Tornadoes in Chiba Prefecture on 11 December 1990

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ABSTRACT

On the evening of 11 December 1990, two supercell storms hit the Chiba Prefecture, southeast of Tokyo, and spawned two tornadoes in Mobara and Kamogawa. The Mobara tornado caused the most severe tornado damage since 1960 in Japan over a damage swath of 6.5 km in length and 500 m in average width. A detailed damage survey revealed that the tornado moved north-northeastward at a speed of about 16 m s⁻¹. The maximum wind speed near the ground, estimated from damage to structures, was more than 78 m s⁻¹.

The storms were initiated in the warm sector of a developing extratropical cyclone, about 6–7 h prior to the tornadoogenesis. They moved straightforwardly northeastward at a speed of about 16 m s⁻¹ throughout their life cycles including their supercell phases.

The mesocyclone in the Mobara storm had been detected by a single-Doppler radar for 44 min. Vertical vorticity of the mesocyclone amplified to 2 × 10⁻² s⁻¹ almost simultaneously between 1 and 5 km AGL, about 20 min prior to the tornadoogenesis. About 4 min before the tornadoogenesis a small mesocyclone formed in the south edge of the major mesocyclone, and this new mesocyclone produced the tornado.

The Kamogawa storm was in supercell phase for more than 2 h. A mesocyclone was detected both by surface wind records and by the Doppler radar. The vorticity of the mesocyclone amplified and weakened at least two times. Traces of surface pressure, temperature, and precipitation rate at Tateyama Observatory in the storm path, about 20 km southwest of Kamogawa, showed that the center of the mesocyclone was located in cooler air behind the gust front and the major precipitation preceded the mesocyclone. A barometer located near the center of the tornado damage path in Kamogawa recorded two pressure dips, indicating that the centers of the mesocyclone and the tornado were about 5 km apart.

1. Introduction

Tornadoes have been considered generally to be one of the less significant natural disasters in Japan. According to statistics between 1961 and 1982 (Mitsuta 1983), tornadoes killed 0.5 person and injured 20.6 persons per year, on average. These numbers are much smaller than those due to typhoons and heavy rainfalls during the rainy season (baiu). The same statistics show an annual average of about 18 tornadoes per year—much fewer than the 30-yr annual mean of 771 tornadoes in the United States (Ferguson et al. 1989).

However, since the United States has about 25 times the area as Japan, this makes the number density in the United States only 1.7 times greater than that in Japan. Furthermore, the number density in the Okinawa Prefecture is more than six tornadoes per 10⁴ km² per year, comparable to that of Oklahoma in the United States.

Tornadoes in Japan are believed to be weaker, in general, than those in the United States. This is mainly due to the fact that there have been no tornadoes ranked F4 or higher in Japan since 1950 (Fujita 1971; Mitsuta 1983). Fujita (1971) showed that the relative frequencies of occurrence of F0, F1, F2, and F3 tornadoes in Japan are fairly similar to those in the United States. Since the annual occurrence of tornadoes in Japan is about 40 times smaller than that in the United States, it would take many more years until one could compare, with confidence, statistics of F4 tornadoes in Japan with those in the United States.

Because of the lower frequency of tornadoes, not many detailed studies of tornadoes and tornadic storms in Japan have been made so far. Shimada (1967) analyzed the Yoshino Valley tornado on 2 April 1961 in the Ishikawa Prefecture and suggested the existence of a tornado cyclone from a limited number of pressure and wind records. He also mentioned briefly a storm
in the Yamagata Prefecture on 1 July 1965 that exhibited a vault-shaped echo on the radar plan position indicator (PPI) image and spawned a tornado in its right-rear flank near the leading edge of a hook-shaped echo but without presenting any specific evidence. Fujita et al. (1972) made a damage survey and a mesoscale analysis of the Omiya tornado on 7 July 1971. They found a mesocyclone that spawned the tornado and two types of suction-vortex swaths in an open field. Muramatsu (1982) analyzed the mesoscale environment of a tornado in Tokyo on 28 February 1978. The parent storm (with a lifetime of more than 4 h) of the tornado developed at the southern end of a squall line. The mesocyclone that spawned the tornado formed when the parent storm moved over a shear line between warm, moist southwesterly flow and relatively cool, dry, stagnant air. The mesocyclone was accompanied by a hook-shaped echo. Omoto (1982) analyzed the parent storm of a typhoon-associated tornado in Nagoya on 4 September 1979 and stated that it was similar to supercell storms in many respects except that it moved in the direction of the mean wind at around 600 hPa. Shirooka and Uyeda (1991) reported that a tornado vortex signature (TVS) was observed by a Doppler radar in the parent cloud near the Chitose Airport, Hokkaido, on 22 September 1988.

Only a few papers on nontornadic supercell storms exist in Japan. Omoto (1970) reported a hailstorm in the Kanto Plain on 7 June 1966 that lasted for more than 6 h and produced a hail damage path of 150 km in length. He concluded that it had several characteristics similar to the classical supercell storm (Browning 1964). Omoto (1979) stated that a supercell storm developed near Mount Tateyama on 19 July 1976 and moved southeastward with a lifetime of 6 h, leaving a hail swath more than 160 km long.

In this paper we describe the results of a mesoscale analysis of two supercell storms on 11 December 1990—each of which spawned a tornado in the Chiba Prefecture, southeast of Tokyo—and the characteristics of the tornadoes as revealed from a damage survey. The first tornado hit Kamogawa, 70 km south-southeast of Tokyo, from 1747 to 1758 JST (Japanese standard time; all reference to time is in JST—0900 JST is 0000 UTC), and the second hit Mobar, 55 km east-southeast of Tokyo, from 1913 to 1920 (see Fig. 1). The damage caused by the latter tornado was the most severe in Japan since 1960: 1 person died; 74 were injured. Eighty-one houses were totally destroyed, 161 severely damaged, and 1504 lightly damaged. In addition to the damage from the tornado, some damage from heavy winds occurred in Maruyama and the cities of Futsu, Kimitu, and Choshi from 1730 to 1940 JST.

Throughout the evening after 1840, the single-Doppler radar at the Meteorological Research Institute
(MRI) in Tsukuba was operating and detected several mesocyclones, two of which produced tornadoes. A considerable amount of meteorological data was also collected from the Japan Meteorological Agency (JMA), prefectural and city governments, schools, fire departments, and elsewhere, and were used to analyze the mesoscale environment that led to the tornado genesis.

In the following section, characteristics of the tornadoes are briefly described. In section 3, data used for the mesoscale analysis are described. Section 4 describes the synoptic conditions. Sections 5 and 6 present and discuss the life cycles and structures of the parent storms of the tornadoes. In the last section, the results are summarized.

2. Characteristics of the tornadoes

a. Kamogawa tornado

Figure 2 shows the distribution of damaged houses surveyed by the Kamogawa government office. The width and length of the damaged area are roughly 1.5 km (2.0 km in the maximum width) and 8.4 km, respectively. The damage path shown in Fig. 2 extends northeastward into the Tokyo University Forest, where the swath of damaged trees was 3 km long and 400 m wide. A resident of Kamogawa, Y. Suzuki, witnessed a funnel cloud with the aid of lightning at point C in Fig. 2.

The electric power supply at the Earthquake Phenomena Observation System (EPOS) of JMA at point B in Fig. 2 was cut off 16 s after 1754 JST. A computer operated by a building security company recorded the occurrence of damage to a building, at point D in Fig. 2, 4 s after 1754. Taking this information as a reference, and assuming that the tornado moved with the parent storm with a speed of 16 m s⁻¹, we estimate that the tornado damage in Kamogawa occurred in the period from 1747 to 1758 JST. This information will be used later in section 3b.

b. Mobara tornado

The characteristics of the Mobara tornado as revealed from the damage survey are described in detail by Niino et al. (1991) and Niino et al. (1993). For readers’ convenience, however, the important features of the tornado characteristics will be summarized in the following.

The Mobara tornado began at 1913 JST (Niino et al. 1991; Niino et al. 1993), when it was already dark. There were frequent lightning flashes, and the tornado itself seemed to emit a blue light occasionally. The funnel cloud (Fig. 3) illuminated by lightning was recorded on a video by K. Sasaki (at point f in Fig. 4) from a distance of 1 km. The video showed that the tornado was rotating cyclonically and moved approximately northward.

1) Damage characteristics and tornado path

Figure 4 shows the distribution of damaged houses based on the data provided by the Mobara government office. Most of the houses in Mobara are made of wood and have one or two stories with tiled roofs. The majority of the damaged houses are in a highly populated area of Mobara, where the area density of houses was more than 30 per 10⁴ m². However, beyond the Japan Railways (JR) Shin-Mobara Station to the north, there are many farms and fewer houses.

![Image of Mobara tornado](https://example.com/mobara_tornado_image.jpg)

**Fig. 2.** Distribution of houses (open circles) damaged by the Kamogawa tornado. The solid squares show the Kamogawa AMeDAS station (A), the Kamogawa EPOS station (B), the location (C) where the tornado was witnessed, and the building (D) at which the initiation time of the tornado damage was monitored by a computer center of a building security company.

**Fig. 3.** The funnel cloud of the Mobara tornado (printed from a video taken by K. Sasaki). A color version of this picture is found in Suzuki and Niino (1991).
curity company. In the second method, the time lag between the stoppage of electric power supply due to a cut of high-voltage line and the onset of direct damage to a house was obtained from eyewitness evidence and was used to estimate the velocity. The damage path extended to the area around the JR Shin-Mobara Station and farther to the north. The length of the damage path was about 6.5 km, and its average width was about 500 m. (The maximum width was 1200 m.)

Most of the severe damage to houses was concentrated in the Takashi area near the corner where the JR Sotobou Line changes direction from northwest to north-northeast. Figure 5 shows the damage in the Takashi area as photographed from a helicopter early in the morning of 12 December.

The dashed lines in Fig. 4 show the wind convergence lines that were derived from the directions of the strongest surface winds as revealed from the damage survey. The wind directions were estimated from the directions toward which road signs, poles, trees, and crops on farms were leaning or fell down. The inferred wind directions on the east side of the dashed lines were south-southeast and those on the west side were west-northeast. It is seen that these lines are located in the

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Fig. 4. Distribution of the damaged houses and the points where meteorological data were recorded. The open circles show lightly damaged houses, the solid triangles severely damaged houses, and the solid circles completely destroyed houses. The dashed lines show the estimated paths of the tornado center: A—Mobara AMeDAS station, B—Mobara Takashi Station of the Environmental Division of Chiba Prefecture, C—Chosei High School, D—Mobara Agricultural High School, E—Fujimi Junior High School, F—Mobara Transformer Station of the Tokyo Electric Power Company, H—Mobara gymnasium, I—K. Sasaki's house where the video of the tornado was taken, and J—the location where the maximum wind was estimated from distortion of a road sign.

The first damage to a house occurred 500 m south of the intersection of Routes 128 and 409. The width of the damage path increased as the tornado moved north-northeastward at a speed of 16 m s⁻¹, where the translational velocity of the tornado was estimated by two independent methods (Niino et al. 1991; Niino et al. 1993); in the first method, velocities were calculated from the time and locations of alarm signals, which signified an interruption of electrical power supply or direct damage to buildings, and were recorded by a single computer at the control center of a building ser...

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Fig. 5. Damage in the Takashi area [courtesy of Kyodo News Service, Niino et al. (1993)].
northwest side of the severely damaged area in Fig. 4. Since one would expect convergent winds near the surface in the tornado vortex and stronger winds in the southeast quadrant of a cyclonic tornado because of the superposition of the general wind on the rotational wind, the center of the tornado vortex is considered to have moved along the dashed lines.

The convergence line becomes obscure and jumps northward near the Mobara gymnasium (point \( H \)), which has dimensions of 77 m in width, 94 m in length, and 21 m in height. The damage distribution in Fig. 4 has an eastward protuberance of about 500 m near the gymnasium, and even several houses were completely destroyed or severely damaged in the east part of the protuberance. Fujita (1992) speculates that a downburst occurred near the east end of the protuberance and undercut the tornado, and the tornado redeveloped on the northwest side of the gymnasium as the microburst weakened. Although this is possible, eyewitness evidence concerning wind directions in the protuberance are not completely compatible with the streamlines of the microburst drawn by Fujita (e.g., Fujita drew northward streamlines near the point where the microburst started; however, an owner of a grocery shop near the point saw a vending machine be blown over and carried eastward). Some structural change of the tornado vortex such as widening of the vortex diameter could be another possibility. However, insufficient data exist to clarify the detailed mechanism that produced the peculiar behavior of the convergence lines and the damage distribution.

2) METEOROLOGICAL RECORDS

Wind and pressure records near the tornado path were taken at several schools and at other observation points shown by the crosses in Fig. 4. Unfortunately, there was no means of calibrating the time marks on the traces at these points. Therefore, the times described in this section will be based on the original time marks on the traces and may not be very accurate.

Figure 6 show the wind records taken at the height of 10 m in the farm field of the Mobara Agricultural High School (at point \( D \) in Fig. 4), at the Mobara Transformer Station of the Tokyo Electric Power Company (point \( F \)), and at the Mobara Takashi Station of the Environmental Division of the Chiba Prefecture (point \( B \)). Maximum southeasterly winds of 30.8 and 29.5 m s\(^{-1}\) were observed at about 1912 JST at point \( D \) and at about 1920 at point \( F \). The wind direction was northeast at point \( B \) at about 1915 (just before the interruption of electrical power). The wind directions at these three points are consistent with a cyclonic tornado passing west of points \( D \) and \( F \) and east of point \( B \).

Figure 7 shows the pressure records at the Chosei High School (point \( C \)), at point \( D \), and at the Fujimi Junior High School (point \( E \)). The barographs had a damping mechanism whose time constant for the response to a sudden pressure change of the order of 10 hPa is about 3 s (Y. Yokota 1993, personal communication). A sudden pressure drop of about 9 hPa was recorded at about 1925 JST at point \( C \), about 400 m east of the tornado path, where the time for the pressure record was calibrated in reference to that at point \( E \) by assuming that the pressure variations at the stations in this area from 11 to 12 December except around the time of the tornadogenesis were identical. The pressure drop at point \( C \) is considered to be caused by a superposition of the pressure decrease due to the tornado and that due to a mesocyclone. At point \( D \), 900 m east of the path, a pressure drop of 2 hPa was recorded at about 1930, and at point \( E \), about 2000 m west of the path, a drop of 2.5 hPa was recorded at about 1920 JST. The pressure drops by the tornado alone at points \( D \) and \( E \) would be about 20% and 6% of the pressure drop at point \( C \), respectively, if a potential vortex is assumed for the tornado. The contribution of the mesocyclone to the pressure drop at point \( C \) is unknown. However, even if the whole pressure drop (9 hPa) was assumed to be caused by the tornado alone, the pressure drops at points \( D \) and \( E \) due to the tornado would be 1.8 and 0.5 hPa, respectively. Thus, the pressure drop at point \( D \) is likely to have been caused by the superposition of the pressure fields of the tornado and the mesocyclone, and that at point \( E \) was mainly caused by the mesocyclone.

3) ESTIMATE OF THE MAXIMUM WIND SPEED

Since no anemometer existed just inside the tornado path, maximum wind speeds were estimated from damage to structures. To this end, a very crude assumption that the structures were damaged only by pressure forces due to steady unsheared winds was made. Thirteen structures of relatively simple shapes were chosen to estimate the wind speeds. The maximum wind speed among them was 78 m s\(^{-1}\) obtained from distortion of a road sign at point \( J \) in Fig. 4. The road sign consisted of a square steel plate of 58.0 cm \( \times \) 58.0 cm and a supporting steel pole of a circular cross section, whose length, diameter, and thickness were 357.0, 7.63, and 0.32 cm, respectively. The supporting pole surrendered at a height of 30.0 cm above the ground. Though the road sign had been painted, no damage on the painted surface due to flying objects was found. The maximum wind speed of 78 m s\(^{-1}\) indicates that the Fujita scale of the Mobara tornado was at least F3. Fujita (1992) analyzed the video of the tornado taken by K. Sasaki (Fig. 3) and obtained wind speeds of 83–91 m s\(^{-1}\) at the height of 70 m AGL. Our estimate of 78 m s\(^{-1}\) at the height of about 2 m is not inconsistent with his analysis. Fujita (1992) used another videotape, filmed by a Japan Broadcasting Cooperation (NHK in Japanese) helicopter early in the morning of 12 December, to obtain the damage
directions of trees, debris, and structures. This information led him to conclude that several suction vortices were generated in the tornado and the one that formed near the Mobara gymnasium produced F4 winds.

3. Data for mesoscale analysis

The following data are used in the mesoscale analysis presented in sections 5 and 6.

a. Radars

The time series data of digitized precipitation intensity from two JMA meteorological radars at Tokyo and Mount Fuji and two JMA airport radars at Haneda and Narita were used for the present analysis. The locations of the radars are shown in Fig. 1. The Tokyo radar is not actually located in Tokyo but in Kashiwa, about 30 km northeast of Tokyo (see Fig. 1b). Mount Fuji radar operates on a wavelength of 10.4 cm, while the rest of the radars operate at a 5.7-cm wavelength.

The Tokyo and Mount Fuji radars cover 250- and 500-km ranges, respectively. The Tokyo radar provides precipitation intensity $R$ defined on 2.5-km $\times$ 2.5-km squares at the level of about 2 km AGL every 7.5 min,
where $R$ was simply calculated from a single $Z-R$ relationship of $Z = 200R^{1.6}$ and was categorized into seven classes separated by the precipitation intensities of 1, 4, 8, 16, 32, and 64 mm h$^{-1}$. The Mount Fuji radar similarly gives the precipitation intensity on 5-km $\times$ 5-km grid squares approximately every 10 min.

The Haneda and Narita radars cover a 100-km range and provide the precipitation intensity $R$ on 1-km $\times$ 1-km grid squares, where $R$ was calculated from the same $Z-R$ relationships and was categorized into five classes separated by 1, 4, 16, and 64 mm h$^{-1}$. The former provided $R$ about every 5 min, and the latter every 10 min. The elevation angles of Haneda and Narita radars were fixed to 1.7° and 1.5°, respectively.

The single-Doppler radar at MRI was in operation from 1840 JST. The operating wavelength of this radar is 5.7 cm, and its pulse repetition frequency was 1120 Hz. This gives an unambiguous velocity of 32 m s$^{-1}$. The resolution in the elevation angle and azimuth was 1.41°, and that in the direction of the beam 500 m. It gives reflectivity and Doppler velocity data at 11 elevation angles, every 7.5 min.

**Fig. 7.** Pressure records from 0600 JST 11 December to 0600 JST 12 December at (a) Chosei High School, (b) Mobra High School, and (c) Fujimi Junior High School.

**b. Surface meteorological data**

Surface meteorological data were collected from local observatories, the Automated Meteorological Data Acquisition System (AMeDAS) and EPOS stations of JMA, environmental monitoring stations of the Chiba Prefecture and local city governments, universities, schools, fire stations, the Japan Defense Agency, and transformer stations of the Tokyo Electric Power Company. The locations and category of the stations are shown in Fig. 8. The AMeDAS stations are distributed with an average density of about one site per 400 km$^2$.

The quality of the data and instruments at these stations varies so widely that a thorough description is not possible. However, a description will be given in the text where it is thought to be essential.

**4. Synoptic conditions**

At 0900 JST 11 December, an extratropical cyclone with central pressure of 1004 mb was located east of the Korean Peninsula (see Fig. 1). It moved eastward while continuing to develop and reached the Noto Peninsula (Fig. 1) by 1500 JST. Figure 9 shows the surface weather map at 1800, which is almost the time when the Kamogawa tornado was on the ground and about 1 h prior to the outbreak of the Mobra tornado. The original cyclone was located to the east of the Noto Peninsula. A new extratropical cyclone was generated to the south of Mount Fuji between 1700 and 1800,
on the cold front that extended southward from the center of the original cyclone. A warm front extended eastward from the center of the new cyclone. It is seen that the tornadoes occurred in the warm sector of the new cyclone.

Because of sparsity of the upper-air soundings in space and time, no upper soundings of thermodynamic quantities or winds were available in the warm sector except the vertical profile of wind at Hachijou Island (Fig. 1), about 200 km south of the storms, at 1500 JST.

5. Life cycle of the parent storms

During the evening of 11 December, several strong storms moved northeastward off the southern coast of Honshu. Figure 10 shows the evolutions of storms as viewed from the Mount Fuji, Tokyo, Haneda, and Narita radars.

Three convective systems are found at 1500 JST in Fig. 10b. The largest convective system (hereafter referred to as CS1), which extends southwestward from the Noto Peninsula, corresponds to the cold front that extended southwestward from the center of the original cyclone.

The second convective system (CS2) near Hamamatsu (see Fig. 10a) had an east–west scale of about 100 km and a northeast–southwest scale of about 250 km. It was located about 200 km east of the cold front. CS2 started to develop rapidly after 1200 JST when it moved east of Nagoya (Fig. 10a).

a. Kamogawa storm

The storm labeled A in Fig. 10b, hereafter referred to as storm A, later spawned the Kamogawa tornado. It was located about 100 km southeast of CS2 at 1500 JST (Fig. 10b) in the third convective system.

The embryo of storm A can be traced back by Mount Fuji radar to a weak echo that was located 120 km southeast of Shionomisaki (see Fig. 10a) at 1100 (not shown). This echo evolved into several weak cells after 1100, but was eventually organized into a single isolated cell by 1319 when it was 80 km south of Hamamatsu (Fig. 10a). It continued to move in a straight line northeastward and intensified (Fig. 10b). It passed over the Izu-Ohshima (Island) at around 1637 (Fig. 10c), producing 1.0-cm-diameter hail.

Storm A reached the Boso Peninsula at around 1720, 4 h after it was organized as an isolated storm. Within the following 2.5 h, it continued to move straightforwardly northeastward and caused damage due to hail, heavy winds, and a tornado along its path in the Chiba Prefecture. At 1737 JST a weak-echo region (WER) started to be identified clearly on Haneda radar in the southwest quadrant of storm A (Fig. 10d). The WER was observed from 1737 to 1742 and from 1800 to 1840 by Haneda radar, from 1800 to 1930 by Narita radar, and from 1734 to 1920 by Tokyo (Kashiwa) radar.

Storm A spawned a tornado in Kamogawa (Fig. 10d) at about 1747. It had crossed the Boso Peninsula by 1840 (Fig. 10e) and moved over the Pacific Ocean. However, it approached Choshi (Fig. 10f) at around 1940 and caused damage due to hail and heavy winds. Narita radar detected a hook-shaped echo pattern accompanied with the storm at 1940 JST (see the tip of the arrow in Fig. 10f). As is evident from Fig. 10, storm A lasted for more than 6 h and was an isolated echo throughout most of its lifetime.

b. Mobara storm

The storm labeled B in Fig. 10b, hereafter referred to as storm B and located about 50 km south of CS2 at 1500 JST (Fig. 10b), produced the Mobara tornado about 4 h later. The initial echo that later developed into storm B can be traced back by Mount Fuji radar to 1146 when it was 70 km south of Shionomisaki. Storm B also moved northeastward and was always located in the southwest of CS2. However, it remained isolated, about 50–70 km from the southwest end of CS2 until 1700, when it crossed the Izu Peninsula. After 1700, it started to interact with storm C in the north (see Fig. 10d). However, it kept its identity and reached the Boso Peninsula at around 1815. Storm C and storm D in CS2 (see Figs. 10d,e) started to dissipate after 1840. Storm B remained strong and caused damage not only by heavy winds in Futtsu and Kimitsu from 1830 to 1850, but also by the tornado in Mobara from 1913 to 1920 JST. Hail was also reported in Futtsu, Kimitsu, and Mobara.

6. Structures of the parent storms

a. Kamogawa storm

Storm A passed over the Tateyama Observatory of JMA located at the southern tip of the Boso Peninsula
Fig. 10. Precipitation intensity distribution of storms A and B observed by Mount Fuji, Tokyo, Haneda, and Narita radars. The times and radar sites are shown in the upper corner box. The vertical and horizontal lines of each box are oriented to the north-south and east-west directions, respectively. The precipitation intensity for (a)–(c) is categorized into four discrete levels according to the notation shown in the upper-right corner of (a), and that for (d)–(f) according to the notation in (d). The location of each box may be determined by consulting Fig. 1.
time of the sharp temperature decrease. This shows that the mesocyclone was located in cooler air behind the gust front in agreement with previous observations (e.g., Ward 1964).

As described in section 4a, storm A began to show a WER in the southwest part of the echo at 2 km AGL on Haneda radar after 1737 (see Fig. 10c). The heavy wind damage track in Maruyama, about 10 km southwest of Kamogawa (see Fig. 1), coincided with the path of the WER.

Figure 12 shows the surface airflow around storm A at 1800. Time-to-space conversion (Fujita 1963) was used to obtain the spatial distribution of the wind, using a uniform 16 m s⁻¹ translation speed for storm A toward the direction of 56.5°. A WER is located just to the east of the meteorological observation site of the Tokyo University Forest (hereafter referred to as TUF). The flow around the WER rotates cyclonically, suggesting the existence of a mesocyclone near the WER.

The Kamogawa tornado was on the ground from 1747 to 1758 JST, as was described in section 2a. To investigate the relative distance between the damage path and that of the WER, the echo pattern observed by Haneda radar at 1800 was shifted southwestward to a position the storm would have occupied at 1754, assuming no change in the echo pattern and using the assumed storm motion (above). Figure 13 clearly shows that the tornado moved almost in the same direction as storm A did and that the tornado was located near the northeastern end of the WER, where the position of the tornado at 1754 JST was assumed to be near points B and D in Fig. 2 (see section 2a). Large hail was observed on the northeastern side of the WER, in agreement with Browning’s (1964) conceptual model of a classical supercell storm.

at around 1730 JST (see Fig. 10d). Figure 11 shows the traces of temperature, dewpoint temperature, pressure, and precipitation rate at the observatory. The precipitation started at 1650, gradually intensified after 1710, reached the maximum intensity of 28 mm h⁻¹ at 1726, and eventually ceased by 1740 JST.

The temperature remained almost constant (17°C) until 1710 when it started to decrease gradually, corresponding to the intensification of the precipitation. It started to decrease sharply after 1722 and attained a minimum of 13.1°C at 1732. It then gradually recovered to 18°C. The dewpoint temperature increased gradually from 1600, and humidity became fairly high after 1720 JST.

The pressure had a general tendency to decrease in response to the approach of the subsynoptic-scale extratropical cyclone (see Fig. 9). However, it showed a sharp drop of about 2 hPa after 1720 and a sharp rise after 1731. This indicates the passage of a mesolow (and the corresponding mesocyclone) with a diameter of about 13 km, assuming the storm speed of 16 m s⁻¹. The minimum pressure occurred at 1728 JST, which is 2 min after the most intense precipitation and at the

FIG. 11. Traces of (a) temperature T, dewpoint temperature Td, pressure P, and (b) precipitation rate at the Tateyama Observatory of JMA. The period of the precipitation is indicated by the bar near the top of (b). The numbers at the top of (b) show total precipitation in the last 1 h.

FIG. 12. The precipitation intensity of storm A at 2 km AGL derived from Haneda radar at 1800 JST, together with the wind data at Kamogawa and Sakahata AMeDAS stations, Katsuura Observatory of JMA, and TUF. One barb corresponds to 2 m s⁻¹. The contour lines for the precipitation intensity are drawn for 1, 4, 16, and 64 mm h⁻¹. Large solid circles depict the locations of the stations where the wind data were obtained. The two AMeDAS stations and Katsuura Observatory recorded 10-min mean winds, and TUF recorded 6-min mean winds. The arrow in the lower-left corner shows the direction of the storm motion.
Fig. 13. Hail and tornado damage path together with the precipitation intensity distribution at 1754 JST, extrapolated from that at 1800 JST by assuming a storm translational speed of 16 m s⁻¹. The tornado damage path in Fig. 2 and that in TUF are shown by the broad lines, and the area with damaging hail of 2–3 cm in diameter is shown by the shaded region. The contour lines for the precipitation intensity are shown by the solid lines. The arrow in the lower-left corner indicates the direction of the storm motion.

Shown in Fig. 14 is an interesting pressure record taken at the Kamogawa EPOS station (point B in Fig. 2), nearby which the tornado passed over. The barometer at the station reports the pressure value digitally to a computer at the Earthquake Prediction Information Center of JMA for every 10 s after the original value is time filtered and analog-to-digital transformation is made. The time filter that prevents aliasing is a second-order Butterworth low-pass filter with a cutoff frequency of 20 s. The analog-to-digital transformation employs the successive approximation technique. The pressure began to dip sharply at 1744 JST and, after having experienced the first minimum of 1.4-hPa dip at 1749, recovered nearly to its original level at 1752. Assuming a translation speed of 16 m s⁻¹ for storm A, the diameter of the mesolow is estimated to be 9 km. The pressure then started to decrease again very abruptly from 1754 due to the approach of the tornado. Unfortunately, the record was discontinued for 2 min and 28 s because of an electric power supply failure.

There have been several reports that a mesolow preceded a pressure fall due to a tornado. Omoto (1982), in his study of the Nagoya tornado on 4 September 1979, found that the tornado, which moved northward, was located about 3 km south-southwest of the center of the mesolow. Barnes (1978) analyzed the Oklahoma City storm and found that a mesolow proceeded about several kilometers ahead of a mesocyclone (defined by the circulation center). Davies-Jones (1985), using a simple theoretical flow model of an updraft rotating at midlevels and a veering environmental wind, showed that the pressure field around a mesocyclone consists of three components: the pressure field due to a tornado, which is likely to be located near the center of the mesocyclone circulation; the nearly axisymmetric pressure field due to a mesocyclone; and the pressure field due to the interaction of the environmental wind with the storm inflow. The Davies-Jones model predicted that the mesolow is collocated with the maximum storm-relative surface wind but that it does not contain a mesocyclone at low levels.

Figure 15 shows the hodograph at 1500 JST at Hachijou Island, about 200 km southeast of the storm (see Fig. 10a,b). The wind at 1 km AGL was southwesterly, and there was a fairly strong westerly shear between 4 and 10 km AGL. The propagation vector of storm A is shown by the arrow and almost coincides with the wind vector at 5 km AGL. A plot of the locations of storm A from 1100 to 1940 on a map (not shown) reveals that the storm moved in a straight line northeastward even after it entered a supercell stage. This behavior differs from that of the classical right-moving supercells. However, since no sounding of winds was available in the near-storm environment except at Hachijou Island, discussions concerning the relationship between the storm motion and the environmental shear are not possible.

As described in section 4a, storm A had crossed the Boso Peninsula by 1840 JST, continued to move northeastward, and reached Choshi City at around 1940 (see Fig. 10f). It produced damage due to heavy winds in Choshi. Figure 16 shows a time–height cross
section of the maximum vertical vorticity in storm A, estimated from the Doppler velocity data, where the vorticity was calculated from the difference between the approaching and receding velocity maxima and the distance between the locations of the maxima. A mesocyclone was detected by the Doppler radar in storm A from 1840 to 1904 and from 1932 to 1953 JST. The maximum vertical vorticity amplified from 1840 and attained $2 \times 10^{-2}$ s$^{-1}$ between 2 and 3 km AGL from 1855 to 1902. It then started to dissipate and became less than $1.0 \times 10^{-2}$ s$^{-1}$ after 1904. After 1932 the vertical vorticity again increased markedly near 1 km AGL and reached $3 \times 10^{-2}$ s$^{-1}$ at 1941, the time around which damaging winds occurred in Choshi. This seems to indicate that the mesocyclone experienced two cycles of amplification and dissipation from 1840 to 1941. However, since the distance between storm A and the Doppler radar was between 70 and 90 km and the resolution in the cross-beam direction was more than 2.0 km, it was not possible to examine if cyclic generations of mesocyclones as found by Brandes (1977), Burgess et al. (1982), Jensen et al. (1983), and Johnson et al. (1987) occurred in storm A. Figure 16 also seems to show that the mesocyclone that produced the Kamogawa tornado at around 1800 did not retain its vorticity by 1840 JST. Thus, there is a possibility that the mesocyclone in storm A experienced one additional cycle of amplification and dissipation before and after the Kamogawa tornado.

The reflectivity data at 1941 JST are superposed on the Doppler velocity data in Fig. 17a. A hook-shaped echo is seen to the southwest of Choshi. This feature is more evident in Narita radar data at 1941 (Fig. 10f). A strong shear associated with the mesocyclone is located near the hook-shaped echo. It is therefore likely that the observed strong wind was associated with a tornado, though no residents witnessed a funnel cloud. House damage in Choshi is found in a zone of 2.5 km in length and 500 m in maximum width (Fig. 17b), though it looks discontinuous over rice farms where there are no houses.

Figure 17b also shows the distribution of damage to farm crops due to hail around Choshi. The hail damage is located on the northwest side of the hook-shaped echo. Hail was also reported in downtown Choshi, but its detailed distribution is not known.

b. Mobaara storm

The Doppler radar detected a region of strong shear, indicative of a mesocyclone, in storm B at 1840 JST, nearly the time when damaging winds occurred in Futtsu and Kimitsu. The mesocyclone was tracked from 1840 to 1924 between 1 and 3 km AGL by the Doppler radar. Figure 18 shows the distributions of the reflectivity and Doppler velocity at 1902, 1909, and 1917 at 1 km AGL. Since the distance between Mobaara and MRI is about 70 km, the Doppler radar resolution
is about 1.7 km both in the vertical and cross-beam directions around Mobara, and 0.5 km in the direction of the beam.

The Doppler velocity field at 1902 JST (Fig. 18a) shows a cyclonic shear region extending in the north-south direction. The width and length of the shear region are 2 and 7 km, respectively. The maximum in the approaching velocity in the east side of the shear layer is evident. However, its minimum in the west side is difficult to identify, because the true minimum was in the region of weak reflectivity where no Doppler velocity data were obtained. Therefore, for this particular time, it was assumed that the center of a mesocyclone (hereafter referred to as mesocyclone B1) was located in the shear layer. The locations of the maximum and minimum in the approaching velocity (shown by solid circles in Fig. 18a) were estimated subjectively by eye. Though the resolution in the cross-beam direction may have prevented the identification of a hook-shaped echo, the reflectivity did reveal a southwestward protuberance. This seems to be consistent with the advection of raindrops due to weak northerly wind relative to the storm in the western side of the shear layer. The center of mesocyclone B1 at 1902 was located about 12 km southwest of the south edge of the Mobara tornado’s damage path, which is located at the top right edge of Fig. 18a.

Figure 18b shows the distributions of the Doppler velocity and reflectivity at 1909 JST together with the tornado damage path. The tornado started at 1913 at the southern end of the damage path. In addition to the mesocyclone B1 at about 3 km west-southwest of
the southern end of the damage path, there is another pair of maximum and minimum in the approaching velocity at about 3 km southwest of mesocyclone B1. The maximum is 31 m s\(^{-1}\) and the minimum is \(-4\) m s\(^{-1}\). The Doppler velocity distribution around the pair indicates a superposition of convergent and rotational flows, the latter of which shows the existence of a mesocyclone (hereafter referred to as mesocyclone B2). Furthermore, the third mesocyclone (mesocyclone B3) can be found about 9 km west-northwest of mesocyclone B1. These mesocyclones moved north-northeastward between 1909 and 1917 JST.

As described in section 2b, the tornado was located approximately at the midpoint of the damage path at 1917 JST (see Fig. 18c). Figure 18c then indicates that the center of mesocyclone B1 at 1917 was located about 4 km north-northeast of the tornado center. Near the location of the tornado a pair of maximum and minimum in the approaching velocity is found, which corresponds to mesocyclone B2. The pair suggests that the flow consists of rotation and convergence. The Doppler velocity distribution at 1.6 km (not shown) shows an almost purely rotating flow in mesocyclone B2 with the velocity difference of 33 m s\(^{-1}\) over a distance of 1.7 km in the cross-beam direction. The center of mesocyclone B2 almost coincides with the tornado center (Fig. 18c). Thus, the tornado evidently formed from mesocyclone B2. The reflectivity map at 1917 near mesocyclones B1 and B2 showed no prominent features such as a hook-shaped echo pattern. Mesocyclone B3 at 1917 was at 7 km west-northwest of mesocyclone B1. Mesocyclones B1 and B2 moved northeastward after 1917 while dissipating. They can be identified on the Doppler velocity data at 1724 but not at 1731.

The diameter of mesocyclone B1 and the location of its center at successive times are shown in Fig. 19. Several witnesses said that the damage in Futtsu and Kimitsu occurred between 1830 and 1850 JST. The location of the mesocyclone at 1841 was close to the damaged area. The mesocyclone moved between north-northeastward and east-northeastward at a speed of about 14 m s\(^{-1}\). Its speed increased markedly between 1903 and 1910, then slowed down after 1910. The damage path due to the tornado is fairly close to the path of the mesocyclone. The mesocyclone moved its path from northeastward to north-northeastward at around 1910, and the tornado moved north-northwestward. The speed of the mesocyclone from 1910 to 1917 was 16 m s\(^{-1}\), which coincides with the translational velocity of the tornado estimated in section 2b (Niino et al. 1991; Niino et al. 1993). The diameter of mesocyclone B1 was relatively small and its vorticity was large when the tornado or the damage due to heavy winds occurred, in agreement with previous observations (e.g., Lemon et al. 1978).

As was described above, the tornado formed from mesocyclone B2, which was located about 3 km south-southwest of mesocyclone B1. However, the movement of mesocyclone B2 is not shown in Fig. 19, since its location was too close to that of mesocyclone B1 on this scale and mesocyclone B2 followed a path almost parallel to that of mesocyclone B1.

The open circles in Fig. 19 show the locations of barometers. The pressure records at Futtsu, Kimitsu (Kimitsu Agricultural High School), Kamo, and Nagara experienced a sudden drop of about 2 hPa at about the time when mesocyclone B1 passed near the barometer.

Mesocyclone B1 developed almost simultaneously after 1847 JST over the depth of 1–5 km (Fig. 20). Then, a rapid intensification of the vorticity at around 2 km occurred. This behavior of a mesocyclone has been observed for the Del City storm of 20 May 1977 by Johnson et al. (1987). The vorticity exceeded 3 \(\times 10^{-2}\) s\(^{-1}\) at 1903, 10 min prior to the tornadogenesis. Mesocyclone B2 started to develop at around 1910.

Figure 21 shows the surface wind flow and the temperature distribution derived from surface meteorological data at 1900. In addition to the observed winds at 1900, winds generated by the time–space conversion
technique (Fujita 1963) from 1850 to 1910 are used to draw the streamlines, using the assumption that the storm moved from 236.5° at a speed of 16 m s⁻¹.

The isotherms were drawn based on the AMeDAS data. Since the altitudes of the AMeDAS stations in the figure are all less than 120 m, the data were not adjusted to a common level such as sea level. Figure 21 shows a fairly steep north–south gradient of temperature over the south part of the Boso Peninsula.

The wind field at 1900 JST clearly shows a distinct convergence line where the warm southerly wind meets the cold northerly wind. The southwest end of the convergence line nearly coincides with the location of mesocyclone B1 at 1902 identified in Fig. 18a.

Figure 22 shows the areal distribution of precipitation intensity derived from Tokyo radar. The position of the convergence line in Fig. 21 nearly coincides with the southeast edge of the region where the precipitation rate is greater than 16 mm h⁻¹. This suggests that the convergence line was produced by precipitation-induced cold air spreading southeastward against relatively warm southerly wind. A region of relatively weak precipitation existed at the southwest end of the convergence line and seems to have corresponded to the location of mesocyclone B1.

The fact that the mesocyclone existed on the southwest edge of the convergence line and the mesocyclone moved along the convergence line seems to be consistent with the hypothesis by Klemp and Rotunno (1983) that the baroclinically generated horizontal vorticity along the convergence line is tilted up and stretched by the storm’s updraft to form a mesocyclone. In the present case, the horizontal temperature gradient across the convergence zone is at least 2 K (10 km)⁻¹ from Fig. 21. However, since the isotherms were drawn based on temperature data at AMeDAS stations, the actual temperature gradient may be confined to a narrower zone and have a larger value (Johnson et al. 1987). Figure 21 shows that most of the streamlines toward the convergence zone had an easterly component or at least an almost vanishing east–west component relative to the ground. Since the storm was moving northeastward at a relatively high speed (16 m s⁻¹), the storm-relative flow along the convergence zone was toward the mesocyclone. If an air parcel moved southwestward relative to the storm at a speed of 16 m s⁻¹

![Storm motion](image)

**Fig. 21.** Surface wind and the surface temperature distribution at 1900 JST 11 December 1990. The thick solid lines with arrows show the streamlines. The surface winds at AMeDAS stations and local governmental stations at 1900 JST are shown, where one barb corresponds to 2 m s⁻¹. The double solid line shows the surface wind convergence line, and the dashed lines are isotherms in degrees centigrade. The solid arrow in the lower-left corner shows the direction of the storm motion.

![Radar echo](image)

**Fig. 22.** The radar echo on Tokyo radar at 1900 JST 11 December 1990. The weak-echo region is located 7.5 km (three grid squares) north of the tip of the arrow labeled “mesocyclone.” The double line indicates the convergence line in Fig. 21.
along the convergence zone (whose length was 12 km; see Fig. 21), the baroclinic generation of vorticity during its travel from the east edge of the convergence zone to the west edge would be about $8 \times 10^{-3}$ s$^{-1}$. The observed vertical vorticity at 2 km AGL increased by $2 \times 10^{-2}$ s$^{-1}$ during 12 min from 1850 to 1902 JST. This would require the vertical velocity gradient of $3.5 \times 10^{-3}$ s$^{-1}$ in the horizontal if the convergence term in the vertical vorticity equation is neglected. If the horizontal temperature gradient had been underestimated by a factor of 2, a vertical velocity change of 9 m s$^{-1}$ over a horizontal distance of 5 km is required. This value of horizontal gradient of the vertical velocity is not uncommon. Thus, the observed vertical vorticity change could be explained in terms of tilting of the baroclinically generated horizontal vorticity along the convergence zone.

Cyclic evolution of mesocyclone cores has been documented by several authors (e.g., Brandes 1977; Burgess et al. 1982; Jensen et al. 1983; Johnson et al. 1987). Johnson et al. (1987) analyzed the Fort Cobb storm on 20 May 1977 and found two maxima in vertical vorticity: one associated with the major mesocyclone, and the other with a newly developing mesocyclone at the leading edge of the gust front. They suggested that convergence is likely to produce the necessary vertical vorticity amplification prior to tornado formation in the vorticity-rich air along the gust front. As shown in Fig. 18, mesocyclone B2 was accompanied with convergence in addition to rotation. Thus, it may be possible that mesocyclone B2 was generated along a gust front in a way similar to the second mesocyclone in the Fort Cobb storm. However, since we do not have data concerning vertical velocity distribution, detailed discussion on this generation mechanism of the new mesocyclone is not possible.

7. Summary and concluding remarks

The characteristics of the tornadoes in Kamogawa and Mobar and their parent supercell storms, which hit Chiba Prefecture on 11 December 1990, were analyzed using data from a damage survey, surface stations, a single-Doppler radar, and four conventional radars.

The Kamogawa tornado moved northeastward, leaving a damage path of 8.4 km in length and 2.0 km in maximum width. Another damage path of 3 km in length and 400 m in maximum width was found in the Tokyo University Forest about 10 km northeast of the damage path in Kamogawa.

The Mobar tornado moved north-northeastward at the speed of 16 m s$^{-1}$. Its damage swath was 6.5 km long with a 1.2-km maximum width. The maximum wind speed near the ground, estimated from distortion of a road sign, was 78 m s$^{-1}$. A sudden widening of the damage path and a discontinuity in the tornado path were found.

The Kamogawa storm exhibited many features similar to a classical supercell storm (Browning 1964; Doswell et al. 1990): it had an isolated structure; it lasted more than 6 h; and it had either a WER or hook-shaped echo pattern in the southwestern edge for about 2 h. Different from the classical supercell, however, it always moved in a straight line northeastward throughout its life from initiation period to the mature supercell phase. A mesocyclone circulation was detected from surface wind data. A surface pressure trace taken about 20 km southwest of the point of the tornadogenesis showed a mesoslow, which was located on the cool side of the gust front. The major precipitation preceded the mesocyclone. Another pressure trace in the tornado damage path showed that a mesoslow preceded the pressure dip due to the tornado by 5 min, indicating that the storm-relative inflow may have helped reduce the pressure in the forward side of the mesocyclone (Davies-Jones 1985). The same storm later hit Choshi where it produced damaging winds. The single-Doppler radar detected a mesocyclone, the vertical vorticity of which exceeded $3 \times 10^{-2}$ s$^{-1}$ at about 1 km AGL right before the damaging winds. At the same time, a hook-shaped echo was observed both by the Doppler radar and the Narita radar. At least two cycles of amplification and weakening of vorticity in the mesocyclone were observed by the Doppler radar.

Three mesocyclones were detected in the Mobar storm by the Doppler radar. The most prominent mesocyclone was first detected at 1 km AGL at least 33 min prior to the tornadogenesis and was traced for 44 min. In Futtsu and Kimitu, several houses were damaged by heavy winds when the mesocyclone passed nearby. Sudden pressure drops were observed along the path of the mesocyclone when it passed near barographs. Analysis of the surface wind field showed that the mesocyclone was located at the southwest end of a convergence line that formed between the warm southerly wind and the cold northwesterly air flowing out of the region of intense precipitation. A region of weak radar reflectivity also seemed to correspond to the location of the mesocyclone. A detailed analysis revealed that the tornado was accompanied by a new mesocyclone that started to develop at the south-southwest edge of the major mesocyclone 4 min prior to the tornado touchdown. The center of the new mesocyclone determined from the Doppler velocity data almost coincides with that of the tornado location determined from the damage survey.

As was mentioned in the Introduction, classical supercell storms have been rarely observed in Japan. The Kamogawa storm provided one of the precious examples of a nearly classical supercell observed in Japan. Observations of tornadic storms by a Doppler radar in Japan started only in late 1980s. Although eight tornadoes have occurred within the detection range of MRI Doppler radar since 1990, the Doppler
radar was operating only for six tornadoes. Five of them were accompanied with mesocyclones identified by the Doppler radar. Before the Doppler radar observations began, the only means to identify supercell storms were conventional radars and surface meteorological instruments. This implies that only classical supercell storms are likely to have been identified. Recent studies (e.g., Doswell et al. 1990) revealed that supercells can have a wide spectrum. Especially some of the characteristics that were believed to be typical of classical supercells (e.g., a long-lived, steady-state cell that propagates to the right of the mean wind) may not necessarily apply to all classes of supercell storms. In fact, the Kamogawa storm studied in this paper exhibited many similarities to classical supercells but did not show a rightward deviation of its movement. As the number of the Doppler radar observations increases, the documentation of supercell storms in Japan may increase in number. It would be of interest then to examine the observed structure of the supercell storms and obtain statistics concerning the types of the supercells (i.e., a classical supercell, a low-precipitation supercell, or a high-precipitation supercell).

The atmospheric environment that produces tornadoes in Japan is not well known, because of sparsity of upper-air soundings with respect to time and space (as was the case in the present study) and the low frequency of tornadoes. A study on the relationship between the environmental parameters (e.g., shear, convective available potential energy, and a bulk Richardson number) and the types of the storms observed in Japan is certainly desired. It would then become possible to discuss, in terms of physical parameters, the characteristics of tornado storms in Japan in comparison with those in the United States.

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