Numerical Modeling on A Hazardous Microburst-Producing Hailstorm

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Abstract A hazardous downburst-producing severe thunderstorm in Saitama Prefecture of Japan was reproduced by a three-dimensional cloud model in which hail/graupel is divided into 21 size categories . The observed characteristics of the storm such as overhang structure, descending precipitation cores, intensity of divergent wind velocities and cooling process near the surface were well simulated with a grid resolution 0.5 km. In addition, the development of multiple microbursts was also captured. The mechanism of the downburst formation process was investigated based on the simulation and it is found that the formation of hail/graupel was important to the production of the microbursts : the microbursts were initiated by hail loading above the melting level and were enhanced primarily by cooling due to the melting of hail/graupel and secondarily by evaporative cooling of rain. The microburst outflow was produced by the increasing of horizontal pressure gradient near surface due to the descending of cool air and was further intensified by surface rain evaporation. An existence of relatively dry middle ambient layers and obvious overhang structure caused by vertical shear is found to be favorite to the production of the microbursts in a large quantity hail-producing thunderstorm.
Key Words Microburst, Numerical Simulation, Hailstorm

1. Introduction

As one of atmospheric phenomena, microburst is of particular concern due to its hazardous strong downdraft and low-level divergent wind shears which have been shown to be responsible for some of the previous aircraft accidents (Fujita and Caracena 1977; Fujita, 1980). After investigating the crash of an aircraft at New York City's John F. Kennedy (JFK) Airport, Fujita (1976) proposed the term “downburst” to describe the wind which affected the airport. It means that a strong downdraft which induces an outburst of damaging winds near the ground .

Because of various temporal and spatial scales of downbursts, they are subdivided into “macroburst” and “microburst”. A microburst is a small downburst with its outflow diameter smaller than 4 km and peak winds lasting only 2 to 5 min, while a macroburst has a horizontal distance between outflow peaks larger than 4 km . Depending upon the amount of precipitation measured at the ground, microbursts were further subdivided into either “wet” or “dry” (Fujita, 1985). Analysis of JAWS (Joint Airport Weather Studies) data has revealed that microburst outflows can be classified into two morphological types individual microbursts and microburst lines (Hjelmfelt and Roberts 1985; Hjelmfelt 1987). Organization of microbursts into microburst lines results in much longer lasting wind shear and creates a much greater potential for hazard to aircraft than an isolated microbursts. Roberts and Wilson (1986) have characterized microburst as either “low reflectivity” (less than 35 dBZe), “moderate reflectivity” (35-55 dBZe), or ”high reflectivity” (greater than 55 dBZe).

A number of field research programs such as the Northern Illinois Meteorological Research on Microburst (NIMROD) project (Fujita 1978, 1985), the Joint Airport Weather Studies (JAWS) project (McCarthy et al. 1982;Wilson et al. 1984), the FAA/Lincoln Laboratory Operational Weather Studies (FLOWS) project (Wolfson et al.1985), the Microbnrst and Severe Thunderstorm (MIST) project (Dodge et al. 1986), and the Convective Initiation and Downburst Experiment (CINDE) project (Wilson et al. 1988) have been conducted in United States to sample hundreds of microbursts with single Doppler radar and in many cases multi-Doppler radar data of the entire microburst-producing storm (e.g., Lee et al.1992ab;Hjelmfelt 1988, Lin and Hughes 1987; Lin and Coover 1988; Wakimoto et al. 1944; Knupp 1996). These projects have provided important observational findings and led to a greater understanding of downburst phenomena. These studies indicate that microbursts are common phenomenon, having been detected on 60% to 70% of the days during which thunderstorms occurred.

The progress of modeling study on downburst has been limited due to requirements of very large memories of computer for a three-dimensional model with a proper grid resolution of downburst. The models used to study microbursts can be generally categorized into two types: subcloud models , in which some sort of forcing is imposed in an elevated region of the domain (e.g., Teske and Lewellen 1977; Srivastava 1985; Proctor 1988,1989;

Anderson et al. 1992; Orf et al. 1996,1999), and full cloud models, in which the life cycle of a microburst-producing storm is modeled (e.g., Tuttle et al. 1989; Knupp 1989;Hjelmfelt et al. 1989; Straka and Anderson 1993; Parsons and Weisman 1993; Proctor and Bowles 1992; Guo et al.1999) .

Cloud models with detailed cloud microphysics are useful to reproduce the full life cycle of a microburst- producing storm and to illustrate the mechanisms of initiation of downburst from dynamical and microphysical processes and their interactions, as well as the effect of environment on formation of microburst. In contrast, subcloud models are used to study the detail structure of microburst itself with higher resolution by relying on some sort of a priori forcing to initialize the microburst rather than the whole storm.

Alahyari and Longmire (1995) simulated microbursts in a laboratory tank by releasing a dense volume of fluid into a less dense ambient fluid.

Although some significant progresses have been made since the creation of the term of downburst, the real mechanisms of downbursts are still unclear, especially for wet downbursts. A distinguishing characteristic of downburst is the continued intensification of downdraft as it descends. The continued intensification of downdraft was thought to be produced mainly by negative buoyancy induced by melting, evaporation and loading of hydrometors under certain circumstances. So the studies on mechanisms of downbursts might be transformed to the studies on the interactions among buoyancy fluid, ambient fluid and hydrometors. Obviously, the stratification state of ambient air such as the distribution of temperature and humidity with height and the type, amount and size of hydrometors dominate the buoyancy fluid.

Some researchers (e.g. Hookings,1965; kamburova and Ludlam, 1966; Lndlam, 1980) have studied the maintenance of downdrafts by the evaporation of falling precipitation as a function of drop size, rain intensity and downdraft speed. An important conclusion of them is that the intensity of rainfall and steepness of environmental lapse rate play an important role in the production of strong downdrafts. However, if the environmental lapse rate was approximately equal to the dry-adiabatic lapse rate, then the microphysics of evaporation have little restriction on downdraft magnitude, and even in a moderate or light rain, strong downdrafts may be generated.

Evaporative cooling of precipitation is thought as being the primary driving mechanism for most dry microbnrsts. And dry microbursts typically occur in an environmental conditions with a dry boundary layer and a deep dry adiabatic profile that extends to about 3 km AGL or higher (Wakimoto 1985). Fujita (1983b) analyzed a wet microburst which struck Andrews Air Force Base, when the recorded peak horizontal wind speeds were greater than 67 m/s. The related vertical wind speeds could not be attained by negative buoyancy alone. Thetemperature profile of nearby sounding was close to the moist-adiabatic lapse rate from the surface to near the tropopause. So for wet microbursts, the dominant physical mechanisms are still not so clear since it seems to be a result of several forcing mechanisms. But it is believed that except for the cooling of precipitation, the wet microburst intensity may also be a function of either upper-level momentum descending to lower levels (Forbes et al. 1980) or strong motions produced by dynamically-induced pressure gradients (Wolfsen, 1983).

The current study was inspired by the need for a more complete description of cloud microphysics than could be offered by existing numerical models in the study of reproducing of microburst- and hail-producing storm as well as the mechanism of wet downbursts. Srivastava's one-dimensional simulations showed that the intensity of downdraft increases with decreasing the size of precipitation particles. Many studies have shown that the rain formed in severe hail-producing thunderstorms mainly comes from the melting of hail and graupel. Downburst-producing storms were always observed to be accompanied with hailfall at the surface. The loading of hail and cooling from melting of hail and graupel are found to be important to the formation of intense downdraft. So the size distribution of hydrometeors like hail/graupel and rain might have some important effects on the formation of intense downdraft. Therefore, a hail/graupel category cloud model was used in this study in order to overcome some too general descriptions of size distribution of graupel and hail (Cheng and English, 1983; Guo etal. 1998) .

A severe long-lasting thunderstorm associated with hail and a strong gusty wind passed over the Gunma and Saitama prefecture on September 8 1994 in Japan which was documented in detail by Takayama and Niino et al.(1997)(hereafter referred to as TN97). The storm produced great damage to window glass in the Misato Junior High School and injured 2 teachers and 71 students. Based on the analysis of available data observed during that time. downbursts were identified and should be responsible for producing this damaging wind.

2. MODEL DESCRIPTION

A three-dimensional, nonhydrostatic hail/graupel category cloud model (Guo ,1997; Guo et al. 1999) is used for the present simulations. The coordinates of the model are the standard Cartesian coordinates (*x,y,z*). The independent variables of the model are the velocity components *u,v* and *w* in the *x,y* and *z* directions, respectively, pressure *p*, potential temperature θ, mixing ratio of water vapor q_v, bulk cloud water q_c, bulk cloud ice q_i, bulk rain water q_r ,bulk snow aggregates q_s and ice concentration N_i, mixing ratio of discrete mass categories of graupel/hail $q_h(I)$ for I=1, L_h, where L_h is the number of graupel/hail categories.

2.1. Governing Equations

The model incorporates time-dependent, non-hydrostatic equations cast in compressible form similar to that proposed by Klemp and Wilhelmson(1978) as

D_u, D_v, D_w, D_θ and D_π are the turbulent fluxes of u,v,w,θ,and π ,respectively. Q_{mf},Q_{ce} and Q_{ds} are the latent heating/cooling terms due to melting/freezing,

$$\frac{du}{dt}+c_p\bar{\theta}_v\frac{\partial\pi'}{\partial x}=D_u, \tag{1}$$

$$\frac{dv}{dt}+c_p\bar{\theta}_v\frac{\partial\pi'}{\partial x}=D_v, \tag{2}$$

condensation/evaporation,and eposition/sublimation

$$\frac{dw}{dt}+c_p\bar{\theta}_v\frac{\partial\pi'}{\partial z}=f_w+D_w, \tag{3}$$

$$\frac{d\pi'}{dt}+\frac{\bar{C}^2}{c_p\bar{\rho}\bar{\theta}_v^{\,2}}\left(\frac{\partial\bar{\rho}\bar{\theta}_v\mu_j}{\partial x_j}\right)=f_\pi+D_{\pi'}, \tag{4}$$

$$\frac{d\theta}{dt}=Q_{mf}+Q_{ce}+Q_{ds}+D_\theta, \tag{5}$$

$$\frac{\partial q_x}{\partial t}=-D_{qx}+W_{qx}+I_{qx}+\frac{\partial}{\partial x_3}(\rho_0 V_{tx}q_x), \tag{6}$$

$$f_w=g\left(\frac{\theta'}{\bar{\theta}}+0.608q_v'-q_c-q_r-q_i-q_s-\sum_{I=1}^{L_h}q_g(I)\right), \tag{7}$$

produced by microphysical processes, Respectively. W_{qx} and I_{qx} are cloud microphiysical sink and

$$f_\pi=-\frac{R_d}{c_v}\pi\frac{\partial\mu_j}{x_j}+\frac{C^2}{c_p\theta_v^{\,2}}\frac{d\theta_v}{dt}. \tag{8}$$

source terms which are related to warm and cold rain processes. V_{tx} is the terminal velocity of a hydrometeor q_x, where q_x is one of the mixing ratios of water vapor q_v, cloud water q_c, rain water q_r ,cloud ice q_i , snow q_s , and hail/graupel category water content $q_g(I)$. In this paper. L_h was assumed to be 21. V_{tx} for cloud water was assumed to be zero and for cloud ice was considered to be the function of its mean diameter(Locatelli and Hobbs,1974).

Cloud ice number concentration N_i was also predicted based on a concentration equation in this study.

The model includes a conventional first-order closure for subgrid turbulence and a diagnostic surface boundary layer based on Monin-Obukhov similarity theory.

2.2. Microphysical Processes

In this study, the size distribution of hail/graupel is not prescribed, but is allowed to evolve naturally through mass category technique proposed by Berry (1967). In addition, considering the important role of ice crystal particles playing in cold cloud, an equation for number concentration of ice crystal is included besides the one for its mixing ratio. A detail of the microphysical parameters and constants used in the model can be found in Guo(1997).

Important microphysical processes included in this model are: the melting of snow and hail/graupel; evaporation of rain; the accretion of rain by snow and hail;the shedding of water from melting of snow and hail; the sublimation of water vapor from snow and hail/graupel and the evaporation of liquid water from melting of snow and hall/graupel. The terminal velocities for rain and snow are computed as mass mean-weighted values while hail/graupel with a certain diameter (or category) falls at a terminal velocity given by Wisner et al. (1972).

2.3. Numerical Method

The equations described in the previous section are solved numerically using finite-difference methods on a rectangular grid. The model variables are staggered using Arakawa C-grid system with scalars defined at the center of the grid boxes and the normal velocity components defined on the corresponding box faces.

Time integration of governing equations of the compressible atmosphere uses a conventional time-splitting technique (Klemp and Wilhelmson, 1978): a short time step 0.0625 seconds is used for acoustically active terms, while a large time step of 5 seconds is used for the remaining terms. The large time-step integration uses a standard second-order leap-frog scheme. To prevent separation of numerical solutions deduced by second-order frog-leap scheme, an Asselin (1972) time filter is used in the model.

With the exception of the advections terms which are fourth-order accurate, the spatial difference terms are second-order accurate. In addition. fourth-order and vertical second-order spatial filters in the horizontal and vertical directions are used for all variables except for pressure to damp grid scale noise due to nonlinear instability.

The lateral boundaries use a radiation boundary scheme suggested by Klemp and Wilhemson (1978). The rigid top and bottom boundary conditions with an upper boundary Rayleigh damping layer which absorb upward propagating wave disturbances and eliminate wave reflection at the top boundary are used in the model.

2.4. Model Initialization

The model atmosphere has an initial profile which was synthesized from the upper air soundings at Maebashi and Tateno stations (Fig.1). Maebashi station is nearer to the area of occurrence of downburst and is to be considered as the most representive environment for bearing this downburst-producing storm. Due to lacking of soundings of wind data in Maebashi. the wind data of Tateno at the same period was used in the simulation. The storm passed over the area between Gumma and Saitania prefectures at 0500JST 8 September 1994.

Fig.1 Skew T-logp diagram for temperature and dewpoint temperature observed 0500JST & September 1994 Japan

The domain size for the simulation was 36 km $\times$ 36 km in the horizontal and 19 km in the vertical. The grid interval was 500m in both horizontal and vertical directions. The simulation was initialized by a thermal bubble with size 8 km $\times$ 8 km in the horizontal and 2 km placed at the center of the model domain (Klemp and Wilhelmson,1978). The peak temperature perturbation at the center of the thermal bubble was 4 K (Klemp and Wilhelmson,1978).

2.5. Storm Environment

The microburst-producing storm occurred while a cold front originally extending from a low over the Sea of Okhotsk passed over the Japan Islands and moved toward the Pacific Ocean. During the moving of the cold front over the northern part of the Japan Sea, a line-shaped cloud area was detected by both an infrared image of GMS-4 and a Mt. Fuji radar after 1200JST on 8 September 1994. At 500 hPa level, a trough moved over the Japan Sea and accompaning cold-air intruded into the north and east parts of Japan in the afternoon of that day. The storm which produced hail and microbursts started to develop along the line-shaped cloud area. When the storm moved to the border of Gunma and Saitania prefectures, the lowest temperature at cloud-top reached about -65.0°C which corresponded to about 15 km AGL according to the sounding data at Maebashi. The damaging wind near the surface estimated from the general damage characteristics in that area reached more than 50 ms^{-1} .The storm was produced in the environment with the surface temperature of about 30°C and surface dewpoints of about 26.8°C. The temperature sounding is characterized by an extremely deep moist-adiabatic layer from . the top of boundary layer to near the tropopause. In the boundary layer, temperature lapse rate was about 7.0°Ckm^{-1} and lapse rates of water vapor mixing ratio were 4$gkg^{-1}km^{-1}$. The vertical profile of moisture exhibits a relatively wet atmospheric boundary layer and relatively dry middle layers. This moisture condition is opposite to that of dry microburst occurrence, and so it would be more conducive to wet-microburst generation. The ambient wind profile indicates variable winds below 2 km AGL, transitioning to southwest winds with speeds greater than 15 m/s above 3 km AGL. The mountain orography along Saitama area might help generate convective thunderstorm in moist sounding condition. The lifted condensation level(LCL) was near 950 hPa and the lifted indices

(LI) was -7°C. The atmosphere was in the most unstable condition with CAPE about 3243 $m^{-2}s^{-2}$ in the prior to the occurrence of the storm, and with a relatively strong vertical wind shear.

The present temperature sounding also includes a very typical upper-air inversion layer. It was found that the temperature profile prior to severe weather formation typically shows a temperature decrease with height from the surface to about 800 hPa followed by a temperature increase with height for a short distance and then a temperature decrease again. The presence of an upper air inversion generally suppresses the development of cloud until a thermal, or small cloud is able to break through the inversion layer. This delay frequently allows the resulting thunderstorm to grow much larger than ordinary because the stable inversion layer prevents initial thunderstorm development. The delayed cloud development lets the sunlight continue to warm the ground and lower atmosphere until a bubble of air is so much warmer than the air above the inversion layer that much more violent thunderstorm form.

3. Results

The simulation produced good quantitative and qualitative agreement with observations of the major features of the storm. Some of the comparisons are listed in Tables 2.

Table 2. A list of sounding parameters, and observed and simulated storm characteristics for this study

mulated (occurred time)	Observed	
Lifetime	Long-lived	Long-lived
Lifted index	-	-7.0
Cloud top (km AGL)	15.0	15.0
Cloud base (km AGL)	0.6	-
Melting level (km AGL)	4.5	-
Overhang	Yes	Yes
Rotation	Yes	-
Microbursts	Multiple	Multiple
Descending precipitation core	Yes	Yes
Max radar reflectivity (dBZ)	>=55	>=55
Max updraft (ms^{-1})	40.0 (16 min)	-
Max downdraft (ms^{-1})	-90.0 (18 min)	-
Max surface ΔV(ms^{-1})	105.0 (20 min)	-
Max outflow speed (ms^{-1})	52.0	50.0
Max surface pressure gradient (mb)	20.0 (18 min)	-
Time between max ΔV and max downdraft	2.0 (min)	-
Minimum surface temperature drop ΔT (℃)	–10.7	-10.0

3.1. Life Cycle of The Storm

Fig. 2 shows time evolutions of maximum updraft, minimum downdraft, maximum outflow velocity, maximum pressure and surface temperature drop. To easily understand the space structure of storm, a time sequence of specific water content of the storm is showed in Fig.3. A persistent and deep storm develops soon after the initial impulse is imposed. After 16 min, the simulated storm enters the mature stage with the maximum updraft strength of about 40 m/s shown in Fig. 2 (thick solid line) and at an altitude of 7 km above the ground. After reaching peak updraft strength, the simulated storm does not collapse immediately instead exhibits some degree of weak updraft redevelopment. This corresponds to the long-lived feature of observed storm. The main updraft, which generates much of the storm's precipitation, is at the rear of the storm and tilts due to strong vertical wind shear.

The precipitation first reaching the ground is rain at 14 min and then hail at 18 min. 2 min after the occurrence of maximum updraft. Strong downdraft occurs with the arrival of hall at the ground. The main precipitation shaft is located on the downshear side of the main storm (Fig.3). The first downburst with a peak downdraft -90 m/s (thin solid line in Fig2) and a peak divergent wind speed of over 50 m/s (dotted line in Fig.2) at the surface occurs accompanied with the strong descending motion of precipitation core (Fig.3). An obvious overhang, located between 2-9 km AGL, extends eastward from the main precipitation shaft (26 min in Fig.3) and creates a favorite condition for precipitation falling from it to get maximum melting and evaporation in a relatively dry middle layers. The storm enters a stage of weak updraft redevelopment after 22 min, and further tilts due to weakening updraft. Following 26 min, precipitation begin to fall from the overhang, reaching the ground at 30 min and producing a second downburst. But the second downburst is much weaker than the first one.

The simulated obvious descending process of precipitation core and the structure of overhang are well consistent with observations by the Mt. Fuji radar. Observations shows that the storm had a marked overhang in the direction of its movement and an obvious descent of reflecting core which was speculated to be caused by falling hail(TN97). The descending speed of precipitation core between 16 min and 18 min is about 2000 m / 120 s =16.7 m/s which is nearly equal to that observed by the radar.

Fig.2 Time series (min) of maximum simulated updraft velocity Wmax (m/s) (thick solid line), minimum downdraft Wmin (m/s)(thin solid line), maximum outflow velocity at the surface Umax (m/s)(dotted line) ,peak pressure deviation at the surface dtaP (mb)(long dashed line) and temperature drop at surface dtaT(short dashed line).Only temperature drop and minimum downdraft correspond negative vertical coordinate in the figure.

Fig.3 The evolution of mixing ratio of total hydrometeors (1.0 g/kg contour) for the simulated microburst-producing storm (light shaded area is 1.0 g/kg contour of total hydrometeor, deep shaded area indicates the hail larger than 1.0 g/kg)

The distribution of perturbation temperature in Fig.2 (short dashed line) shows that the maximum surface perturbation temperature is -10.7°C which is quite close to the observed value of - 10.0°C (TN97) . In the simulation, a dome of high pressure begin to develop at 16 min at the surface 2 min prior to the development of divergent outflows. The peak pressure deviation at the surface reaches near 20 hPa (long dashed line in Fig.2) and occurs behind 2 min of the time that the storm reaching the maximum updraft .

3.2. Downdraft Development

As noted above, a distinguishing characteristic of downburst-producing storm is its downdraft continued intensification as it descends. This nature is also obvious in present simulation. Fig.4 shows the time-height cross section of peak downdraft (dotted line) and the mixing ratio of hail (shaded area) in downdraft area. It shows that downdraft originates at the upper part above the 0°C level and intensifies rapidly as it descends after passing through the 0°C level. The size of low-level downdraft in the simulated storm was about 3-4 km in diameter. The simulated peak downdraft velocity of -90 m/s occurred in the layer between 1.5-2.0 km AGL at 18 min, 2 min after the storm attained maximum updraft (see the thin solid line in Fig.2).

Fig.4 also shows that the formation of the simulated downdraft well coincides with the descending hail core, and the arrival of the downdrafts at the surface with that of hail. The fact that the maximum downdraft located below the melting level means that the other processes such as melting and evaporation might also play important roles in forming the strong downdraft. The downdraft reaches maximum when the time-integrated forcing reaches maximum. It has been noted that the hail core became much weaker after passing through the 0°C level. So the peak value of downdraft caused by loading of hail above melting level is much less than that caused by cooling of melting of hail/graupel and evaporation of rain below melting level. In fact from the result of section 3.6, we will know that the maximum downdraft located at about 2 km AGL was mainly caused by hail/graupel melting process and second, by rain evaporation.

The formation of rain water in the downdraft region is mainly due to the melting of hail and the maximum rain water content located near surface will be very important to enhance the intensity of divergent wind speed through strong evaporational cooling process (Fig.5). In the updraft region, rain water can be formed through autoconversion process of condensed cloud water and grow up through the collection of cloud water. The rain drops will be falling into the downdraft region when their sizes are too large to be supported by the updraft. So there are also rain water above 0°C level.

Fig.4 The height-time distribution of maximum downdraft (m/s)(dotted line) and hail specific water content (kg/kg)(shaded area larger than 0.02 kg/kg)

The time evolution of space structure of downdrafts (less than -10 m/s) in Fig.6 shows a general process of downbursts development as well as tilting and multiple characteristics of the simulated downbursts. It shows that a weak downdraft forms in association with hail at the height of 8 km AGL at 16 min. With the growth of hail, the downdraft caused by loading of hail increases and descends with a Peak speed -40 m/s . After passing through the 0°C level. the downdraft is further intensified and attains its peak velocity about -90 m/s at 1.5-2.0 km AGL. As an indication of occurrence of first microburst. There is an outflow peak wind speed reaching more than 50 m/s at 18 min near surface. It takes about 4 minutes for the microburst to form and the maximum outflow wind speed lasts for 2 min until to 20 min. The downdraft has a diameter of about 3-4 km and shows an obvious tilting. After producing the first microburst. the downdraft intensity become weak after 20 min and enters a dissipating stage. At 24 min, the second downburst starts to develop, and after 6 min, it touches the surface with a relatively weak outflow wind speed. It should he noted in here that both of these downbursts are formed from overhang due to the tilting of the storm. So the relatively dry ambient air at the middle layer of the case has fundamental importance to the formation of the downbursts. The dryer ambient air creates a favorite condition to the melting of ice particles as well as the evaporation of rain. The occurrence of multiple microbursts also agrees with the observation (TN97). The dotted line at the surface is the temperature deviation which indicates the cooling process in association with downbursts.

Fig.5 The height-time distribution of maximum downdraft (m/s)(dotted line) and rain specific water content

3.3. Surface Outflows

In the simulation, the development of a dome of high pressure is closely related to the development of divergent surface outflows. Maximum high pressure at the surface occurred at 18 min corresponding to the time of the strongest downdraft. The maximum amplitude of the surface high pressure was about 20 mb (short broken line in Fig.2) and it is much larger than that previous cases of microbursts (Straka and Anderson,1993)

Fig.6 The evolution of downdraft velocities and temperature deviation (dashed line) at the surface for the simulated microburst-producing storm. The region where the downdraft is less than -10m/s is shaded, streamline is that near surface.

time=20(min) z=0(km)

Fig.7 The wind vector distribution associated with downbursts at the surface.

knupp (1989) and others have shown that microbursts are forced by strong horizontal pressure gradients near surface. Based on damage characteristics, the maximum wind speed was estimated to be about 50 m/s (FI) in present case (TN97). So the maximum amplitude of surface pressure should be much larger than that of ordinary cases. The simulated surface outflow was confined to the lowest one layer in the model (500m), and peak differential velocity also occurred in the same layer.

The structure and degree of symmetry in the simulated surface outflows can be seen in the surface wind vector fields in Fig.7. It shows that the simulated surface outflows are roughly symmetric at early stage, but become more asymmetric as it expands with time. Hjelmfelt (1988) found that some microbursts asymmetries might be apparent in nature and result from the superposition of a subcloud mean flow on a symmetric outflow. Straka and Anderson (1993) studied the correlation between the aspect ratio of surface outflows and vertical shear of low levels of the environmental winds and suggested that a vertical shear of the environmental winds, through the layers where hydrometeors form, might produce an asymmetry in a storm's low-level outflow. For the current simulation, a downshear tilting of downdraft due to vertical shear in the environmental wind can be seen clearly. So characteristic of asymmetry of the simulated downburst may also relate to the environmental vertical shear and the interaction between subcloud mean flow and tilting descending flow. A more detailed analysis of the asymmetrical surface outflows is outside the scope of this study.

A very strong convergent inflow is formed at middle level (2-6 km) of the storm in order to compensate the strong divergent outflow at the surface .

Fig.8 The vertical cross section distribution of deviation pressure in association with downburst. Solid lines are the positive deviation and dashed lines are negative deviation.

From the cross section distribution of pressure perturbation (Fig.8), it shows that the strong divergent outflow near the surface closely corresponds with the dome of high pressure at same level while the strong convergent inflow corresponds well with the low pressure region at middle level.

3.4. Radar Reflectivity

Hail category model can provide a unique condition to more accurately calculate the reflectivity factor Z caused by hail than that generally used in hail parameterization model. Because the reflectivity factor Z has been defined as the sixth power of the drop diameter summed over all drops in a unit volume and is thus given in terms of the particle size distribution. The exponential particle diameter distribution with parameters deduced by Marshall and Palmer was generally used in hail parameterization model.

The Marshall-Palmer size distribution poorly estimates the actrual particle size density at too small and too large particle diameters. The contribution to Z from the smallest and numerous particles is small, but the sixth power of diameter causes the fewer larger diameter Particles to be the most important contributors to Z. Therefore, if the exponential distribution fails to estimate the actual particle size density at large particle diameters, this should produce much error in Z. In contrast, the size distribution of hail particles is not prescribed, but is predicted through mass-category technique in hail category model.

The evolution of descending precipitation core in association with downbursts in the simulated storm is depicted in X-Z cross section plot of reflectivity factor Z (Fig.9).

Fig.9 The X-Z cross section distribution of radar reflectivity and wind vector distribution associated with downbursts.

Fig.10 The horizontal distribution (z=2 km AGL) of radar reflectivity and wind vector associated with downbursts (shaded area is larger than 55 dBZ).

Fig. 10 shows the horizontal distribution of radar reflectivities at 2 km AGL. At the initial stage, a weak bowshaped echo is first detected at low layer. At 18 min, a strong bow-shaped echo is formed and soon turns into the shape of spearhead at 20 min. Corresponding to the formation of the second downburst. there is an obvious change of radar reflectivity at 28 min, which is also consistant with two remarkable increase in width of damage observed.

Initially, there is a rapid increase in the intensity of radar reflectivity . With the development and descending of radar reflectivity core, the intense downdraft is produced. However. the maximum simulated reflectivities are 5-10 dBZ larger than observed. This may be caused by using different resolution for observation and simulation. The radar refiectivities only in 2.5 km resolution were given in observation. The locations of the reflectivity maximums in the simulations well corresponded to regions with larger mixing ratios of hail/graupel.

3.5. Hail Category Contributions

In the study of sensitivity of microbursts to different precipitation types, Srivastava (1987) suggested that the rapid cooling associated with the melting of small ice particles could be important in the production of intense downdrafts, especially when the atmosphere is stably stratified. And some researchers thought that large size ice particles could be more important.

To easily check the characteristics of hail/g-raupel size distribution in the downburst-producing storm, four sub-categories divisions are made based on general definition of hailstone sizes: graupel and hail embryos (1 <D<5*mm*), small hailstones (5<D<10*mm*),typical hailstones (10<D<25*mm*).and large hailstones (D> 25*mm*).

Mixing ratio and number concentration of hail/graupel in four sub-categories divisions at the time of downburst occurrence are shown in the Fig. 11. Although the water content of graupel/hail embryos (1 < D < 5*mm*) is small (0.2-1.8 $g\ kg^{-1}$), its number concentration is relatively very large (50-450 m^{-3}). And the particles distribute in a thick layer ranging from near surface to 8-10 km AGL. It is relatively small both in mixing ratio (0.2-0.8 $g\ kg^{-1}$) and number concentration (less than 2 m^{-3}) for small hails (5 < D < 10*mm*). But there are a large number of typical hails (10 < D < 25*mm*) with maximum value of mixing ratio and number concentration reaching up to over 20 $g\ kg^{-1}$ and 18 m^{-3}, respectively. The location of maximum value of typical halls well coincides with that of maximum downdraft. This is also well consistent with hail size observed at the surface. Large hails (D>25*mm*) mainly locate at the surface and the top of updraft area and occupy relatively small part.

Generally, graupel/hail embryos having a high number concentration and a low density will melt and sublimate faster than typical hail and large hail which have low number concentrations and high density. But typical hail and large hail due to their higher terminal velocities and particle density, as well as lower number concentration will tend to melt through a deeper layer than small hail and graupel. Further, being relatively inefficient at cooling the air through sublimation, downdrafts generated by typical hail and large hail are generally confined to levels below the melting level (Proctor,1988b). Thus, typical hail and large hail may produce stronger microbursts in more stable environments than other precipitation forms, since the downdraft is maintained only at lower elevations where it is less likely to be depleted of negative buoyancy due to compressional heating. However, in environments with deep, dry adiabatic lapse rates, snow and graupel may produce microbursts of greater intensity than those produced by typical and large hail, since the snow and graupel more effectively cool the air at higher altitude.

Fig11. The evolution of the simulated X-Z cross section hail category mixing ratio and number concentration associated with downburst on & September 1994 in Japan.

3.6. Microphysical Forcing of Down Drafts

Many researchers such as Srivastava (1985,1987),Krueger et al (1986), Proctor (1988,1989,1992), Knupp (1989), Tuttle et al(1989), Orville et al. (1989), Hjelmfelt et al. (1989), and Straka et al. (1993) have demonstrated that the importance of hydrometeors in the development of downdrafts through changes in buoyanc3y due to loading and heat loss associated with water phase changes.

It was shown from the studies by Orville et al. (1989) and Knupp (1989) that loading effects of hail were largest above the melting level, cooling by melting of hail was greatest at a height midway between the ground and the melting level, and cooling by evaporation of rain was greatest at 1 km of the ground. The numerical simulation described by Krueger et al. (1986) and Proctor (1989) indicated that the cooling associated with phase changes of various hydrometeor types can influence surface outflow strength and low-level downdraft strength. However it should be noted here that the sensitivity tests done by Krueger et al. (1986) and Proctor (1989) were concerned with downdraft production below cloud base and that the influence of ice physics on the evolution of the parent storm was not considered.

In a numerical simulation of a microburst-producing storm, Hjelmfelt et al.(1989) showed that switching off the cooling associated with evaporation of rain and melting of hail in regions of low-level downdrafts substantially reduced surface outflow strength. He also showed that switching off the formation of graupel/hail significantly influenced the evolution of the storm. the formation of rain, the intensity of low-level downdrafts, and the strength of the surface outflow. The simulations of microburst-producing storms by Straka et al.(1993) showed that low-level downdrafts are in some cases stronger and deeper in simulations with the ice phase than in those without the ice phase.

To examine the influence of microphysics on the production of downdrafts in the simulations of thunderstorm occurred on September 8 1994 in Japan, downward acceleration rates due to hydrometeor loading and cooling associated with phase changes are calculated following the formulation by Hjelmfelt (1989) and Orville et al. (1989). A simplified form of vertical equation of

$$\left(\frac{\partial \theta'}{\partial t}\right)_{melting} = \frac{L_f}{c_p}\left(\frac{\partial q_h}{\partial t}\right)_{melting} \quad (10)$$

motion is differentiated with respect to time, and is

$$\left(\frac{\partial \theta'}{\partial t}\right)_{evaporation} = \frac{L_v}{c_p}\left(\frac{\partial q_h}{\partial t}\right)_{evaporation} \quad (11)$$

$$\frac{\partial}{\partial t}\left(\frac{dw}{dt}\right) = \frac{g}{\Theta}\left(\frac{\partial \theta'}{\partial t} - \Theta\frac{\partial q}{\partial t}\right) + \ldots \quad (9)$$

written as

where w is the vertical motion, g gravity, t time, q the total condensate mixing ratio, θ′ temperature or

and

potential temperature deviation from a base state Θ, respectively

The temperature changes due to melting and evaporation are given by andrespectively, where Lv and Lf are the latent heats of fusion and evaporation, respectively; qh is the mixing ratio of graupel/hail and snow: qr is the mixing ratio of rain. and cp is the specific heat at constant pressure.

To analyze the temperature change term in (10) and (11), it is necessary to know the rates of changes in hail/graupel, snow and rain due to melting (or freezing) and evaporation. Similarly, it is also necessary to know the time rate of change in the loading of hail/graupel, snow and rain and multiply by Θ to evaluate the effect of that term in (11). As an example, a constant cooling rate of l°Cmin-1 causes an acceleration of -0.096 ms-2 in 3 min and a vertical velocity of -8.64 ms-1 assuming zero initial acceleration and vertical velocity Temperature deficits of 1°C and precipitation loads of 1 gkg-1 produce accelerations of -0.033 and -0.01 ms-2. If constant. and with the same assumptions as above, these values produce vertical velocities of -6.0 and -1.8 ms-1 respcctively, in 3 min.

The time-height distribution of maximum cooling rates is given in Fig.12. All variables shown on figures are the maximum values in the downdraft region. At the time of downdraft initiation (16-18 min),the largest acceleration rate which is equivalent to about 14.0°*C/min* is due to graupel/hail loading near the height of 6.5 km AGL (Fig.12a). This illustrates the role of loading by the graupel/hail field in initiating the downdraft above the melting level. Below the melting level, graupel/hail melting acceleration rates become the most important term to the strong downdraft producing

Fig.12 The time-height distribution of maximum magnitude of effective cooling rates for forcing mechanisms due to a) hail loading, b)hail melting, c)snow melting , d)rain evaporation, e)rain loading, and f) the sum hydrometeor loading.

The maximum acceleration rates due to hail/graupel melting is located near 2-2.5 km AGL near the time of the maximum downdraft occurrence. So the primary mechanism for the downdraft acceleration in the layer between 2 and 3 km AGL immediately beneath the 0°C level is the cooling produced by melting of graupel/hall. There are two maximum centers for hall/graupel melting acceleration rates the first one of about 16.0°C/*min* is located near 2-2.5 km AGL and occurred at 18 min. about the time of the maximum downdraft (Fig.12b). The second one is located near the surface at 20 min. the time of maximum outflow velocity at the surface.

The snow melting acceleration rates is very small compared with hail/graupel (Fig.12c). The second largest acceleration mechanism between 1 and 3 km AGL is the evaporative cooling. Evaporation increases as rain reaches the ground (Fig.12d). At 20 minutes, evaporation near surface attains the maximum as the rain falls out. Although the cooling due to evaporation at low levels is not able to produce a strong downdraft, it may contribute to strength of the outflow through enhanced horizontal pressure gradient force (Krueger et al. 1986).

The rain loading acceleration rates was only important near the surface (Fig.12e). This is because the most rain is produced from hail melting. Comparing hail loading acceleration rates (Fig.12a) with total hydrometeors loading rates (Fig.12f), one can find that a majority of loading is due to hail.

In summary, the downdraft in the storm is initiated primarily as a result of graupel/hail loading. Melting of graupel/hail into rain below 0` level then importantly contribute to the development and acceleration of the downdraft. A relatively large cooling rates due to evaporation occurs at low altitudes and enhances the microburst outflow intensity The maximum cooling rates due to evaporation just occurs at the time of microburst. So graupel/hail is of vital importance in microburst producing storm.

4. Summary

From the results of simulation, some of key factors which induced the hazardous wet microbursts may be summarised as three the first is the formation of a great quantity of graupel/hails in the upper part of the severe thunderstorm, the second is existence of a relatively dry middle layers and the third is relatively strong vertical shear which causes an obvious overhang structure. The formation of a large number of graupel/hail particles has tw0 meaning: one is the strong dragging force to the fluid and another is the strong cooling to the fluid due to melting of enough hail. The overhang structure of the storm caused by vertical shear can create a very favorite environment for the precipitation falling from it get the maximum melting and evaporation in a relatively dry ambient air. Simulation results suggested that the production of downburst of the storm on & September 1994 in Japan can be explained by tbe effects of negative buoyancy initiated by hail loading and enhanced by the cooling due primarily to the melting of graupel/hail and evaporation of rain. which is similar to the results of Orville et al.(1989), Knunpp (1989), Hjelmfelt et al. (1989) and Straka et al. (1993). The low-level strong hazardous outflows are mainly due to the present of high pressure deviation produced by intense downdraft and enhanced by strong surface evaporative cooling of rain.

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