Impact of the Cold Pool on Mesoscale Convective System–Produced Extreme Rainfall over Southeastern South Korea: 7 July 2009

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ABSTRACT

In this study, an extreme rainfall-producing quasi-stationary mesoscale convective system (MCS) associated with the Changma front in southeastern South Korea is investigated using numerical simulations and sensitivity tests. A record-breaking rainfall amount was recorded in response to repeated initiation of new cells (i.e., back-building) over the same area for several hours. The aim of this study is to realistically simulate and analyze this extreme rainfall event to better understand an impact of the cold pool that leads to the quasi-stationary MCS over southeastern South Korea by using a convection-allowing-resolution (2 km) non-hydrostatic atmospheric model.

The control experiment (CNTL) was successfully performed, yielding the quasi-stationary, back-building MCS at approximately the correct location and time. In the CNTL run, diabatic cooling due to evaporation of raindrops was responsible for the formation of the cold pool. The development of the cold pool was responsible for the deceleration of the propagating convective line, which played a role in the stalling of the MCS over southeastern South Korea. Moreover, new convective cells were repeatedly initiated in the region where an oncoming warm inflow met the leading edge of the cold pool and was uplifted. In an experiment without evaporative cooling (NOEVA), the simulated precipitation pattern was shifted to the northeast because the MCS became nonstationary without the cold pool. The cold pool had an essential role in the stationarity of the MCS, which resulted in extreme rainfall over the Busan metropolitan area.

1. Introduction

Extreme rainfall is arguably the most widespread weather-related hazard worldwide. Most extreme rainfall events are the result of several convective cells moving successively over the same area while the overall convective system is nearly stationary (Chappell 1986; Doswell et al. 1996). This process often produces extreme local rainfall associated with “back-building/quasi-stationary” mesoscale convective systems (MCSs), which consist of a line or cluster of deep convection and occur when convective cells repeatedly form upstream of their predecessors (Schumacher and Johnson 2005). For the continuous development of upstream convection in a specific area, a continuous supply of instability and moisture is first required, and a triggering mechanism is then needed to produce a relatively small area of deep convection. Factors that can contribute to the convective triggering and/or maintenance of the convection over the a given region include frontal lifting, orographic barriers, vertical wind shear, and outflow boundaries or density currents (e.g., Weisman and Klemp 1982; Hane et al. 1987; Corfidi et al. 1996; Corfidi 2003; Kirshbaum et al. 2007; Schumacher 2009; Houston and Wilhelmson 2011; Coniglio et al. 2012).
Hane et al. (1987) reported that cooling owing to evaporation and melting within an MCS leads to a near-surface cold pool. An outflow boundary exists at the interface between this cold pool and the ambient air. This evaporative cooling acts to increase the air density within a cold pool and consequently strengthens the downdraft, which is coupled with the low-level inflow (Corfidi 2003). New convective cells develop along the leading edge of the cold pool, and cold pools are known to be the primary mechanism for the maintenance of convective lines in two- and three-dimensional numerical simulations (Rotunno et al. 1988; Fovell and Tan 1998; Lin et al. 1998).

In East Asia, it is generally recognized that cold pools are rarely observed in the Changma (mei-yu or baiu) frontal region in which the lowest layer of the atmosphere is very moist. Recent studies have shown the importance of cold pools near frontal precipitation during the East Asian monsoon period. For example, Ishihara et al. (1995) showed that a baiu frontal rainband was maintained for 7 h by a self-sustaining mechanism due to cold pool formation. Davis and Lee (2012) and Xu et al. (2012) also found that the cold pool in the lowest layer (below 500 m) interacted with oncoming warm and moist air, which repeatedly produced heavy rainfall over southwestern Taiwan. In a numerical study, Nagata and Ogura (1991) demonstrated cold pool formation in a stratiform precipitation region, and showed that the cold pool formation had a local effect of enhancing the frontogenesis along the baiu front. Zhang and Zhang (2012) reported the occurrence of convective initiation as southwesterly monsoonal air lifted above the cold pool. Previous studies suggest that cold pools are not uncommon in frontal precipitation systems, but their impacts can be highly variable in response to the terms of maintaining a frontal precipitation system (Kato 1998; Kato and Goda 2001). However, previous studies have been unable to fully investigate the impact of the cold pool on the frontal precipitation.

On 7 July 2009, an extreme rainfall-producing quasi-stationary MCS developed along the Changma...
front, producing 310 mm of rain in less than 12 h over Busan metropolitan area (see Fig. 1 for location) in southeastern South Korea. The rainfall broke the maximum daily rainfall record at Busan station in July and is ranked as the second-highest all-time daily rainfall total at the station since 1905 (Jeong et al. 2016). This study focuses on the stationary stage of the convective system, with the goal of identifying the mechanisms that result in such heavy precipitating systems becoming stationary over a given location. In particular, we use a high-resolution (2 km) non-hydrostatic atmospheric model to simulate the unusual 7 July 2009 high-rainfall case in the Busan metropolitan area.

The remainder of this paper is organized as follows. In section 2, we describe the data and methodology, and section 3 provides a brief overview of this extreme case with a synoptic environment. Section 4 describes the numerical model and the design of the sensitivity study, and section 5 presents the model results. Finally, the summary and conclusions are given in section 6.

2. Data and methodology

The National Centers for Environmental Prediction (NCEP) gridded final (FNL) analyses, produced by the Global Forecast System (GFS) every 6 h at 0300, 0900, 1500, and 2100 LST (LST = UTC + 9 h) with a horizontal resolution of $1^\circ \times 1^\circ$ and 26 vertical levels (from 1000 to 10 hPa) on 7 July 2009, are used to describe the synoptic conditions. Data from 11 operational S-band
(10 cm) Doppler radars operated by the Korea Meteorological Administration (KMA) are used to examine the evolution and structure of the quasi-stationary MCS over South Korea at 10-min intervals. The Doppler radar data are composited into a reflectivity mosaic that is then interpolated to horizontal Cartesian coordinates with a 1-km grid spacing, covering the whole Korean Peninsula. The reflectivity mosaic used for this study was only extracted at 1.5-km height due to excessive noise in the radar measurements at lower layers. Data derived from 61 rain gauges that are part of the automatic weather system (AWS) operated by the KMA are used to examine the rainfall amount and rate. The rain gauge data are updated at 1-min intervals. The observational data mentioned above are compared with model results in order to validate the simulations.

3. Synoptic and case overview

a. Synoptic environment

The daily mean NCEP FNL gridded analysis of geopotential height and horizontal winds on 7 July 2009 is presented in Fig. 2. The Changma front (dashed line in Fig. 2a) extended from eastern China to southern South Korea, and the heavy precipitation occurred to the north of the front. The geopotential height gradient suggests a strong low-level flow over the Korean Peninsula. A southerly to southwesterly flow prevailed south of the front, corresponding to the meso-α-scale cyclone in the lower troposphere, with stronger winds (up to approximately 10 m s⁻¹) over the East China Sea. East of the front, warm air transported along the northwestern periphery of the subtropical high (Fig. 2a). The region of strong winds at 850 hPa remained stationary, and the southwesterly low-level jet (LLJ; ≥12 m s⁻¹) appeared ahead of the front over southern South Korea (Fig. 2b). The LLJ continued to provide a source of low-level moisture toward southern South Korea. Frontal precipitation systems over East Asia are frequently observed with low-tropospheric warming and moistening (e.g., Akiyama 1973; Ninomiya 1980; Ninomiya and Muraki 1986; Jeong et al. 2012). Low-tropospheric warming and moistening enhanced by LLJ can also often present in the environments of quasi-stationary MCSs over the United States (e.g., Maddox 1983; Moore et al. 2003; Schumacher and Johnson 2005). A trough was located just off the west coast of South Korea at 500 hPa (Fig. 2c). At 300 hPa, the center of South Korea was under a diffuent upper-flow regime along the southern edge of a westerly upper-level jet (ULJ; Fig. 2d). This coupling of the LLJ and ULJ is associated with upward motion (Uccellini and Johnson 1979), which in this case favored the development of a broad area of convection over Korea.

b. Case overview

Figure 3 presents the reflectivity mosaic (dBZ) over South Korea for 0600–1000 LST 7 July 2009. During this period, a widespread area of convection occurred to the north of the Changma front over the Korean Peninsula. The convective system was primarily made
up of three convective cells (indicated by numbers in Fig. 3 and referring to the northern, western, and eastern convective cells, respectively). The convective cells were different in their organization and propagation. The northern cell C1 moved northeastward and gradually dissipated due to the moisture supply being reduced after it moved inland at 1000 LST. The western cell C2 migrated onshore over southwestern South Korea at 0600 LST, after which it moved southeastward to the coast of southern South Korea. At 0600 LST, the eastern cell C3 moved slightly northeastward, and at 0800 LST it gradually organized into a squall line along the coastline of southeastern South Korea. This resulted in a well-defined convective line oriented northeast–southwest (NE–SW) and located from 35.0°N, 129.0°E to 36°N, 130.5°E.

The Busan (PSN) radar images further illustrate C3 during its stationary stage near the coast of southeastern South Korea at 10-min intervals from 0700 to 0730 LST 7 July 2009 (Fig. 4). Here, the stationary stage is defined as the period when the convective line is remaining nearly stationary. During the stationary stage, the convective system formed north of the warm front over the coast of southeastern South Korea. The convective cells gradually intensified while moving slowly northeastward. The convective cells became organized and moved parallel to the convective line (i.e., training line). The convective line was associated with a region of adjoining stratiform precipitation in addition to the relatively small leading intense convective region. The leading edge of the convective line showed reflectivity in excess of 40 dBZ and attained a length of roughly 180 km and a width of 30 km. The environment wind showed veering winds at low levels, and the midlevel (700–500 hPa) wind shear had a large component parallel to the convective line (Fig. 5). These structures identified by Schumacher and Johnson (2005; Fig. 3) as trailing-line/adjoining-stratiform (TL/AS). The convective line was maintained by continuing upstream convective redevelopment [the back-building (BB) process; e.g., Bluestein and Jain (1985); Schumacher and Johnson (2005)]. During this process, new cells continued to develop at the southwestern end of the line, which remained in the same position during the stationary stage (Fig. 4). The organizational structure of this system resembles a blend between the TL/AS and BB types of the convective system. The surface winds, cold pool outline, and radar reflectivity at 2 km were superimposed in Fig. 6. A cold pool (temperature depression of 4 K) developed over the stratiform precipitation region behind the convective line. The role of the cold pool is discussed in detail in section 5. The primary results in the synoptic environment and features of quasi-stationary MCS are illustrated schematically in Fig. 7.

The overview presented above shows that the continuous formation of convective cells upstream of the

![Fig. 4. Series of radar reflectivity (dBZ, color shading) at a height of 2 km observed by PSN radar at 10-min interval from 0700 to 0730 LST 7 Jul 2009. Selected convective cells are labeled as A–C.](image-url)
convective line and the dissipation of older cells downstream caused the MCS to become stationary. This process maintains convection over the same region, but it requires a triggering mechanism. To understand this process, we performed 3D cloud-resolving model (CRM) simulations at convection-allowing-resolution with full physics, as well as a sensitivity test.

4. Numerical model configuration

a. Description of the CReSS model

The Cloud-Resolving Storm Simulator (CReSS) model of Nagoya University, Japan (Tsuboki and Sakakibara 2002), was used to simulate the development and evolution of the quasi-stationary MCS in the present case. The CReSS model has successfully simulated various types of atmospheric phenomena, including heavy rainfall (Wang et al. 2005), snowstorms (Liu et al. 2004; Maesaka et al. 2006), and supercell-type storms (Shimizu et al. 2008). Although the CReSS model has been used successfully for numerous types of precipitation systems, it has not previously been used to simulate the phenomena of interest in this study (i.e., an extreme rainfall-producing quasi-stationary MCS).

The CReSS model is nonhydrostatic and compressible, and is designed to simulate mesoscale and rainfall systems realistically at high resolution using explicit cloud microphysics without cumulus parameterization. The microphysics parameterization is 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).

This model employs a height-based terrain-following vertical coordinate, with prognostic equations for the three velocity components \(u, v, w\), perturbation
potential temperature $\theta'$, and pressure $p$. A total of six species are included in the microphysics scheme via the mixing ratios of vapor $q_v$, cloud water $q_c$, rain $q_r$, cloud ice $q_i$, snow $q_s$, and graupel $q_g$, and the number concentrations of cloud ice $N_i$, snow $N_s$, and graupel $N_g$. The microphysical processes of nucleation (condensation), sublimation, evaporation, deposition, freezing, melting, falling, conversion, collection, aggregation, and liquid shedding are included.

Subgrid-scale turbulent mixing is parameterized using a 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 radiative processes at the surface are also included with a one-dimensional heat diffusion model for ground temperature prediction (Kondo 1976; Louis et al. 1981; Segami et al. 1989).

In the CReSS model, the Arakawa C staggered and Lorenz grids are used in the horizontal and vertical grid directions, respectively, with no nesting. A time-splitting scheme (Klemp and Wilhelmson 1978) is adopted to separately integrate fast and slow modes. The filtered leapfrog method (Asselin 1972) is used for integration at large time steps (Δτ), while the implicit Crank–Nicolson scheme is used in the vertical at small time steps (Δτ).

b. Model output and experiment design

A control experiment (CNTL) with a full physical model was performed with a horizontal grid spacing of 2 km (Fig. 1) and 70 vertically stretched layers with the model top at 20.8 km (the minimum Δz at the surface is 100 m). The Japan Meteorological Agency mesoscale model (JMA-MSM) at a horizontal resolution of 5 km and at 16 levels (Saito et al. 2006) was used for the model initial and boundary conditions. The start time of the simulation is 1800 LST 6 July 2009, with an integration time of 30 h and an output frequency of 10 min.

The sensitivity experiment for no evaporative cooling (NOEVA) is designed to examine the effect of evaporative cooling of rainwater on an extreme rainfall event. The simulation for microphysical sensitivity is configured within the CNTL run (i.e., the cooling associated with the evaporation of rainwater is turned off, while cooling associated with the evaporation of cloud water remains unchanged). The microphysical sensitivity is only configured during the extreme rainfall over the Busan area, which occurred for 12 h starting at 0550 LST 7 July 2009. Therefore, the impact of evaporative cooling is applied in the NOEVA run from 0600 LST.

The other features of the numerical model are summarized in Table 1.

5. Numerical model results

a. Validation of CNTL and comparison of NOEVA experiment

The simulated rainfall distributions on 7 July 2009 from the CNTL and NOEVA runs are first compared

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**Table 1. A summary of the CReSS model and the design for the two experiments. Configuration of basic setup, physics, and numerical methods were used.**

<table>
<thead>
<tr>
<th>Expts</th>
<th>CNTL</th>
<th>NOEVA</th>
</tr>
</thead>
<tbody>
<tr>
<td>Description</td>
<td>Full physics</td>
<td>No cooling from evaporation of rainwater</td>
</tr>
<tr>
<td>Basic setup</td>
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<td>Lambert conformal, center at 128.5°E, secant at 10°/40°N</td>
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<tr>
<td>Projection</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Grid spacing (km × km × km)</td>
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<td>2 × 2 × 0.1-0.3</td>
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<tr>
<td>Grid dimension (x, y, z)</td>
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<td>500 × 560 × 70</td>
</tr>
<tr>
<td>Min Δz (m)</td>
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<td>100</td>
</tr>
<tr>
<td>Topography and SST</td>
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<td>Real at (1/120)° and daily SST with 0.25° resolution</td>
</tr>
<tr>
<td>IC/BCs</td>
<td>JMA- MSM every 6 h</td>
<td>JMA- MSM every 6 h</td>
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<tr>
<td>Initial time</td>
<td>1800 LST 6 Jul 2009</td>
<td>As in CNTL, but sensitivity starting at 0550 LST 7 Jul 2009</td>
</tr>
<tr>
<td>Integration length</td>
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<td>24 h</td>
</tr>
<tr>
<td>Output frequency</td>
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<td>10 min</td>
</tr>
<tr>
<td>Model physics</td>
<td></td>
<td></td>
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<td>Both fourth order in horizontal and vertical</td>
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<td>Cloud microphysics</td>
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<td>Bulk cold rain scheme (mixed phase with six species)</td>
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<td>Cumulus parameterization</td>
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<td>PBL parameterization</td>
<td>1.5-order closure with TKE prediction</td>
<td>1.5-order closure with TKE prediction</td>
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<td>Energy and momentum fluxes, shortwave and longwave radiation</td>
</tr>
<tr>
<td>Soil model</td>
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<td>41 levels, every 5 m to 2 m deep</td>
</tr>
<tr>
<td>Numerical methods</td>
<td></td>
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<td>Δτ = 4 s, Δτ = 2 s</td>
<td>Δτ = 4 s, Δτ = 2 s</td>
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<tr>
<td>Integration method</td>
<td>Filtered leapfrog for Δτ (HE-VE), leapfrog and Crank–Nicolson for Δτ (HE-VI)</td>
<td>Filtered leapfrog for Δτ (HE-VE), leapfrog and Crank–Nicolson for Δτ (HE-VI)</td>
</tr>
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</table>
with observations. Figure 8 compares the 24-h accumulated precipitation obtained from the observations, CNTL, and NOEVA runs for 6–30 h of simulation time. Comparing Figs. 8a and 8b, we see that the observed rainfall intensity and distribution (Fig. 8a) are well captured by the CNTL simulation [i.e., the strong narrow rainband of substantial precipitation (≥150 mm) along southern South Korea is reasonably well simulated]. The simulated rainfall is concentrated over southern South Korea, and localized heavy rainfall in southeastern South Korea is well represented. The extreme rainfall amount observed near the coast of Busan at 35.1°N, 129.0°E (marked by a cross in Fig. 8a) is 310 mm. The location of the simulated precipitation maxima near and just offshore the coast of Busan is 35.0°N, 129.0°E (cross in Fig. 8b), only 10 km from the observed precipitation maxima, with a simulated rainfall maxima of 322 mm.

To further demonstrate the CNTL run’s capability in reproducing the timing and amount of extreme rainfall, Fig. 9 compares the evolution of the hourly precipitation in the observations and the CNTL experiment at the location of the precipitation maxima, and shows that the general pattern of hourly rainfall is fairly well captured. In particular, the model
produces a peak rainfall rate of 68.3 mm h\(^{-1}\), which corresponds closely to observations from the convective triggering phase during 0700–1000 LST 7 July 2009. The simulated peak rainfall is seen to result from continuous back-building of the convective line, which produced a quasi-stationary MCS. These results show that the CNTL experiment is capable of simulating realistic locally observed precipitation maxima during the stationary stage of the MCS.

In this case, operational forecast models were unable to predict the magnitude of the extreme rainfall (not shown). The KMA use Regional Data Assimilation and Prediction System (RDAPS) and KMA Weather Research and Forecasting (KWRF) Model for short-range regional forecasting. The operational forecast models underpredicted rainfall amounts by approximately 60% at a 30-km horizontal grid spacing (RDAPS) and 35% at a 10-km grid spacing (KWRF), which highlights the difficulty in resolving localized deep convection at relatively coarse grid spacing (Randall et al. 2003a; Arakawa 2004). To improve forecasts for these types of extreme rainfall events, a higher-resolution model configured as a convection-allowing resolution is required (Davis and Trier 2002; Schwartz et al. 2009; Kain et al. 2013; Schwartz et al. 2015; Wang et al. 2016).

Figure 8c presents the simulated 24-h accumulated rainfall for the NOEVA run and the differences between the two simulations. The simulated precipitation pattern was shifted northeastward, with a maximum rainfall amount onshore in Busan less than 190 mm. The simulated rainfall is mainly concentrated over the inland Busan region due to the non-stationarity of the MCS in the NOEVA run. As the simulation with evaporation turned off in the full

Fig. 9. Time series of hourly rainfall (mm h\(^{-1}\)) at the cross symbol in Fig. 8 comparing the observation (35.1°N, 129.0°E) and CNTL simulation (35.0°N, 129.0°E).

Fig. 10. Time–latitude Hovmöller diagrams (transect A–A’ in Fig. 8d) of model-simulated reflectivity for (a) CNTL and (b) NOEVA runs at 10-min intervals from 0700 to 1200 LST 7 Jul 2009. The black dashed line shows the backward propagation. The white dashed arrows indicate the northward phase propagation.
model integration, the simulated rainfall moved farther northward (not shown). Figure 8d also shows the differences in accumulated rainfall amount between the two simulations. The CNTL run shows a narrow rainband of substantial precipitation along the coast of southern South Korea, while the NOEVA run shows the simulated rainfall concentrated farther inland over southern South Korea. The difference in maximum rainfall amount between the two simulations is about 30 mm.

The main difference between the CNTL and NOEVA runs was the stationarity of the MCS over the Busan region. Figure 10 compares the motion and propagation of precipitation by using time–latitude sections (Hovmöller diagrams) between two simulations. The simulated reflectivity was used to create Hovmöller diagrams along precipitation maxima (33.75°–37.3°N, 129°E, A–A’ in Fig. 8d) during the stationary period of 0700–1200 LST 7 July 2009. For CNTL (Fig. 10a), continuous precipitation only occurs near the coast of Busan (34.75°N, black dashed line) from 0730 to 1100 LST with northward propagation (dashed arrow). The average phase speed of the northward-propagating rain belt of the CNTL run was about 2.3 m s⁻¹. Precipitation for the CNTL run was different than for the NOEVA run in terms of backward propagation (Fig. 10b). The NOEVA run has only moved toward a northward-propagating
The modeled rainfall pattern from the CNTL run suggests that cold pool dynamics influence the backward propagation.

b. CNTL experiment results

Figure 11 presents the evolution of the simulated quasi-linear MCS during its stationary stage until 0900 LST 7 July 2009. The stationary stage of this event reflects the convective triggering in the vicinity of Busan, in close agreement with observations (Figs. 4 and 6). By 0600 LST, the primary area of convection had become organized into a line, with convective cells on the leading edge (Fig. 6a). Deep convection with vertical velocities of up to 3 m s\(^{-1}\) developed upstream on the southwest flank of the convective line. During the next 2–3 h, these cells gradually intensified while slowly moving northeastward. They became organized into a linear convective system as a result of the convective trigger. The overall convective line showed slow system motion for several hours, resulting in localized heavy rainfall.

The continuous convective initiation primarily resulted from two features in the simulation. First, the perturbation potential temperature \(\theta'\) showed a warm perturbation ahead of the convective line due to adiabatic advection (cf. Figs. 11 and 12). The
perturbation potential temperature is the deviation from the initial state (i.e., basic state) of potential temperature. During the stationary stage, warm perturbations (≥3 K) gradually intensified upstream of the convective line, over the ocean (Fig. 12). Warm perturbation was due to a southwesterly monsoonal flow (i.e., Jeong et al. 2014). Strong southwesterly low-level flow (≥12 m s⁻¹) was dominant ahead of the convective line in the simulation (Fig. 11). Figure 13 depicts the presence of moisture advection, illustrated by simulated vertically integrated horizontal water vapor fluxes between 1000 and 300 hPa [hereafter integrated vapor transport (IVT); see Neiman et al. (2008); Moore et al. (2012)]. The widespread moisture advection of IVT values of 200–800 kg m⁻² s⁻¹ appeared over southern South Korea at 0600 LST 7 July 2009 (Fig. 13a). By 0900 LST, IVT intensified over the area south of Busan to approximately 1000 kg m⁻² s⁻¹ (Fig. 13b). The sounding within a representative thermodynamic environment of extreme rainfall is shown in Fig. 14. This sounding at 34.5°N, 128.5°E (marked by an asterisk in Fig. 13b) about 113 km upstream from the quasi-stationary MCS at 0900 LST 7 July 2009, indicates that it is conditionally unstable with a convective available potential energy (CAPE) of about 541 J kg⁻¹ and a negligible convective inhibition (CIN) of 6 J kg⁻¹. The lifting condensation level (LCL) showed quite low at 988 hPa, and the level of free convection (LFC) was also low at 943 hPa. This environment indicates that the unstable air mass over the ocean provided a continuing moisture source, which induced conditional instability at upstream environment of the quasi-stationary MCS.

Second, during the organization of the simulated convective cells into a quasi-stationary MCS, the perturbation potential temperature showed a cold perturbation behind the convective line (Fig. 12). This convective line produced a cold pool, which also showed in the surface observations during the stationary stage of the MCS (Fig. 6). In CNTL, the cold pools (i.e., isopleth of θ' = 2 K) due to diabatic/latent cooling from rainfall continued to intensify and expand. After the cold pools reached over the coastline, the motion was nearly stationary. By this time, intense convection at low levels is found along the leading edges of the spreading cold pool, which in turn helps to intensify the low-level lifting (Figs. 11 and 12).

To further investigate the MCS during its stationary stage, we examine a vertical cross section across the convective line (35.45°N, 128.5°E to 34.2°N, 128.5°E).
129.6°E, line A–A’ in Fig. 12). Prior to the stationary stage, at 0400–0500 LST 7 July 2009, warm perturbations are seen in the lower troposphere (from the surface to 3-km height). Relatively cold perturbations were present in the midtroposphere, extending down to the surface in the southeast due to weak precipitation (Figs. 15a,b).

At 0600 LST, wind speeds in the upstream flow above 2 km are between 10 and 20 m s⁻¹. The opposing downstream flow below 2 km was weak in comparison (≤10 m s⁻¹). Where the upstream and downstream flows met (see the short vertical arrow), a convective updraft (vertical velocity of >1 m s⁻¹, \( X' = 128.8^\circ \text{E} \)) driven by thermal energy and latent heat of convection is released (Fig. 15c). North of the convective line, at \( X' = 128.6^\circ \text{E} \), cloud-scale weak downward motion (≤1 m s⁻¹) is seen due to falling precipitation.

One hour later, at 0700 LST, the upstream flow strengthened to 15–23 m s⁻¹ and the maximum \( \theta' \) range increased to 2–7 K (Fig. 15d). Surrounding air cooled due to the evaporation of precipitation, and then accelerated downward due to its acquired negative buoyancy (i.e., Emanuel 1981; Wakimoto 1982). The spreading cold pool associated with the downdraft eventually cut off the potentially unstable air behind the main updraft (at \( X' = 128.65^\circ \text{E} \)), and new convection formed repeatedly at the same location (Figs. 15c–f). Evaporative cooling associated with the convective system produced a density current, which then triggered new convective updrafts at \( X' = 128.85^\circ \text{E} \). As described in many previous studies (e.g., Lafore and Moncrieff 1989; Fovell and Tan 1998; Lin et al. 1998), long-lived squall lines and MCS consist of several convective updrafts that initiated at the leading edge of the cold pool. In the present case,
the convectively generated cold outflow acted like a gravity current (Simpson 1997) and progressively pushed southeastward toward the upstream oncoming (southwesterly) flow. This process is thought to be an important factor in allowing the convective system to remain stationary. The impacts of evaporative cooling of rainfall will be further examined when comparing the NOEVA run.

At 0800 LST, the convection ascended over subsequent convective development along the edge of the

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**Fig. 15.** Vertical cross section (transect A–A’ in Fig. 12d) of model-simulated $\theta’$ (K, color shading), vertical velocity (m s$^{-1}$, dotted for negative values), and wind vectors (m s$^{-1}$) along the section plane in CNTL run at 1-h intervals from (a) 0400 to (f) 0900 LST 7 Jul 2009. The locations of onshore/offshore at low-level flow are marked by the short vertical arrows. The topography is masked in black in bottom panels.
cold pool, due to lifting caused by warm diabatic advection (Fig. 15e). The convection collided with warmer, moister, less dense air, and forced that air to rise up over the cold pool. The convective cold pool appeared to weaken and the low-level lifting also became weaker after that time (Fig. 15f). The updraft/downdraft couplet in the leading edge of cold pool continued to weaken until it moved northeastward with the cold pool. Similar multicell behaviors associated with extreme rainfall events have also been observed or simulated in previous studies (e.g., Schumacher 2009; Zhang and Zhang 2012).

c. NOEVA experiment results

We now discuss the results of the NOEVA run, in which evaporation cooling of rainwater was disabled in the CNTL run at 0550 LST 7 July 2009 following the formation of convective cells near Busan. This was done to focus on the stationary stage of the convective systems leading to the extreme rainfall-producing quasi-stationary MCS over Busan. In this subsection, the NOEVA results are compared directly with those from the CNTL run (which are shown in Figs. 11, 12, and 15).

At 0600 LST, the convection became organized upstream of the convective cells, and deep convection on the southwest flank of the system is shown in Fig. 16a, similar to that in the CNTL run. The impact of evaporative cooling is apparent in the NOEVA run at 0600 LST. At 0700 LST, the convective region associated with weak precipitation extended northward and the convective cells strengthened compared with...
those in the CNTL run (Fig. 16b). However, the low-level upward motion at the leading edge of convective line was weaker in the NOEVA experiment. By comparing Figs. 11c and 16c at 0800 LST, we see that without the formation of the cold pool, the convective line was not stationary at a single location, and new cells were not triggered upstream. In other words, the convective system migrated to the northeast, farther inland (Fig. 16c). This suggests that evaporative cooling associated with precipitation caused upstream deceleration (of approximately 3.3 m s$^{-1}$) in the propagation of the convective line. This also highlights the important role of evaporative cooling by rain produced in the convective line, in terms of initiating upstream convection. The convective line continued to move northeastward due to a strong oncoming southwesterly flow (Fig. 16d).

The distribution of $\theta'$ in the NOEVA run also reveals some detailed features of the convective system. The cold perturbation was significantly weaker than in the CNTL run, and the warm perturbation was displaced behind the convective line (cf. Figs. 12b and 17b). In particular, a strong warm perturbation (35.2°N, 128.8°E) appeared within the convective line. After 2 h, the warm perturbation increased and extended toward the northeast.

Figure 18 shows a vertical cross section through the convective line in the NOEVA run (line B–B'; 35.45°N, 128°E to 34.2°N, 129.6°E). At 0600 LST, the vertical cross section from the NOEVA run is nearly...
indistinguishable from that of the CNTL run, and the cold pool remained in the convective region (Fig. 18a). At 0700 LST, the cold perturbations at low levels had been displaced by a warm perturbation (Fig. 18b). The updraft was slightly moved toward the northwest (vertical velocity of $>1 \text{ m s}^{-1}$, $X' = 128.8^\circ \text{E}$). By 0900 LST, the warm perturbations at low levels below 1 km increased with intensified oncoming flow (Figs. 18c,d).

Figure 19 summarizes and compares the evolution of each simulation in terms of vertically averaged low-level (below 1 km) values of $q_y$, $u_0$, and simulated rainfall along the cross section. For the NOEVA run, the mean $\theta$ is about 3 K warmer upstream due to the effect of evaporative cooling by precipitation (Fig. 19a). The mixing ratio of water vapor increased due to warm and moist air from oncoming flow. The NOEVA run was also convectively updraft, which enhanced the cloud-scale upward motion from the moist air mass. Crook and Moncrieff (1988) and Schumacher (2009) removed the evaporation of rain in their simulations, and found that in an environment with large-scale lifting (i.e., frontal uplift), a cold pool (produced by evaporative cooling) was not necessary to maintain a convective system. A similar simulation (which did not produce a surface cold pool) by Peters and Schumacher (2016) also showed that the cold pool was not essential for organizing and maintaining the convection. In this study, the cold pool played an essential role in the stalling of the MCS over southeastern South Korea. The synoptic-scale conditions were favorable for organized convection despite the lack of the cold pool, but the cold pool was instrumental in determining where the heaviest rains fell.

6. Summary and conclusions

In this study, an extreme rainfall event over southeastern South Korea was studied to investigate
the stationary stage of a quasi-stationary MCS over the Busan metropolitan area. High-resolution simulations of this extreme event using the CReSS model enabled a detailed investigation of the finescale processes relative to the system-scale dynamics and microphysics of the convective system. The mechanisms leading to the stationary stage of the MCS were highlighted through a sensitivity study as well. The results of this study are summarized below and illustrated schematically in Fig. 20.

Using the CReSS model at a horizontal grid spacing of 2 km (with 70 vertical levels), the observed extreme rainfall during the stationary stage of the MCS by the simulation is especially notable. The convective line was maintained by continuing upstream convective redevelopment near the coast of southeastern South Korea. The simulated convective line reproduced the result of the convective trigger in the control experiment (CNTL) fairly well. The initiation of a continuous upstream convective resulted from mesoscale and cloud-scale factors. The simulated mesoscale environment in the CNTL run showed strong southwesterly low-level flow strengthening ahead of the convective line. Warming and moistening upstream flows induced a warm perturbation through the adiabatic advection of warm air, which could be lifted above the cold pool. Meanwhile, a cold perturbation was located behind the convective line, under the precipitation system. Diabatic cooling due to the evaporation of raindrops was responsible for the formation of the cold pool. The cold pool acted like a density current and pushed against the oncoming flow from the south. This suggests that evaporative cooling associated with precipitation can slow the upstream propagation of a convective line. Thus, evaporative cooling was responsible for pushing the convective line southward and progressively toward an oncoming flow from the edge of the cold pool downstream. The extended cold pool eventually cut off the internal circulation (Fovell and Tan 1998) and acted as a mechanism for advection (Lin et al. 1998). The repeated development of new cells occurred in a similar location at the leading edge of the cold pool. Cooling induced by the evaporation of rain, therefore, played a role in the formation of the convective line itself by initiating upstream convection. The cold pool is hardly observed in the frontal region because the lower layer is a nearly moist environment. It is difficult to predict the location and strength of cold pools. To improve predictability of extreme rainfall events, one should be aware of the characteristics of the low-level cold pool development.

In an experiment without evaporative cooling (NOEVA), the cold perturbation was significantly weakened and was displaced by a warm perturbation behind the convective line. The simulated precipitation pattern was shifted northeast over the Busan area. The simulated rainfall amount was mainly concentrated over the inland Busan region due to the nonstationarity of the convective system. Thus, the cold pool played a role in the extreme rainfall-producing quasi-stationary MCS over the Busan metropolitan area.

![Fig. 19. Vertically averaged (a) $\theta$ (K, black color, scale on the left) and mixing ratio of water vapor (g kg$^{-1}$, gray color, scale on the right), onshore/offshore flow boundary (dotted line) below 1 km, and (b) 24-h accumulated rainfall (mm) comparing between the CNTL (solid line) and NOEVA (dashed line) runs. The topography is masked in black in the bottom panels.](image-url)
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FIG. 20. Schematic illustration of the cold pool effects, development of convection, and rainfall distribution near Busan as simulated by the CReSS model using (a) the CNTL run and (b) the NOEVA run. The crimson shading shows the regions of warm temperature, while the deep blue shaded region represents cold temperatures. Cross circles (○) denote the wind direction into the page. Shading at the bottom indicates the amount and location of rainfall, with a darker shade signifying a greater amount.
Schwartz, C. S., and Coauthors, 2009: Next-day convection-


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