Thermal Stratification in Baiu Frontal Medium-Scale Disturbances with Heavy Rainfalls

By T. Akiyama

Meteorological Research Institute, Tokyo 166, Japan
(Manuscript received 6 February 1979, in revised form 29 August 1979)

Abstract

Thermal stratification in medium-scale (wave length of ~1,000 km and period of ~20 hour) disturbances in the Asian subtropical humid region is studied based on dense upper observation data. Attention is focussed on thermodynamic process associating with variation of instability in the lower troposphere (600~900 mb).

Generation of instability in the southeast quadrant of the disturbances was mainly due to the rise of equivalent potential temperature, which was primarily attributed to advection of high \( \theta_e \) with southerly wind in the lower layer (~900 mb). While the release of the generated instability is mainly due to differential heating in the narrow (~200 km) area of active convections in the disturbances.

Cumulus mass flux, associated stability change and the height of cloud top in the unstable area of the disturbances are estimated by applying the cumulus parameterization scheme to the observed soundings.

Soundings in the Asian subtropical humid region are compared with those in tropical squall lines in VIMHEX and GATE areas.

1. Introduction

Over many regions of the middle latitudes as well as over the tropics, organized convective systems sometimes develop. Since the development of convection is strongly influenced by thermal stratification of air mass, many meteorologists studied the stratification and its time variation relating to convective activity.

In the marine tropical region, potential instability in the lower troposphere is a common situation. Organized convections and/or tropical squall lines are usually initiated when the instability is realized by large-scale upward motion associated with tropical wave disturbances (Holton 1972, Betts 1974, Reed et al.; 1977, Houze 1977). Betts et al. (1976) and Mansfield (1977) studied the time change of stratification through the developing and decaying stages of tropical squall lines by comparing the soundings “before” squall lines with those “after” squall lines.

On the other hand, strong potential instability in the lower troposphere is not a common situation in the middle latitude. Strong instability which causes the organized convection is sporadically generated through differential advection under the particular large-scale situations. For example, great instability causing severe storms in U.S.A. is generated by the northward intrusion of the low-level warm-moist air from the Gulf of Mexico and the upper-level cold-dry advection associated with cold trough aloft (Palmen and Newton 1969).

Now we describe features in “the Asian subtropical humid region”. This region spreads from the southeastern part of China to the southwestern part of Japan, which are located in the northwestern periphery of the Pacific subtropical anticyclone in the warm season. A stationary front (Baiu front) is formed along the periphery in the presummer rainy season and sometimes organized convective systems develop in association with frontal depressions. In an observational study on heavy rainfalls in the warm-sector of a Baiu frontal depression, Ninomiya (1978) computed trajectories of air mass and showed that the generation of instability within the warm sector was primarily due to the northward intrusion of
the low-level warm-moist air from the tropical–subtropical Pacific. In the present paper, it is intended to examine the time variation of instability within such frontal depressions.

Akiyama (1978) made a detailed study on radar echo and rainfall features in three medium-scale disturbances† (wave length of ~1,000 km and period of ~20 hour) developed successively in Baiu front in 08~12 July 1968. She revealed that the organized convective echo systems producing heavy rainfalls were found in the limited areas within the disturbances, where instability was realized. In this paper, we treat the case of these three medium-scale disturbances studied in the previous paper. We will focus our attention on the generation of instability before the occurrence of heavy rainfalls and the release of instability in convective areas. In the section 7, the stratification around the heavy rainfall area in the Baiu frontal disturbance will be compared with that around tropical squall lines in VIMHEX and GATE areas.

2. Data source and description of the frontal situation

Data source

Severe Rainstorms Research Project of MRI performed the First Field Experiment over southwest Japan for 4 day period of 00Z July 8~00Z July 12, 1968. The original records of aerological observations made 6-hourly at four rawinsonde and one radiosonde stations are used for the present analysis. The locations of the stations are shown in Fig. 1. The dense (in time and space) observations enabled the detailed analysis of the time variation of thermal stratification associated with the heavy rainfalls in the medium-scale disturbances.

Description of the frontal situation

The Baiu front extended east to west stationarily over the southwestern Japan for the period of 08~12 July 1968 (see Fig. 3 in Akiyama 1978). Three medium-scale disturbances developed successively within the frontal zone with a period of~one day and brought heavy rainfalls over Kyushu (see Fig. 18 in Akiyama 1978). The map of total amount of rainfall for the 4 day period is presented in Fig. 2. Since the north-south displacement of Baiu front was small during the period, the total rainfalls concentrated in a narrow zone of~300 km width over Kyushu. The area of heavy convective rain was found along the southern part of the frontal zone and the area of light continuous (uniform) rain was along the northern part of the frontal zone.

Four stations, 47807, 47843, 47827 and Ryofu (research ship) were located within the frontal (rainfall) zone. Two of them, 47827 and Ryofu, were in the area of heavy convective rain, while

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† See Ninomiya (1973, GARP Publication Series NO 13) for the scale classification of large, medium and mesoscale. The medium and mesoscale will correspond to meso β and meso γ scale in the scale classification of Orlanski (1975).
the other two, 47807 and 47843 were in the area of light continuous rain. The southernmost station 47909 was located in the northwestern rim of the Pacific subtropical anticyclone. At this station, rainfall did not occur, except the last day (July 12) of the experimental period.

3. Variations of rainfall and of stratification associated with medium-scale disturbances

The time variation of hourly rainfalls averaged over all gage stations in Kyushu \((\sim(300\text{ km})^2)\) is presented in the upper part of Fig. 3. Three large peaks of the hourly rainfall (around 09 LST 09, 03 LST 10 and 03 LST 11) were associated with the passage of three medium-scale disturbances mentioned before. It should be noted the large values of averaged rainfall were mainly attributed to the heavy convective rainfalls in the southern half of Kyushu.

Next we will examine the time change in stratification during the analyzed period. The difference between \(\theta_e\) (equivalent potential temperature) in the lower troposphere and \(\theta_e^*\) (saturated equivalent potential temperature) in the middle troposphere is widely adopted as an index of instability of the air mass. The time variations of \(\theta_e\) \((\theta_e \text{ at } 900 \text{ mb})\) and \(\theta_e^*\) \((\theta_e^* \text{ at } 700 \text{ mb})\) at 5 stations for the analyzed period are shown in Fig. 3. The graphs of \(\theta_e\) and \(\theta_e^*\) indicate two kinds of variations; a long-term trend and medium-scale variations superimposed on the trend. The long-term trend is notable at the northern stations (47807 and 47843), where \(\theta_e\) increased gradually while \(\theta_e^*\) decreased gradually. This trend was associated with the passage of a long wave trough in the middle latitude westerly zone.

The medium-scale variations of \(\theta_e^*\) were evident at all stations. The \(\theta_e^*\) indicated the decrease of \(\sim2^\circ\text{K}/12\text{ hour}\) around the peak period of heavy rainfalls. The appearances of minimum \(\theta_e^*\) at Ryofu seemed to precede those at 47827 about 6 hours.

The medium-scale variations of \(\theta_e\) were large at the southern stations (47827 and Ryofu). The \(\theta_e\) at the stations increased by more than \(5^\circ\text{K}/12\text{ hour}\) around the peak period of heavy rainfalls. On the other hand, the medium-scale variations of \(\theta_e\) at the northern station (47807) were very small. The horizontal gradient of \(\theta_e\) in north-south direction was, consequently, greatly enhanced around the peak period of rainfalls. It is found in Fig. 3 that the maximum \(\theta_e\) at 47909 evidently appeared about 12 hour ahead of those at 47827. This implies that the low-level air with high \(\theta_e\) \((345\sim350^\circ\text{K})\) was transported from the south toward the central portion of the medium-scale disturbances in the frontal zone.

Potentially unstable stratification \((i.e., \theta_e^*\sim700)\), indicated by stippled portions in Fig. 3) appeared at stations 47827 and Ryofu for the period from the beginning to the peak of heavy rainfalls. It is an important fact that the generation of instability is mainly due to the rise of \(\theta_e\) at 900 mb, and partially due to the decrease of \(\theta_e^*\) at 700 mb.

Potentially unstable layers also appeared at the southernmost station 47909 with \(\sim\text{one day period}\), though rainfalls did not occur there, except July 12. As described before, this station was not located in the frontal zone but in the northwestern
part of the Pacific subtropical anticyclone, and the potential instability in this region was not realized because the upward motion did not predominate there.

4. Contribution of low-level warm-moist advection to generation of instability

In the observational study on the air-mass transformation over the Kuroshio region (AMTEX area), Ninomiya and Akiyama (1976) analyzed the stability change of air mass by estimating each term in the following equation,

\[
\frac{\partial}{\partial t} \left( - \frac{\partial \theta_e}{\partial p} \right) = \frac{\partial}{\partial p} \left( \nabla \theta_e \right) + \frac{\partial}{\partial p} \left( \bar{u} \frac{\partial \theta_e}{\partial t} \right) \tag{1}
\]

Terms (a), (b), (c) and (d) in eq. (1) express local time change of stability, differential (horizontal) advection, differential (vertical) advection and differential heating of \( \theta_e \) respectively.

The similar analysis will be useful to understand the process of stability change within the frontal zone of the present study. In the present case, however, upper wind observation was not made at ship Ryofu and there were some missing data of wind aloft observation at 47843. Therefore complete estimation of these terms in eq. (1) is impossible. We will only discuss the contribution of low-level warm-moist advection due to meridional wind component \( (v \theta_e / \partial y) \) to the generation of instability within the frontal zone.

Fig. 4 indicates the time variations of meridional wind component at 900 mb averaged on 47807 and 47827 \( (\psi_{900}) \), difference of \( \theta_{e900} \) between 47807 and 47827 \( (\Delta \theta_{e900}) \) and \( \theta_e \)-advection due to meridional wind component \( (v \theta_e / \partial y) \). The distance between the both stations, \( \Delta y \), is \( \sim 260 \) km (see the locations of stations in Fig. 1).

It is an important fact that \( \theta_{e900} \) increased significantly at the southern station around the peak period of areal averaged rainfall, while the variation of \( \theta_{e900} \) was small at the northern station. The difference of \( \theta_{e900} \) between 47807 and 47827, therefore, rapidly increased around the peak period of rainfall (see Fig. 4). It is also found that the variations of \( -\Delta \theta_{e900} \) and of \( \psi_{900} \) were in phase. This indicates that the air with high \( \theta_e \) was transported from the south into the frontal zone. Consequently the three predominant peaks of \( -(v \theta_e / \partial y)_{900} \) appear almost concurrently with the peaks of areal averaged rainfall. It is found that the largest peak of rainfalls (around 03 LST July 10) followed the strongest peak of \( -(v \theta_e / \partial y)_{900} \).

We note that the estimated magnitude of \( (v \theta_e / \partial y)_{900} \) is \( -0.5 \sim -1.0^\circ K/hr \). This negative value of \( (v \theta_e / \partial y)_{900} \) works to destabilize the stratification in 600~900 mb layer at the rate of \( -0.2 \sim -0.3 (^\circ K/100 mb)/hr \), and seems to explain a large part of the observed stability change \( \{\partial/\partial t(\theta_e/\partial p)\} \).

Of course, there are contributions from the other terms in eq. (1) (i.e., \( u \theta_e / \partial x \), \( w(\theta_e / \partial p) \) and \( \theta_e / \partial t \)) to change of stability, and the evaluation of them is necessary to describe the process of the stability change conclusively. Although we were unable to describe the thermodynamic process of stability change completely because of the lack of data, it is still possible to say, from the results shown in Fig. 4, that the intrusion of the warm-moist tropical air by the southerly wind plays an important role for the generation of instability in the medium-scale disturbances.

\[\text{Fig. 4 Upper: Same as the upper part of Fig. 3. Lower: Time variation of } v \theta_e / \partial y, \Delta \theta_e \text{ (difference of } \theta_e \text{ between 47807 and 47827) and } v \text{ (meridional component of wind averaged on the both stations) at 900 mb. Shaded area indicates contribution to destabilization of stratification.}\]

\[\text{\small \# Since the horizontal gradient in the north-south direction of } \theta_e \text{ in 500~700 mb layer is very small (see profiles in Fig. 6), } v \theta_e / \partial y \text{ at the middle level is also small.}\]
5. Details of stratification change in the medium-scale disturbance

As seen in Figs. 3 and 4, the three medium-scale disturbances are different each other, to some degree, in the features of instability, low-level $\theta_e$ advection and rainfalls. The second disturbance, on July 10, is characterized by the most intense rainfalls, most unstable stratification and largest value of $-(\nu \partial \theta_e / \partial y)_{900}$. Therefore detailed analysis on stratification change will be made on this disturbance.

The density of aerological observations is neither sufficient in time to make reliable time section analysis at a certain station, nor sufficient in space to make detailed analysis on a certain synoptic map. We therefore, made “composite time section” by using the data at the stations in the experimental area together. For this purpose, we arranged the observations in the positions relative to the disturbance by means of time-to-space conversion technique. Since the disturbance propagated eastward at a phase speed of $\sim 50$ km/hour, the horizontal distance of 100 km in east-west direction was converted into time lag (or lead) of 2 hour. In Fig. 5 are illustrated the positions of soundings ($k, l, m, n, K, L, M$ and $N$) with the time lapse composite echo map.

We should note again, that Fig. 5 is a “time section over the experimental area” obtained by composite technique. Though the medium-scale disturbance propagated from the west, the rainfall and radar echo associated with the disturbance were remarkably enhanced when it approached to Kyushu (Akiyama 1978). Therefore, Fig. 5 should not be considered as “a synoptic map”.

The features of radar echoes will be briefly described. Stratus echoes which brought light continuous rain spread in the northern portion of the disturbance. Organized convective echoes which brought intense convective rain over the southern half of Kyushu appeared at the central portion of the disturbance. It is notable fact that the domain of active convection was narrow ($\sim 200$ km)$^2$ as compared with the horizontal scale of the medium-scale disturbance (see Akiyama, 1978).

Potential instability is measured by $-(\partial \theta_e / \partial p)$. To estimate buoyancy of the parcel, however, $\theta_e$ in the lower layer should be compared with $\theta_e^*$ in the higher layer. The realization of potential instability requires also a saturation. Therefore we should examine the vertical profile of $\theta$, $\theta_e$ and $\theta_e^*$ to understand completely the instability of the air mass (see Bates et al. (1972)). Vertical profiles of $\theta$ (potential temperature), $\theta_e$ (equivalent potential temperature) and $\theta_e^*$ (saturated equivalent potential temperature) at points $k, l, m, n, K, L, M$ and $N$ are presented in Fig. 6.

The profiles at points, $k, l$, and $n$ indicate the stratification in the area of stratus echoes in the northern portion of the disturbance. The nearly saturated shallow ($\sim 600$ mb) layer with neutral stratification ($-(\partial \theta_e / \partial p) \sim 0$) is found around 600 mb at $k$ and $l$. The layer above the neutral layer is moist and stable, while the layer below is dry and stable and $\theta_e$ in the lower layer does not exceed $\theta_e^*$ in the higher layer. This stable stratification accounts for the stratus echo and light continuous rain.

Profiles at point $L$ represent the situation before the passage of active convective echoes. The observed value of $-(\partial \theta_e / \partial p)$ in 600–900 mb layer was as much as $-3^\circ$K/100 mb. The analysis in section 4 indicated that $\nu \partial \theta_e / \partial y$ of $\sim -1^\circ$K/hour contributed to generation of instability. The intense convective rain initiated at point $L$, $\sim 90$ min later than the sounding time. In other words, the sounding $L$ indicates the conditions just before the initiation of intense convective rainfalls. It is an important fact that the heavy rainfalls are not initiated until instability is accumulated up to a certain level. From the analysis of the present case, the critical level of instability ($-(\partial \theta_e / \partial p)$ for the initiation of deep convection is estimated to be $-2 \sim -3^\circ$K/100 mb in the nearly saturated lower layer.†

Profiles at point $M$ represent the situation in the southern boundary (inflow side) of the active system developed in the warm-sector of a Baiu frontal depression of June 27, 1972 (Ninomiya and Yamazaki, 1979).

† Similar features were found for the convective system developed in the warm-sector of a Baiu frontal depression of June 27, 1972 (Ninomiya and Yamazaki, 1979).
Fig. 6 Upper and middle: Vertical profiles of $\theta_c$, $\theta_e$, and $\theta_e^*$ at sounding points k, l, m, n, K, L, M and N (see Fig. 5). Lower: Vertical profiles of static energy $S$, moist static energy $H$ and saturated moist static energy $H^*$ at points K, L, M and N. Cumulus mass flux and rainfall amount estimated by applying Arakawa-Schubert cumulus parameterization scheme are also illustrated. In the figure, $H_M$ indicates mean moist static energy in planetary boundary layer (PBL), $MB$ and $R$ indicate total mass flux at cloud base and total rainfalls estimated from the scheme.

convective echo area, while profiles at point $m$ represent the situation in the northern boundary (outflow side). The deep saturated layer (400~1,000 mb) with unstable stratifications ($-\frac{\partial \theta_e}{\partial p} \approx -3^\circ K/100 mb$) was found at $M$, while the deep nearly neutral ($-\frac{\partial \theta_e}{\partial p} \approx 0^\circ K/100 mb$) saturated layer was found at $m$ except the stable lowest 100 mb. The difference between soundings $M$ and $m$ indicates the release of instability due to active convection, or the convective mixing of $\theta_e$ within the convective area, except in the lowest 100 mb.

Profiles at $N$ show the stratification after the passage of active convective echoes in the central portion of the medium-scale disturbance. Though $\theta_e$ is slightly decreasing with height ($-\frac{\partial \theta_e}{\partial p} \approx -2^\circ K/100 mb$, convectively unstable), the lower troposphere is dry and $\theta_e$ in the lower layer hardly exceeds $\theta_e^*$ in the higher layer. The rise of $\theta_e^*$ and decrease of relative humidity around 800 mb would suggest the downward motion behind the convective area.

There was a weak convective echo zone along the trailing portion of the disturbance (Fig. 5). We can not give any description about this situation, because no aerological data were available there.

6. Analysis based on a cumulus parameterization scheme

Estimation of cumulus mass flux and rainfall

In this section we estimate rainfall amount, cloud height and cumulus mass flux induced from the release of instability, by applying Arakawa-Schubert (1974) cumulus parameterization scheme to soundings in Fig. 6. We use the “subroutine of 10-layer discrete model” prepared by Yamazaki† for the general circulation model of Meteorological Research Institute. The obtained results for soundings $K$, $L$, $M$ and $N$ are illustrated in the lower part of Fig. 6. The vertical profiles of static energy $S$, moist static energy $H$ and saturated moist static energy $H^*$ are also presented. In the figure, $H_M$ indicates mean moist

† The analysis of the heavy rainfall using this subroutine was made by Ninomiya and Yamazaki (1979).
static energy in PBL (planetary boundary layer††). By this subroutine, the cumulus mass flux at cloud base and associated rainfall amount are calculated for each cloud identified by detrainment level. (The detrainment level corresponds to the height of cloud top.)

The results indicate that mass flux, rainfall amount and cloud height increase from point K towards point M, as the instability increases. The largest mass flux of $\sim 210$ kg/m$^2$ is obtained at points L and M. The highest cloud top (240 mb) and the largest rainfall amount (5.4 mm) are obtained at point M.

It is a notable fact that no substantial rainfall reached the ground at point M (see Fig. 14 in Akiyama, 1978), though strong convective echoes were observed there and the largest rainfall was also estimated from the cumulus parameterization scheme. This suggests that the area of the most intense rainfall (reaching the ground) shifted leeward from the area of the maximum instability.

Any substantial cumulus mass flux and rainfall amount are not calculated at point N, since the convective instability had been almost released. In actual, the heavy rainfall did not occurred there.

It should be noted that the estimated mass flux and rainfall amount are not for "per unit time" but for "per an adjustment". "Adjustment" means the adjustment of thermodynamic situation from the given unstable stratification to the neutral one. We will discuss later the meteorological meaning of "per an adjustment".

The changes in equivalent potential temperatures ($\Delta \theta_e$) caused by an "adjustment" for soundings L and M are presented in Fig. 7. The increase of $\theta_e$ in 300~600 mb and the decrease of $\theta_e$ in 600~1,000 mb indicate the release of instability, though it is not given in "per unit time" but "per an adjustment". As seen in Fig. 7, the values of $\Delta \theta_e$ at 880 mb and 560 mb are about $-0.4 ^o K$/adjustment and $0.8 ^o K$/adjustment respectively. The magnitude of differential heating $\partial \theta_e / \partial p (\Delta \theta_e)$ in 560~880 mb layer is then evaluated as

$$\frac{\partial}{\partial p} (-\Delta \theta_e) = \frac{0.4 ^o K \rightarrow (-0.8 ^o K)}{320 \text{ mb}} \approx 0.4 (^o K/100 \text{ mb})/\text{adjustment}.$$  

Positive sign means, of course, the stabilization of stratification.

**Discussion**

It was noted previously that the estimated cumulus mass flux and rainfall amount shown in Fig. 6 are not for "per unit time" but "per an adjustment". In the general circulation model or numerical prediction model, the "adjustment time" is equal to the time interval between time steps on which the parameterization scheme is employed. In the model, the instability generated by the large-scale process during the interval is released by the cumulus mass flux.

In the present analysis, the generation of instability due to the large-scale process is not evaluated (see the discussion in section 4). Therefore we will try to make the crude estimation of the cumulus mass flux and associated stability change "per unit time" by the following speculations.

We described previously that the soundings at L and M represent the thermal situations in the convective area within the central portion of medium-scale disturbance. The mean amount of...
Table 1. The rainfall amount, cumulus mass flux and changes in $\theta_e$ averaged over L and M.

<table>
<thead>
<tr>
<th>Rainfall amount</th>
<th>4 mm/adjustment</th>
</tr>
</thead>
<tbody>
<tr>
<td>Cumulus mass flux</td>
<td>210 kg m$^{-2}$/adjustment</td>
</tr>
<tr>
<td>$\Delta \theta_e$ at 560 mb</td>
<td>0.8 K/adjustment</td>
</tr>
<tr>
<td>$\Delta \theta_e$ at 880 mb</td>
<td>-0.4 K/adjustment</td>
</tr>
</tbody>
</table>

This indicates the differential heating of $\theta_e$ due to the convective motion contributes greatly to the stabilization of the stratification in the heavy rainfall area.

Table 2. The amount of cumulus mass flux and changes in $\theta_e$ for 1 mm of rainfall calculated from Table 1.

<table>
<thead>
<tr>
<th>Cumulus mass flux</th>
<th>50 kg m$^{-2}$/1 mm of rainfall</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\Delta \theta_e$ at 560 mb</td>
<td>0.2 K/1 mm of rainfall</td>
</tr>
<tr>
<td>$\Delta \theta_e$ at 880 mb</td>
<td>-0.1 K/1 mm of rainfall</td>
</tr>
</tbody>
</table>

Comparison with tropical squall lines

The stratification around the convective area in subtropical medium-scale disturbances of the present study is compared with stratification around tropical squall lines. This comparison will make the characteristics of organized convective systems in the Asian subtropical humid region more clear.

Comparison of stratifications

We refer to the study by Betts (1974) based on VIMHEX II. In the analysis on soundings at Carrizal (9°23’N, 66°55’W), Venezuela, he classified them into four categories according to daily rainfall amount. The “disturbed day” was defined as the day with rainfall exceeding 5 mm. From his paper, averaged profiles of $\theta$, $\theta_e$ and $\theta_e^*$ for disturbed 21 days (91 soundings) are presented in Fig. 8. Although the stratification in the lowest 150 mb (850~SFC) is potentially unstable, relative humidity there is considerably low.

Fig. 8 Tropical sounding averaged on “disturbed” 21 days (sounding number 91) Carrizal, Venezuela in the rainy season (after Betts, 1974).
Fig. 9 Soundings around tropical squall lines. Averaged profiles of mixing ratio $q$ and equivalent potential temperature $\theta_e$ "before" and "after" passage of squall lines on 4 cases in VIMHEX (upper figure, after Betts et al., 1976) and on 3 cases in GATE (lower figure, after Mansfield, 1977). Wet bulb potential temperature $\theta_w$ is converted into equivalent potential temperature $\theta_e$.

We also refer averaged soundings around tropical squall lines in VIMHEX II (Betts et al., 1976) and GATE phase III (Mansfield, 1977). They made averaged soundings "before" and "after" passage of squall lines. ("Before" and "after" soundings correspond mostly to "inflow" and "outflow" ones relative to squall lines respectively in their cases.) These soundings are presented in Fig. 9. For comparison, two soundings in the present case are also illustrated in Fig. 9. The one is the sounding at ship Ryofu, which is a typical one "in an active convective area within the medium-scale disturbance in Baiu front". The other is the sounding at 47909 (Naze), which is a typical one "in the inflow air to Baiu front". (Naze is located at the northwestern rim of Pacific subtropical anticyclone, along which the moist unstable air mass is transported toward the frontal zone from tropical–subtropical moisture source regions.)

Comparisons are made for the following pairs; (1) before squall line vs. inflow to Baiu front, (2) tropical disturbed day vs. in Baiu front and (3) after squall line vs. after heavy rainfall (profile N in Fig. 6). (1) The vertical gradients of $\theta_e$ of "inflow to Baiu front" and "before squall line" are nearly equal in the lower troposphere while $\theta_e$ of the former is significantly higher ($\sim 10^1 K$) than that of the latter. This difference is accounted for mainly by the difference of mixing ratio $q$ ($\sim 5 g/kg$). It is found that the low-level inflow air to the Baiu front is extremely moist as compared with the "before squall" tropical air mass in VIMHEX and GATE areas. (2) The tropical air mass in "disturbed day" (Fig. 8) is relatively dry (not saturated) while the "in Baiu front" air is very moist (almost saturated) throughout the whole troposphere. (3) Though the vertical gradient of $\theta_e$ of sounding N is nearly equal to that of "after squall line" soundings in the lower troposphere, the mixing ratio $q$ and $\theta_e$ of the former are still higher ($q, 2\sim 4 g/kg; \theta_e, 5\sim 10^1 K$) than those of the latter (though mixing ratio at N is not shown here).

From the above comparisons, it is emphasized that the low-level "inflow" air to Baiu front is extremely moist and unstable, and that the air in the subtropical convective system associated with the medium-scale disturbance is nearly saturated through the troposphere. This nature of the air mass in the Asian subtropical humid region suggests that the moist updraft air in convective systems is not diluted by entrainment of dry air and the cooling of downdraft air due to the evaporation of falling water substances is very weak.

Comparison of the changes in stratifications

We examine the changes in stratification caused by tropical squall line by comparing the "after squall line" sounding with "before squall line" one in Fig. 9. The "after squall line" sounding indicates the large decrease of $\theta_e$ ($\sim 10^1 K$) in 700~1,000 mb layer and the large increase of $\theta_e$ ($\sim 10^1 K$) in 700~400 mb layer. It is an important fact that the absolute value of increased
\( \theta_e \) in the upper layer was approximately equal to that of decreased \( \theta_e \) in the lower layer. This indicates the redistribution of heat energy in "an air column" due to the convective transport of heat energy. It is reasonable to infer that, in the region of tropical squall lines, the decrease of \( \theta_e \) in the lower layer is not immediately counterbalanced by the horizontal advection. This seems to be one of characteristics of the tropical system. Therefore we may mentioned that "tropical squall line" is a convective system in an air mass where the large-scale advection is not dominant.

Next we examine the changes in stratification caused by heavy convective rainfalls in the Baiu frontal zone by comparing the sounding \( N \) with the sounding \( L \) (see Fig. 6). The sounding \( N \) indicates the large increase of \( \theta_e \) (~10°K) in 800~500 mb layer. The low-level (1000~800 mb) \( \theta_e \), however, does not indicate large decrease. This feature in the lower layer is evidently different from the feature of the tropical squall line. A possible explanation is that the decrease of \( \theta_e \) in the lower layer, which is due to the convective transport of heat energy, is immediately counterbalanced by the large-scale advective process and therefore the value of low-level \( \theta_e \) does not change significantly. Then it may be noted that "heavy rainfalls" of the present case are convective systems in the frontal zone, where the large-scale low-level advection is always dominant.

8. Concluding remarks

Three medium-scale disturbances (with wave length of \(~1,000 \text{ km and period of } \sim 20 \text{ hour}\)) developed successively in Baiu front (the stationary front in the Asian subtropical humid region) in 08~12 July 1968. Heavy convective rainfalls were brought about over Kyushu (the southwestern part of Japan) when the disturbances passed over. The time variation of stratification around the heavy rainfall areas was investigated by using dense radiosonde observation data. We focussed our attention on the thermodynamic process in generation and release of instability in the lower troposphere (900~600 mb).

Results of the analysis are summarized as follows:

1) Convective instability in the lower layer was generated in the southeastern part of medium-scale disturbances.

2) The generation of convective instability was mainly due to the rise of \( \theta_e \) in the lower layer (900 mb) and partly to the decrease of \( \theta_e \) in the middle layer (700 mb). The rise of \( \theta_e \) in the lower layer was confined within the southern part of disturbances.

3) The maximum value of \(- \partial \theta_e / \partial y\) was found around the peak period of heavy convective rainfalls. The large (negative) value of \(- \partial \theta_e / \partial y\) seems to account for a large part of observed instability, though we did not make complete evaluation of thermodynamic equation.

4) The instability \((- \partial \theta_e / \partial p\) reached to the value of \(-3°K/100 \text{ mb}\) in front and south of the convective area. The intense convective rain did not initiate until the instability exceeded a certain (critical) level (e.g., \(- \partial \theta_e / \partial p \sim -3°K/100 \text{ mb}\)).

5) By comparing the "sounding before heavy rainfall" with the "sounding after heavy rainfall", we found the significant rise of \( \theta_e \) (~10°K) in 800~500 mb layer of the latter. This indicates the release of instability or convective transport of heat energy in the heavy rainfall area.

6) Cloud height, cumulus mass flux, rainfall amount and change of \(- \partial \theta_e / \partial p\) within the unstable layer were estimated by applying Arakawa-Schubert cumulus parameterization scheme to the observed soundings within the central part of the disturbance. The estimated height of cloud top was \(~240 \text{ mb}\). The calculations indicated that when convective rainfall of 1 mm is brought, cumulus mass flux and change in \(- \partial \theta_e / \partial p\) are \(~50 \text{ kg/m}^2\) and \(~0.1°K/100 \text{ mb}\) respectively. As the observed peak value of the areal averaged rainfall over the central part of the disturbance was \(~8 \text{ mm/hour}\), the cumulus mass flux and the change in \(- \partial \theta_e / \partial p\) there were estimated to be \(~400 \text{ kg/m}^2\cdot\text{hour}^{-1}\) and \(~0.8°K/100 \text{ mb}\cdot\text{hour}^{-1}\) respectively. The estimated value of \( \partial/\partial t \) (\(- \partial \theta_e / \partial p\)) seems to account for the large part of observed change in stability in the convective area.

7) Soundings of the heavy rainfalls were compared with these of tropical squall lines. After the passage of convective systems, the large increase of \( \theta_e \) in the middle troposphere is a common feature for both heavy rainfalls and tropical squall lines. However the large decrease of \( \theta_e \) in the lower troposphere was not found for the heavy rainfall in the present study, but for the tropical squall lines. This suggests that in Asian subtropical humid region the decrease of \( \theta_e \) in the lower layer associated with convective transport of heat energy is immediately counterbalanced by the large-scale warm-moist advection.
Acknowledgements

The author is grateful to Dr. K. Ninomiya, Electronic Computation Center, JMA, for his helpful suggestions and discussions on this study. She also thanks Mr. K. Yamazaki, MRI, for the permission to use the "subroutine of cumulus parameterization scheme". She acknowledges the comments of the reviewers in preparing the final paper.

References


を，〜40 mb/hour および 〜0.8 (°K/100 mb) hour⁻¹ と評価した。評価された安定度の変化量は，観測された安定度の変化をよく説明しうるものである。

6. このケース（豪雨）の成層変動状況と，tropical squall line のそれとを比較した。対流活動後の中層での \( \theta_e \) の著しい増加は両域に共通している。一方，下層での \( \theta_e \) の著しい減少は，豪雨の場合には見られない。梅雨前線帯では，対流活動による下層の \( \theta_e \) の減少は，前線帯への \( \theta_e \) の大規模水平移流によって埋合されるものと推測される。