The structure and formation mechanism of a supercell-like storm in a moist environment below a melting layer were investigated using dual-Doppler radar analysis and a cloud-resolving storm simulator (CReSS). The supercell-like storm developed over the Kanto Plain, Japan, on 24 May 2000. The environment of the supercell-like storm possessed large convective available potential energy (1000 J kg\(^{-1}\)), strong vertical wind shear (4.2 \(\times 10^{−3}\) s\(^{-1}\) between the surface and 5 km above sea level), and a moist layer (the relative humidity was 60%–90% below a melting layer at 3 km in height). The dual-Doppler radar analysis with a variational method revealed that the supercell-like storm had similar structures to those of a typical supercell in a dry environment below a melting layer, such as that in the Great Plains in the United States. The structures included a hook echo, an overhanging echo structure, and a strong updraft with strong vertical vorticity. However, some of the characteristics of the supercell-like storm differed from those of a typical supercell. For example, a weak downdraft, a weak outflow, a weak inflow, and a short time maintenance of a single cyclonically rotating updraft (about 30 min) were noted. Dual-Doppler radar analysis revealed that the convergence between the weak outflow and the weak inflow kept its location just under the updraft for about 30 min; in other words, the strength of the outflow balanced the strength of the inflow. The observed features were simulated well using CReSS, and the thermodynamical features of the formation mechanism were revealed. The weak downdraft with a small evaporative cooling rate was simulated in a moist layer below the melting layer at 3 km in height. The small evaporation cooling was a major cause of the weak downdraft and the weak outflow. Because the outflow was weak and did not cut off the initial updraft, the weak inflow was able to keep supplying warm air to the initial updraft for about 30 min. Therefore, the present supercell-like storm could form as a result of the balance of the strengths of the weak inflow and the weak outflow in a moist environment.

1. Introduction

Most of the work relating to the structure and formation mechanism of a supercell thunderstorm is based on observations in a dry environment, such as that found in the U.S. Great Plains. The observations of a supercell in a moist environment, such as a subtropical humid region, would afford the opportunity to expand our understanding of the formation mechanism of a supercell thunderstorm. In past studies, a suitable environment for determining the formation mechanism of a supercell has been well established. Supercells in a dry environment often form in environments possessing large instability and strong vertical wind shear (Weisman and Klemp 1982, 1984). In recent years, the focus has been placed on midtropospheric humidity, which is an environmental factor involved in the formation of a supercell (Gilmore and Wicker 1998). However, it is difficult to observe the impact of humidity on the formation of a supercell using observation data obtained over the Great Plains because the midlevel air below the melting layer over the Great Plains is usually dry (Carlson et al. 1983). Therefore, the observation and
Supercells have been generally defined as storms with significantly persistent spatial collocation between updraft centers and vorticity centers (Weisman and Klemp 1984). Many observation studies (Browning 1964; Marwitz 1972; Browning and Foote 1976; Klemp et al. 1981; Ray et al. 1981; Dowell and Bluestein 1997) have reported that supercells generally have characteristic features, such as a propensity for steadiness and longevity (storm lifetimes of 2 h or more), a principal intense and cyclonically rotating updraft, large hail, and a characteristic echo structure (hook echo and overhanging). In addition, many modeling studies have been used to investigate the formation mechanism of supercells and suitable environments for supercell formation.

Two mechanisms have been found to be necessary for the formation of supercells in a dry environment. First, a strong inflow prevents a cold outflow from moving ahead of a storm (Klemp 1987). In an ordinary multicell thunderstorm (Browning et al. 1976), the cold outflow from a dissipating cell often moves ahead of the storm and cuts off the supply of warm and moist air to the cell. Second, lifting pressure gradients induce an updraft growth almost directly above a surface gust front (Schlesinger 1980; Rotunno and Klemp 1982).

With regard to the environment around supercells, Weisman and Klemp (1982) revealed that supercells are well formed in environments of large instability and strong vertical wind shears. For distinguishing among storm types, they used the bulk Richardson number (BRN) as a nondimensional parameter that combines the effects of environmental instability and vertical wind shears. They concluded that supercells form and are stably maintained when $5 \leq \text{BRN} \leq 50$. Davies-Jones et al. (1990) proposed that the storm relative helicity (SRH) could be used as a parameter for estimating the potential for vertical vorticity derived from environmental vertical wind shears. They described that $\text{SRH} \geq 150 \text{ m}^2 \text{s}^{-2}$ is a necessary condition for the organization of a supercell.

Recently, in addition to the instability and vertical wind shears, midtropospheric humidity was proposed as an environmental parameter for the formation of a supercell in a numerical study by Gilmore and Wicker (1998). They numerically simulated a supercell with the vertical profiles of parameters representative of a pre-storm environment near Del City, Oklahoma, where a supercell formed on 20 May 1977 (Ray et al. 1981). They used a three-dimensional, nonhydrostatic cloud model, which is similar in design and construction to the model used by Wicker and Wilhelmson (1995). They conducted sensitivity experiments of relative humidity (RH) below the melting layer on supercell formation without changing the convective available potential energy (CAPE) and vertical wind shear. Gilmore and Wicker (1998) demonstrated that, when the initial air below the melting layer around a modeled storm was made drier, the downdraft was stronger, and the modeled storm lost its supercell characteristics. Therefore, they proposed that humidity is an important factor, as are instability and vertical wind shear, for the formation of supercells. However, their proposition needs to be verified with the observation of supercells in environments possessing different humidity profiles (e.g., moist environments).

In a moist environment, such as a subtropical humid region, some supercells have been observed (Niino et al. 1993; Suzuki et al. 2000). Using single-Doppler radar, Niino et al. (1993) provided one rare example of a storm similar to a typical supercell observed in Japan. Suzuki et al. (2000) investigated a minisupercell (McCaul 1987) associated with a typhoon observed in Japan by using single-Doppler radar as well. Suzuki et al. (2000) and Niino and Noda (2000) pointed out that a supercell in a moist environment had several features that were different from those observed during a typical supercell (Browning 1964; Browning et al. 1976) in a dry environment. However, the number of case studies of supercells in moist environments is inadequate to clarify the environmental factors that determine the formation mechanism of supercells in a moist environment.

To clarify the factors, observations and numerical simulations that are sophisticated enough to reveal the fine dynamic and thermodynamical structure of supercells in a moist environment need to be conducted. Fortunately, in Japan, finescale data can be obtained with the two C-band Doppler radars installed in 1999 at Narita and Haneda airports. A storm was observed by the two Doppler radars on 24 May 2000 over the Kanto Plain, Japan (Fig. 1). This storm had several characteristics of a typical supercell, such as a hook echo, an overhanging echo structure, and a spatial collocation between the updraft centers and vorticity centers (except that the collocation was not persistent). Therefore, this storm was identified as a “supercell-like storm” in this study. It was investigated by using dual-Doppler analysis and numerical simulations with a cloud-resolving storm simulator (CReSS; Tsuboki and Sakakibara 2002).

The objective of this paper is to reveal the structure and formation mechanism of a supercell-like storm that
develops in a moist environment below the melting layer and to suggest which environmental factors determine the formation mechanism. A brief description of the data used for the present analysis is presented in section 2. The synoptic situation and the storm environment are presented in section 3. The storm evolution investigated by dual-Doppler analysis and surface data is described in section 4. The thermodynamical structures simulated by CRess are described in section 5. Section 6 is a summary of the results and a presentation of a conceptual model for the formation mechanism of a supercell-like storm in a moist environment; it is compared with one in a dry environment, such as the type that occurs in the Great Plains. The conclusion and proposal for a formation mechanism applicable to a moist environment are presented in section 7.

2. Observation data and numerical models

A storm suitable for the purpose of this study was observed over the Kanto Plain on 24 May 2000. The observation provided three-dimensional information with fine spatial and temporal resolution needed for a cloud-resolving numerical simulation.

a. Observation data and analysis method

As shown in Fig. 1, there are two C-band Doppler radars located at the Narita and Haneda airports and an upper-air sounding at the Tateno Aerological Observatory of the Japan Meteorological Agency (JMA). We used the surface station data of the Automated Meteorological Data Acquisition System (AMeDAS) of JMA at Sakura and Koga. The two Doppler radars, covering a radius of 120 km over the Kanto Plain (Fig. 1, upper right) recorded sets of volume scans of reflectivity and Doppler velocity every 6 min. The sampling resolution of the radar data was 150 m in the radial direction and 0.7° in the azimuthal direction. One volume scan consisted of 16 elevation angles (0.7°, 1.0°, 1.3°, 1.7°, 2.1°, 2.7°, 3.4°, 4.3°, 5.4°, 6.8°, 8.6°, 10.9°, 13.7°, 17.4°, 22.2°, and 28.5°), but the volume scans of the Narita radar after 1224 Japan standard time (JST = UTC + 9 h) consisted of 15 elevation angles (0.7°, 1.1°, 1.5°, 1.7°, 2.1°, 2.8°, 3.8°, 5.1°, 6.9°, 9.2°, 12.5°, 17.0°, 23.1°, 32.1°, and 45.9°).

The Doppler radar data were interpolated in a Cartesian coordinate system with a 1-km horizontal grid interval and a 0.5-km vertical grid interval (CAPPI dataset). The correction method introduced by Gal-Chen (1982) was applied to cancel the advection of radar echoes during one volume scan. For the interpolation, a Cressman-type weighting function was used. The horizontal effective radius of influence was fixed to 1.5 km. The vertical effective radius of influence increased linearly from 0.5 to 1.0 km in proportion to its distance from the radar. The interpolation for the velocity data from a small region at high altitude at 1224 JST (X = 10 to 15 km and Z = 11 to 13.5 km in Fig. 9j) was exceptionally conducted with a relatively large horizontal effective radius of influence (3 km) without changing the vertical radius. This made it possible to obtain sufficient continuity of the CAPPI dataset for the dual-Doppler radar analysis at 1224 JST.

Three components of wind in the Cartesian coordinate system were calculated within the dashed circles, except in the hatched area in Fig. 1 (the intersection angle is less than 30°). We conducted dual-Doppler analysis with the variational method introduced by Gao et al. (1999). The boundary condition of Gao et al. (1999) were \( \lambda_0 = 1 \), \( \lambda_y = 4.0 \times 10^6 \), \( \lambda_u = \lambda_v = \lambda_w = 5.0 \times 10^5 \), and...
The terminal fall velocities of rain and snow were adapted from Foote and du Toit (1969), and the temperature fall velocity of “graupel and small hail” was adapted from Straka et al. (2000), as shown below:

\[ w_i = \begin{cases} 
-0.75 \left( \frac{\rho}{\rho_0} \right)^{0.4} Z_e^{0.0714} & \text{for snow} \\
-3.80 \left( \frac{\rho}{\rho_0} \right)^{0.4} Z_e^{0.0714} & \text{for rain} \\
-1.23 \left( \frac{\rho}{\rho_0} \right)^{0.4} Z_e^{0.103} & \text{for graupel}
\end{cases} \]  

Here, \( Z_e \) (mm\(^6\) m\(^{-3}\)) indicates the equivalent reflectivity factor; \( \rho \) and \( \rho_0 \) indicate the air density and surface air density, respectively. The precipitation particles were identified as snow if \( Z_e < 45 \) dBZ\(_e \) above the 0°C level. The particles were identified as graupel if \( Z_e \approx 45 \) dBZ\(_e \) above the 0°C level (Straka et al. 2000). The diameters of frozen precipitation particles were 0.5–1.0 cm before 1224 JST from surface observations (Mito et al. 2000). Around 1242 JST, hailstones of 3-cm diameter were observed at Sakura City (Mori and Takaya 2004). Therefore, in the dual-Doppler radar analysis for a storm at a time equal to and before 1224 JST, the terminal fall velocity of graupel and small hail (diameter less than 2 cm) from Straka et al. (2000) was applied.

The iteration process using the steepest descent method was continued until a satisfactory solution was found. The analysis was terminated after 200 iterations because the horizontal and vertical wind remained essentially unchanged. The mean error of wind velocity was evaluated using the root-mean-squared error (RMSE) between the analyzed radial velocity and observed radial velocity, which is defined as follows:

\[ \text{RMSE} = \sqrt{\frac{\sum_{i,j,k,m} (V_{ri} - V_{Ri})^2}{RN \times DN}}. \]  

Here, \( V_{ri} \) is the observed radial velocity, and \( V_{Ri} \) is the analyzed radial velocity calculated from three components of wind velocity after 200 iterations. \( RN \) is the number of Doppler radars and DN is the number of grid points where three components of wind are calculated. The subscripts \((i, j, k, m)\) denote the observed grid points and the number of radars, respectively. In this study, the RMSE was roughly 1.0 m s\(^{-1}\) at both 1206 and 1224 JST. An RMSE equal to and below 4 km in height was 0.5 m s\(^{-1}\), while the RMSE in the small region around the storm top at 1224 JST was relatively large (3.0 m s\(^{-1}\)). As Gao et al. (1999) noted, the RMSE of the vertical velocity is generally larger than that of horizontal velocity. Therefore, the 1.0 m s\(^{-1}\) of the horizontal RMSE in the radial wind velocity would be largely attributed to vertical components. In this study, our quantitative descriptions of wind velocity are limited to wind velocity equal to and below 4 km in height.

To estimate the storm environment, upper-air soundings of wind, temperature, and humidity at 0900 JST were obtained from the Tateno sounding. Air density obtained from sounding is also used for the estimation of precipitation terminal velocity [Eq. (1)]. The surface temperature, wind speed and direction, and precipitation amounts were available every 10 min from the surface stations of the AMeDAS.

b. Model description

To reveal the thermodynamical structures of the present storm, a simulation using CReSS with a 1-km resolution (1km-CReSS) was conducted.

CReSS is a three-dimensional nonhydrostatic model developed by the Hydrospheric Atmospheric Research Center (HyARC) of Nagoya University, Japan (Tsuboki and Sakakibara 2002). CReSS uses the quasi-aneleastic Navier–Stokes equations and includes a bulk cold rain parameterization and a 1.5-order closure with a turbulent kinematic energy prediction (Tsuboki and Sakakibara 2001). The prognostic variables are the three components of wind velocity, perturbation of potential temperature and pressure, subgrid-scale turbulent kinetic energy, mixing ratios of vapor (\( q_v \)), cloud water (\( q_i \)), rain (\( q_r \)), cloud ice (\( q_s \)), snow (\( q_s \)), and graupel (\( q_g \)), and the number concentrations of cloud ice (\( N_i \)), snow (\( N_s \)), and graupel (\( N_g \)).

The microphysics in CReSS is based on Lin et al. (1983), Cotton et al. (1986), Murakami (1990), Ikawa and Saito (1991), and Murakami et al. (1994). In the microphysics of CReSS, six species (water vapor, cloud, ice, rain, snow, and graupel) are considered. Their mixing ratio and number concentrations of ice, snow, and graupel are predicted. The microphysical processes considered in CReSS are described in Tsuboki and Sakakibara (2002).

The surface fluxes of the momentum and energy and surface radiation processes (Kondo 1976; Louis et al. 1981; Segami et al. 1989) are included with a one-dimensional heat diffusion model in the underground layer for ground temperature prediction. The sea surface temperature is fixed to a climatological value (296 K). The other features of CReSS are summarized in Table 1.

Initial and lateral boundary conditions to the CReSS model were provided by outputs of the Japan Meteorological Agency regional spectrum model (JMA-
The RSM contains a horizontal resolution of 20 km with 129 × 129 grid points and 20 vertical σ levels. Details of the RSM are shown in Saito et al. (2001) and Segami et al. (1989). The domains of the 5-km resolution version of CReSS (5km-CReSS) and 1km-CReSS were nested in the RSM domain as shown in Fig. 2. The 5km-CReSS was integrated forward in time using the RSM output data at 0900 JST on 24 May 2000 as the initial data. Time-dependent one-way boundary conditions were used for the integration of the 5km-CReSS. The initial and lateral boundary data of the 1km-CReSS were provided from the results of the 5km-CReSS.

In the simulation with 5km-CReSS, the horizontal grid size was 5 km, while the vertical grid contained 70 levels with variable grid intervals (Δz = 200 m near the surface, Δz = 370 m at the top level, which is at 20.8 km). The horizontal domain has 200 × 200 grid points, and a time step interval of Δt = 6 s was used.

In the simulation with 1km-CReSS, the horizontal grid size was 1 km, and the vertical grid was similar to that of the 5km-CReSS simulation. The horizontal domain has 300 × 300 grid points, and a time step interval of Δt = 2 s is used.

The parameterization of hail was not included in this simulation. There is considerable validity in this simulation without hail because the focus in this paper is on the formation mechanism of the storm in its early stage, that is, before ice particles grow large (corresponding to hail with a diameter of more than 1 cm) and influence the development of the storm (before 1230 JST).

3. Synoptic condition and storm environment

A surface weather map at 0900 JST is shown in Fig. 3. A weak pressure trough was seen over Honshu in Japan. This pressure trough was accompanied with an upper cold core at 500 hPa (not shown). In the pressure trough, an isolated storm was observed by the Geostationary Meteorological Satellite (GMS) at 1200 JST (Fig. 4). This isolated storm began to appear from 1000...
JST and moved southeastward at a speed of approximately 55 km h\(^{-1}\), developing quickly from 1100 to 1300 JST. After 1300 JST, the storm moved over the Pacific Ocean.

The upper-air sounding at Tateno a few hours before the development of the storm is shown in Fig. 5. A dry layer (where the RH was less than 60%) existed from 650 to 500 hPa. A melting layer existed at 680 hPa [about 3.3 km above sea level (ASL)]. Below the melting layer, there was a relatively moist layer (where the RH was 60% to 90%). CAPE was approximately 1000 J kg\(^{-1}\). This value is smaller than that of a typical supercell environment over the Great Plains (e.g., 2542 J kg\(^{-1}\) for Oklahoma supercells; Bluestein and Jain 1985). However, the frequency distribution of CAPE calculated at Tateno in May from 1990 to 1999 shows that the 1000 J kg\(^{-1}\) is classified as the 10th highest value in 10 yr (not shown). Strong vertical wind shear of more than 4.2 \(\text{m s}^{-2}\) existed from the surface to 5 km ASL. The BRN was 46, and the value was at the higher end of a possible range for simulated and observed supercell formation. The SRH was 195 m\(^2\) s\(^{-2}\), assuming storm motion of 14.0 m s\(^{-1}\) to the southeast. This value is higher than the threshold value (150 m\(^2\) s\(^{-2}\)) for mesocyclone formation noted by Davies-Jones et al. (1990). The estimated BRN and SRH val-
ues indicate that the environment would support the formation and maintenance of supercells.

4. Observation results

The present storm that developed over the Kanto Plain was observed by two Doppler radars from 1112 to 1354 JST (Fig. 1). The storm moved at a speed of 14.0 m s⁻¹ toward the southeast from 1200 to 1230 JST. The storm motion slightly accelerated (15.2 m s⁻¹) toward the southeast after 1230 JST and passed over the Narita Doppler radar from 1230 to 1248 JST. The 15.2 m s⁻¹ (55 km h⁻¹) of the storm motion corresponded with the cloud motion in GMS image (Fig. 4). In this study, the reflectivity obtained by the Haneda Doppler radar is used because the radar almost always detected the three-dimensional structures of the present storm.

The structures of the storm at 2 km ASL from 1154 to 1254 JST are shown in Fig. 6. This storm had a remarkable hook echo structure from 1154 to 1218 JST (Figs. 6a–c). Before 1154 JST, the storm was a single convective cell with the shape of an ellipse inclined in a north-easterly direction. At 1154 JST, a hook-shaped structure appeared in the south of the convective cell (Fig. 6a) and was seen until 1218 JST (Fig. 6c). From 1224 to 1236 JST, the echo region over 25 dBZₑ began to spread east-southeast from the strong echo region over 50 dBZₑ, and the strong echo over 45 dBZₑ was aligned from west to east (Fig. 6d). From 1236 to 1248 JST, a new convective cell over 55 dBZₑ (X = −1 km and Y = −13 km in Fig. 6e) developed 5 km south of the strongest echo region over 60 dBZₑ. At 1254 JST, a strong echo region was aligned in the north-south direction (Fig. 6f).

To investigate the time evolution of the reflectivity profile in the strongest echo region of the storm, a vertical profile of the maximum reflectivity was calculated in a 10 km × 10 km analysis area in Fig. 6. The location of the analysis area at each time was selected so that the strongest echo at each height could be included in the analysis area. Figure 7 shows the time series of the vertical profile of the maximum reflectivity from 1112 to 1254 JST. The height of an echo of 54 dBZₑ (Fig. 7) increased from 5 to 10 km around 1200 JST. A strong echo over 62 dBZₑ was observed at 5–6 km ASL just before 1200 JST. This strong echo descended to the ground continuously and reached a height below 1 km ASL from 1230 to 1248 JST.

Corresponding to the transformation of the reflectivity structure from 1200 to 1254 JST (Fig. 6), there was a remarkable difference in the temperature and wind obtained at the surface stations of AMeDAS (Fig. 8). The surface weather data at Koga and Sakura are shown in Fig. 8 with the reflectivity field at a height of 1 km ASL from 1124 to 1306 JST. The reflectivity core of the present storm moved over Koga at 1136 JST and over Sakura at 1242 JST. By 1136 JST, when the storm passed over Koga station, the surface temperature had fallen only 3°C, and the wind speed had barely changed. However, by 1242 JST, when the storm passed over Sakura station, the surface temperature had fallen 9°C, and the wind speed had strengthened. Furthermore, the direction of the wind had changed significantly by around 1242 JST. The surface data obtained at the Koga and Sakura sites would represent the dynamical and thermodynamical characteristics in the strongest outflow region because the reflectivity core at a height of 1 km ASL passed over the two sites.

The three-dimensional wind structures at 1206 JST, when the present storm had a hook-shaped structure, were investigated by dual-Doppler radar analysis and were compared with those at 1224 JST, when the hook echo began to be unclear below a height of 2 km ASL (Fig. 9).

The radar reflectivity overlaid with storm-relative wind vectors and the vertical velocities at 1206 and 1224 JST at 3 km ASL are indicated in Figs. 9a and 9b, respectively. The storm motion from 1206 to 1224 JST was 14.0 m s⁻¹ toward the southeast. The reason that 3 km ASL was selected is that it was sufficient to indicate where a major updraft at 1206 JST and a new updraft at 1224 JST were located. The vertical vorticity at 1206 and 1224 JST at 4 km ASL is indicated in Figs. 9c and 9d, respectively. The reason that 4 km ASL was selected is that the vertical vorticity at 4 km ASL was the largest at heights equal to and below 4 km ASL at both 1206 and 1224 JST. The distribution patterns of vertical vorticity below 4 km ASL were similar to those at 4 km ASL at both 1206 and 1224 JST.

At 1206 JST (Fig. 9a), there was an updraft center (U1) in the center of the hook echo. The maximum updraft at 1206 JST was 10 m s⁻¹. A cyclonic flow with strong vertical vorticity (6.0 × 10⁻³ s⁻¹) was observed in the strong updraft (Fig. 9a). The location of the vorticity center at 3 km ASL was just under that of the vorticity center at 4 km ASL (Fig. 9c). The vorticity center was collocated with the updraft center below 5 km ASL. The maximum vertical vorticity at 4 km ASL (Fig. 9c) exceeded 1 × 10⁻² s⁻¹. These wind structures observed at 1206 JST had been observed from 1154 JST. Before 1154 JST, the maximum vertical vorticity was weaker than 4 × 10⁻³ s⁻¹ below 3 km ASL (not shown). From 1212 to 1218 JST, the maximum vorticity remained in the order of 1 × 10⁻² s⁻¹ at a height of 4 km ASL (not shown). At 1224 JST (Fig. 9b), the maximum updraft at 3 km ASL increased to
12 m s$^{-1}$, and the maximum vorticity at 4 km ASL was over $1.2 \times 10^{-3}$ s$^{-1}$ (Fig. 9d). A new updraft core (denoted by “U2” in Fig. 9b) to the east-southeast of the U1 was observed at 3 km ASL.

The vertical velocity and divergence distributions of the storm at 1 km ASL at 1206 and 1224 JST are indicated in Figs. 9e–h. The reason that 1 km ASL was selected is that it was sufficient to indicate the location...
of the major downdraft and divergence points. A low-
level environmental storm-relative inflow blew from
east-southeast from 1206 to 1224 JST (Figs. 9e,f). The
low-level storm-relative inflow at 1 km ASL was about
10 m s\(^{-1}\) and became a cyclonic flow around an updraft
region of the storm at both 1206 and 1224 JST. At 1206
JST (Fig. 9e), a downdraft core existed in the rear por-
tion of the storm (northwest of the strong echo core).
The downdraft velocity in almost all regions was less
than 2 m s\(^{-1}\). Both the major downdraft and updraft
regions (≥2 m s\(^{-1}\)) occurred in the vicinity of the re-
flexivity core at 1206 JST (Fig. 9e). The divergence in
almost all regions at 1206 JST was 3 \times 10^{-3} \text{ s}^{-1} or less
(Fig. 9g). As well as the vertical velocity, major conver-
gence and divergence regions occurred in the vicinity of
the reflectivity core at 1206 JST (Fig. 9g). Weak
downdraft and narrow divergence had been observed from
1154 to 1212 JST. At 1224 JST (Fig. 9f), the downdraft
velocity in almost all regions was more than 2 m s\(^{-1}\),
and the maximum downdraft increased to 4 m s\(^{-1}\). A
major downdraft region expanded in the rear side of
the storm, and a new updraft (denoted by “U2” in Fig.
9f) was generated in an east-southeast direction of the
major updraft region. The divergence in almost all re-
gions at 1224 JST was more than 2\times 10^{-3} \text{ s}^{-1} (Fig. 9h).

![Fig. 7](image1)

**Fig. 7.** Time–height cross section of the maximum reflectivity in
the analysis area in Fig. 6 from 1112 to 1254 JST. The contours
show the reflectivity every 1 dB\(Z_e\) starting from 54 dB\(Z_e\). The
hatched area indicates no data.

![Fig. 8](image2)

**Fig. 8.** Time-sequential reflectivity fields at 1 km ASL from 1124 to 1306 JST. The contours
indicate the reflectivity every 5 dB\(Z_e\) starting from 45 dB\(Z_e\). The time series of the surface
temperature (solid line), wind speed and direction, and precipitation amounts every 10 min at
Sakura and Koga (indicated by solid triangles) are indicated in the boxes at the upper-right
and lower-left corners.
The maximum divergence increased to $5 \times 10^{-3}$ s$^{-1}$ at 1224 JST. In addition to a new updraft region (Fig. 9f), a new core of convergence ($3 \times 10^{-3}$ s$^{-1}$) was generated in an east-southeasterly direction of the major convergence region (Fig. 9h). Around the new core of convergence, a new updraft core was observed at a height of 1 km ASL, which was a value greater than RMSE (0.5 m s$^{-1}$).

The vertical cross sections along the A–A’ and B–B’ lines (Figs. 9a–h) through the center of the downdraft core below 3 km ASL are shown in Figs. 9i and 9j, respectively. Each line is selected to be parallel to the direction of the outflow, which was judged from a ground-relative wind near the downdraft core at 1 km ASL. At 1206 JST (Fig. 9i), a strong echo over 50 dBZ$_e$ was overhanging ahead of the storm. A southeasterly inflow turned into a strong updraft over 12 m s$^{-1}$. The new updraft core was located near a weak echo region, which corresponded to the edge region of 40 dBZ$_e$. The maximum downdraft at 1206 JST, which was located in the rear portion of the storm, was 2 m s$^{-1}$. The storm-relative outflow from the downdraft was less than 5 m s$^{-1}$. A gust front, whose location was determined by a technique developed by Uyeda and Zrnic (1986), was located around the strong echo region at a
height of 1 km ASL (denoted by “open arrow” in Fig. 9i). Therefore, the storm-relative inflow could supply warm air to the updraft (U1) just over the gust front because the outflow was weak. On the contrary, at 1224 JST (Fig. 9j), the downdraft accelerated to 6 m s⁻¹, and the downdraft region expanded toward east-southeast. Associated with this expansion of the downdraft region at 1224 JST, the convergence and updraft regions expanded to the southeast, and an echo area in excess of 25 dBZₑ spread to the southeast.

These results of dual-Doppler radar analysis reveal that this storm had characteristics similar to those of a typical supercell storm (from 1154 to 1218 JST): a hook echo, an overhanging echo structure, and an updraft core were collocated with the vorticity center (about $1.0 \times 10^{-3}$ s⁻¹ at 4 km ASL). However, this storm had two different characteristics from those of a typical supercell storm; a single principal intense and cyclonically rotating updraft was not maintained for a long time (another new updraft was generated), and the downdraft in the rear side of the storm was weak. At 1224 JST, this storm was losing the narrow hook echo struc-
ture associated with the strengthening of the downdraft below 3 km ASL. From 1154 to 1224 JST, this storm satisfied the lowest value of the vorticity in a mesocyclone (1.0 \times 10^{-2} \text{s}^{-2}) reported by Burgess and Lemon (1990) and almost satisfied the definition of a supercell of Weisman and Klemp (1984) except for the single persistent rotating updraft. Therefore, in this study, this single storm with strong vertical vorticity was identified as a “supercell-like storm.”

Another remarkable feature of the present supercell-like storm was a slight drop in temperature at the surface layer associated with the storm passage (at Koga in Fig. 8) when the downdraft at 1 km ASL was weak before 1212 JST (Fig. 9e). The temperature decline at the surface layer corresponded with the strength of the downdraft in the reflectivity core region at 1 km ASL. In fact, the decline in temperature increased (at Sakura in Fig. 8) when the downdraft became stronger after 1224 JST (Fig. 9f).

The small reduction in temperature and the weak downdraft at the reflectivity core in the lower layer were considered to be the most important factors for the formation of the present supercell-like storm because the present storm began to lose its supercell-like structure (shown by the hook echo in Fig. 6) when the stronger downdraft and larger reduction in temperature were observed after 1224 JST (Figs. 9f and 8). To reveal the cause of the weak downdraft with a small decline in temperature, thermodynamical structures simulated by CReSS are investigated in the next section.

5. Numerical results

The thermodynamical environment and microphysics processes related to the downdraft acceleration in the present supercell-like storm were investigated using a 1km-CReSS simulation in order to reveal the cause of the weak downdraft of the present supercell-like storm.

The simulated horizontal structures of the present supercell-like storm at a height of 0.97 km at $t = 230$ and 260 min are shown in Fig. 10. The distributions of the mixing ratio of rain ($q_r$) over 2 g kg$^{-1}$ at $t = 230$ and 260 min (Figs. 10a,b) were similar to the hook-shaped echo structures observed by radars at 1206 and 1224 JST (Figs. 9e,f), respectively. Corresponding with the east-southeastward spread of the echo region observed at 1224 JST (Fig. 9f), the region of $q_r$ over 0.1 g kg$^{-1}$ at $t = 260$ min spread toward southeast (Fig. 10b). The cyclonic flow with strong vertical vorticity around the hook at $t = 230$ and 260 min was simulated well (Figs. 10a,b, respectively). The maximum vertical vorticity at 0.97 km ASL was $1.1 \times 10^{-2}$ s$^{-2}$ at $t = 260$ min. A weak downdraft less than 4.0 m s$^{-1}$ was simulated in the rear side of the vertical vorticity center at $t = 230$ min (Fig. 10a). The weak downdraft velocity increased from 4.0 to 6.0 m s$^{-1}$ during 30 min (Fig. 10b). The expansion of the downdraft region at $t = 260$ min was simulated in the rear portion of the simulated storm (Fig. 10b).

Although the simulated storm shifted 30 km west and developed 1 h after the observed storm, the basic characteristics of the observed supercell-like storm were well simulated by 1km-CReSS: 1) hook-shaped rain distribution at $t = 230$ and 260 min, 2) southeastward spread of rain distribution at $t = 260$ min, 3) strong vertical vorticity collocated with a strong updraft from $t = 230$ to 260 min, 4) weak downdraft to the rear of the storm at $t = 230$ min, and 5) strengthening of the downdraft in the rear side of the storm at $t = 260$ min. Therefore, we assumed that the storm structures simulated at $t = 230$ and 260 min corresponded to those observed at 1206 and 1224 JST, respectively (hereafter, 1206 and 1224 JST indicate the time at $t = 230$ and 260 min, respectively).

The distributions of the RH at 0.97 km ASL at 1206 and 1224 JST are shown in Figs. 10c and 10d, respectively. The environmental RH in the northeast side of the present storm (the mean value of the RH in the 15 km \times 15 km region including the downdraft region and excluding the updraft region) was about 70% at 1206 JST (Fig. 10c). The minimum value of the RH in the downdraft core at 1206 JST was 55%. On the contrary, at 1224 JST, the environmental RH to the northeast side of the storm decreased to about 50%, and the minimum peak of the RH descended to 32% (Fig. 10d).

The distributions of the evaporative cooling rate (potential temperature decrease by rain evaporation per unit second) at 1206 and 1224 JST at 0.97 km ASL are shown in Figs. 10e and 10f, respectively. Corresponding to the decrease in the RH below the melting layer (Figs. 10c,d), the maximum evaporative cooling rate at a height of 0.97 km increased from $-0.023$ to $-0.06$ K s$^{-1}$ (Figs. 10e,f) from 1206 to 1224 JST. The increase in the evaporative cooling rate during 30 min ($t = 230$ to 260 min) led to a decrease in the potential temperature perturbation from a hydrostatic balance at a height of 100 m from $-2.0$ to $-5.5$ K around the downdraft region (not shown). This reduction in the potential temperature during 30 min ($t = 230$ to 260 min) corresponded to the surface temperature observations at Koga and Sakura (Fig. 8).

The vertical cross sections of the wind field, the RH, the evaporative cooling rate, and the mixing ratios of precipitation calculated by the summation of the mixing ratio of rain, snow, and graupel along the A–A’ line in Fig. 10 are shown in Fig. 11. Each cross section crosses the updraft and downdraft core below the melting
Fig. 10. Horizontal structures of the supercell-like storm simulated by 1km-CReSS (at a height of 970 m): (left) 1206 and (right) 1224 JST. (a), (b) Mixing ratios of rain (gray shades), winds (arrows), and downdraft velocities (contours); (c), (d) RH (gray shades), winds (arrows), and mixing ratios (contours); and (e), (f) RH (gray shades), winds (arrows), and evaporative cooling. The contours in (a) and (b) indicate the downdraft velocity every 2 m s\(^{-1}\) starting from \(-2\) m s\(^{-1}\). The contours in (c) and (d) indicate the mixing ratios of rain every 0.1 g kg\(^{-1}\) starting from 0.1 g kg\(^{-1}\). The contours in (e) and (f) indicate the evaporative cooling rate every 0.02 K s\(^{-1}\) starting from 0.02 K s\(^{-1}\). The vectors indicate storm-relative wind. Storm motion is indicated by the vector in a box in the lower-right corner. The vertical cross sections along the A–A' line or the B–B' line are shown in Fig. 11.
layer. The mixing ratios of precipitation at 1206 and 1224 JST are shown in Figs. 11a and 11b, respectively. The heavy precipitation above 3 km ASL mainly consisted of graupel, while the precipitation below 2.5 km in height corresponded to rain (not shown). The vertical distribution of the heavy precipitation was similar to the observed vertical reflectivity structure with regard to the following: 1) overhanging structure of heavy precipitation, 2) storm height of about 12 km at 1206 JST, and 3) small amount of precipitation in the strong updraft core corresponding to the weak echo region. As in the case of the precipitation structure, the weak downdraft in the rain shaft was well simulated (Figs. 11a,b). The downdrafts at 1206 and 1224 JST were located to the rear of the storm and below the melting layer, respectively (3.3 km ASL). The maximum downdraft velocity at 1206 JST was 4 m s\(^{-1}\) (Fig. 11a), which was consistent with the observed weak downdraft (Fig. 9). The weak downdraft produced a weak storm-relative surface outflow (0.2 m s\(^{-1}\)) in the rain shaft (Fig. 11a). A southeasterly warm storm-relative inflow (high equivalent potential temperature \(\theta_e\) over 328 K) existed.
at less than approximately 1.6 km ASL ahead of the storm. The storm-relative inflow was weak (less than 10 ms$^{-1}$) ahead of the downdraft core. The maximum downdraft velocity at 1224 JST was stronger (up to 6 ms$^{-1}$) (Fig. 11b). The outflow at 1224 JST (10–15 ms$^{-1}$) was stronger than that at 1206 JST (less than 5 ms$^{-1}$) in Figs. 11a and 11b. Corresponding to the strengthening of the downdraft, the precipitation area at a height of 970 m ASL spread to the southeast (Fig. 11b).

The vertical cross sections of the RH at 1206 and 1224 JST are shown in Figs. 11c and 11d, respectively. At 1206 JST, a relatively moist environment (the RH was 60%–75%) existed to the rear of the rain shaft below the melting layer (Fig. 11c). Corresponding to the downdraft acceleration below the melting layer in the rain shaft from 1206 to 1224 JST, the dry air (the RH was less than 30%) above the melting layer descended to the layer below the melting layer at 1224 JST (Fig. 11d). The descended dry air was limited to a major downdraft region.

The vertical cross sections of the evaporative cooling rate at 1206 and 1224 JST are shown in Figs. 11e and 11f, respectively. At 1206 JST, the evaporative cooling rate was less than 0.02 K s$^{-1}$ (Fig. 11e). Corresponding to the descent of the dry air to the rain shaft at 1224 JST (Fig. 11d), the evaporative cooling rate at 1224 JST increased by approximately 2 times that at 1206 JST (Fig. 11f). The region with evaporative cooling of less than –0.02 K s$^{-1}$ at 1206 and 1224 JST (Figs. 11e,f) corresponded to the downdraft region over 2 m s$^{-1}$ at 1206 and 1224 JST (Figs. 11a,b), respectively.

To reveal how the evaporative cooling acted as downward acceleration at 1206 and 1224 JST, the core locations of the evaporative cooling rate at 970 m ASL, negative buoyancy of precipitation loading at 970 m ASL, melting cooling rate of graupel at 2450 m ASL, and sublimation cooling of graupel at 4620 m ASL at 1206 and 1224 JST are shown in Fig. 12. The relative importance for the downward acceleration among these four terms was investigated. We did not include the pressure gradient in our analysis, which generally acts to maintain the updraft ["lifting pressure gradient" proposed by Rotunno and Klemp (1982)] because there is no major vertical vorticity in the downdraft region below 4 km ASL and because the vertical pressure gradient force in the downdraft region at lower layer was positive. The vertical pressure gradient force in the downdraft region acted as compensation for the major negative buoyancy, which is estimated in this study, to maintain the hydrostatic balance. Furthermore, we did not consider the buoyancy of the virtual temperature difference proposed by Srivastava (1985). Srivastava (1985) revealed that the moist environment leads to stronger downdrafts, since the virtual temperature difference between the downdraft parcel and the environment is greater using a simple one-dimensional model, which neglects pressure gradient force. However, the effect of the vapor perturbation is much smaller than that of the evaporative cooling in the simulation using three-dimensional model, which includes the vertical...
pressure gradient force preventing a downdraft from being too strong.

The cores of melting cooling and sublimation cooling were 5–10 km away from the focusing downdraft core (Fig. 12a) and located on the leeward side of the downdraft core above 3 km ASL. The core of the precipitation loading term was located in the region between the downdraft and updraft cores (Fig. 12a), while the core of the evaporative cooling rate collocated with the downdraft core at 1206 JST (Fig. 12a). The collocation between the downdraft core and evaporative cooling rate core had been simulated before 1218 JST. At 1224 JST, the core of the evaporative cooling rate was located on the northwest edge of the downdraft core, and the core of water loading was located in the southern edge of the downdraft core (Fig. 12b). The peak value of the mixing ratio of rain at 970 m did not change significantly (approximately 1 g kg\(^{-1}\) increase) from 1206 to 1224 JST (Figs. 12a,b). The 1 g kg\(^{-1}\) of the mixing ratio of rain is roughly equivalent to 0.3 K of the negative thermal buoyancy (Cotton and Anthes 1989), which is negligibly small relative to the observed 3- and 7-K temperature drop at the Koga and Sakura surface stations.

The vertical profiles of the four terms averaged in the analysis region in Fig. 12 at 1206 and 1224 JST are compared in Fig. 13. The maximum peak value of the evaporative cooling rate was larger than those of the melting and sublimation cooling rates of graupel at both 1206 and 1224 JST. The contribution to the downdraft acceleration of evaporative cooling was larger than those of melting and sublimation cooling below the melting layer. The evaporative cooling rate at 1206 JST was half the size of those at 1224 JST at each height below the melting layer. Judging from Figs. 12 and 13, the locations of downdraft cores before 1218 JST corresponded well with the locations of the cores of the evaporative cooling rate. The weak downdraft around 1206 JST was strongly related with the small evaporative cooling rate in the moist environment (Fig. 11e). Indeed, when the evaporative cooling rate increased by about 2 times (Figs. 11f and 13) in the drier environment (Fig. 11d), the downdraft at the rear of the storm became stronger (Fig. 11b). The weak evaporative cooling consisted of a small potential temperature perturbation at 100 m ASL, which corresponded well with the surface temperature observations. The weak downdraft caused by the small evaporative cooling did not cut off the weak inflow to the updraft of the supercell-like storm for 30 min.

6. Discussion

In this section, the results of the observation and numerical simulation of the 24 May 2000 storm with strong vertical vorticity before 1224 JST are summarized, and the formation mechanism is discussed with the presentation of conceptual models.

The present storm was formed in an environment possessing a large value of CAPE, strong vertical wind shear (4.2 \(\times\) 10\(^{-3}\) s\(^{-1}\) between the surface and 5 km ASL), and a relatively moist layer below the melting layer (Fig. 5). The structure and formation mechanism of the present storm were revealed from 1154 to 1224 JST. Before 1218 JST, the structure of this storm was similar to that of a typical supercell. The present storm had a hook echo (Fig. 9a) and an overhanging echo structure (Fig. 9i). The vertical vorticity at 4 km ASL reached an order of 1.0 \(\times\) 10\(^{-2}\) s\(^{-1}\), which is the lowest value of vertical vorticity in a mesocyclone (Burgess and Lemon 1990). The present storm would satisfy the definition of a supercell (Weisman and Klemp 1984), were it not for a single long-lived updraft with strong vertical vorticity. Therefore, the time when a single up-
draft with strong vertical vorticity was observed from 1154 to 1218 JST was identified as a “supercell-like stage” in this study. In addition, the time when another updraft core was observed to the southeast of the old updraft from 1218 to 1230 JST was identified as a “transition stage.” A conceptual model of the present storm in the supercell is shown in Fig. 14. Judging from the results of the numerical simulation, the precipitation particles that made up the overhanging echo structure were mainly graupel and hail (above the melting layer) and rain (below the melting layer). A weak southeastward storm-relative inflow (about 10 m s\(^{-1}\)) converged with an outflow from the present storm. Above a gust front, the inflow turned to a strong updraft with strong midlevel vorticity (Figs. 9a,c). As Schlesinger (1980) and Rotunno and Klemp (1982) suggested, the lifting pressure gradients derived from the strong midlevel vertical vorticity would induce updraft growth almost directly above a surface gust front. At the rear portion of the present storm, the northwesterly wind blew into a rain shaft and turned to a weak downdraft below the melting layer (Figs. 9i and 11a). This weak downdraft caused a weak surface divergence (about 2 \(\times\) 10\(^{-3}\) s\(^{-1}\)) and a weak outflow (about 10 m s\(^{-1}\)), which converged with the weak storm-relative inflow. The storm-relative inflow and the outflow were weaker than those of a typical supercell (20 m s\(^{-1}\); Ray et al. 1981). The major cause of the weak downdraft was the small evaporative cooling by rain below 3 km ASL at 1206 JST (Fig. 13). The small evaporative cooling was simulated in a relatively moist environment (the RH was 60%–75%) below the melting layer (Fig. 11c). Because of the weak inflow and outflow, the convergence location of the outflow and the inflow was just under the updraft location for about 30 min; that is, the strengths of the inflow and outflow retained their balance for 30 min. Even the weak storm-relative inflow was able to prevent the weak outflow from moving out of the storm and continue to supply warm air to the updraft (Figs. 9i and 11a).

The balance of the inflow and outflow strengths was gradually being lost after 1224 JST, when the RH below the melting layer became drier (less than 30%). As the downdraft, divergence, and outflow became stronger at 1 km ASL at 1224 JST (Figs. 9f and 10b), a convergence region at 1 km ASL (Fig. 9h) expanded to the southeast, and a new updraft developed to the southeast of the major updraft (Figs. 9b,d). In addition, the single hook-shaped echo became indistinct after 1236 JST (Fig. 6).

To retain the balance of the weak outflow and weak inflow, the RH below the melting layer was a crucial parameter. As Gilmore and Wicker (1998) proposed, the RH below the melting layer, as well as the BRN and SRH, was demonstrated to be an important environmental parameter to determine a moderate outflow for the formation of a single convective cell with strong vertical vorticity. For a general understanding of the formation mechanism of a storm with strong vertical vorticity and a suitable environment for the formation, more observation and numerical studies on supercells and supercell-like storms in moist and dry environments should be conducted with a focus on the effects of the RH distribution.

7. Conclusions

The structure and formation mechanism of a storm that developed over the Kanto Plain, Japan, on 24 May 2000 were investigated using dual-Doppler radar analysis and the cloud-resolving storm simulator (CReSS).

The storm formed in an environment possessing a large value of CAPE (1000 J Kg\(^{-1}\)), strong vertical wind shear (4.2 \(\times\) 10\(^{-3}\) s\(^{-1}\) between the surface and a height of 5 km), and a moist layer (the RH below a melting layer was 60%–90%). The BRN was 47, and the SRH was 195 m\(^2\) s\(^{-2}\).

Using dual-Doppler analysis with a variational method, it was revealed that the structure of the present storm had many characteristics of a typical supercell (Browning 1964; Browning and Foote 1976). A hook echo, an overhanging echo structure, and a strong updraft with strong vertical vorticity were observed in the

---

**Fig. 14.** Conceptual model of the present storm during the supercell-like stage. The storm formed in an environment possessing large vertical wind shear and a moist layer below the melting layer. In the supercell stage, the weak (warm) inflow balanced the weak (cold) outflow. The updraft with strong vertical vorticity formed just above the gust front. Heavy graupel and hail (reflectivity more than 60 dBZ) did not descend to less than 3 km in height. The main cause of the weak downdraft was small evaporative cooling due to a moist environment below 3 km in height.
present storm. However, this supercell-like storm had a weak downdraft (weak outflow), a weak inflow, and short maintenance of a single cyclonically rotating updraft (about 30 min). The convergence location of the outflow and inflow was just under the initial updraft location for about 30 min. The weak inflow with horizontal vorticity turned into an updraft with strong midlevel vertical vorticity (on the order of $1.0 \times 10^{-2}$ s$^{-1}$).

The observed features were simulated well using CReSS, and the thermodynamical features of the formation mechanism were revealed. Corresponding to the observed hook echo, a hook-shaped distribution of precipitation was also well simulated. In correspondence with the observed overhanging echo structure, graupel was distributed mainly above a melting layer (3 km ASL), and rain was distributed mainly below that level. A weak downdraft core at a height of 2 km was located at the edge of the rain shaft. The downdraft core was located in the moist layer below the melting layer (where the environmental RH was 70%). This moist layer made the evaporative cooling inactive and caused a weak downdraft and a weak outflow. The weak outflow was able to balance the weak inflow, and the convergence of the outflow and the inflow kept its location just under the initial updraft. Therefore, the weak outflow did not cut off the initial updraft, and the weak inflow was able to keep supplying warm air to the initial updraft for 30 min. The present supercell-like storm formed as a result of the balance of the strengths of a weak inflow and a weak outflow in a moist and strong vertical shear environment.

RH below a melting layer was demonstrated to be an important environmental parameter to determine the outflow strength, in agreement with what Gilmore and Wicker (1998) had proposed. More observation and numerical studies on the relationship between environmental factors and supercell formation would provide a systematic understanding of the formation mechanism of supercells that develop in various environments.

Acknowledgments. We would like to express our appreciation to the Haneda Airport Meteorological Observatory and the Narita Airport Meteorological Observatory for providing radar data. We would also like to express our thanks to associate professor K. Tsuboki and assistant professor T. Shinoda of the Hydropheric Atmospheric Research Center of the Nagoya University for providing useful suggestions and comments. We thank Mr. A. Sakakibara of Chuden CTI Co., Ltd. for giving useful advice for CReSS running. We are grateful to Prof. H. Niino of the Ocean Research Institute of the University of Tokyo for providing helpful suggestions and comments. Thanks are also due Mr. Y. Yuuki of ITOCHU Techno-Solutions Corporation for providing helpful comments. This study was partly supported by a Grant-in-Aid for Scientific Research of the Japan Society for the Promotion of Science. All figures are created using Generic Mapping Tools.

REFERENCES


——, R. B. Wilhelmson, and P. S. Ray, 1981: Observed and nu-


