Maintenance Mechanism and Thermodynamic Structure of a Baiu Frontal Rainband Retrieved from Dual Doppler Radar Observations

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Abstract

During the intensive observation period of the Baiu Front Heavy Rainfall Experiment in July 1988, a Baiu frontal rainband passed over the Kyushu District, and its kinematic structure was observed by dual Doppler radar. The purposes of this paper are to investigate the thermodynamic structure of the rainband using a retrieval method and to clarify its maintenance mechanism.

The rainband had many similarities to those for tropical and mid-latitude squall lines. The dry rear inflow from the northeast merged with the convective downdrafts and resulted in low-level divergent outflows. The warm and moist southwesterly converged with the outflow directed toward the leading edge and caused the rearward sloping convective updrafts.

The retrieved perturbation temperature field showed a positive anomaly in the convective updrafts above a height of 4 km, and a negative anomaly due to evaporation of raindrops in the dry rear inflow behind the gust front. The perturbation pressure pattern was consistent with the temperature and flow patterns and showed a meso-high behind the gust front and a meso-low beneath the warm anomaly of the convective updraft. In the low-level of the stratiform region, a meso-low was generated owing to the subsidence warming of the descending rear inflow. The distributions of temperature and pressure perturbations suggest that the convective updraft was maintained by the upward pressure gradient force below a height of 4 km, while it was maintained by a positive buoyancy above this level. The horizontal pressure gradient force associated with the mid-level meso-low in the convective region accelerated the lifted air to the rear, and the convective updraft tilted to the rear of the system. The raindrops were transferred to the rear of the system and fell into the convective downdraft region, which did not prevent the convective updraft from developing and the intense rainband was maintained.

1. Introduction

Seasonal change from winter to summer in East Asia is characterized by a rainy season. It is referred to as the “Baiu” in Japan and as “Mei-Yu” in China. In this season, a quasi-stationary front is formed between polar and tropical air masses and extends from the eastern part of China to Japan. The Baiu front is one of the most significant characteristic features of the water circulation in that region, and much research has been performed to study the disturbances associated with the Baiu front.

Matsumoto *et al.* (1970), Akiyama (1973), and Ninomiya and Akiyama (1971, 1972, 1973) showed that sub-synoptic-scale cyclones develop along the Baiu front. They have a typical horizontal scale of 1000–3000 km, and are termed meso- α -scale or

medium-scale cyclones. Heavy rainfalls are generally caused by organized meso- β - or meso- γ -scale convective systems within a meso- α -scale cyclone (Matsumoto and Akiyama, 1969, 1970; Akiyama, 1978, 1984; Ninomiya *et al.*, 1988ab). Takeda and Seko (1986) analyzed the structure of meso- β -scale organized convective systems using radar data, and discussed their formation and maintenance.

The synoptic analysis and satellite studies clarified the structure of the meso- α -scale disturbance and radar studies showed the movement and development process of the meso- β -scale disturbance. Previous studies, however, have not revealed the kinematic structure of the meso- β -disturbance because no observation of velocity field in the disturbance in Baiu front by dual Doppler radars has been performed. Therefore, the internal dynamic processes of the meso- β and meso- γ -scale convective systems are still unknown as are the cloud physical processes in the organized convective systems. This

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makes it difficult to discuss the organization process of cumulus cloud and to forecast the localized heavy rainfalls associated with the Baiu front, which bring severe damage to the human society.

An intensive observation called “the Baiu Front Heavy Rainfall Experiment” was performed in the Kyushu District and its surrounding area during a period from 6 to 19 July 1988 in order to clarify the mechanism of localized heavy rainfalls associated with the Baiu front (Asai, 1990). During the observation, a meso- β -scale rainband which developed along the Baiu front passed over the observation sites, and its internal kinematic structure was deduced from dual Doppler observations. Ishihara *et al.* (1995) found that the overall structure of the rainband was similar to those of mid-latitude or tropical squall lines. The rainband was composed of a convective region forming the leading edge of the system and a trailing stratiform region. The dry rear inflow from the northeast merged with the convective downdrafts and resulted in low-level divergent outflows toward and away from the leading edge.

Although the kinematic structure of the rainband was revealed, it still remained to study the temperature and pressure perturbation fields in the rainband. Knowledge of these thermodynamic variables are essential for the understanding of the physical processes maintaining the observed air circulation and precipitation. In recent years, some indirect methods of recovering thermodynamic variables from wind and reflectivity fields derived from Doppler radar have been developed. Gal-Chen (1978) and Hane and Scott (1978) proposed dynamic retrieval techniques, where horizontal gradients of pressure and temperature are deduced from the velocity field through

Fig. 1. The surface weather map at 2100 JST 17 July 1988.

a variational analysis of the equations of motion. These methods have been applied to real storms with strong updrafts, including squall lines (Roux *et al.*, 1984; Lin *et al.*, 1986), tornadic storms (Brandes, 1984; Hane and Ray, 1985), a cold-frontal rainband (Parsons *et al.*, 1987), as well as to a planetary boundary layer inversion (Gal-Chen and Kropfli, 1984) in order to investigate the physical processes responsible for the observed structures.

Further, Roux (1985, 1988) and Roux and Sun (1990) proposed a method for the retrieval of three-dimensional fields of pressure and temperature perturbations instead of their relative perturbation values on each horizontal plane from the combined use of the equations of motion and a simplified thermodynamic equation. More recently, in order to study

Fig. 2. The GMS (Geostationary meteorological satellite) IR image at 18 JST 17 July 1988. The cloud cluster is indicated by an arrow.

Fig. 3. Radar reflectivity of the rainband at a height of 1.0 km measured by the MRI radar at 1855 JST 17 July 1988. The contour interval is 10 dBZ, and the minimum level is 10 dBZ. Shaded areas represent reflectivities > 30 dBZ. The thick lines are coastlines of Kyushu. The rectangular box in the lower panel indicates the horizontal domain of dual Doppler analysis and retrieval. The location of the MRI and ILTS Doppler radars are indicated. System motion and north direction are also indicated by arrows.

the precipitation processes in the convective systems, Hauser *et al.* (1988) and Greets and Hobbs (1991) retrieved distributions of water contents by a combined use of the thermodynamic retrieval method and a parameterized microphysical model.

In this paper, a retrieval method proposed by Roux (1988) is applied to the wind and reflectiv-

ity data of the Baiu frontal rainband. The thermodynamic and dynamic structure of the rainband is analyzed quantitatively to clarify the maintenance mechanism of the rainband. The synoptic conditions and the internal kinematic structure of this rainband have already been investigated in other papers (Asai, 1990; Ishihara *et al.*, 1995). Therefore, only the main points are recalled in Section 2. In Section 3, a detailed description of the retrieved thermodynamic structure and the dynamic processes of the rainband are presented. This is followed by the discussion of the results in Section 4. Finally, Section 5 presents the summary and conclusions of the study.

2. General description of the rainband

The Baiu frontal rainband studied here was observed on 17 July 1988 (Asai, 1990; Ishihara *et al.*, 1995). We summarize hereafter the main features of the rainband, particularly its kinematic structure.

2.1 Characteristics of the environment

The surface weather map at 21 JST (Japan Standard Time, used throughout this paper; 1200 UTC = 2100 JST) 17 July 1988 (Fig. 1) shows that a Baiu front extended from the Yellow Sea to Kyushu. There were no medium-scale cyclones near Kyushu. A southwesterly wind blew from the East China Sea to the Baiu front.

The satellite image at 18 JST 17 July 1988 shows a cloud cluster over Kyushu (Fig. 2). The cloud cluster was generated over the southern part of the Korean peninsula at 03 JST on 17 July. The cloud cluster moved southeastward along the surface Baiu front and traversed Kyushu from the northwest to the southeast. The rainband studied here was embedded in the southwestern part of the cloud cluster.

Dual Doppler radar observation was performed during the passage of the rainband by the research groups of the Meteorological Research Institute (MRI) and the Institute of Low Temperature Science, Hokkaido University (ILTS). The MRI Doppler radar and the ILTS Doppler radar were installed at Takada and Kumamoto, respectively. Figure 3 shows a radar reflectivity pattern at a height of 1 km derived from the MRI Doppler radar at 1855 JST on 17 July. The convective region signified by the intense echo (more than 30 dBZ) extended from the northwest to the southeast. The stratiform rain region with relatively uniform echo (about 20 dBZ) is located on the northeast side of the convective region. The ground-relative speed of the rainband deduced from the successive reflectivity images was 8.3 m s^{-1} toward the south-southeast (155° in azimuth). Heavy rain with a maximum rainfall rate of over 100 mm hr^{-1} coincided with the passage of the rainband along the west coast of Kyushu.

Figure 4 shows the vertical profiles of the potential temperature (θ), equivalent potential tempera-

Fig. 4. Vertical profiles of potential temperature θ (thin line), equivalent potential temperature θ_e (thick line), and saturated equivalent potential temperature θ_e^* (broken line) from soundings at (a) Kagoshima at 2030 JST 17 July 1988 and (b) Takada at 2119 JST 17 July 1988.

ture (θ_e), and saturated equivalent potential temperature (θ_e^*) derived from soundings at Kagoshima (2030 JST, 30 km ahead of the rainband) and Takada (2119 JST, 30 km behind the rainband). At Kagoshima (Fig. 4a), the stratification was as a whole convectively unstable from the surface to 6 km in height. However, a stable layer between 1.5 and 2 km suppressed the development of convection. θ_e had a maximum value (356 K) at the surface and a minimum (338 K) at 6 km. If the air parcels near the surface are lifted from the ground up to 0.8 km (level of free convection), and penetrate through the stable layer, they will acquire a positive buoyancy up to 12 km. At Takada (Fig. 4b), the stratification was characterized by an absolute stability below 1 km, while the layer 1.5–6 km was still convectively unstable. Low θ_e values were observed in the layers between 5 and 6 km and between 6.7 and 8.5 km.

Figure 5 shows the vertical profiles of the parallel components of the winds relative to the direction of the system movement (155°). Positive (negative) U denotes south-southeastward (north-northwestward), representing system-relative rear-to-front (front-to-rear) airflow. In front of the rainband (Kagoshima), U was negative below 3.5 km. This indicates that moisture-rich environmental air flowed toward the leading edge of the rainband. Significant positive U between 12 and 14.5 km suggests that the air flowed out of the rainband to the front of the system. Behind the rainband (Takada), positive U appear in layers between 3 and 6 km and between

Fig. 5. Vertical profiles of the system-relative horizontal wind velocity along the direction of the system movement (155° in azimuth).

Fig. 6. Horizontal cross sections of system-relative wind and reflectivity at a height of (a) 1.0 km, (b) 2.0 km, (c) 4.0 km, and (d) 6.0 km at 1855 JST 17 July 1988. The contour interval is 5 dBZ. The shaded areas represent reflectivities > 30 dBZ. In the panel (a), the solid line indicates the position where the vertical cross sections as presented in Figs. 8, 11, 12 are made.

6.5 and 12 km. Relatively low θ_e values were observed in these layer. This wind profile, therefore, represents an influx of dry mid-level environmental air from the rear of the rainband.

2.2 Kinematic structure

When the convective region arrived above the experimental site, the dual Doppler radar observations

were made, and the three-dimensional wind recovery was performed by Ishihara *et al.* (1995).

RHI display of single Doppler radar showed that some convective cells of the rainband extended to 16 km in height (not shown). However, three-dimensional velocity data were obtained only for levels lower than 10 km and the area of analysis decreased with height. For these reasons, the re-

Fig. 7. As in Fig. 6, except for vertical velocity (w). The contour interval is 2 m s^{-1} . Areas of negative values are shaded.

gion from 1 km to 7.5 km in vertical was adopted for the thermodynamic retrieval. The data were interpolated into an $16 \times 31 \times 13$ Cartesian grid with 1 km grid spacing in the horizontal and 0.5 km in the vertical.

Figure 6 shows system-relative horizontal winds and the reflectivity patterns at the heights of 1.0, 2.0, 4.0 and 6.0 km. Horizontal distributions of vertical velocity at each height are shown in Fig. 7. The thick dashed lines in Figs. 6a and 6b denote

the gust front. Comparison between Fig. 6 and Fig. 7 reveals that the updrafts and downdrafts were well organized along the rainband and downward motion was significant in the intense echo region. At 1 km in height (Fig. 6a), there was intense divergence along the high-reflectivity zone ($Z > 30 \text{ dBZ}$). The gust front is clearly seen at 2 km in height (Fig. 6b). On the south side of the gust front, system-relative winds were southerly and transferred warm and moist low-level air toward the rainband. At

Fig. 8. Vertical cross section of (a) system-relative wind and reflectivity, (b) system-relative horizontal velocity, and (c) vertical velocity at 1855 JST 17 July 1988. The contour intervals are 5 dBZ, 2 m s^{-1} , and 2 m s^{-1} , respectively. In panel (a), shaded areas represent reflectivities > 30 dBZ. Areas of negative values are shaded in (b) and (c).

4 km in height (Fig. 6c), the convergence of two distinct flows are evident; one is the aforementioned southerly and the other is a northeasterly coming from the rear of the rainband. The northeasterly corresponds to the "rear inflow" which is frequently observed in mid-latitude and tropical squall lines (Smull and Houze, 1987b), so that we refer to it as the "rear inflow" hereafter. The rear inflow is considered to be an inflow of the environment air characterized by low θ_e . The maximum value of the upward velocity was 7 m s^{-1} , while the speed of downdrafts was slightly larger than that of the up-

drafts. A zone of descending motion of 1~2 m s^{-1} is prevalent to the north of the convective region, indicating that the rear inflow air is gradually descending as it approaches the intense echo region. The southerly wind components were predominant at 6 km (Fig. 6d), and the convergence zone was located to the north of the high-reflectivity area.

Figure 8 shows system-relative winds with reflectivity, horizontal velocity and vertical velocity along the NNE-SSW cross section (line NNE-SSW in Fig. 6a). This vertical cross section was chosen because it is representative of the region where the vertical mo-

Fig. 9. Horizontal distribution of temperature perturbation at a height of (a) 1.0 km, (b) 2.0 km, (c) 4.0 km, and (d) 6.0 km at 1855 JST 17 July 1988. The contour interval is 1 K, and areas of negative values are shaded.

tions and the rear inflow are the most intense. Five distinct airflows can be identified in Fig. 8a. The southerly is lifted upward by the gust front, forming a convective updraft (Flow A) with the maximum vertical velocity of 7 m s^{-1} (Fig. 8c). The updraft tilts toward the rear of the system. The rear inflow (Flow B) with a speed of 10 m s^{-1} enters the system at a height of 2–7 km (Fig. 8b). The convective downdraft (Flow C) is located in the high reflectivity

region. The rear inflow merges with the convective downdraft and results in the low-level divergent outflows toward and away from the leading edge (Flows D, E). Flow D collided with the environmental warm and moist southerly (Flow A), forming a gust front at a distance of 30 km. A part of the Flow A merged into the convective downdraft. Thus a distinct rotor circulation was centered at $D = 25 \text{ km}$, $Z = 3.5 \text{ km}$.

Fig. 10. Horizontal distribution of pressure perturbation at a height of (a) 1.0 km, (b) 2.0 km, (c) 4.0 km and (d) 6.0 km at 1855 JST 17 July 1988. The contour interval is 0.2 hPa with negative values shaded.

3. Thermodynamic structure of the rainband

The three-dimensional wind and reflectivity data are used to derive the thermodynamic structure of the rainband using the thermodynamic retrieval method (see Appendix A). In addition, sounding data at Kagoshima and Takada presented in Section 2 are used to determine the stratification in the retrieval domain. This section describes the retrieved thermodynamic structure and the dynamic

processes of the rainband.

3.1 Retrieved thermodynamic structure

Figures 9 and 10 show the horizontal cross section, at the heights of 1.0, 2.0, 4.0, and 6.0 km, of the retrieved virtual-cloud potential temperature perturbation θ_{C1} and pressure perturbation, respectively. θ_{C1} is defined as

$$\theta_{C1} = \theta_1 + 0.61\theta_0q_{V1} - \theta_0q_C, \quad (1)$$

where θ is the potential temperature and q_V and q_C are the mixing ratios of the water vapor and the cloud water, respectively. The subscript “0” refers to basic fields derived from the linear interpolation of two sounding data at Kagoshima and Takada, and the subscript “1” refers to first-order perturbations. θ_{C1} includes contributions from the water vapor perturbation and the cloud water content. However, since these terms have opposite effects in θ_{C1} , their total influence should not be very large. Therefore, we will refer to θ_{C1} as the “temperature perturbation” for simplicity.

a. Horizontal view at 1.0 km (Fig. 9a, Fig. 10a)

At 1 km, a band-shaped negative temperature area is found behind the gust front. This indicates that the divergent outflow toward the gust front is colder than the environment. The perturbation pressure field has a band-shaped high pressure area with a maximum of 1.2 hPa, and a low pressure to the north of it. The distribution of the high and low pressure perturbations is closely related to the temperature perturbation. The high pressure with its maximum located at $X = 4$ km, $Y = 12$ km is associated with the cold pool in the southern part of the domain, and the low pressure area is associated with the warm air above (Fig. 9b).

b. Horizontal view at 2.0 km (Fig. 9b, Fig. 10b)

At 2 km, the convective updraft region in the southern part of the domain is associated with a negative temperature perturbation. The distribution of pressure is similar to that at 1 km, but the pressure excess in the southern part of the domain is much smaller than that at 1 km. This indicates that the upward pressure gradient force is prevalent in the lowlevels of the convective region.

c. Horizontal view at 4.0 km (Fig. 9c, Fig. 10c)

At 4 km, the positive temperature perturbation is found around $X = 7$ km, $Y = 4$ km, while the temperature deficit reaching -3.5 K is found at $X = 7$ km, $Y = 9$ km. Comparison between Fig. 7c and Fig. 9c reveals that centers of the maximum warming (cooling) do not exactly correspond to centers of the updrafts (downdrafts) in the convective region. A zone of positive temperature perturbation is prevalent behind the convective region. It may be attributed to the subsidence warming of the descending rear inflow air. The predominant feature of perturbation pressure is a band of negative anomaly reaching -1.2 hPa along the wind shift zone. The meso-low is associated with the warm air above (Fig. 9d).

d. Horizontal view at 6.0 km (Fig. 9d, Fig. 10d)

At 6 km, the overall perturbation temperature pattern corresponds to the distribution of the vertical velocity with warming in updrafts and cooling in downdrafts. The maximum warming up to 3 K

corresponds well to the position of the maximum upward velocity at $X = 7$ km, $Y = 7$ km. The pressure deficit in the convective region is smaller than that observed at 4 km, indicating that downward pressure gradient force is acting there.

e. Vertical cross section

Figure 11 shows vertical cross section (along the line NNE-SSW in Fig. 6a) of the perturbation temperature, the perturbation pressure, and the production rate of rain $F(q_R)$. Large negative temperature perturbations are found at the level of 2–4 km in the convective region. The perturbation pressure shows a meso-high around $D = 25$ km in the lowlevels, and a meso-low with a minimum of -1.2 hPa above. The meso-high mainly results from the presence of cold air. In addition, a dynamic effect due to the convergence between the divergent outflow and the lowlevel southerly also increases the pressure. The meso-low in the midlevels of the convective region ($22 < D < 30$ km, $3 < Z < 6$ km) mainly results from the hydrostatic effect owing to the presence of warm air above. In addition, the dynamic effect of the rotor circulation also reduces the pressure there. The positive perturbation temperature area in the stratiform region ($6 < D < 20$ km, $2 < Z < 4$ km) is considered to be the result of subsidence warming of the rear inflow. In response to the warm air above, the perturbation pressure field shows a small negative anomaly in the lowlevels of the stratiform region.

The production rate of rain $F(q_R)$ is calculated from the rain water mixing ratio q_R and fall speed of raindrops V_T derived from reflectivity values (the empirical formulas are given in Appendix B):

$$F(q_R) = (\mathbf{V} \cdot \nabla)q_R + \frac{1}{\rho} \frac{\partial}{\partial z}(\rho V_T q_R) \quad (2)$$

where $\mathbf{V} = (u, v, w)$ is the system-relative air velocity vector and ρ is the air density. As mentioned in Roux and Sun (1990), positive values of $F(q_R)$ indicate that rain is generated from cloud in saturated air, while negative values indicate evaporation of rain in the subsaturated air. Note, however, that $F(q_R)$ calculated here could be a rough estimate, because the reflectivity increase due to the melting of solid hydrometeors or the rapid aggregation of precipitating particles brings uncertainties into q_R - Z and V_T - Z relationships. So we will discuss here the microphysical processes occurring in the rainband only qualitatively.

Positive values of $F(q_R)$ are found above 3.5 km in the convective region, where the condensed cloud water is considered to be converted into rain. Large negative $F(q_R)$ values in the lowlevels of the convective region indicate intense evaporation of raindrops in the dry rear inflow air. Comparison between Fig. 11a and Fig. 11c reveals that the cold pool behind the gust front was produced by the evaporation of rain.

Fig. 11. Vertical cross sections of (a) temperature perturbation and (b) pressure perturbation, and (c) production rate of rain at 1855 JST 17 July 1988. The contour intervals are 1 K, 0.2 hPa, and $1.0 \text{ g kg}^{-1} \text{ hr}^{-1}$, respectively. Areas of negative values are shaded.

3.2 Dynamic process

The dynamic structure of the rainband can be explained further by analyzing the force balance. In the present case we will suppose that the contributions of evolution and sub-grid scale dissipation are negligible. Then, the wind and precipitation fields are assumed to obey the following equation of motion:

$$(\mathbf{V} \cdot \nabla)\mathbf{V} = -C_p \theta_{V0} \nabla \Pi_1 + g \left(\frac{\theta_{C1}}{\theta_0} - q_R \right) \mathbf{k}, \quad (3)$$

where C_p is the specific heat capacity of moist air at constant pressure, Π the Exner function, g the ac-

celeration due to gravity, $\mathbf{k}$ the vertical unit vector, θ_{V0} the virtual potential temperature. The horizontal acceleration results from the horizontal pressure-gradient force $[-C_p \theta_{V0} \nabla_H \Pi_1]$, where subscript "H" denotes two-dimensional horizontal vectors. The vertical acceleration results from the combination of the vertical pressure gradient force $[-C_p \theta_{V0} \partial \Pi_1 / \partial z]$, the thermal buoyancy $[g(\theta_{C1}/\theta_0)\mathbf{k}]$, and the rain loading $[-g q_R \mathbf{k}]$. These components of force are shown in Fig. 12 for the same cross section as in Fig. 11.

The horizontal pressure gradient force (Fig. 12a) is directed toward the rear of the system at

Fig. 12. Vertical cross sections of (a) horizontal pressure gradient force, (b) vertical acceleration, (c) vertical pressure gradient force, (d) thermal buoyancy, and (e) rainwater loading at 1855 JST 17 July 1988. The contour intervals are 10^{-2} m s^{-2} for (a), (b), (e) and $2 \times 10^{-2} \text{ m s}^{-2}$ for (c) and (d), respectively. Areas of negative values are shaded.

Fig. 12. (Continued)

$26 < D < 33$ km, $Z > 2.5$ km. The horizontal pressure gradient force and the environmental wind shear cause the rearward tilt of the convective updraft. On the contrary, the positive horizontal pressure gradient force in the convective downdraft at $20 < D < 25$ km, $Z > 2.5$ km accelerates the downdraft toward the front of the system. As a consequence, the rear inflow smoothly merges with the downdraft, increasing its horizontal speed, and penetrates into the high reflectivity region (Fig. 8b). The horizontal pressure gradient force below 2.5 km in the convective region is directed toward the front at $25 < D < 33$ km, while it is directed toward the rear at $20 < D < 25$ km. Such a distribution of the horizontal pressure gradient force causes the low-level divergent outflows.

Figure 12b shows the vertical acceleration calculated from the wind data. The positive maximum value up to $6 \times 10^{-2} \text{ m s}^{-2}$ is found in the convective updraft. The negative maximum of $-7 \times 10^{-2} \text{ m s}^{-2}$ is present in the transition zone from the convective updraft to the downdraft. A comparison between the vertical pressure gradient and the effective buoyancy fields (Figs. 12c and 12d) shows that they are closely related. In particular, pressure-buoyancy relation is nearly a hydrostatic balance in the stratiform region. Figures 12b, 12c and 12d show that upward acceleration in the convective updraft is due to the upward pressure gradient force below 4 km, while it is due to the positive thermal buoyancy above 4 km. Likewise, the downward acceleration in the convective region is related to the downward pressure gradient force above 5.0 km, while it is related to the negative thermal buoyancy below. The magnitude of the rain loading (Fig. 12e) is considerably smaller than the vertical pressure gradient and the thermal buoyancy. The influence of the rain loading on the dynamics of the rainband, however, cannot be neglected in the high reflectivity cores because the vertical pressure gradient force and the thermal buoyancy generally have opposite contribu-

tions to the vertical acceleration. The maximum rain loading of $-2.0 \times 10^{-2} \text{ m s}^{-2}$, which is found in the convective downdraft region, has 30 % contributions to the downward acceleration.

4. Discussions

4.1 Internal consistency checks

Because the temperature and pressure perturbation fields are obtained in a least-square sense in the thermodynamic retrieval, their validity must be evaluated by checking the balance of the original equations. Following Roux (1988), four parameters are defined to check the reliability of the results:

$$E_{\theta}^H(z) = \frac{\iint_{\mathcal{H}(z)} [\nabla_H \theta_{C1}^* - \mathbf{B}_H]^2 dx dy}{\iint_{\mathcal{H}(z)} \mathbf{B}_H^2 dx dy}, \quad (4)$$

$$E_{\theta}^V(z) = \frac{\iint_{\mathcal{H}(z)} \left[w \frac{\partial \theta_{C1}^*}{\partial z} - (B_T - \mathbf{V}_H \cdot \mathbf{B}_H) \right]^2 dx dy}{\iint_{\mathcal{H}(z)} [B_T - \mathbf{V}_H \cdot \mathbf{B}_H]^2 dx dy}, \quad (5)$$

$$E_{\Pi}^H(z) = \frac{\iint_{\mathcal{H}(z)} [\nabla_H \Pi_1^* - \mathbf{A}_H]^2 dx dy}{\iint_{\mathcal{H}(z)} \mathbf{A}_H^2 dx dy}, \quad (6)$$

and

$$E_{\Pi}^V(z) = \frac{\iint_{\mathcal{H}(z)} \left[\frac{\partial \Pi_1^*}{\partial z} - A_z - \frac{g \theta_{C1}^*}{C_p \theta_0 \theta_{V0}} \right]^2 dx dy}{\iint_{\mathcal{H}(z)} \left[A_z + \frac{g \theta_{C1}^*}{C_p \theta_0 \theta_{V0}} \right]^2 dx dy}, \quad (7)$$

where θ_{C1}^* and Π_1^* are the retrieved temperature and pressure perturbations, respectively, $\mathcal{H}(z)$ is the horizontal domain at the height z where wind and reflectivity data are available. These parameters

Fig. 13. Vertical profiles of the checking parameters.

are used to check how the retrieved temperature and pressure perturbations fit the experimental estimates (B_X, B_Y, B_T) and (A_X, A_Y, A_Z). When all these parameters are zero, the retrieved variables satisfy the individual momentum equation and thermodynamic equation exactly. As suggested by Gal-Chen and Hane (1981), the retrieved variables are virtually meaningless if these parameters are larger than 0.5.

Figure 13 shows the vertical profiles of these parameters. The values of $E_\theta^V(z)$ and $E_\theta^H(z)$ are within the interval of [0.3, 0.4] below 6 km, indicating that the retrieved temperature fields are consistent with the input wind field, except near the highest level. The large values of $E_\theta^V(z)$ and $E_\theta^H(z)$ in the highest level mainly come from the errors in the vertical velocity estimated by the upward integration of the continuity equation. The values of $E_\pi^H(z)$ and $E_\pi^V(z)$ are less than 0.3 above 2 km, indicating the reasonably retrieved pressure field.

The reliability of the retrieved data can also be tested through comparisons with the observed surface temperature and pressure variations. As reported in Ishihara *et al.* (1995), the passage of the precipitation core of the rainband was accompanied by a temperature drop about -4 K and a pressure rise about 1.3 hPa. The low-level retrieved temperature and pressure fields produced a high-pressure cold pool in the lowlevels of the convective region, and are consistent with the observed variations of surface temperature and pressure.

4.2 Maintenance mechanism of the rainband

Figure 14 shows a conceptual model constructed from the results to describe the kinematic and thermodynamic structures of the rainband. The maintenance mechanism of the rainband is considered as a feedback of the following processes (1)~(5);

(1) Dry mid-level air enters from the rear of the rainband, and it merges with the convective downdraft which developed in the heavy rain region. The evaporative cooling in the dry air causes a cold downdraft in the lowlevels of the convective region.

(2) Owing to the large pressure gradient at the surface, the cold downdraft air spread out in the lowlevels toward and away from the leading edge. The gust front is formed between the cold divergent flow and the ambient southwesterly on the southwest side of the rainband.

(3) The warm and moist southwesterly air is lifted up by the gust front and is forced upward by the vertical pressure gradient force. Above 4 km, the effective buoyancy becomes positive and reinforces the convective updrafts.

(4) Owing to the presence of warm air above and cold air below, a low pressure area is induced immediately beneath the warm anomaly in the convective updraft. The resultant horizontal pressure gradient force and environmental wind shear cause the rearward tilt of the convective updraft.

(5) The raindrops are carried toward the rear of the system and fall into the convective downdraft. The raindrops accelerate the downdraft through its loading and evaporative cooling.

In the Baiu frontal zone, the atmosphere is generally moist through the entire troposphere. This makes evaporation of rain weak and a cold air pool is not usually developed in the precipitation area. A squall-line type convective system, therefore, is not often observed in the Baiu frontal zone. In the present case, however, the dry rear inflow caused evaporative cooling and developed the cold air pool at the ground, which lifted up the low-level moist and warm southwesterly air. Consequently, the intense rainband developed with some characteristic features of a squall-line type convective system: the organized precipitation band, the cold pool due to evaporation, advancing gust front, rearward-tilted convective updraft.

4.3 Comparison with other mesoscale convective systems

As reported in Ishihara *et al.* (1995), the rainband in this study was different from typical convective systems which are accompanied by medium-scale lows and bring about localized heavy rainfalls in the Baiu season of Kyushu (*e.g.* Ogura *et al.*, 1985). Rather, it showed many similarities in its kinematic structures to those for mid-latitude and tropical squall lines. The rainband was composed

Fig. 14. Schematic vertical cross section of the rainband. Arrows denote air flows relative to the rainband. Open arrows denote air flows with positive θ_{C1} , while solid arrows denote air flows with negative θ_{C1} . H and L denote the high and low pressure region, respectively.

of a convective region forming the leading edge of the system and a trailing stratiform region. The mid-level rear inflow, which is often observed in mid-latitude squall lines, was also identified in the present case.

The retrieved thermodynamic structure also showed some characteristic features similar to those for mid-latitude and tropical squall lines. The perturbation temperature field showed a warming due to condensation in the convective updraft above 4 km, and a cooling due to evaporation of rain in the dry rear inflow behind the gust front. The perturbation pressure field showed a meso-high behind the gust front and a meso-low immediately beneath the warm convective updraft. The meso-low accelerates the convective updraft front-to-rear, and the convective downdraft rear-to-front. These thermodynamic structure and associated acceleration of updrafts and downdrafts have also been cited by LeMone (1983), LeMone *et al.* (1984), for tropical squall lines during GATE, Roux (1985, 1988) for tropical squall lines during COPT81, and Smull and Houze (1985, 1987ab) for mid-latitude squall lines.

5. Conclusions

The thermodynamic structure of a Baiu frontal rainband on 17 July 1988 was studied using a thermodynamic retrieval method.

The important characteristic feature of the environment was the influx of mid-level dry air from the

rear of the system, which merged with the convective downdraft and resulted in divergent outflows behind the gust front. The warm and moist southwesterly flowed toward the leading edge of the rainband and the rearward sloping convective updrafts developed.

Temperature and pressure perturbations are recovered from a three-dimensional wind field using a thermodynamic retrieval method. The perturbation temperature field shows a warming due to condensation in the convective updraft above 4 km, and a cooling due to evaporation of rain in the dry rear inflow behind the gust front. Corresponding to the temperature and flow pattern, the perturbation pressure pattern shows a meso-high behind the gust front and a meso-low immediately beneath the warm anomaly in the convective updraft. In the low levels of the stratiform region, a meso-low is generated reflecting the subsidence warming of the descending rear inflow.

The examination of the force balance showed that the convective updraft was maintained by the upward pressure gradient force below 4 km, while it was maintained by the positive buoyancy above. In response to the perturbation pressure minimum below the convective updraft, the lifted high- θ_e air is accelerated front-to-rear, and the convective downdraft rear-to-front. The rain drops are carried to the rear of the system and fall into the convective downdraft. The rainwater causes a downward drag force and supplies water to the dry rear inflow.

In this study, complicated microphysical processes including the ice phase are not taken into account in the retrieval. This is a shortcoming, since the ice phase may play an important role in severe convective storms. For the further understanding of the precipitating processes and the dynamics of convective systems, it will be necessary to retrieve the distribution of hydrometeors as well as the thermodynamic fields using a retrieval method including parameterized microphysical processes.

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Appendix

A. The thermodynamic retrieval method

The thermodynamic retrieval method used in this study was basically outlined by Roux (1988), and the retrieved temperature and pressure fields are in balance with both the velocity fields and the heat source.

Two equations for θ_{C1} and three equations for Π_1 can be deduced from the horizontal components of the curl of Eq. (3) and from the components of Eq. (3), respectively, as

$$\nabla_H \theta_{C1} = \mathbf{B}_H \quad (\text{A.1})$$

and

$$\nabla \Pi_1 = \mathbf{A} + \left(\frac{g}{C_p \theta_0 \theta_{V0}} \right) \theta_{C1} \mathbf{k}, \quad (\text{A.2})$$

with

$$\mathbf{B}_H = \frac{\theta_0}{g} \left\{ \nabla_H [(\mathbf{V} \cdot \nabla)w + gq_R] - \theta_{V0} \frac{\partial}{\partial z} \left[\frac{1}{\theta_{V0}} ((\mathbf{V} \cdot \nabla) \mathbf{V}_H) \right] \right\} \quad (\text{A.3})$$

and

$$\mathbf{A} = -\frac{1}{C_p \theta_{V0}} [(\mathbf{V} \cdot \nabla) \mathbf{V} + gq_R \mathbf{k}]. \quad (\text{A.4})$$

$\mathbf{B}_H$ and $\mathbf{A}$ can be calculated from the Doppler-derived velocity and reflectivity fields. This set of equations is incomplete since the vertical gradient of θ_{C1} remains unknown. Following Roux and Sun (1990), a simplified thermodynamic equation can provide an additional condition for θ_{C1} :

$$\begin{aligned} (\mathbf{V} \cdot \nabla \theta_{C1}) &= B_T \\ &= B_0 + \delta_S B_S + (1 - \delta_S) B_{NS}, \end{aligned} \quad (\text{A.5})$$

where

$$B_0 = w \left(-0.61 \theta_0 \frac{dq_{V0}}{dz} - \frac{d\theta_0}{dz} \right), \quad (\text{A.6})$$

$$B_S = w \left(1.61 \theta_0 - \frac{L}{C_p \Pi_0} \right) \left(\frac{dq_S}{dz} \right) + \theta_0 F(q_R), \quad (\text{A.7})$$

$$B_{NS} = \left(\frac{L}{C_p \Pi_0} - 0.61 \theta_0 \right) F(q_R), \quad (\text{A.8})$$

and L is the latent heat of condensation. δ_S is a parameter which is equal to 1 when air is saturated, and to 0 when it is not. q_S is the saturation mixing ratio deduced from the basic temperature and pressure fields using Tetens' (1930) formulation. Owing to the complexity of the microphysical process involved in the net production of ice particles, their contribution to the heat source or sink is not taken into account here. Using (A.1), (A.5) can be rewritten as

$$w \frac{\partial \theta_{C1}}{\partial z} = B_T - \mathbf{V}_H \cdot \mathbf{B}_H \quad (\text{A.9})$$

The derived winds contain errors from a number of different sources and the assumptions used here may also be in error, thus (A.1), (A.9) are solved in a least-square sense. A solution for θ_{C1} is given by the minimization of the functional I_θ which is defined as

$$\begin{aligned} I_\theta &= \iiint_{\mathcal{D}} \left\{ [\nabla_H \theta_{C1} - \mathbf{B}_H]^2 \right. \\ &\quad \left. + \left[\frac{w}{\nu} \frac{\partial \theta_{C1}}{\partial z} - \frac{1}{\nu} (B_T - \mathbf{V}_H \cdot \mathbf{B}_H) \right]^2 \right\} dx dy dz, \end{aligned} \quad (\text{A.10})$$

where ν is a reference velocity amplitude, $\mathcal{D} = [(x_1, x_2), (y_1, y_2), (z_1, z_2)]$ is the three-dimensional analysis domain where wind data are available. $x = x_1$ and x_2 denote western and eastern boundaries, respectively, while $y = y_1$ and y_2 denote southern and northern boundaries, respectively. $z = z_1$ and z_2 denote the lower and upper boundary, respectively. Here, ν is chosen to be large (20 m s^{-1}) compared to the r.m.s. value of the vertical velocity, because (A.1) yields a more accurate buoyancy distribution than (A.9), and because the value improved the internal consistency of the equations. Then, using the retrieved values of θ_{C1} , a solution for Π_1 is obtained by minimizing the functional

$$\begin{aligned} I_\Pi &= \iiint_{\mathcal{D}} \left\{ [\nabla_H \Pi_1 - \mathbf{A}_H]^2 \right. \\ &\quad \left. + \mu \left[\frac{\partial \Pi_1}{\partial z} - \left(A_z + \frac{g \theta_{C1}}{C_p \theta_0 \theta_{V0}} \right) \right]^2 \right\} dx dy dz. \end{aligned} \quad (\text{A.11})$$

A weighting factor μ (here 0.05) is used because the vertical gradient of pressure perturbation is much

larger than the horizontal one. (A.10) and (A.11) may be solved by direct integration of the associated Euler equation with proper boundary conditions. In this study, θ_{C1} and Π_1 are set to zero at the southern boundary ($y = y_1$) of the calculation domain, because the vertical velocity was small and the flow pattern was relatively uniform there (see Fig. 8), and associated temperature and pressure perturbations are assumed to be small. The Neumann boundary conditions are adopted for the other boundaries. For θ_{C1} , these may be written

$$\frac{\partial \theta_{C1}}{\partial x} = B_X, \text{ at } x = x_1, x_2; \quad (\text{A.12})$$

$$\frac{\partial \theta_{C1}}{\partial y} = B_Y, \text{ at } y = y_2; \quad (\text{A.13})$$

$$\frac{\partial \theta_{C1}}{\partial z} = w(B_T - \mathbf{V}_H \cdot \mathbf{B}_H)/(w^2 + w_0^2), \quad (\text{A.14})$$

at $z = z_1, z_2$.

w_0 (here 0.2 m s^{-1}) in (A.14) is a constraint to avoid a divergence of vertical gradient of θ_{C1} when w is close to 0. For Π_1 , the boundary conditions are

$$\frac{\partial \Pi_1}{\partial x} = A_X, \text{ at } x = x_1, x_2; \quad (\text{A.15})$$

$$\frac{\partial \Pi_1}{\partial y} = A_Y, \text{ at } y = y_2; \quad (\text{A.16})$$

$$\frac{\partial \Pi_1}{\partial z} = A_Z + \frac{g\theta_{C1}}{C_p\theta_0\theta_{V0}}, \text{ at } z = z_1, z_2. \quad (\text{A.17})$$

B. Empirical q_R - Z and V_T - Z relationships

The rain water mixing ratio q_R in Eq. (2) and Eq. (3) is estimated from the radar reflectivity after Hauser and Amayenc (1986), while the terminal fall velocity V_T in Eq. (2) is estimated after Rogers (1964) with correction for air density (Foote and du Toit, 1969):

$$q_R = 1.73 \times 10^{-6} Z^{0.613} \rho^{-1} \quad (\text{B.1})$$

and

$$V_T = -3.8 \left(\frac{\rho_0}{\rho} \right)^{0.4} Z^{0.0714}, \quad (\text{B.2})$$

where Z is expressed in $\text{mm}^6 \text{ m}^{-3}$, and ρ_0 is the air density at the surface. As stated in Ishihara *et al.* (1995), it is appropriate to add 7 dBZ to the reflectivity value of the MRI radar when we discuss its absolute value. So we used the modified reflectivity value in calculating q_R and V_T .

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2 台のドップラーレーダー観測からリトリートされた 梅雨前線に伴うレインバンドの熱力学的構造と維持機構

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1988 年 7 月に実施された梅雨末期集中豪雨の特別観測期間中、梅雨前線に伴ったレインバンドが九州を通過し、その内部の風と降水のデータがドップラーレーダーのデュアルモード観測により得られた。この論文の目的は、そのデータにリトリート法を適用することによりレインバンドの熱力学的構造を調べ、維持機構を明らかにすることである。

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このレインバンドは中緯度や熱帯のスコールラインと似た構造を持ち、北東から流入した乾燥空気が強雨域に発達した対流規模下降流と合流し、下層で発散していた。南西からの暖湿な流れは leading-edge に向かった発散流と収束し、後方に傾いた対流規模の上昇流を形成していた。

リトリーバルの結果、対流規模上昇流の高度 4 km 以上で気温の正の偏差域が示され、ガストフロントの後方に乾燥空気中での雨滴の蒸発により形成された冷氣プールが示された。気温場と流れの場を反映し、下層のガストフロント後方には気圧の正の偏差域が、対流規模上昇流の下側には気圧の負の偏差域が示された。また層状性領域の下層には、後部から流入した乾燥空気の下降による断熱加熱を反映し、低圧域が示された。この気温と気圧の偏差の分布から、対流規模上昇流は高度 4 km 付近まで上向きの気圧傾度力によって、それ以上の高度では正の浮力によって維持されていたといえる。また対流性領域中層の低圧域に伴う水平気圧傾度力は上昇流を後方に傾ける役割をしていた。後方に運ばれた雨滴は上昇流を妨げることなく対流規模下降流中に落下し、強いレインバンドが維持された。