage TC is analogous to a midlatitude mesoscale convective system with trailing stratiform

505   (hereafter MCS-TS, Parker and Johnson 2000) in both morphology and organization, with

506   ascending flow in the front (radially inward in a TC) of the moving MCS-TS and descending

507   inflow from the rear (radially outward in a TC) and below the freezing level (Fig. 18d).

508   However, the CPS in a sheared TC differs from MCS-TS due to its imposed swirling

509   circulation of the TC. In the mature stage of MCS-TS, convective updrafts are sustained by the

510   high-θ_e inflow in the front, and the cold pool (i.e., low-θ_e) region remains in the rear flank of the

511   convective system that induces low-level convergence for the convective updrafts. In a sheared

512   TC, however, the downdraft-induced low-θ_e parcels are "recycled" in the TC circulation,

513   gradually recovered by positive enthalpy fluxes, slowly ascend in the boundary layer (e.g., see

514   the newly-formed spiral rainband along the trajectory in Fig. 18a) during their advection toward

515   the downshear convergence zone, and ultimately become a part of the ascending branch radially

516   inward of the downwind part of the CPS if their entropy has been sufficiently recovered. This

517   scenario is seen in both TCs (Figs. 18b-e), while the newly-developed convection in the inward

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518   rebuilding events is weaker in the S1 TC. This conceptual model in Fig. 18 highlights the critical

519   role of warm SST and boundary layer recovery in replenishing these low-θ_e parcels and favoring

520   the symmetrization of the CPS in the sheared TCs prior to the formation of a closed eyewall.

521   An important implication of this study is that the low-level ventilation does not necessarily

522   weaken the early-stage TCs over warm SSTs; instead, the competition between the low-level

523   ventilation and boundary layer recovery matters to the subsequent convective activity and

524   structural/intensity change of the early-stage TCs in VWS.

525   **7. Concluding remarks**

526   Precipitation symmetrization or eyewall formation preceding the rapid intensification (RI)

527   of tropical cyclones (TCs) under moderate vertical wind shear (VWS) has been documented in

528   recent studies (e.g., Zagrodnik and Jiang 2014; Tao and Jiang 2015; Alvey et al. 2015; Tao et al.

529   2017; Chen et al. 2017; Fischer et al. 2018). Understanding thermodynamic/dynamical

530   mechanisms controlling the precipitation symmetrization is the central question addressed in this

531   study. By analyzing two representative numerical simulations of Typhoon Mujigae (2015),

532   initialized with warm (i.e., CTL) and relatively cool (i.e., S1) SSTs, respectively, key results are

533   summarized as follows:

534   1) A downshear convergence zone forms due to the differential advection by the VWS. The

535   convective precipitation shield (CPS) is initially embedded in this convergence zone and

536   subsequently propagates into the upshear side before RI onset. Downdraft-cooled parcels beneath

537   the CPS are advected downwind by the swirling winds and their entropy is gradually recovered

538   by positive enthalpy fluxes (i.e., boundary layer recovery).

539   2) The boundary layer recovery is key to the convective development and precipitation

540   symmetrization before RI onset. Trajectory analyses of the downdraft-cooled parcels and the

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along-trajectory vertical momentum budget demonstrate that the boundary layer recovery is more efficient in CTL due to the warmer SST, and the resulting stronger buoyancy acceleration are responsible for the development of much more deep convection in CTL than in S1, particularly in the upshear quadrants. Additionally, downdraft-cooled parcels are lifted out of the boundary layer by both dynamical and buoyancy acceleration in the convergence zone (i.e., convective initiation).

3) The precipitation symmetrization before RI onset in both experiments is maintained by the continuous development of deep convection radially inward of the azimuthally propagating CPS (i.e., inward rebuilding event), as deep convection matures in the downwind part of the CPS. In CTL, precipitation symmetrization is sustained by stronger newly-developed deep convection in successive inward rebuilding events, and the associated stronger microphysics diabatic heating inside/near the radius of the maximum wind (RMW) aids in the earlier RMW contraction of the CTL TC (see discussions in Chen et al. 2018a). In contrast, precipitation symmetrization is delayed in S1 due to the weaker newly-developed convection radially inward of the CPS, particularly in the upshear-left quadrant, and the RMW contraction is also delayed.

These above processes form a positive feedback between boundary layer recovery, *inward rebuilding of the CPS*, precipitation symmetrization, and RMW contraction under the warmer SST, and highlight the key role of the boundary layer recovery of the downdraft-cooled parcels in alleviating the low-level ventilation and organizing the CPS during precipitation symmetrization. Additionally, these results provide an explanation for the frequently observed deep convection in the upshear quadrant and inside the RMW of the sheared TCs before RI onset.

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743   Table 1: Experiment descriptions.

	Description
CTL	Initialized with the SST at 0000 UTC October 2, 2015.
S1	Initialized with the 1990-2013 monthly-mean SST.

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Figure captions

Fig. 1. Verification of simulated 10-m maximum wind speed (m s^{-1}) in the CTL (red) and S1 (black) experiments against the best track data of China Meteorological Administration (gray line with circles). (b) Evolution of the 450-850 hPa vortex tilt magnitude (km) for the CTL and S1 TCs from 1200 UTC 2 October to 1200 UTC 4 October. The three dashed lines mark the downshear-left (DL), upshear-left (UL), and upshear-right (UR) quadrants that the tilt vector points toward. Evolution of (c) local Rossby number R_o and (d) RMW (km) at 10-m height for the CTL and S1 TCs. The gray line in (c) denotes $R_o = 12$. The red (black) arrows in each panel denote the RI onset timing of CTL (S1) TC. Adapted from Chen et al. (2018a).

Fig. 2. (a)-(c) The 1.5-km radar reflectivity (shading; dBZ), background horizontal convergence (dashed contours with values of -2.0, -1.0, -0.5×10^{-4} s^{-1}), and storm-relative winds (vector, m s^{-1}) at (a) 1600 UTC 2 October, (b) 1800 UTC 2 October, and (c) 2100 UTC 2 October in CTL. (d)-(f) As in (a)-(c), but for S1.

Fig. 3. (a)-(b) The location of deep convection within $r = 100$ km relative to the surface TC center (black star) occurred over 1200-1500 UTC 2 October (black plus), 1500-1800 UTC 2 October (blue plus), and 1800-2100 UTC 2 October (red plus) in CTL and S1, respectively. (c)-(d) As in (a)-(b), but for the location of CBs. The black arrow in (a)-(d) denotes the 300-850 hPa vertical wind shear. Azimuthal distribution of the number of (e) grid points of deep convection and (f) CBs in CTL (red) and S1 (black) within $r = 60$ km accumulated over 1200-2100 UTC 2 October at a 10-min interval. The total number of deep convection and CBs in CTL and S1 are shown as texts. The shear-relative quadrants are labeled in each panel. DR, UR, UL, and DL denote downshear-right, upshear-right, upshear-left and downshear-left quadrant, respectively.

Fig. 4. (a)-(f) Evolution of radar reflectivity (shading, dBZ) and storm-relative wind (vector, m s^{-1}) at the lowest model level for the CTL TC from 1710 UTC 2 October to 1830 UTC 2 October. (g)-(l) As in (a)-(f), but for the S1 TC from 1920 UTC 2 October to 2030 UTC 4 October. The solid black circle denote the 50-km reference radius. The dashed black ellipse tracks the propagation of the developing convection inside of the RMW. The reference vector is shown in (a). The gray arrow on the top right corner denotes the heading of 200-850 hPa VWS.

Fig. 5. (a)-(b) Plan view of radar reflectivity (shading, dBZ) and storm-relative wind (vector, m s^{-1}) at the lowest model level at 1800 UTC 2 October for the CTL and at 2000 UTC 2 October for the S1 TC, respectively. (c)-(d) Composite vertical slice of microphysics diabatic heating (shading, K h^{-1}) and absolute vorticity (contoured at 0.5, 1, and 2×10^{-3} s^{-1}) over 1800-1850 UTC 2 October and over 2000-2050 UTC 2 October for the CTL and S1 TCs, respectively; (e)-(f) As in (c)-(d), but for θ_e (shading, K) and storm-relative wind (streamline). The position of the vertical slice in (c)-(f) is marked as black dash line in (a)-(b). The white arrow in (a)-(d) marks the location of the newly-formed deep convection. The red streamline in (e)-(f) is related to the newly-developed deep convection inside the RMW.

Fig. 6. (a)-(b) Time-radius plot of radar reflectivity (shading, dBZ) averaged azimuthally within the downshear quadrants and below 500 m height from 1200 UTC 2 October to 0000 UTC 3 October for the CTL and S1 TCs, respectively. The black dashed lines in each panel

792 denote the visually trackable outward propagation of newly-formed convection within the
793 RMW.

794 Fig. 7. (a)-(b) Time-azimuthal plot of radar reflectivity (shading, dBZ) for the CTL and S1 TCs,
795 respectively. (c)-(d) As in (a)-(b), but for θ_e (shading, K) and vertical velocity (contours
796 with values of –0.2, 0.2, and 0.5 m s^{-1} and negative values are dashed) for the two TCs. The
797 radar reflectivity and θ_e are averaged within r = 20-50 km and below 500-m height. The
798 vertical velocity is averaged within r = 20-50 km at 1-km height. The solid white line
799 denotes the heading direction of 200-850 hPa VWS. The dashed white line marks the shear-
800 relative quadrants, as labeled at the bottom of each panel. DR, UR, UL, and DL denote
801 downshear-right, upshear-right, upshear-left and downshear-left quadrant, respectively.

802 Fig. 8. Plan view of (a) radar reflectivity (shading, dBZ) at 3-km height and (b) precipitation
803 mode at 1830 UTC 2 October. The red, yellow, and purple area in (b) denote convective,
804 stratiform, and other type precipitation, respectively. (c)-(d) As in (a)-(b), but for S1 TC.
805 The solid black arrow in (a)-(d) denotes the heading direction of 200-850 hPa VWS.

806 Fig. 9. Plan view of (a) radar reflectivity (shading, dBZ) at the 3rd lowest model level, θ_e
807 (shading, K) at the (b) 3rd and (c) 7th lowest model levels at 1700 UTC 2 October for the
808 CTL TC. (d)-(f) As in (a)-(c), but for the S1 TC at 1740 UTC. The location of the initial
809 points of the trajectories are shown as black dots in (a) and (d), and are shown as colored
810 dots based on the maximum height of the subsequent 4-h forward trajectory in (b)-(c) and
811 (e)-(f). Black cross (×) denotes the boundary layer parcels with the maximum height <1.5
812 km. Pink, red, and violet dots denote the maximum height of these parcels within 1.5-4 km,
813 4-8 km, and >8 km, respectively. The large black dot at (0, 0) marks the surface TC center.
814 The black circle represents the RMW near the surface. The orange box in (a) and (e)
815 denotes the same area of (b)-(c) and (e)-(f), respectively. The mean height of each model
816 level is shown in the title of each panel.

817 Fig. 10. Bar plot of the ratio of the track parcels that remain in the boundary layer (gray) or
818 develop into the shallow (pink), midlevel (red) and deep (purple) convection in the CTL and
819 S1 experiments.

820 Fig. 11. The evolution of (a) parcel height (km) and (b) vertical velocity (m s^{-1}) along the 4-h
821 trajectory for the top 20% of the parcels that are stratified by their maximum height of the
822 4-h trajectory. The red and gray lines denote the trajectories in CTL and S1, respectively.

823 Fig. 12. Statistics for the top 20% of the parcels stratified by their maximum height of the 4-h
824 trajectory. (a) Mean parcels height during t = 0-2 h and differences in the mean θ_e from t =
825 0 h to t = 2 h in CTL (red) and S1 (gray). (b) Differences in the mean θ_e between CTL and
826 S1 (CTL–S1) at t = 0 h and t = 2 h.

827 Fig. 13. Plan view of the (a) a_i (dynamic acceleration) + a_b (buoyancy acceleration) (shading,
828 ×10^{-3} m s^{-2}), (b) a_i, and (c) a_b averaged in the lowest 1.5 km layer and over t = 0-2 h in
829 CTL. Contours denote 1.5-km vertical velocity with values of –1, –0.5, –0.2, 0.5, 1.0, 1.5,
830 and 2.0 m s^{-1} (negative values dashed) averaged over the same period. (d)-(f) As in (a)-(c)
831 but in S1. The 0-2 h storm-relative trajectories of the top 20% of the parcels that are
832 stratified by their maximum height of the 4-h trajectory are overlaid. The black arrow in the
833 upper-right corner denotes the heading direction of the 200-850 hPa VWS. The red crosses
834 in each panel denote the starting points of these trajectories.

835 Fig. 14. Vertical velocity budget terms a_i, a_b, and a_i+ a_b (shading, ×10^{-3} m s^{-2}) averaged over
836 (a) t = 0-2 h and (b) t = 2-3 h for the top 20% of the parcels that are stratified by their
837 maximum height of the 4-h trajectory.

32

838   Fig. 15. As in Fig. 13, but for the results over t=2-3 h.

839   Fig. 16. Evolution of vertical velocity budget terms a_i (blue), a_b (red), and $a_i+ a_b$ (black) over t
840   = 2-3 h for the top 20% the parcels that are stratified by their maximum height of the 4-h
841   trajectory in (a) CTL and (b) S1. Lines represent mean values; shading extends from
842   minimum to maximum at each minute. The unit of the budget terms is 10^{-3} m s^{-2}.

843   Fig. 17. Azimuthal-height plot of radar reflectivity (shading, dBZ) and vertical velocity (contours
844   with values of –0.4, –0.2, –0.1, 0.2, 0.5, 1.0, and 2.0 m s$^{-1,}$ negative values dashed)
845   averaged within r =20-60 km and over 1800-1850 UTC 2 October for the (a) CTL and (b)
846   S1 TCs. (c)-(d) As in (a)-(b), but shading denotes relative humidity (%). The vertical dashed
847   white line marks the shear-relative quadrants, as labeled at the top of each panel. DR, UR,
848   UL, and DL denote downshear-right, upshear-right, upshear-left and downshear-left
849   quadrant, respectively.

850   Fig. 18. Conceptual model for the inward rebuilding and precipitation symmetrization under
851   different SSTs. (a) Plan view of the CPS that propagates into the upshear-left (UL)
852   quadrant. Beneath the CPS, downdraft-cooled parcels in the boundary layer subsequently
853   undergo boundary layer recovery and develop into convection during their propagation
854   toward the downshear quadrants. (b)-(c) As in (a), but at a later time before the RI onset of
855   the CTL TC over warm SSTs. In (b), the more efficient boundary layer recovery and more
856   notable inward building of deep convection in UL maintain the precipitation symmetrization
857   over warm SSTs; the stratiform in the right-of-shear semicircle in (b) indicates a nearly-
858   saturated layer above the freezing level. (c) Over relatively cool SSTs, newly-developed
859   convection in the inward rebuilding events is much weaker, which hinders precipitation
860   symmetrization. Reflectivity contours represent the CPS and convective cells. The red
861   dashed arrow in (a)-(c) denotes the trajectory along which boundary layer recovery and the
862   subsequent inward rebuilding occur. (d)-(e) Composite vertical cross-sections of reflectivity
863   and streamlines over warm and relatively cool SSTs, respectively. Locations of the cross
864   sections are marked as thick black lines in (b)-(c).

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Fig. 1. Verification of simulated 10-m maximum wind speed (m s^{-1}) in the CTL (red) and S1 (black) experiments against the best track data of China Meteorological Administration (gray line with circles). (b) Evolution of the 450-850 hPa vortex tilt magnitude (km) for the CTL and S1 TCs from 1200 UTC 2 October to 1200 UTC 4 October. The three dashed lines mark the downshear-left (DL), upshear-left (UL), and upshear-right (UR) quadrants that the tilt vector points toward. Evolution of (c) local Rossby number R_o and (d) RMW (km) at 10-m height for the CTL and S1 TCs. The gray line in (c) denotes $R_o = 12$. The red (black) arrows in each panel denote the RI onset timing of CTL (S1) TC. Adapted from Chen et al. (2018a).

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881 (d)-(f) As in (a)-(c), but for S1.

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882

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892

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896 solid black circle denotes the 50-km reference radius. The dashed black ellipse tracks the
897 propagation of the developing convection inside of the RMW. The reference vector is shown in
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899

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911
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915 visually trackable outward propagation of newly-formed convection within the RMW.
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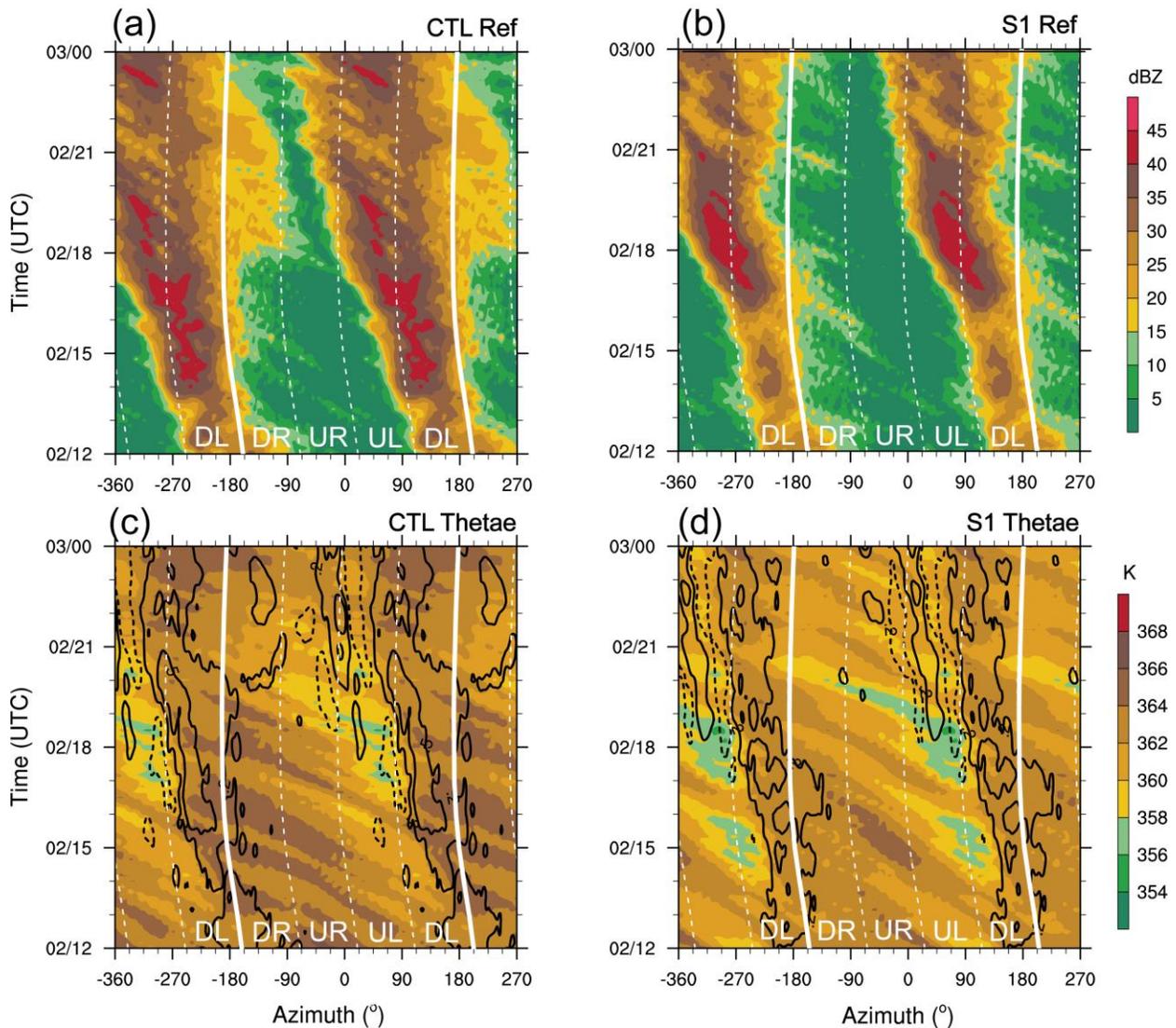
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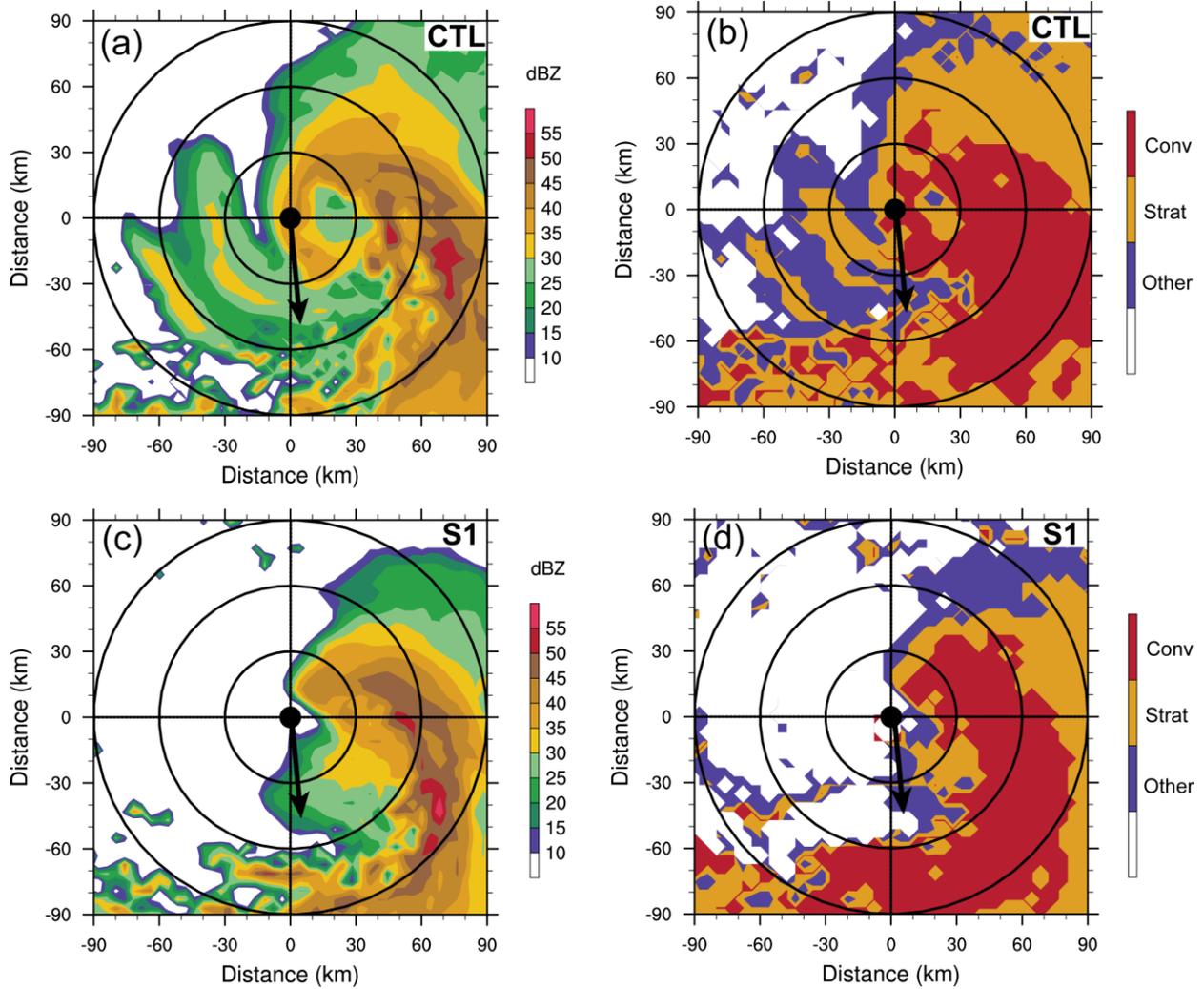
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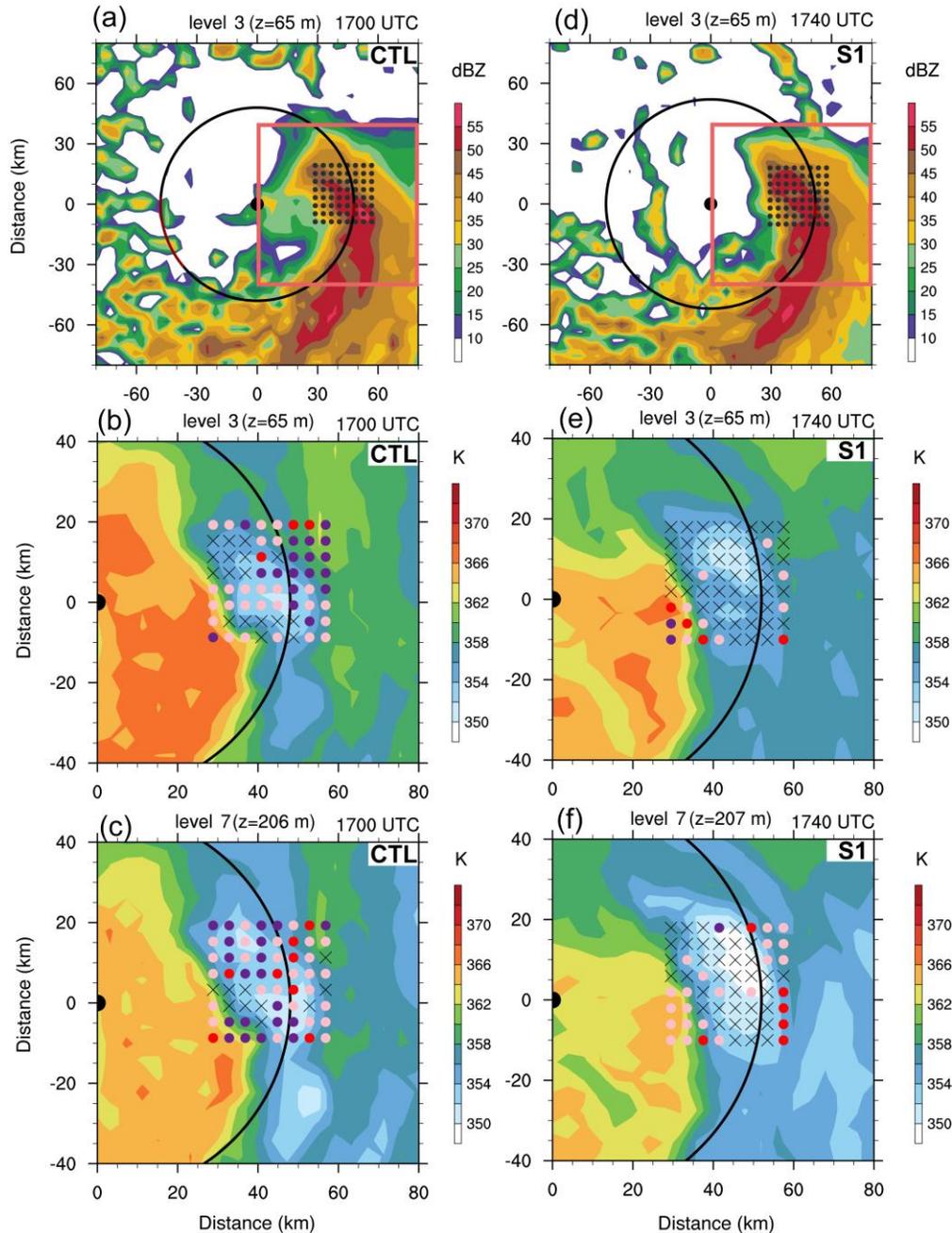
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Fig. 9. Plan view of (a) radar reflectivity (shading, dBZ) at the 3rd lowest model level, θ_e (shading, K) at the (b) 3rd and (c) 7th lowest model levels at 1700 UTC 2 October for the CTL TC. (d)-(f) As in (a)-(c), but for the S1 TC at 1740 UTC. The location of the initial points of the trajectories are shown as black dots in (a) and (d), and are shown as colored dots based on the maximum height of the subsequent 4-h forward trajectory in (b)-(c) and (e)-(f). Black cross (×) denotes the boundary layer parcels with the maximum height <1.5 km. Pink, red, and violet dots denote the maximum height of these parcels within 1.5-4 km, 4-8 km, and >8 km, respectively. The large black dot at (0, 0) marks the surface TC center. The black circle represents the RMW near the surface. The orange box in (a) and (e) denotes the same area of (b)-(c) and (e)-(f), respectively. The mean height of each model level is shown in the title of each panel.
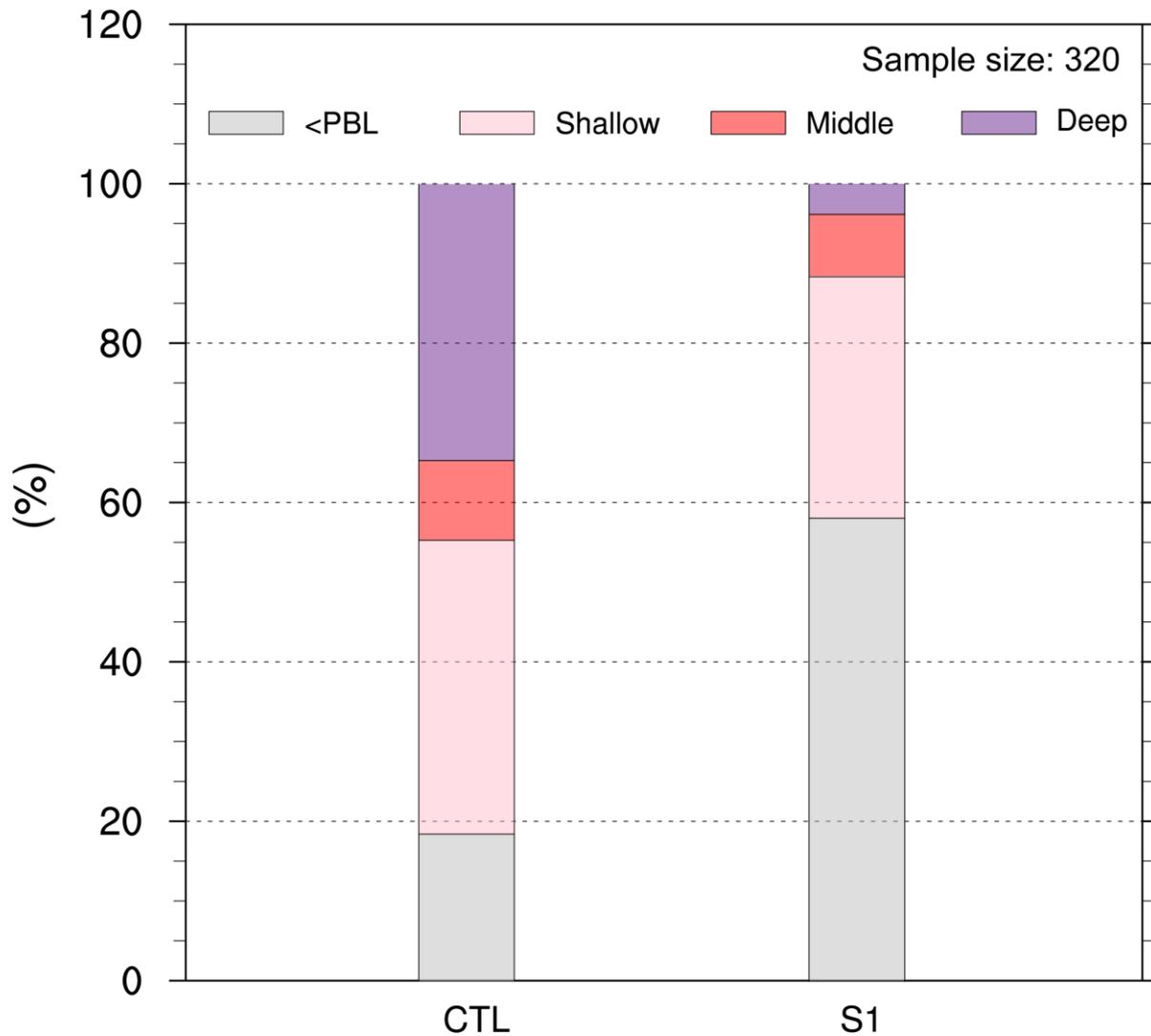
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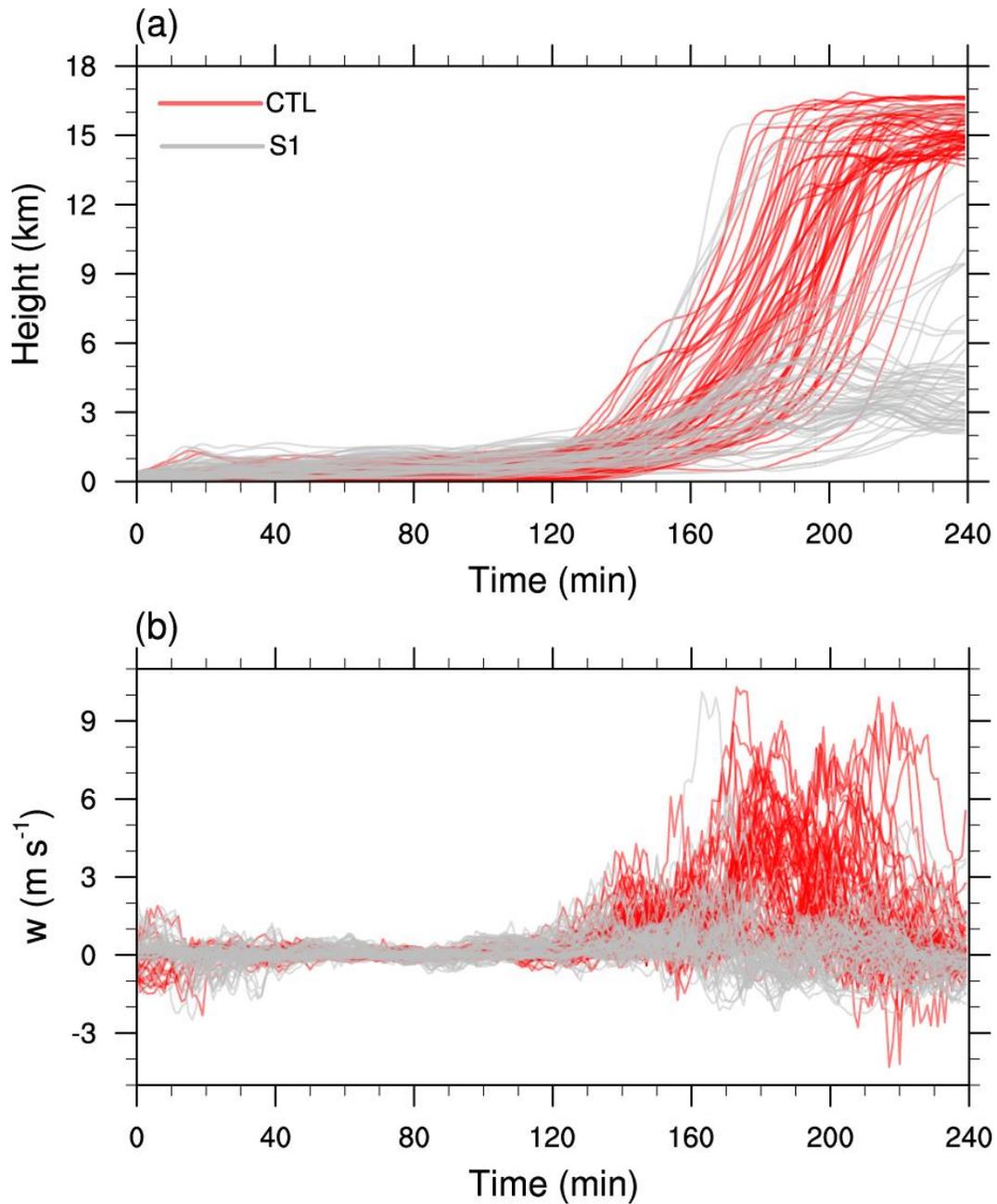
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Fig. 11. The evolution of (a) parcel height (km) and (b) vertical velocity (m s^{-1}) along the 4-h
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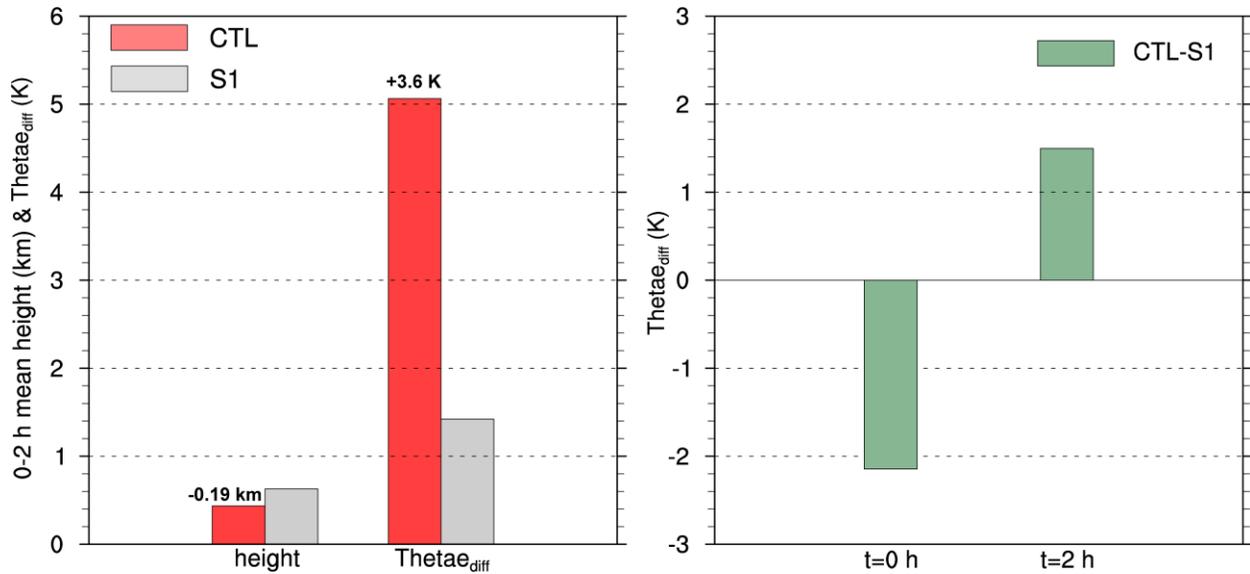
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957
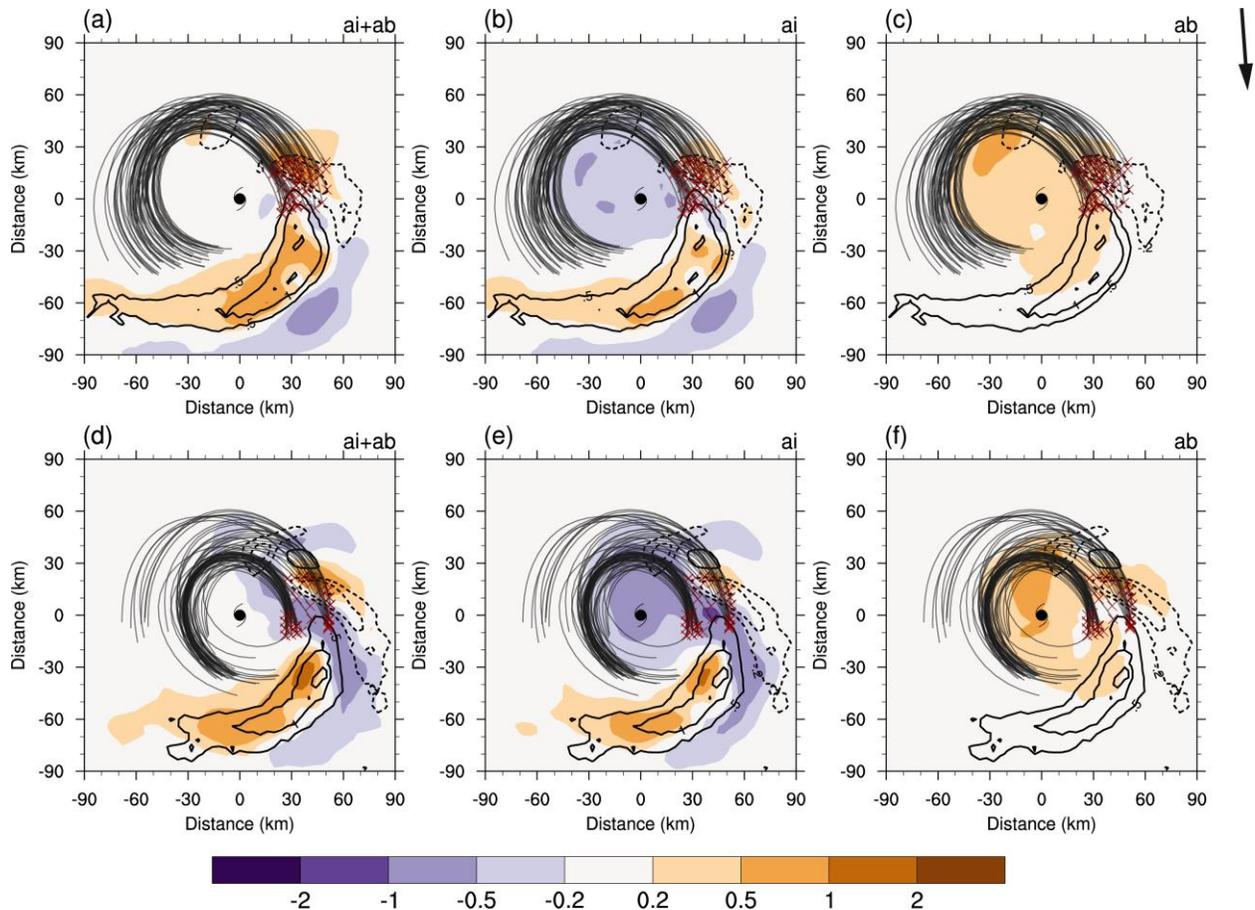
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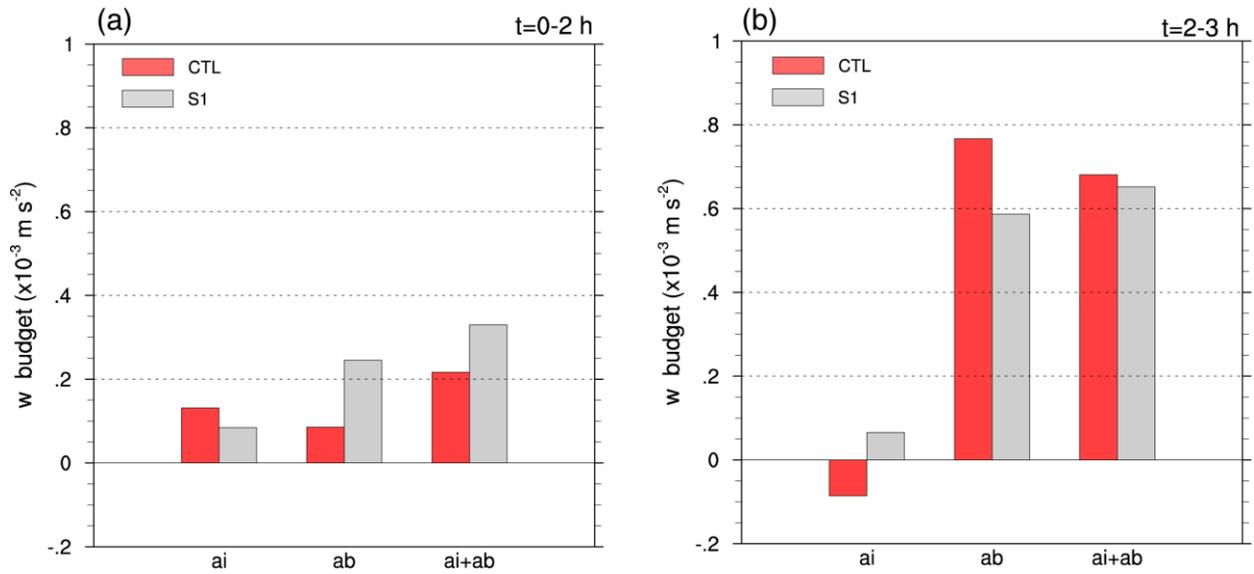
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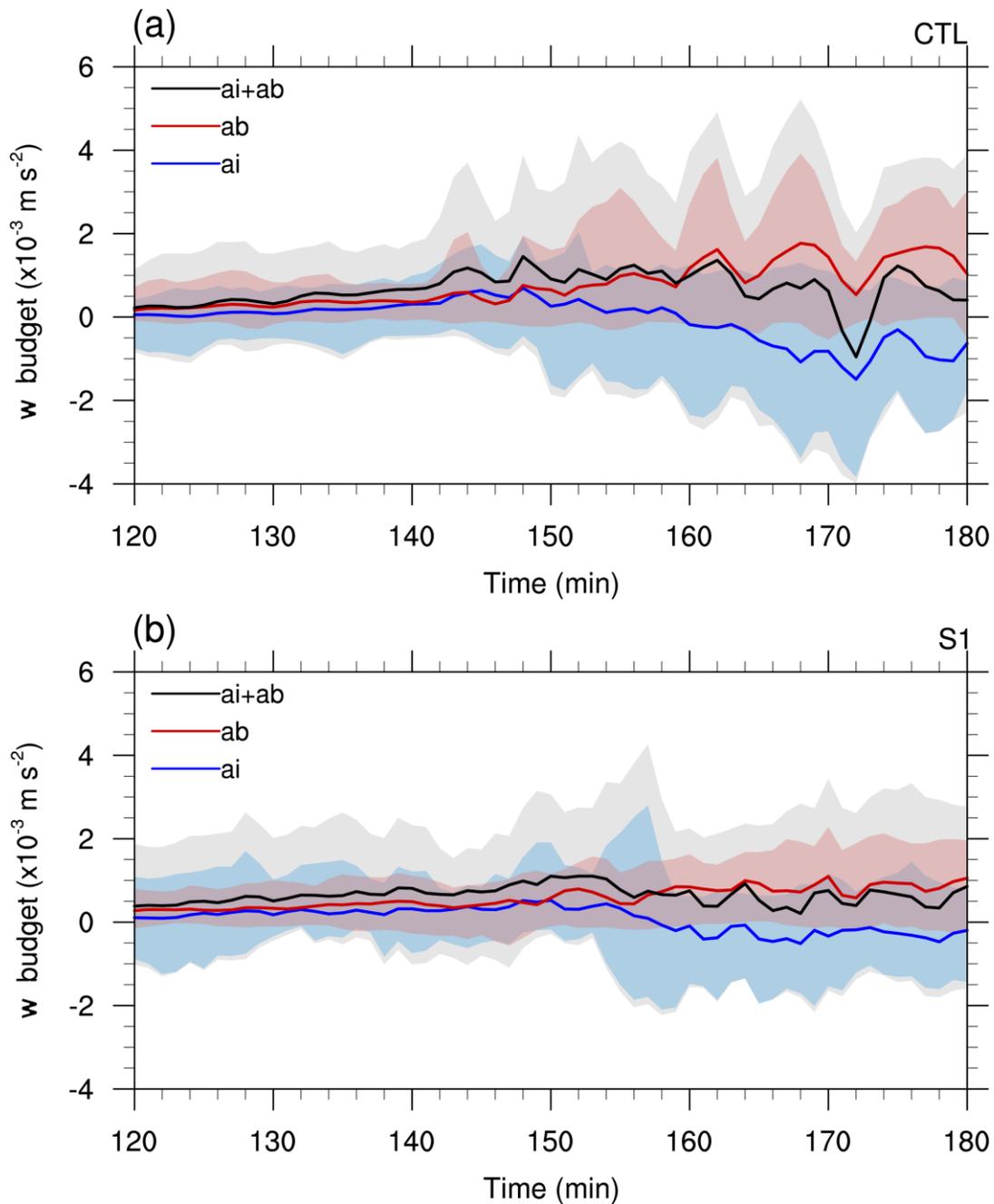
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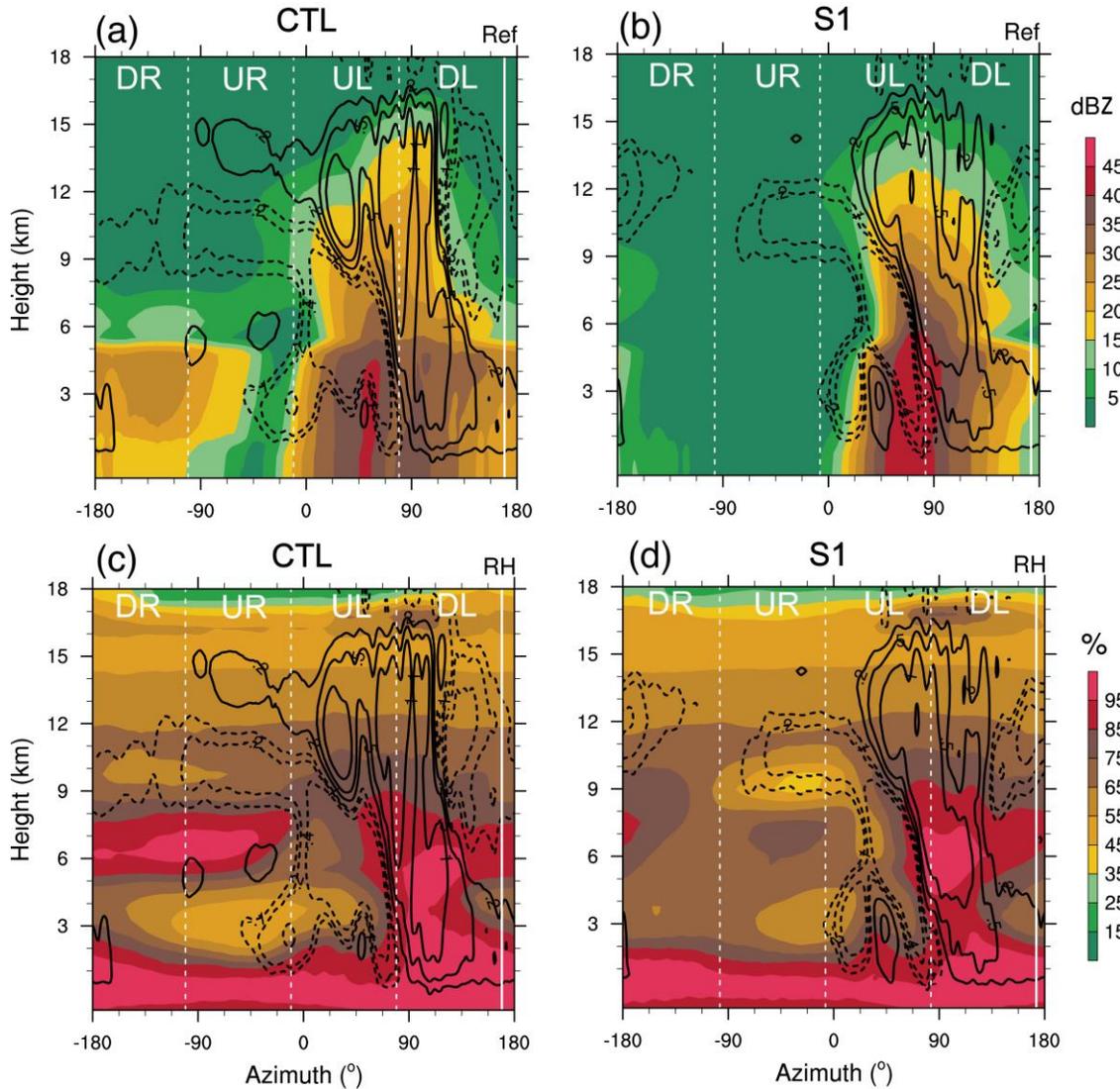
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Fig. 17. Azimuthal-height plot of radar reflectivity (shading, dBZ) and vertical velocity (contours
with values of −0.4, −0.2, −0.1, 0.2, 0.5, 1.0, and 2.0 m s^{-1}, negative values dashed) averaged
within r =20-60 km and over 1800-1850 UTC 2 October for the (a) CTL and (b) S1 TCs. (c)-(d)
As in (a)-(b), but shading denotes relative humidity (%). The vertical dashed white line marks the
shear-relative quadrants, as labeled at the top of each panel. DR, UR, UL, and DL denote
downshear-right, upshear-right, upshear-left and downshear-left quadrant, respectively.

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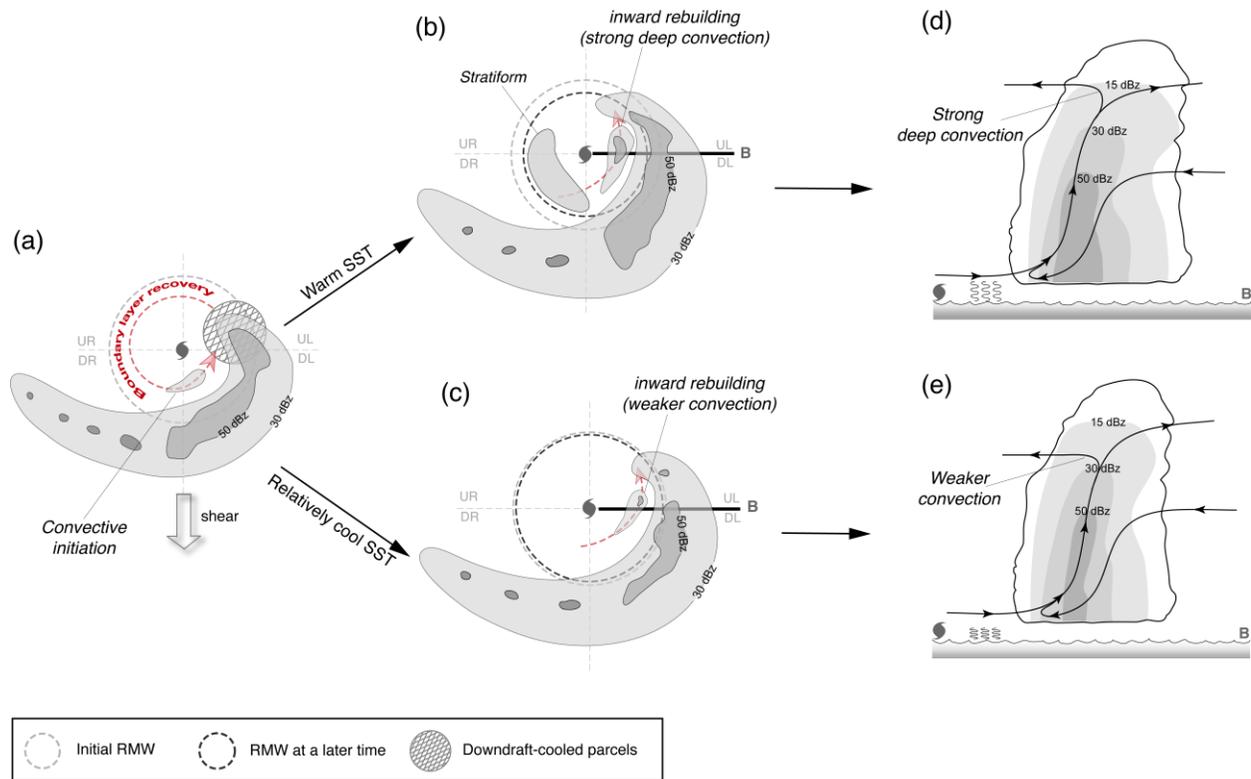
Fig. 18. Conceptual model for the inward rebuilding and precipitation symmetrization under different SSTs. (a) Plan view of the CPS that propagates into the upshear-left (UL) quadrant. Beneath the CPS, downdraft-cooled parcels in the boundary layer subsequently undergo boundary layer recovery and develop into convection during their propagation toward the downshear quadrants. (b)-(c) As in (a), but at a later time before the RI onset of the CTL TC over warm SSTs. In (b), the more efficient boundary layer recovery and more notable inward building of deep convection in UL maintain the precipitation symmetrization over warm SSTs; the stratiform in the right-of-shear semicircle in (b) indicates a nearly-saturated layer above the freezing level. (c) Over relatively cool SSTs, newly-developed convection in the inward rebuilding events is much weaker, which hinders precipitation symmetrization. Reflectivity contours represent the CPS and convective cells. The red dashed arrow in (a)-(c) denotes the trajectory along which boundary layer recovery and the subsequent inward rebuilding occur. (d)-(e) Composite vertical cross-sections of reflectivity and streamlines over warm and relatively cool SSTs, respectively. Locations of the cross sections are marked as thick black lines in (b)-(c).