Wintertime Supercell Thunderstorms in a Subtropical Environment: Numerical Simulation

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ABSTRACT

Following an earlier diagnostic study, the present paper performs numerical simulations of the rare wintertime supercell storms during 19–20 December 2002 in a subtropical environment near Taiwan. Using Japan Meteorology Agency (JMA) 20-km analyses and horizontal grid spacing of 1.5 and 0.5 km, the Cloud-Resolving Storm Simulator (CReSS) of Nagoya University successfully reproduced the three major storms at the correct time and location, but the southern storm decayed too early over the Taiwan Strait. The two experiments produce similar overall results, suggesting that the 1.5-km grid spacing is sufficient even for storm dynamics. Model results are further used to examine the storm structure, kinematics, splitting process, and the variation in the mesoscale environment. Over the Taiwan Strait, the strong surface northeasterly flow enhanced low-level vertical shear and helped the storms evolve into isolated supercells. Consistent with previous studies, the vorticity budget analysis indicates that midlevel updraft rotation arose mainly from the tilting effect, and was reinforced by vertical stretching at the supercell stage. As the ultimate source of vorticity generation, the horizontal vorticity (vertical shear) was altered by the baroclinic (solenoidal) effect around the warm-core updraft, as well as the tilting of vertical vorticity onto, and rotation of vortex tubes in the x–y plane, forming a counterclockwise pattern that pointed generally northward (westward) at the right (left) flanks of the updraft. In both runs, model storms travel about 15°–20° to the left of the actual storms, and they are found to be quite sensitive to the detailed low-level thermodynamic structure of the postfrontal atmosphere and the intensity of the storms themselves, in particular whether or not the existing instability can be released by forced uplift at the gust front. In this regard, the finer 0.5-km grid did produce stronger storms that maintained longer across the strait. The disagreement in propagation direction between the model and real storms is partially attributed to the differences in environment, while the remaining part is most likely due to differences not reflected in gridded analyses. Since the conditions (in both the model and real atmosphere) over the Taiwan Strait are not uniform and depend on many detailed factors, it is anticipated that a successful simulation that agrees with the observation in all aspects over data-sparse regions like this one will remain a challenging task in the foreseeable future.

1. Introduction

Supercell thunderstorms, characterized by a rotating and persistent updraft, are isolated cumulonimbi capable of producing damaging winds, large hail, and tornadoes (e.g., Browning 1964; Lemon and Doswell 1979;
Considerable success has been achieved through idealized modeling in explaining the behavior of supercells (e.g., Weisman and Klemp 1986; Klemp 1987). Studies of real events using single sounding and initial thermal perturbation have also proven useful, such as Kulie and Lin (1998), Atkins et al. (1999), van den Heever and Cotton (2004), and Sun (2005). However, successful simulations employing three-dimensional (3D), horizontally varying observations and real terrain without any perturbation have been few in number and remain quite challenging (e.g., Johnson et al. 1993; Finley et al. 2001; Chancibault et al. 2003). For example, Finley et al. (2001) utilized interior nudging during the integration in order to better capture the mesoscale environment in which the storm developed. The study of a storm over France by Chancibault et al. (2003), although with a location error of some 200 km, is also valuable and represents one of the few studies outside the United States.

In the afternoon of 19 December 2002, severe storms broke out near the coast of China and subsequently evolved into supercells that tracked eastward. In a subtropical environment south of 25°N, these rare wintertime supercells were the first near Taiwan to be reported in the literature. Their environment, initiation, and evolution are documented by Wang et al. (2009, hereafter W09). In this study, the storms are shown to take place behind the surface cold front, which provided strong low-level vertical wind shear (6.4 × 10^{-3} s^{-1} at 0–3 km) to combine with a weak-to-moderate convective available potential energy (CAPE) of 887 J kg^{-1} above the shallow cold air, thus yielding a suitable environment (e.g., Weisman and Klemp 1982; Rasmussen and Blanchard 1998; Thompson et al. 2003, 2007). An approaching upper-level jet (ULJ) also provided deep shear and drier conditions aloft. Obviously, supercell storms can develop near Taiwan in winter when both instability and vertical shear are present.

Near 1400 LST (LST = UTC + 8 h) 19 December 2002, convective storms initiated about 80 km inland along the southeastern coast of China, where significant daytime solar heating occurred over the mountain slopes (cf. Fig. 1; W09). The heating induced upslope/onshore winds and led to convergence and uplifting important to storm initiation. The storms evolved into three isolated supercells about 120 km apart with a northeast (NE)–southwest (SW) alignment, and traveled eastward across the Taiwan Strait at about 18 m s^{-1} (Figs. 1 and 2). The three primary storms were right movers and each experienced multiple splits and eventually made landfall over Taiwan (Figs. 2h,i and 3). Before they decayed, all three storms had lasted for about 10 h and traveled more than 550 km (cf. Fig. 2).

As a follow-up study to W09, the present paper addresses several issues from a modeling aspect. First, can this rare event in the maritime subtropics be reasonably simulated (i.e., forecast) in a high-resolution model using analysis data and real terrain without initial thermal perturbation? If so, which aspects of the simulation are more successful and which are less successful? Second, the storms apparently became isolated supercells over the Taiwan Strait where data were sparse. Through modeling, the mesoscale environment can be reproduced to shed light on their maintenance and the transition process to supercells. Third, the detailed structure, kinematics, and splitting process of storms can be further investigated using model results, if supercells are indeed reproduced. To answer these questions, numerical experiments are performed using the Nagoya University (NU) Cloud-Resolving Storm Simulator (CReSS; Tsuboki and Sakakibara 2002). This model has been applied in the past to study high-latitude cloud streets during cold-air outbreak (Liu et al. 2004), local heavy rainfall

![Fig. 1. Model domains of run 1 (light shaded) and run 2 (thin line), topography (m; shaded as indicated with additional white contours at 1 and 3 km), and the tracks (dotted) of the three primary storms (northern, central, and southern) during 1455–2300 LST 19 Dec 2002. The locations of Lungyen Doppler radar (cross), sounding stations (triangle), and the three storms when they reached maximum reflectivity (solid dot) are marked.](image-url)
FIG. 2. Base reflectivity (dBZ) observed by the Doppler radar at Lungyen (25.06°N, 117.19°E; cf. Fig. 1), with an elevation angle of 0.5°, at roughly 1-h intervals from (a) 1455 to (f) 1957 LST, and composite VMI radar reflectivity over the Taiwan area at (g) 2000, (h) 2100, and (i) 2200 LST 19 Dec 2002. Shading scales are indicated (a)–(f) to the left and (g)–(i) at the top.
FIG. 3. Radar reflectivity (dBZ, contours) and tracks (dotted lines) of the (a) northern, (b) central, and (c) southern storms and nearby cells at approximately 1-h intervals between 1455 and 2300 LST 19 Dec 2002. The reflectivity contours are drawn at 30, 40, and 50 dBZ (areas ≥50 dBZ blackened), and hatched every other time for clarity. The scale of distance, time (LST), cell numbers, and starting and ending positions of the primary right-moving storms are indicated (adapted from W09).
(Tsuboki 2004), and terrain-induced convective lines in the mei-yu season (Wang et al. 2005). It is perhaps worthwhile to note that past modeling studies on convective storms over the Taiwan Strait are also few, with one such example being Zhang et al. (2003).

This paper is arranged in the following manner. Section 2 describes the data used to give a brief overview of the case and to carry out the numerical study, as well as the CReSS model and its configuration. The overall results of model simulation are presented in section 3. A vorticity budget analysis at the storm scale and the related discussion on the transition to supercells are given in section 4. Following an investigation on the mesoscale environment over the Taiwan Strait and its effects on storm evolution in section 5, conclusions are given in section 6.

2. Data and CReSS model

The data used to review the supercell storms on 19 December 2002 in section 1 include base reflectivity by the Doppler radar at Lungyen, China [25.06°N, 117.19°E; 1504.9 m above ground level (AGL); cf. Fig. 1], at roughly 6-min intervals, and vertical maximum-echo indicator (VMI) radar reflectivity composites over the Taiwan area every 1 h. To verify model results, surface and upper-level analyses, Quick Scatterometer (QuikSCAT) oceanic winds, Geostationary Metereological Satellite-5 (GMS-5) cloud images, and rainfall data in Taiwan as used in W09 are employed. The objective analyses from the European Centre for Medium-Range Weather Forecasts (ECMWF), available every 6 h with a resolution of 1.125° latitude–longitude at 11 pressure (p) levels, are also compared with model outputs.

For model simulation, the Japan Meteorological Agency (JMA) 6-hourly (at 0200, 0800, 1400, and 2000 LST) gridded regional analyses at 20-km horizontal resolution at 20 p-levels (1000–10 hPa) during the case period are used as initial and lateral boundary conditions (IC and LBC, respectively). The variables provided are geopotential height, temperature (T), u and v wind components, and specific humidity. The JMA objective regional analysis is performed by the regional spectral model using multivariate 3D optimum interpolation (OI) to combine first-guess fields with observations that are free of internal/external quality control problems from a variety of platforms (Segami et al. 1989; Onogi 1998; Tsuyuki and Fujita 2002).

The CReSS model used in this study (v.2.2) is a non-hydrostatic, fully compressible, cloud-resolving model developed at the Hydrospheric Atmospheric Research Center of NU, Japan (Tsuboki and Sakakibara 2002, 2007). This model employs a terrain-following vertical coordinate ζ, defined as ζ = z[z - z_s(x, y)]/[z_t - z_s(x, y)], where z_t and z_s are heights at the model top and surface, respectively. Prognostic equations for 3D momentum, potential temperature (θ), p, and mixing ratios of water vapor (q_v) and other hydrometeors (q_w, where x denotes a species) are formulated, and the system includes all wave modes in the atmosphere. To properly simulate clouds at high resolution, an explicit bulk cold rain scheme based on Lin et al. (1983), Cotton et al. (1986), Murakami (1990), Ikawa and Saito (1991), and Murakami et al. (1994) are used without any cumulus parameterization (Table 1). A total of six species (water vapor, cloud water, cloud ice, rain, snow, and graupel) with microphysical processes of nucleation (condensation), sublimation, evaporation, deposition, freezing, melting, falling, conversion, collection, aggregation, and liquid shedding are included (Tsuboki and Sakakibara 2002). A bulk warm rain scheme that considers only liquid and gas phases is also available but not used here. Subgrid-scale turbulent mixing is parameterized using 1.5-order closure with turbulent kinetic energy (TKE) prediction (Tsuboki and Sakakibara 2007), and planetary boundary layer (PBL) processes are parameterized following Mellor and Yamada (1974) and Segami et al. (1989). The momentum and energy fluxes and radiation at the surface are also considered with a substrate model (Kondo 1976; Louis et al. 1981; Segami et al. 1989) but cloud radiation is neglected.

In the CReSS model, the Arakawa-C staggered and Lorenz grids are used in the horizontal and vertical with no nesting. For computational efficiency, a time-splitting scheme (Klemp and Wilhelmson 1978a) is adopted to separately integrate fast and slow modes. The filtered leap-frog method (Asselin 1972) is used for integration at large time steps (Δt), while the implicit Crank–Nicholson scheme is used in the vertical at small steps (Δz) by choice (Table 1). For parallel computing, data exchange between processing elements (PEs) is performed through the Message Passing Interface (MPI) and/or Open MP. Since the version 1.4 used in Wang et al. (2005), options for different map projections, nudging of radar data, and specification of sea surface temperature (SST), sea ice, and land-use types were implemented, augmenting the model’s applicability to larger domain, more sophisticated lower boundary, and data assimilation.

In this study, two experiments were performed and named run 1 and run 2. Run 1 had a horizontal grid spacing of 1.5 km and 65 vertical levels (stretched from a spacing of 100 m at the bottom to 475 m at the top) and used the JMA regional analyses as IC/LBCs, and the integration was from 0800 LST 19 December 2002 for 24 h without initial thermal perturbation (Table 1). At the lower boundary, terrain elevation at 30 s (~900 m)
and weekly SST at $1^\circ \times 1^\circ$ resolution (Reynolds et al. 2002) were provided. No data nudging at model interior was used and outputs were produced every 15 min. Afterward, a high-resolution run, run 2, was carried out with a grid spacing of 0.5 km from 1100 LST for a length of 14 h, using outputs of run 1 as IC/LBCs also without nudging (Table 1). The model top of run 2 was slightly lower, with a total of 63 vertical levels at nearly identical heights as run 1. Output frequency was also every 15 min, but increased to 5 min over 1400–1900 LST 19 December 2002 for analyses at shorter time intervals. The model configurations are summarized in Table 1 while domains are shown in Fig. 1. Later in this article, run 2 results are presented for those aspects at storm-scale (sections 3b to 4), while the coarser run 1 results are used in relation to mesoscale environment of storm initiation and propagation (sections 3a and 5).

3. Model results

a. Storm initiation

In W09, the convergence and uplifting associated with the upslope (onshore) flow were found to be vital in the initiation of storms in our case (section 1). Prior to 1400 LST, solar heating raised surface $T$ and $\theta$ by about 3 K over the sloping terrain near the coast under clear-sky conditions (their Figs. 15 and 17). Here, similar conditions were also reproduced in both runs using the IC/LBC and setting described in section 2. In the vertical cross section along AA’ that passed through one storm (storm n1, see next subsection for details) at 1400 LST in run 1 (cf. Fig. 6a), clear onshore flow of about 3–4 m s$^{-1}$ appears below about 1 km over the coastal region and convective cloud forms at its leading edge (Fig. 4a). The updraft reaches over 5 m s$^{-1}$ near 3–6 km (Fig. 4b), which is much stronger than revealed by the coarse JMA gridded analysis in W09. The increase in near-surface $\theta$ since 0800 LST (i.e., $\theta^*$), on the other hand, is also about 2–2.5 K and quite comparable to the analysis.

b. Overall storm evolution

Both run 1 and run 2 successfully reproduced convective storms over the area of interest on 19 December 2002. The overall storm evolution in run 2 is first shown by column-maximum mixing ratio of precipitating hydrometeors ($q_r + q_s + q_g$) in Fig. 5. Here, storms are labeled in a way similar to W09 (and Fig. 3), with odd (even) numbers for right- (left-) movers but using lower-case letters. The cells best corresponding to the three primary storms (Figs. 1–3) are “n1”, “c1”, and “s1”, respectively, and they were initiated over about 23°–23.5°N in a NE–SW alignment by 1400 LST among other cells with only small location errors (Fig. 5a; cf.
Fig. 4. Vertical cross sections from 24.10°N, 116.65°E to 23.63°N, 117.25°E (AA’ in Fig. 6a) at 0–6 km at 1400 LST 19 Dec 2002. (a) The 2D vectors on the section plane ($V_t$ and $w$; m s$^{-1}$) with zero-speed contour (thick dash line) and total mixing ratio of cloud hydrometeors ($q_v + q_i$; g kg$^{-1}$; shaded). (b) Speed of horizontal wind normal to section plane ($V_n$; m s$^{-1}$; thin contour, zero-line thickened), vertical velocity (m s$^{-1}$; shaded), positive potential temperature perturbation ($\theta'$; K; thick dash lines), and outline of cloud approximated by the 0.05 g kg$^{-1}$ value of $q_v + q_i$ (thick dash–dot gray lines). In (a), reference vector length of 20 m s$^{-1}$ for $V_t$ are indicated, and a length equivalent to 1 km in the vertical is 4 m s$^{-1}$ for $w$. In (b), contour intervals are 4 m s$^{-1}$ for $V_n$ and 0.5 K for $\theta'$ (starting at 0.5 K).
FIG. 5. Topography (km; shaded) and model-simulated column-maximum total mixing ratio of precipitating hydrometeors \( (q_r + q_s + q_g) \text{ g kg}^{-1} \) and 10-m horizontal wind (m s\(^{-1}\)) in run 2 at selected times from (a) 1400 to (l) 2300 LST 19 Dec 2002. The total mixing ratio is analyzed at 2, 5, 8, 11, and 14 g kg\(^{-1}\) with contours for 2, 8, and 14 g kg\(^{-1}\) thickened. For winds, full (half) barbs represent 5 (2.5) m s\(^{-1}\), while selected cells are labeled.
Table 2. Comparison between major characteristics of (top) observed and simulated storms by the CReSS model in (middle) run 2 and (bottom) run 1. For observed cells, radar data were unavailable prior to 1425 and after 2300 LST, 19 Dec 2002.

<table>
<thead>
<tr>
<th>Storm cell</th>
<th>Time of initiation (LST)</th>
<th>Life span (h)</th>
<th>Distance traveled (km)</th>
<th>Mean direction and speed* (°/m s⁻¹)</th>
<th>Size ≥ 40 dBZ (km)</th>
<th>Split</th>
</tr>
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<td></td>
<td></td>
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<tr>
<td>N1</td>
<td>By 1425</td>
<td>≥8.6</td>
<td>≥570</td>
<td>260°/19.6</td>
<td>15–35</td>
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<td>C1</td>
<td>By 1425</td>
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<td>≥518</td>
<td>268°/17.8</td>
<td>15–35</td>
<td>Yes</td>
</tr>
<tr>
<td>S1</td>
<td>By 1455</td>
<td>≥8.6</td>
<td>≥507</td>
<td>271°/17.4</td>
<td>15–40</td>
<td>Yes</td>
</tr>
<tr>
<td>N2</td>
<td>1525</td>
<td></td>
<td>247</td>
<td>215°/19.2</td>
<td>&lt;20</td>
<td>Yes</td>
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<tr>
<td>Others</td>
<td>–</td>
<td>~2–6</td>
<td>~150–400</td>
<td>~243°/-17–19</td>
<td>≤15</td>
<td>–</td>
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<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
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<tr>
<td>n1</td>
<td>1200</td>
<td>≥11.0</td>
<td>≥633</td>
<td>243°/16.0</td>
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<tr>
<td>s1</td>
<td>1245</td>
<td>≥12.3</td>
<td>≥570</td>
<td>248°/15.7</td>
<td>10–20</td>
<td>Yes</td>
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<tr>
<td>h1</td>
<td>1600</td>
<td>≥9.0</td>
<td>≥539</td>
<td>259°/12.1</td>
<td>10–25</td>
<td>Yes</td>
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<tr>
<td>h2</td>
<td>1830</td>
<td>4.8</td>
<td>237</td>
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<td>6.0</td>
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<tr>
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<td>623</td>
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<tr>
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<td>1500</td>
<td>9.0</td>
<td>410</td>
<td>258°/15.4</td>
<td>10–15</td>
<td>No</td>
</tr>
</tbody>
</table>

* The 500–700-hPa environmental wind was from 243° at 18.5 m s⁻¹ (W09).

Figs. 1 and 2). After formation, the three storms also went through splitting (e.g., n1 at 1645 LST, c1 at 1500 LST, and s1 at 1430 LST) then moved offshore during 1600–1730 LST near 24.5°, 23.9°, and 23.3°N, respectively (Figs. 5b–k), in good agreement with observation as well (Figs. 1 and 2). After 1700 LST, n1 and c1 continued to track toward the ENE over the Taiwan Strait, where the latter also strengthened to exhibit clear supercell characteristics, while s1 gradually weakened and decayed (Figs. 5g–k). As their propagation direction was not toward the east as observed, n1 missed Taiwan entirely, and c1 passed through the northern part of the island (Fig. 5l). Behind the major storms, several other cells in run 2 also experienced splitting that produced left movers of appreciable strength, most notably h1 and i1 (Figs. 5h–l) that also agree with radar data (see Figs. 8 and 9 of W09). Thus, although one cannot expect the model to simulate all individual storms correctly on such local scale, the overall storm evolution and morphology in run 2 bear close resemblance to the actual event on 19 December 2002 (Figs. 1–3 and 5).

While the model output does provide an appropriate analog for our case, especially for c1, differences still exist in storm tracks and other characteristics (Table 2, top and middle). Compared to real storms, right movers in run 2 generally travel about 15°–20° too much to the left (i.e., not right enough) and the southern storm s1 decays too early (Fig. 6b). In run 1, although the major storms tend to be weaker and less compact as expected, they are also reproduced with similar tracks (Fig. 6a and Table 2, bottom). The most notable difference in evolution is that c1 also weakens too early at 2000 LST (Fig. 6a). The similar overall results of the two runs suggest that a horizontal grid spacing of 1.5 km is sufficient to capture the high resolution in run 2 is essential in the maintenance of appreciable strength, most notably h1 and i1 (Figs. 5h–l) that also agree with radar data (see Figs. 8 and 9 of W09).

Figure 7 presents the peak updraft/downdraft strength ($w_{\text{max}}/w_{\text{min}}$) and relative vorticity ($\zeta_{\text{max}}$ or $\zeta_{\text{min}}$) of selected cells in run 2. For the three major storms, the peak updraft intensity (22–26 m s⁻¹) is reached by 1500 LST, within 3–4 h after initiation, then the splitting starts (Figs. 7a,b). After moving offshore, n1 and c1 experienced more splits, especially the latter whose updraft remained stronger, while the left movers (h2 and i2) over land also exhibit splitting behavior. Consistent with earlier studies (e.g., Klemp and Wilhelmson 1978b; Wilhelmson and Klemp 1981), storms split near or shortly after their updraft intensifies. The peak downdraft strength of these storms is about −9 to −14 m s⁻¹. The corresponding $\zeta_{\text{max}}$ (or $\zeta_{\text{min}}$) in run 2 also shows variation similar to the updrafts, and the peak values can exceed 2–4 × 10⁻² s⁻¹ (Figs. 7c,d). The above $w_{\text{max}}$ and $\zeta_{\text{max}}$ values in run 2 are comparable to those found in U.S. cases at similar grid resolutions (e.g., Weisman and Klemp 1984; Finley et al. 2001; van den Heever and Cotton 2004), and are about 1.5–2 times for $w_{\text{max}}/w_{\text{min}}$ and can be 2–4 times for $\zeta_{\text{max}}$ (or $\zeta_{\text{min}}$) greater than those in run 1.
Fig. 6. Tracks of major storm cells simulated by the CReSS model during 19–20 Dec 2002 in (a) run 1 and (b) run 2 with topography (km; shaded). Cell locations are marked as open dots at every full hour, and labeled with their assigned name (as in Fig. 5) and time (LST) every 3 h. End points are marked as solid dots and also labeled. Dotted line AA’ in (a) depicts the vertical cross section used in Fig. 4. The domain of run 2 is also plotted in (b).
c. Storm structure and morphology

In section 3b, it was shown that c1 is reproduced more successfully among the three major storms, and it undergoes a transition into an isolated supercell after moving offshore (Fig. 5). The structure of this storm on land at 1445 LST is revealed by Fig. 8 for an area of roughly 50 x 50 km². At this time, \( w_{\text{max}} \) of c1 reaches its peak (up to 26 m s\(^{-1}\)) after initiation (cf. Fig. 7b) and the storm is traveling at 243°/13.9 m s\(^{-1}\). From low to mid-levels (Figs. 8a,b), multiple updraft centers exist slightly to the east and southeast (at the front to right-front side) of the precipitation (shown by \( qr,qs,q_g \)). At 2.4 km, the rear-flank downdraft (RFD) is also visible north of the updraft (reaching \(-25\) m s\(^{-1}\)) and coincides with the precipitation (Fig. 8a). Although not evident at low levels, it is clear that the main updraft of c1 exhibits cyclonic rotation at 4.8 km (with \( \zeta \) reaching \(10^{-2}\) s\(^{-1}\)) while \( \zeta < 0 \) appears to the north (Figs. 8c,d). In a relative sense, the updrafts are also associated with negative pressure perturbation (\( p' \)) at the front side (to the E-NE) and positive \( p' \) at the rear [to the west-southwest (W-SW)], a configuration consistent with the horizontal gradient of vertical pressure gradient force that leads to downstream propagation of supercell storms (e.g., Rotunno and Klemp 1985; Klemp 1987; Cai and Wakimoto 2001).

The vertical cross sections through the main updraft of c1 at 1445 LST along lines BB’ and CC’ in Fig. 8b are presented in Fig. 9. Along BB’, strong storm-relative low-level inflow (from the E-NE) and upper-level outflow (from the W-SW) both exceed 20 m s\(^{-1}\) (Fig. 9a), in accordance to the strong vertical shear. The rain (\( qr \)) peaking near 7 g kg\(^{-1}\)) mainly appears below 5 km, snow (\( qs \)) above 7 km, and graupel over 4–9 km (also \( \approx 7\) g kg\(^{-1}\)). While the main updraft is over 15 m s\(^{-1}\) at 4–7 km, several other updrafts are also developing nearby with the closest only 12 km ahead (Fig. 9a). Thus, c1 still displays some multicell characteristics and has yet to become an isolated and quasi-steady supercell at 1445 LST. Associated with the updraft and condensational heating, increase in \( u \) (downward intrusion of adiabats) is seen at 4–9 km in section CC’, while some cooling exists near cloud edges (Fig. 9b).

At 1730 LST, storm c1 is moving at 259°/17.3 m s\(^{-1}\) over the Taiwan Strait and has become more isolated in run 2 (cf. Fig. 5). From 2.4 to 7.2 km, the relative flow changes from about 25 m s\(^{-1}\) from the east to nearly 20 m s\(^{-1}\) from the southwest (Figs. 10a–c), veering clockwise and consistent with the right-moving behavior of c1. At low levels, the strong inflow is collocated with the main updraft with an inflow notch at the immediate
right-front side of the precipitation area. At 7.2 km, however, the updraft becomes coincident with precipitation (Fig. 10c). The RFD at 2.4 km and the forward-flank downdraft (FFD) at 4.8 km (near 24.01°N, 118.31°E) are both visible (Figs. 10a,b). At this time, a single, strong updraft is associated with cl while a left-moving cell c4 is under development (near 24.05°N; Figs. 10b,c). As expected, both the updrafts of cl and c4 are rotating, with $\zeta$ reaching $10^{-2}$ s$^{-1}$ for the former and $-5 \times 10^{-3}$ s$^{-1}$ (anticyclonic) for the latter (Figs. 10d–f). Again, the forward-directed pressure gradient force is clear, especially at 4.8 km, with a horizontal gradient of about 2 hPa over only 3 km (0.67 Pa km$^{-1}$) across the updraft. The pressure gradient also rotates clockwise with height as observed (Figs. 10d–f), which favors right movers (e.g., Weisman and Klemp 1984; Klemp 1987).

In the vertical cross section along line DD’ (Fig. 10b), again, a single and strong updraft is present (Figs. 11a,b). The maximum concentrations of graupel and rain are about 10 and 4 g kg$^{-1}$, respectively, and snow particles

FIG. 8. Plain views of storm cl in run 2 at 1445 LST 19 Dec 2002, of storm-relative horizontal winds (m s$^{-1}$; vectors), vertical velocity ($w$; m s$^{-1}$; thick contours), total mixing ratio of rain, snow, and graupel ($q_r + q_s + q_g$; g kg$^{-1}$; shaded), and cloud boundary (thick dash–dot lines) approximated by 0.5 g kg$^{-1}$ of the column-maximum total mixing ratio of all five species of hydrometeors, at (a) 2366 and (b) 4829 m. (c),(d) Same as (a),(b) but for pressure perturbation (hPa; shaded), the vertical component of relative vorticity ($\zeta$; $10^{-3}$ s$^{-1}$; thin contours), positive $w$ (thick contours), and cloud boundary. Shading scales and reference vectors of 20 m s$^{-1}$ are indicated. Contours for $w$ are every 5 m s$^{-1}$ plus one additional level at $-2$ m s$^{-1}$, and those for $\zeta$ are every $5 \times 10^{-3}$ s$^{-1}$ plus $\pm 2 \times 10^{-3}$ s$^{-1}$ (dashed for negative values and zero-line omitted for both). Straight dash lines BB’ and CC’ in (b) are used to construct the vertical cross sections shown in Fig. 9.
are mostly within the anvil farther downstream. Within the updraft, the warming as shown by $u$ (perturbation since 1100 LST in run 2) can reach almost 4 K in upper levels. On the other hand, evaporative cooling also occurs in association with rain at low levels as well as below the anvil (both over $q_r$; Fig. 11b). Along section EE’ (Fig. 11c), it can be seen that the updraft of c1 tilts toward the NNW (left) with height and keeps a large amount of graupel suspended aloft, forming a clear vault that corresponds to the bounded weak echo region (BWER). The rain falls below the strongest ascent, over a region of transition from updraft to downdraft. The developing updraft of c4 ($\sim 5$ m s$^{-1}$) from a splitting process is also depicted (Figs. 11c,d).

d. Storm splitting process

The storm splitting process is also well reproduced in both runs, as predicted when significant vertical shear is present from previous studies (e.g., Klemp and Wilhelmson 1978b; Wilhelmson and Klemp 1981; Rotunno 1981; Rotunno and Klemp 1982; Klemp 1987). Distributions of 1-h rainfall for four selected periods and areas are shown in Fig. 12. With a 500–700-hPa environmental wind of 243°/18.5 m s$^{-1}$ (cf. Table 2), the separation of storm tracks into right and left branches is clearly depicted during the split of s1 and s2 and h1 and h2 (Figs. 12a,c). The rainfall resulted from c1 has significantly wider swath (over 20 km for amount $\geq$5mm) after it becomes more isolated (Fig. 12b), in agreement with Fig. 10a of W09. In Fig. 12d, the split of h3 from left-moving h2 during 2000–2100 LST is also visible (cf. Figs. 5j,k). The splitting process for storm c1 during a 25-min period in its mature stage from 1715 to 1740 LST at midlevel in run 2 is presented in Fig. 13. At 1715 LST, the main updraft is located at the right flank (southeast side) of the storm complex and clearly associated with cyclonic rotation (Fig. 13a). North of the updraft, an area of anticyclonic rotation exists where ascending motion gradually forms at 1725 LST (Fig. 13b). This updraft and its accompanying precipitation north of 24°N continue to develop, and become well established as c4 ($\sim 5$ m s$^{-1}$) at 1730 LST (Fig. 13c). With $w_{\text{max}} > 5$ m s$^{-1}$ and $q_r + q_g > 5$ g kg$^{-1}$, cell c4 further separates from c1 at 1740 LST after the split (Fig. 13d).

The splitting of h1 during 1820–1855 LST at 4.8 km is shown in Fig. 14. Not long after its initiation (Figs. 5i,j), this storm is mainly composed of a single updraft at 1820 LST (Fig. 14a). At 1830 LST, the precipitation grows and extends toward the rear of the storm, and a new updraft h2 develops (Fig. 14b). The distance between old and new updrafts gradually increases through 1855 LST and the split is completed (Figs. 14c,d). Afterward, both h1 and h2 continue to intensify (Figs. 5i–l) and the left-moving h2 also experiences splitting (Fig. 12d). In Figs. 13 and 14, it is quite clear that storm updrafts tend to strengthen at the forward flank of the precipitation,
FIG. 10. (a)–(c) Same as Figs. 8a,b, but for storm c1 at 1730 LST 19 Dec 2002 at (a) 2366, (c) 4829, and (e) 7105 m. (d)–(f) Same as Figs. 8c,d, but for c1 at 1730 LST at 2366, 4829, and 7105 m, respectively. Straight dash lines DD' and EE' in (b) are used to construct the vertical cross sections shown in Fig. 11.
which in turn often intensifies within several minutes after the updraft enhanced, in agreement with earlier studies (e.g., Weisman and Klemp 1986).

4. Storm vorticity budget analysis

To understand whether supercell-like structure can be replicated by the CReSS model, a vorticity budget analysis is performed for the period of 1600–1800 LST, where storm c1 evolved into an isolated supercell (cf. Fig. 5) and outputs at 5-min intervals are available in run 2. The vorticity equations commonly used to interpret supercell dynamics are obtained by taking the curl of the Boussinesq equations of motion (e.g., Klemp 1987; Weisman and Rotunno 2000; Chancibault et al. 2003). When transformed into a storm-relative (quasi-Lagrangian) frame for interpretation (e.g., Lee and Wilhelmson 1997), they can be written as

\[
\frac{\partial \zeta}{\partial t} = - \left[ (\mathbf{v}_H - \mathbf{c}) \cdot \nabla \right] \zeta - w \frac{\partial \zeta}{\partial z} + (Îô_H \cdot \nabla H)w + Îô \frac{\partial w}{\partial z} + F_z, \tag{1}
\]
\[
\left( \frac{\partial \omega_H}{\partial t} \right)_{\text{SR}} = -[(v_H - c) \cdot \nabla_H] \omega_H - w \frac{\partial \omega_H}{\partial z} + (\omega \cdot \nabla) V_H + \nabla \times (Bk) + F_{\text{vort}}, \quad \text{and (2)}
\]

\[
\nabla \times (Bk) = \frac{R}{\rho} \nabla_p \times \nabla T, \quad \text{(3)}
\]

where \((\partial/\partial t)_{\text{SR}}\) is the local derivative in the storm-relative frame, \(v_H\) is the horizontal velocity vector (relative to the ground), \(c\) is the (constant) storm-motion vector, \(v_H - c\) is the storm-relative horizontal velocity, \(w\) is vertical velocity, \(\omega = (\xi, \eta, \zeta)\) is the 3D vorticity vector while \(\omega_H = (\xi, \eta)\) is the 2D vector on the plane, and \(V_H\) is \(u\) or \(v\). The terms \(F_p\) and \(F_{\text{vort}}\) represent mixing, and \(B\) is buoyancy, \(R\) the gas constant, and \(k\) the unit vector in \(z\) direction. In both equations, the first two terms on the rhs are horizontal and vertical advection, respectively. In Eq. (1), the third and forth terms are the generation of vertical vorticity by tilting of horizontal vorticity into the vertical \([\omega_H \cdot \nabla] w = \xi (\partial w/\partial x) + \eta (\partial w/\partial y)\) and by stretching of existing vertical vorticity. Equation (2) has two components \((\partial \xi/\partial t, \partial \eta/\partial t)\), and the third term is the combined effect of stretching \([\xi (\partial u/\partial x), \eta (\partial u/\partial y)]\) and rotation \([\eta (\partial u/\partial y), \xi (\partial u/\partial x)]\) of vortex tubes in the \(x-y\) plane, and tilting of vertical vorticity onto the \(x-y\) plane \([\xi (\partial u/\partial z), \eta (\partial u/\partial z)]\). The fourth term on the rhs of Eq. (2) is the baroclinic (or solenoidal) generation due to a buoyancy gradient, and can be expressed as Eq. (3).

Figure 15 presents the results of the vorticity budget analysis for the major terms in Eq. (1) for \(c_1\) near 4 km at

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**Fig. 12.** Model-simulated 1-h accumulative rainfall over selected areas for (a) 1415–1515, (b) 1715–1815, (c) 1815–1915, and (d) 2000–2100 LST 19 Dec 2002 in run 2. Shading scale is indicated to the lower right of (c),(d), and isohyets at 1, 10, and 30 mm are drawn. Maximum rainfall areas caused by different cells are also marked.
1630 LST, when the storm is moving offshore (cf. Fig. 5). In Fig. 15a, the main updraft is elongated and peaks at about 7 m s$^{-1}$. Since the vorticity vector $\omega_H$ usually has a northward component associated with westerly shear, tilting from updrafts generates positive (negative) $\zeta$ at the southern (northern) flank (and the opposite for downdrafts) as expected (Fig. 15a). The peak value just south of the main updraft is almost $4 \times 10^{-5}$ s$^{-2}$. Near updraft centers of c1 (within $\sim$3 km) and also to their south, the stretching effect has opposite sign to vorticity; that is, $\partial \zeta / \partial t < 0$ over regions of $\zeta > 0$ and vice versa, with a maximum rate of about $-4 \times 10^{-5}$ s$^{-2}$ (Fig. 15b). This is because the updraft tilts northward with height and $\partial w / \partial z$ at 4 km AGL is in fact negative, resulting in vertical shrinking of vortex tubes. When the two effects are combined (Fig. 15c), the pattern of $\zeta$ generation agrees more with that in Fig. 15a (and the pattern of $\zeta$ itself), confirming that the tilting effect is the primary

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Fig. 13. Same as Fig. 8b, but for storm relative winds (m s$^{-1}$; vectors), vertical velocity ($w$; m s$^{-1}$; thick lines), column-maximum mixing ratio of rain and graupel ($qr + qg$; kg$^{-1}$; shaded), and cloud boundary (thick dash–dot lines) approximated by the 0.5 kg$^{-1}$ value of the column-maximum total mixing ratio of all five species of hydrometeors for storm c1 at 4829 m at (a) 1715, (b) 1725, (c) 1730, and (d) 1740 LST 19 Dec 2002. For $w$, contour intervals are drawn every 5 m s$^{-1}$ (starting from 5 m s$^{-1}$), while shading scales and reference vectors of 20 m s$^{-1}$ are indicated.
source for the storm’s overall rotation at midlevels. Once positive $\zeta$ is generated, it is advected by the storm-relative flow toward the updraft from the south/southeast (Fig. 15d). On the other hand, vertical advection tends to cancel with stretching (Fig. 15e). When all effects in Eq. (1) are combined (except friction), the total tendency $(\partial \zeta / \partial t)_{SR}$ is mostly positive at updraft centers and thus tends to enhance existing $\zeta$ (Fig. 15f). At a
maximum rate of $6 \times 10^{-3}$ $s^{-2}$, only 2.5 min are needed to generate a vertical vorticity close to $10^{-2}$ $s^{-1}$, which is roughly the peak value in Fig. 15b.

At 1730 LST when c1 has moved offshore and becomes an isolated supercell (Fig. 16a; cf. Figs. 5, 10, and 13), its updraft reaches 14 m $s^{-1}$ near 4 km (Fig. 11), twice than that at 1630 LST. The tilting effect is also more effective and can reach $9 \times 10^{-5}$ $s^{-2}$, consistent with the acceleration in storm rotation with a peak $\zeta$ of $\sim 1.4 \times 10^{-2}$ $s^{-1}$ (Fig. 16b). The stretching effect continues to destroy $\zeta$ along the southern flank because the updraft still tilts northward with height. However, it starts to reinforce existing $\zeta$ near the center and along the northern flank (Fig. 16b), as the updraft becomes significantly wider (cf. Figs. 10 and 11). Thus, the combined effect of tilting and stretching becomes strongly positive near the updraft center (Fig. 16c). The horizontal advection effect near 4 km is now somewhat reduced because of less penetration by the relative flow through the updraft, while vertical advection continues to counteract the stretching term (Figs. 16e,f). Finally, the resulted total $\zeta$ tendency in the storm-relative frame, still dominant by tilting among all terms, is positive over the region of $\zeta > 0$, at the updraft center, and along its right flank (Fig. 16f).

From low levels up to 9 km, the budget terms in Eq. (1) reveal that the tilting effect is significant above 2.5 km and becomes strong at midlevels (up to 6 km, not shown). In the upper troposphere where $\mathbf{V}_H \mathbf{w}$ is large (cf. Figs. 11b,d), it can also be significant. The positive stretching effect close to the updraft center, once $\zeta$ is produced, is typically maximized near 3.5 km (not shown), where the upward acceleration is more intense (cf. Fig. 11). The combined effect from both terms, therefore, usually peaks near 3–4 km.

As the source for $\zeta$ generation, the horizontal vorticity vector $\mathbf{\omega}_{H}$ (i.e., vertical shear) is clearly not uniformly distributed on the $x$-$y$ plane and typically points northward (westward) at the right- (left-) front flank of the updraft (Figs. 15a and 16a). This counterclockwise (cyclic-rotating) pattern indicates that westerly (southerly) shear is enhanced at the right (left) flank. To a lesser degree, $\mathbf{\omega}_{H}$ generally has an anticyclonic pattern near downdrafts. The rates of $\mathbf{\omega}_{H}$ generation from individual effects in Eq. (2) at 1730 LST are shown in Fig. 17. The combined effect from the third and forth terms (Fig. 17a) shows that $[(\partial \mathbf{\omega}_{H}/\partial t)_{SR}]$ usually has an eastward or forward (westward to southward or backward) component at the right (left) flank of the updraft, over the area where $\zeta > 0$ (<0), also forming a counterclockwise pattern for $\mathbf{\omega}_{H}$ tendency. This distribution of $[(\partial \mathbf{\omega}_{H}/\partial t]$, thus, tends to reinforce existing vortex tubes at both flanks (gray shades), in agreement with Brandes et al. (1988), who noted a similar $\mathbf{\omega}_{H}$ pattern near the updrafts. For individual components, the horizontal stretching effect $[\xi(\partial \mathbf{u}/\partial x), \eta(\partial \mathbf{u}/\partial y)]$, produced by speed change in the direction of vortex tubes, tends to be slightly negative (i.e., weakening existing $\mathbf{\omega}_{H}$) since air usually decelerates (on the $x$-$y$ plane) when entering the updraft (Fig. 17b). The combined effect of horizontal rotation $[\eta(\partial \mathbf{u}/\partial y), \xi(\partial \mathbf{u}/\partial x)]$ and the tilting of $\zeta$ onto the $x$-$y$ plane $[\dot{\zeta}(\partial \mathbf{u}/\partial z), \zeta(\partial \mathbf{u}/\partial z)]$ is dominated by the latter. At the right flank, the westerly vertical shear tends to twist the positive (upward pointing) $\zeta$ into (eastward pointing) $\xi$, which is then rotated cyclonically toward the north (because $\zeta > 0$), thus strengthening existing $\mathbf{\omega}_{H}$ (Fig. 17c). At the left flank, likewise, the southerly shear tends to twist negative $\zeta$ (pointing down) into negative $\eta$ (pointing south) that is then rotated toward the west. Therefore, this combined effect corresponds to a cyclonic pattern for $\dot{\mathbf{\omega}}_{H}/\dot{t}$ that bears similarity to the total effect in Fig. 17a. The baroclinic effect, with strong ascent within a warm-core updraft and descent (or weaker ascent) farther away, also produces cyclonic $\mathbf{\omega}_{H}$ turning tendency (Fig. 17d; Brandes et al. 1988) that contributes toward the total effect. Considering also horizontal advection, which tends to be positive near the updraft center (Fig. 17e), and vertical advection (not shown), the total tendency $(\dot{\mathbf{\omega}}_{H}/\dot{t})_{SR}$ exhibits a cyclonic pattern over the region of $\zeta > 0$ near the main updraft (Fig. 17f). It is noted that all terms in Fig. 17 have similar peak magnitude of about $5\sim 10 \times 10^{-5}$ $s^{-2}$, in the same order as the terms in Eq. (1). Consistent with previous studies (e.g., Kulie and Lin 1998; Chancibault et al. 2003), the baroclinic generation of $\mathbf{\omega}_{H}$ is identified to be an important source for local enhancement of vertical shear, which then can be tilted near and stretched inside the updraft to produce storm rotation. Based on Fig. 17, however, once appreciable $\zeta$ is present, regardless of its sign, it can also be tilted onto the $x$-$y$ plane and then rotated, and contributes toward the modification of vertical shear (i.e., $\mathbf{\omega}_{H}$) into the horizontally varying, cyclonic pattern as seen in Fig. 16a.

In other words, there exists complex interaction among the three $\mathbf{\omega}$ components ($\xi, \eta, \zeta$) through kinematics that readily affect one another near the rotating updraft of supercell thunderstorms. The roles of tilting effect (from $\zeta$ into $\xi$ or $\eta$) and horizontal rotation in mature storms have not been emphasized recently in the literature.

5. Mesoscale environment over the Taiwan Strait

One of the major differences between modeled and observed storms is the shorter lifespan of “s1” in the model. While “n1” and “c1” last much longer in run 2, one wonders why these model storms behave differently.
FIG. 16. Same as Fig. 15, but for storm c1 at 3984 m at 1730 LST 19 Dec 2002.
As c1 improved in total lifespan using a finer grid, further increase in resolution from run 2 may help s1 last longer, but some reasons related to the variation in mesoscale environment must also exist. Using QuikSCAT oceanic winds at 1825 LST and largely land-based observations at 2000 LST, the surface mesoscale environment over the Taiwan Strait was manually analyzed and compared with the model fields in run 1 in Fig. 18. Only subtle differences exist between the two, as the observed winds appeared slightly stronger over the strait while the model surface air of 19°–22°C seemed to have advanced more to the south. The strong northeasterly flow over the strait, in contrast to the weak onshore flow along the coast of China, indicates that the low-level vertical shear was much stronger over the ocean. This explains why the storms evolved into isolated supercells once they moved offshore in the current case.

To further shed light on the differences in mesoscale environment in the model, five points over the strait, marked “A” through “E” from north to south in Fig. 18b, are selected to compute several thermodynamic and shear parameters. These include CAPE and convective inhibition [CIN (J kg⁻¹); Weisman and Klemp 1982; Colby 1984] at the most unstable level, the distance to the level of free convection (LFC), 0–3 km mean shear (Thompson et al. 2007) and storm relative helicity (SRH; Davies-Jones et al. 1990), bulk Richardson number (Moncrieff and Green 1972), energy–helicity index (Hart and Korotky 1991; Davies 1993), and supercell composite parameter (SCP; Thompson et al. 2003). As shown in Fig. 6b, storm n1 passed near point B (25°N, 120°E) at around 1945 LST and point A (25.5°N, 121°E) at around 2115 LST, while c1 passed near point C (24°N, 119°E) around 1845 LST in run 2. On the other hand, s1 dissipated when approaching point D (23.5°N, 118.5°E) at about 2000 LST, and h1 passed near the same point later at 2330 LST. Point E (22.5°N, 118°E), slightly farther to the south, is chosen for comparison. Time series of selected parameters, computed from run 1 results, are shown in Fig. 19, in which the occurrences noted above are also marked.

In the late afternoon/early evening of 19 December, maximum CAPE increased from north to south (Fig. 19a) as the surface cold air was thicker (and colder) to the north (cf. Fig. 18). At points A and B, CAPE was almost nonexistent initially but increased after 2100–2300 LST. Similarly, CAPE at C and D decreased before 1700 LST to about 200 J kg⁻¹ and increased after 2000 LST. The CAPE values decreased because of the southward advance of cold air, but were slowly restored as the near-surface air was gradually warmed and moistened by ocean fluxes. However, since the southern storm s1 dissipated near D at 2000 LST (CAPE ~170 J kg⁻¹) while n1 passed A and B during 1945–2115 LST (CAPE <100 J kg⁻¹), there appears to be a paradox. Therefore, CIN and the distance to LFC are also examined to yield information about how easily CAPE can be released (Figs. 19b,c). Consistent with earlier discussion, both CIN and the distance to LFC at A and B were very large in the afternoon with no chance for convection, but decreased drastically to within 50 J kg⁻¹ and 100 hPa prior to 2000 LST. For this reason, convection could occur, though with only limited CAPE, given the strength of storm n1 around 2000 LST (cf. Fig. 7a). When other storms appeared near A and B later around 2300 LST (cf. Fig. 5l), CIN became small (<40 J kg⁻¹) while CAPE grew further, and the distance to LFC reduced to ≤200 hPa (Figs. 19a–c). This indicates only weak stability with respect to saturated ascent. At point C at 1845 LST and point D at 2000 LST, some CAPE (~200 J kg⁻¹) existed and CIN was less than 75 J kg⁻¹, and the distance to LFC was about 120 hPa at C but almost 200 hPa at D (Figs. 19b,c). Thus, whether the limited CAPE can be released at C and D through forced uplift at the gust front depends heavily on the strength of the passing storm. Therefore, the stronger storm c1 in run 2, with a finer grid and an inflow layer at least 3 km in depth (cf. Figs. 10 and 11), could be maintained when it propagated across the strait. On the contrary, storms s1 in both runs and c1 in run 1, once becoming too weak, cannot sustain themselves in such an environment with a long path to LFC. Slightly farther south at point E, CAPE decreased from 1400 LST but still remained at 400–600 J kg⁻¹ after 1900 LST, and CIN increased to 50–70 J kg⁻¹. The distance to LFC was also large (>200 hPa) during 1600–2300 LST, but
FIG. 18. (a) QuikSCAT oceanic winds with isotach (kt; thick solid lines) over the Taiwan area at 1825 LST and manual mesoscale surface analysis of MSL temperature (°C; dashed) at 2000 LST 19 Dec 2002. (b) Model-simulated surface winds (barbs in knots) with isotach (m s⁻¹; shading) at 10-m height at 1830 LST, and temperature (°C; contour) at the lowest model level of 50 m at 2000 LST 19 Dec 2002 in run 1. Isotherm intervals are 2°C in (a) and 1°C in (b), and solid dots mark the position of the three primary storms at 1830 LST. Letters A through E indicate the locations of model sounding used in Fig. 19.
Fig. 19. Selected thermodynamic and shear parameters at points A through E (cf. Fig. 18b for locations) in run 1 during 1100 LST 19 to 0200 LST 20 Dec 2002. 
(a) Maximum CAPE (J kg$^{-1}$), (b) CIN (J kg$^{-1}$) for the most unstable level, (c) distance to LFC (hPa), (d) 0–3 km (or surface to 700 hPa) mean vertical shear ($10^{-3}$ s$^{-1}$), (e) 0–3-km SRH (m$^2$ s$^{-2}$), and (f) supercell composite parameter. All variables were computed at times labeled by ▽ at the top, while only SRH was interpolated at times labeled by ▼. At other times not labeled, all variables were interpolated at 30-min intervals. Values of CIN and the distance to LFC are limited to 200 J kg$^{-1}$ and 500 hPa, respectively. Letters P and D inside the figures represent passage and dissipation of a storm nearby, while the subscript indicates the point involved.
gradually lowered afterward also because of surface fluxes.

In contrast to CAPE, low-level mean vertical wind shear generally decreased from north to south, but was significant ($5-10 \times 10^{-3} \text{ s}^{-1}$) over the Taiwan Strait at all five points (Fig. 19d). The perturbation in shear caused by n1 at point B at 2000 LST was also clear. The 0–3 km SRH generally peaked at 73–82 m$^2$ s$^{-2}$ during 1700–2000 LST at points A through D, but the same level was not reached at point E until 0100 LST 20 December (Fig. 19e). The time series of SCP shows that its value increased to 0.7 at point B, but only 0.3 at point A as n1 moved close. At C and D, SCP was about 0.6–1.0 when c1, s1, and h1 moved through or dissipated nearby. Thus, it is impossible to distinguish which storm environment among the five points was suitable for supercell maintenance based merely on low-level shear, SRH, or SCP values (Figs. 19d–f). Rather, the more detailed structure of sounding profile, especially at low levels, must be taken into account to give information about not only whether instability exists, but also how easily it can be released as shown here. From Figs. 18 and 19, for the present case, it can be identified that each point over the Taiwan Strait only experienced a certain period of time from the afternoon of 19 to early morning of 20 December 2002, perhaps no longer than 6–10 h, that was favorable for supercells. Given sufficient low-level shear (and SRH), the CAPE must exist and its release was achievable; that is, the surface cold air was still shallow enough (so that convection could occur above it), or a sufficient period of time had elapsed to allow the ocean fluxes to modify the near-surface air inside the marine PBL and reduce CIN to a level that could be overcome.

The previous analysis shows that in this case, in addition to the common ingredients of sufficient shear and instability, the evolution of model storms depends heavily on the detailed low-level vertical structure of storm environment, which varies horizontally and is linked to the rapid evolution of surface cold air. Over the Taiwan Strait, the environmental vertical profile at any given point and time was dictated by a number of factors, including the depth and advancing speed of the cold air, and the modification rate (in both temperature and moisture) from the underlying ocean. Thus, inaccurate representation or simulation in any of these factors or processes in the model from reality can alter the time window in the environment suitable for supercells and hamper storm evolution, thereby presenting serious challenges for a successful simulation that agrees with the observation in almost all aspects.

The averaged environmental wind near 600 hPa among points B through D in the model is from 237°, which is about 6° to the left of the observed wind at Shantou (from 243°; cf. Table 1). This only partially accounts for the difference in storm propagation direction. When Figs. 1 and 6 are compared in more detail, it is found that storm h1 travels more to the right, almost eastward, during 1900–2200 LST in run 2 when it develops into an isolated supercell (cf. Figs. 5i–l). However, the tracks of the weaker n1 and stronger c1 over the Taiwan Strait in run 2, as well as those in run 1 using a coarser grid, show virtually no difference. Therefore, it is unlikely that the propagation direction is dictated by the intensity of the storm in the model. It can only be hypothesized that the remaining part of the difference between the observed and modeled propagation direction is due to a deviation of low-level shear profile in the model from the real atmosphere, which was by no means adequately represented over the data-sparse, or even data-lacking, Taiwan Strait in this case. Thus, if detailed observations were available and assimilated into the model, it is quite likely that the overall results will further improve.

6. Conclusions

The present paper is a follow-up numerical study on the rare wintertime supercell thunderstorms that occurred during 19–20 December 2002 near Taiwan in a maritime subtropical environment, which were documented and studied by Wang et al. (2009). Using the NU CReSS model at grid spacing of 1.5 (run 1) and 0.5 km (run 2), both simulations successfully reproduced the three primary storms at correct time and location with multiple splits as observed, using real terrain and JMA (20 km) regional analyses as IC/LBC with no initial perturbation. However, when compared to observation of the actual event, the model storms travel about 15°–20° too much to the left and the southern storm diminishes too early over the Taiwan Strait. The model results are described and used to perform a vorticity budget analysis on a storm during its transition to an isolated supercell, and to diagnose on the mesoscale environment over the strait.

The vorticity budget analysis suggests that the tilting from horizontal into vertical vorticity is the major source of midlevel storm rotation, producing positive (negative) $\zeta$ at the right (left) flanks, while vertical stretching also strengthens existing $\zeta$ in the lower parts of updrafts at supercell stage, consistent with previous studies. As the source of $\zeta$ generation, the horizontal vorticity (i.e., vertical shear) usually points northward (westward) [i.e., enhancing westerly (southerly) shear] at the right (left) flanks, forming a counterclockwise pattern across the updrafts. This pattern is largely attributed
to the combined effect of baroclinic (solenoidal) generation, tilting of vertical vorticity onto, and rotation of vortex tubes in the $x$–$y$ plane. Except for the baroclinic generation, roles of the other two kinematic effects in mature storms have not been emphasized recently in the literature.

Over the Taiwan Strait, the strong surface north-easterly flow enhanced low-level vertical shear, and helped the storms evolve into isolated supercells once they moved offshore in the present case. Through the diagnosis on storm environment over the strait, it is shown that the evolution of model storms is quite sensitive to the detailed low-level structure of the post-frontal environment and the intensity of the storms themselves. While traveling, storms are sustained only when CAPE exists and can be released through forced uplift at the gust front, in addition to the common conditions of sufficient vertical shear and instability. The horizontal heterogeneity in mesoscale conditions over the strait caused the storms to behave differently, and potentially explains the differences between modeled and observed storms. Furthermore, the fact that these conditions are dictated by a number of factors not resolved by the observation (and inadequately represented in gridded data) suggests that a successful simulation of supercell storms that agrees with observations in almost every aspect will likely remain difficult in the near future, at least in cases over data-lacking oceans similar to the present one.

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