Large- and Meso-α-scale Characteristics of Meiyu/Baiu Front
Associated with Intense Rainfalls in 1–10 July 1991

By Kozo Ninomiya

Center for Climate System Research, University of Tokyo, Tokyo, Japan

(Manuscript received 16 August 1999, in revised form 19 January 2000)

Abstract

The large- and meso-α-scale features of the active Meiyu/Baiu front 1–10 July 1991 are studied by utilizing mainly the Geostationary Meteorological Satellite (GMS) IR data and ECMWF re-analysis data. The intense Meiyu/Baiu frontal precipitation occurs over the Yangtze River Basin in association with the westward elongation of the North Pacific subtropical anticyclone. The intense precipitation zone of ~500 km width extends from the eastern foot of the Tibetan Plateau to the western North Pacific. The frontal zone is characterized by intense precipitation, low-level jet stream, nearly moist neutral stratification and strong gradient of equivalent potential temperature θe.

The large-scale confluence/convergence in the frontal zone sustains strong gradient of θe due to the convective transport. The differential advection of θe generates the convective instability against the stabilizing effect of the convection, and thus, the moist neutral stratification is sustained during the period of intense convective rainfalls. The strong condensation heating is one of the factors to sustain the ascent motion in the frontal zone.

The strong low-level convergence in the frontal zone is accompanied by the northward strong ageostrophic wind, which is associated with the strong acceleration along the northwestern periphery of the westward protruding subtropical anticyclone.

The intrusion of the mid-latitude disturbances into the frontal zone, which occurs in the vicinity of the cut off vortex, enhances the precipitation in the frontal zone by inducing the ascent motion, the cold and dry advection.

The largest diurnal variation of convective clouds is found in 90–100°E. Within the heavy rainfall zone, such significant diurnal variation is not seen, whereas the eastward passage of meso-α-scale cloud systems is evident. They form in the heavy rainfall area, and develop into meso-α-scale frontal depressions during propagation along the frontal zone, where the significant baroclinicity is seen within the nearly moist neutral layer in the lower troposphere.

1. Introduction

The large- and meso-α-scale features of the Meiyu/Baiu front and associated severe rainfalls in 1–10 July 1991 over China and the Japan Islands are studied by mainly utilizing the Geostationary Meteorological Satellite (GMS) IR cloud data and ECMWF re-analysis data.

The background of the present study should be noted first. The summer monsoon rainfalls over the East Asia are identified as Meiyu in China and Baiu in Japan, respectively. The Meiyu and Baiu rainfalls are signified by the intense rainfalls concentrated in the west to east oriented narrow rain band forms with the stationary front, which is called by Meiyu/Baiu front.

Many studies have been made on the Meiyu/Baiu rainfalls from the various viewpoints (see review papers, e.g., Ninomiya and Murakami 1987; Tao and Chen 1987; Ding 1991; Ninomiya and Akiyama 1992). Roughly speaking, these studies will be classified into three categories according to the scales discussed.

Studies of the first category mainly discussed the East Asia summer monsoon as a subsystem of the planetary-scale monsoon circulation. These studies paid special attentions on the variation of the precipitation over the whole monsoon region, influence of the heating over the Tibetan Plateau on the monsoon circulation, relation between the Indian and the East Asia monsoon, moisture transport from the southern hemisphere and the 30–50 days oscillation...
in the monsoon (e.g., Murakami et al. 1984; Tao and Chen 1987; Yanai et al. 1992; Nakazawa 1992; Lau et al. 1988). However, such planetary-scale aspects are not the main subjects in the present study.

The researches in the second category paid attention to the structure and formation processes of the Mei/y/Baiu front and influence of larger-scale circulation systems on the front. Several authors (e.g., Matsumoto et al. 1971; Akiyama 1973 and 1975; Ninomiya 1984 and 1999; Ninomiya and Murakami 1987; Ninomiya and Muraki 1986) have shown that the Baiu front is generally characterized by the low-level southwesterly jet stream, strong moisture gradient and the moist neutral stratification. The important influence of the moisture transport around the western periphery of the North Pacific subtropical anticyclone, the confluence/convergence of the westerly monsoon with the easterly over the South China Sea area, and the mid-latitude disturbances on the Baiu front were also pointed out in these papers. Many authors (e.g., Tao and Chen 1987; Ding 1991 and 1992) also pointed out that the confluence/convergence of the westerly monsoon and the easterly over the South China Sea area is an important condition for the occurrence of the intense rainfalls over China. They also pointed out the southwestern and northwestern vortices as the rain producing meso-scale circulation systems over China.

The studies of the third group have been made with special interests on the meso-scale precipitation systems in the frontal zone. Many authors (e.g., Akiyama 1984 and 1989; Kato et al. 1995; Ninomiya and Akiyama 1992; Ninomiya et al. 1988; Takeda and Iwasaki 1987) pointed out that the intense rainfall in the Baiu/Meiyu front is accompanied by the meso-scale disturbances developed under the influence of the various larger circulation systems. Ninomiya and Akiyama (1992) further stressed the multi-scale features of the summer monsoon rainfall.

However, the synthetic and quantitative studies on the meso-α-scale features of whole Meiyu/Baiu front, influence of high-mid latitude and the tropical-subtropical circulation systems on the Meiyu/Baiu front, and the interaction process among the multi-scale disturbances in the Meiyu/Baiu front have not been performed yet. The present paper will study the aforementioned remaining problems in detail.

2. Features of the Asian summer monsoon precipitation in 1991

The Meiyu season in 1991 was characterized by the extremely large precipitation over the Yangtze River basin. The maximum total precipitation in June and July over this region exceeds 1000 mm, which is more than 200% of the climatological amount of precipitation. About 2/3 of the total precipitation is concentrated in short period between 30 June and 12 July 1991 (Ding et al. 1993; Si et al. 1995).

Figure 1 presents the total precipitation in 1–10 July 1991. The main data sources for this map are “Monthly Report of the Surface Meteorological Observations of China; National Climate Center,” “Monthly Report of Meteorological Observation; Korea Meteorological Administration” and “Monthly Weather Report; Japan Meteorological Agency.” In this map, the precipitation over the East China Sea is estimated by subjective interpolation, referring GMS IR cloud data. The rain band extends from the eastern foot of the Tibetan Plateau to the East China Sea and further to the Japan Islands. Its total extension reaches ~4000 km. The precipitation is concentrated within a narrow rain band of ~500 km width with a maximum precipitation of ~600 mm in 10-day over China (~30°N, 115°E). It is a notable fact that the area of very small precipitation less than 10 mm/10day extends to the south of the intense rainfall zone.

The large-scale monsoon circulation in 1–10 July 1991 was characterized by the westward protrusion of the North Pacific subtropical anticyclone. The variation of the monsoon trough, which forms in the southwestern periphery of the Pacific subtropical anticyclone, is closely related with the west-east oscillation of the North Pacific subtropical anticyclone and the variations of the monsoon westerly front (Ninomiya and Kobayashi 1998 and 1999). The variation of the North Pacific subtropical anticyclone is examined by using Geostationary Meteorological Satellite IR data. The “clear area” is identified by small cloud cover of Ch (amount of cloud higher than 400 hPa level, estimated from GMS IR data) less than 0.1, whereas the active convective regions are identified by the large cloud cover of Ch. The “clear area” will coincide with the area of subsidence in the anticyclone. Figure 2 presents...
the maps of 10-day averaged Ch for 21–30 June, 1–10 July and 11–20 July 1991. The clear area of the North Pacific subtropical anticyclone protrudes westward in 1–10 July 1991 and simultaneously the very active convective region, which forms along the western–northwestern rim of the clear area, moves westward. The southerly moisture transport across 20°N increased significantly around 1–10 July 1991 in the monsoon trough region, when the zonal moisture flux convergence is enhanced in the monsoon trough at 10°N in association with the westward protrusion of the North Pacific subtropical anticyclone (Ninomiya 1999).

The feature of the sea surface temperature (SST) during the summer monsoon in 1991 is briefly described. Although El Nino event becomes significant in 1991 autumn, the maximum SST anomaly in July 1991 over the equatorial Pacific in 130–100°W is no more than ~1.5°C. Any significant anomaly in SST is not seen over the Indian and Pacific Ocean in July 1991 except the western equatorial Pacific.

3. Large-scale environment of the Meiyu/Baiu front

The map of 10-day averaged “cloud top equivalent black body temperature” TBB and the topography of the analyzed area is shown Fig. 3A and Fig. 3B, respectively. We identify five major cloud areas within the tropical and subtropical parts of East Asia. The first one (C-1; the Meiyu/Baiu front
cloud zone) is the quasi-stationary long cloud zone elongating from the southeastern foot of the Tibetan Plateau to the western North Pacific. The second one (C-2) extends northeastward from the northern part of Thailand toward the southwestern tip of the Meiyu cloud zone. The third one (C-3) extends from the southern rim of the Tibetan Plateau toward its southeastern foot to merge with the Meiyu cloud zone. Very large diurnal variation is seen in C-3 as discussed in the later section. The forth (C-4) is the cloud area around Bangladesh, while the fifth (C-5) is the cloud zone in the monsoon westerly zone over the Bay of Bengal. Among them, the cloud area C-1, C-2 and C-3 are the main object of the present study.

The 10-day averaged wind velocity at 850 and 700 hPa are shown in Fig. 4A and B, respectively. In the lower troposphere, the Indian monsoon westerly, the circulation around the North Pacific subtropical anticyclone are predominant circulation systems in the lower latitudes. The monsoon trough forms over the South China Sea between these circulation systems.

At 850 hPa, the monsoon westerly from the Bay of Bengal is split into two branches over the Indochina Peninsula. The northern branch becomes southwesterly toward the cloud area C-3 over the southern periphery of the Tibetan Plateau. The southern branch extends toward the South China Sea as the westerly. It meets with the southeasterly wind around the western rim of the North Pacific subtropical anticyclone, and yields strong southerly over the monsoon trough region. Cloud areas C-1 and C-2 form in association with this strong southerly. This southerly turns into the southwesterly along the Meiyu front and is intensified along the northwestern rim of the North Pacific anticyclone. The strong southwesterly zone, which is identified as the Meiyu/Baiu low level jet stream, extends along the south side of the Meiyu/Baiu frontal cloud zone (C-1). While the Indian monsoon westerly over the Bay of Bengal and the Indochina Peninsula is confined in the lower layer below ~650 hPa, the southwesterly circulation around the northwestern periphery of North Pacific subtropical anticyclone exhibits deep vertical extension up to 400 hPa.

Over the Asian Continent, the middle latitude strong wind belt at 700 hPa, which corresponds to the lower portion of the polar jet stream, runs along 40–45°N latitude. Around 110°E, the north-south distance between the polar jet stream and the Meiyu/Baiu low-level jet stream is ~2000 km. These two jet streams merge gradually to form a strong wind zone along ~35°N latitude circle in 130–160°E.

Within the northern latitudes, two predominating stationary cut-off cyclones are centered over the Bering Sea (~50°N, 160°E) and over Siberia (~50°N, 115°E), respectively. A blocking ridge (anticyclone) is located between these stationary cyclones. The northerly current from the middle latitudes invades southward in the southwestern side of these stationary cyclones. Another zone of northerly wind is seen along the eastern rim of the Tibetan Plateau. Owing to the influence of the stationary cyclone and the Tibetan Plateau, significant confluence and convergence between the tropical and polar air occurs along the north side of the Meiyu front in 100–120°E. The heavy rainfall zone over the Yangtze River Basin is characterized by a distinct shear between the southwesterly and northeasterly winds.

The 10-day averaged wind field at 200 hPa (Fig. 4C) is characterized by predominant circulation of the Tibetan anticyclone centered over ~27.5°N and 85°E. Its eastern extension, margining

**Fig. 4.** The 10-day averaged wind velocity at 850 hPa (A), 700 hPa (B) and 200 hPa (C) for 1–10 July 1991 (unit in m s⁻¹). In panel A and B, the shading shows the area higher than 3000 m above the sea level.
with the North Pacific subtropical anticyclone, protrudes toward the western North Pacific as the west–east extending ridge axis over ~27.5°N. The easterly wind covers large part of the tropical–subtropical areas south to the ridge axis. The upper tropospheric easterly jet with maximum wind speed of ~25 m s⁻¹ elongates along ~17.5°N. In the northern latitudes to the north of ~30°N, westerly covers large area of the middle latitudes. A maximum westerly zone of the subtropical jet stream with a speed of ~40 m s⁻¹ runs over ~40°N.

The location of the cloud area C-1 over China and C-3 coincides approximately with the latitude of the east–west elongating 200 hPa ridge axis, while in 130–160°E the Baiu frontal cloud zone (eastern portion of C-1) exists just along the southern side of the subtropical jet stream.

The 10-day averaged field of 850 hPa temperature T, 700 hPa specific humidity q and 700 hPa equivalent potential temperature \( \theta_e \) are presented in Fig. 5A, B and C, respectively. At 850 hPa (Fig. 5A), a warm area higher than 24°C spreads to the north of the Tibetan Plateau. The high temperature in this area is due to the strong insulation around summer solstice. Over the Continent, the strong meridional thermal gradient is thus sustained in 40–50°N zone, which approximately coincide with the latitude of the polar jet stream in 700–500 hPa. Over the Japan Sea and the western North Pacific, the large meridional thermal gradient in 35–40°N is sustained between the cold polar maritime air-mass and the warm tropical maritime air-mass. The heavy Meiyu rainfall area over the China is characterized by relatively low temperature less than 21°C, which will be owing to both the shading effect of the thick cloud and the cooling effect of intense rainfall.

At 700 hPa (Fig. 5B), the area of low specific humidity less than 4 g kg⁻¹ spreads to the north of the Plateau. The low specific humidity in this area with high temperature (~9°C at 700 hPa) is owing to low relative humidity of ~40%. The low specific humidity in the polar maritime air-mass over the Okhotsk Sea and the Bering Sea is due to the low temperature (~ −3°C at 700 hPa). The very large specific humidity of ~10 g kg⁻¹ at 700 hPa is seen over the subtropical part of the Continent, especially around the Meiyu heavy rainfall area. The area of high specific humidity protrudes as the moist tongue from the Meiyu heavy rainfall area toward the western Japan along the Meiyu/Baiu frontal zone. The largest meridional q gradient of ~5 (g kg⁻¹) (500 km)⁻¹ is sustained along the frontal zone in 105–120°E.

The map of \( \theta_e \) at 700 hPa (Fig. 5C) indicates similar pattern as seen in the map of specific humidity. The largest meridional \( \theta_e \) gradient of ~15 K (500 km)⁻¹ is found to the north of the heavy rainfall area, where \( \theta_e \) at 700 hPa exceeds ~344 K.

### 4. Features of the Meiyu/Baiu front

The 10-day averaged vorticity field at 850 hPa and 200 hPa are presented in Fig. 6A and B, respectively. At 850 hPa, an arch-shaped zone of large positive vorticity extends from Vietnam to the Yangtze River Basin and further to the western North Pacific. The location of cloud zone C-1 (Meiyu/Baiu cloud zone) and C-2 coincides with this maximum vorticity zone. A zone of strong anticyclonic vorticity appears within northern part of the North Pacific subtropical anticyclone. The especially large anticyclonic vorticity to the south of the heavy rainfall area is because of both the strong curvature of the stream along the westward protruding anticyclone and the large lateral wind shear.

At 200 hPa (Fig. 6B), the zone of large cyclonic vorticity elongates at ~45°N along the north side of
the subtropical jet stream axis. The cyclonic vorticity is maximized in 105°-120°E in association with a large-scale trough. The west–east elongating ridge axis of the Tibetan anticyclone is recognized as the zone of the strong anticyclonic vorticity over ~30°N. It is pointed out in many papers (e.g., Matsumoto et al. 1971; Akiyama 1975; Ninomiya 1984) that the Baiu frontal rainfalls are, in general, maximized to the close south of the upper level jet stream. Although it is true for the Baiu front in this case, the zone of maximum rainfall over China is apart from the upper jet stream axis and located just under the ridge axis at 200 hPa.

The 10-day averaged vertical-p-velocity $\omega$ (Fig. 6C) indicates the strong upward motion concentrated in the Meiyu/Baiu cloud zone. Especially strong upward velocity of $\sim -10$ hPa h$^{-1}$ is found in the heavy rainfalls zone over China. In the present study, the vertical velocity given by ECMWF re-analysis is utilized without any modification.

The meridional vertical cross section along 115°E of the 10-day averaged zonal wind velocity, meridional wind velocity, the vertical-p-velocity $\omega$ and the equivalent potential temperature $\theta_e$ are presented in Fig. 7. This cross section cuts across the zone of intense rainfall, which is confined within 29°-33°N. In the cross section of zonal wind velocity (Fig. 7A), the Meiyu/Baiu low level jet stream is seen as a maximum at $\sim 700$ hPa over $\sim 27.5°$N, while middle latitude polar jet stream is found as another maximum at $\sim 400$ hPa over $\sim 40°$N. The core of the subtropical jet stream and that of the upper monsoon easterly jet is situated at $\sim 200$ hPa and $\sim 100$ hPa/17.5°N, respectively.

In Fig. 7B, the largest southerly wind appears in 925–600 hPa over $\sim 27.5°$N (to the south of the intense rainfall zone), while the maximum northerly wind is seen in 700–500 hPa over $\sim 37.5°$N (to the north of the intense rainfall zone). The most predominate meridional flow in the upper troposphere is the northern current centered at 150 hPa over 22.5–25°N, which corresponds to the outflow from the upper level Tibetan anticyclone. Thus, the strong low-level meridional convergence capped by the strong upper level meridional divergence appears in the intense rainfall zone. The vertical cross section of $\omega$ (Fig. 7C) indicates strong ascent motion concentrated within the intense rainfall zone and significant subsidence in both the north and south side of the ascending motion.

In Fig. 7D, a thick layer with nearly moist neural stratification extends from 950 to 300 hPa within the intense rainfall zone at 30°N. The stratification of the tropical-subtropical air-mass south to the rainfall zone is characterized by the high $\theta_e$ in the lower layer capped by the low $\theta_e$ in 700–600 hPa, which is due to the low specific humidity above the stable layer. In this region, convection in the potentially unstable stratification is inhibited by the subsidence. There is also thin potentially unstable layer capped by the low $\theta_e$ layer to the north of the Meiyu front. The convective instability is released selectively only in the Meiyu/Baiu frontal zone, where the ascending motion predominates.

We next study variations in association with the development and propagation of the frontal disturbances of various kinds. In order to reveal variations within this 10-day period, the map of the standard deviation of the 850 hPa wind velocity, 500 hPa vertical-p-velocity $\omega$ and 850 hPa temperature are presented in Fig. 8A, B and C, respectively. The standard deviation of $T$, for example, is defined at each grid point by

$$\text{STD of } T = \left[ \Sigma (T - T_m)^2 / N \right]^{1/2} \quad (1)$$
Here, $T_m$ is the averaged value within the period and $N$ is the number of the sample (in this case $N = 20$, 12-hourly data in 10-day period).

A zone of large STD of 850 hPa wind velocity and that of 500 hPa $\omega$ are seen in the Meiyu/Baiu frontal zone. In 115–160°E, the zone of maximum STD of $V$ and $\omega$ coincide with that of the 850 hPa relative vorticity. It is a notable fact that STD of $\omega$ already attains its maximum over the intense rain-fall area at $\sim$31°N, 115°E, while STD of $V$ begins to increase gradually in the down stream zone of the intense rainfall zone over China. The STD of $V$ increases eastward and maximum of $\sim$8 m s$^{-1}$ appears over the western part of the Japan Islands. This indicates that some meso-$\alpha$-scale disturbances, which generate in the vicinity of the intense rainfall zone as the convective systems, develop into meso-$\alpha$-scale cyclonic circulation systems as they propagate eastward along the frontal zone. As the propagation speed of these disturbances is $\sim$1000 km d$^{-1}$, in general, the disturbance seems to reach its developed stage within $\sim$1 day after the generation in the intense rainfall region. We will study the features of these disturbances in Section 9 in more detail.

It is also an interesting fact that neither the cloud area C-2 nor C-3 is not accompanied by large STD of $V$, although they are associated with significantly large STD of $\omega$. This indicates that the convective disturbances in these zones do not necessary develop into the disturbance accompanied by the cyclonic circulation.

There are zones of secondary maximum of STD of $V$ and STD of $\omega$ in the northern latitude, especially in the vicinity of the cut-off cyclones located over the Bering Sea and over the Asian Continent. The secondary maximum zone of STD of $\omega$, which elongates from northwest to south east around 115°E suggests the approach of some mid- latitude disturbances toward the Meiyu/Baiu frontal zone.

The distribution of STD of $T$ at 850 hPa indicates different features from that of $V$ or $\omega$. The STD of $T$ is very small in the tropical and subtropical areas, including the intense rainfall zone, where the horizontal thermal gradient itself is small. The largest

Fig. 7. The meridional-vertical section along 115°E of 10-day average u-component of wind velocity (A; in m s$^{-1}$), v-component of wind velocity (B; in m s$^{-1}$), vertical-p-velocity $\omega$ (C; in hPa h$^{-1}$) and the equivalent potential temperature $\theta_e$ (D; in K) for 1–10 July 1991.
value of STD of T appears along the polar frontal zone around 45°N. The area of relatively large STD of T elongates along the eastern rim of the Tibetan Plateau from the mid-latitude to ~35°N. This indicates the influence of the disturbances propagating southward along the eastern rim of the Plateau. Relatively large STD of T in the Baiu frontal zone in 135-160°E indicates the temperature variation associated with the frontal disturbances.

5. Moisture transport and moisture balance in the intense rainfall zone

The 10-day averaged moisture flux $V_q$ at 850 hPa and that of the horizontal moisture flux divergence $\nabla \cdot V_q$ at 850 hPa are presented in Fig. 9A and B, respectively. The moisture transport in this period is characterized by the westerly moisture transport within 10–20°N zone in 80–105°E, and the anticyclonic moisture flux around the North Pacific subtropical anticyclone. The convergence of the Indian westerly moisture flux with the easterly moisture flux along the northwestern rim of the North Pacific subtropical anticyclone results in very strong low-level southerly moisture transport toward the intense rainfall zone from the monsoon trough region.

The strong moisture flux convergence over ~25°N and 95°E is due to the blocking effect of the Himalayan Mountains, while the large convergence in the intense rainfall zone over China is associated with the large influx along the northwestern rim of the North Pacific subtropical anticyclone. There is strong moisture flux divergence over ~22.5°N and 110°E, where strong subsidence takes place.

The 10-day averaged zonal moisture flux divergence ($\partial V_q/\partial x$) and meridional moisture flux divergence ($\partial V_q/\partial y$) at 850 hPa are presented in Fig. 10A and B, respectively. In actual evaluation, such terms are calculated in the spherical coordinate system. The zonal moisture flux convergence is very strong over the South China Sea, where the low-level moist Indian monsoon westerly meets with the easterly moisture flux along the southwestern rim of the North Pacific subtropical anticyclone. This strong zonal moisture flux convergence is accompanied by strong meridional moisture flux divergence over the monsoon trough region and the large southerly mois-

Fig. 8. The map of the standard deviation of 850 hPa wind velocity (A; in m s$^{-1}$), 500 hPa vertical-p-velocity $\omega$ (B; in hPa h$^{-1}$) and the 850 hPa temperature (C; in K) for 1–10 July 1991.

Fig. 9. A: The 10-day averaged moisture flux ($V_q$) at 850 hPa (unit in g kg$^{-1}$m s$^{-1}$). B: The 10-day averaged (div $V_q$) at 850 hPa (unit in 10$^{-5}$ g kg$^{-1}$ s$^{-1}$).
ture flux toward the frontal zone. Thus, the monsoon trough region plays the role of the channel transporting moisture from the tropical oceanic zone to the Meiyu/Baiu frontal zone (Ninomiya 1999).

On the contrary, the meridional moisture flux convergence is very strong within the intense rainfall zone over China. Large parts of the meridional moisture flux convergence are consumed as precipitation, but remaining parts are transported eastward and thus very large zonal moisture flux divergence is found over the intense rainfall zone. In this, the intense rainfall zone (i.e., Meiyu front) has the role of transporting the moisture from the monsoon trough region to the Baiu frontal zone.

The moisture balance and the condensation heating in the intense rainfall area are next studied by evaluating the apparent sensible heat source $Q_S$ and the apparent latent heat sink $Q_M$ by:

$$ Q_S \equiv c_p (p/p_0)^{\gamma} \times (\partial \theta/\partial t + \omega \partial \theta/\partial z + \nu \partial \theta/\partial y + \omega \partial \theta/\partial p) \quad (2) $$

and

$$ Q_M \equiv -L(\partial q/\partial t + \nu \partial q/\partial x + \nu \partial q/\partial y + \omega \partial q/\partial p) \quad (3) $$

The distribution of the 10-day averaged value of $Q_M/c_p$ at 600 hPa is presented in Fig. 11. The zone of maximum $Q_M/c_p$ extends over the intense rainfall zone. The vertically integrated apparent moisture sink $-Q_M/L$ (map is not presented) reaches ~35 mm/day at the center of the intense rainfall zone (~30°N and 115°E).

The vertical distribution of 10-day averaged $-Q_M/L$ (apparent moisture sink) and that of $Q_S/c_p$ obtained at 30°N in 110–120°E (the intense rainfall zone over China) are presented in Fig. 12A and B, respectively. The maximum of the $-Q_M/L$ (~7 g kg$^{-1}$ d$^{-1}$) is seen at 600 hPa, whereas the maximum of $Q_S/c_p$ (~12 K d$^{-1}$) is seen in 500–400 hPa. Consequently, the vertical profile of the 10-day averaged $(Q_S - Q_M)/c_p$ (Fig. 12C) shows negative value (apparent sink of moist static energy) in the lower troposphere and positive value (apparent source) in the upper troposphere. This is due to the convective transport of moist static energy from the lower troposphere to the upper troposphere as discussed in many energy budget analyses (e.g., Johnson 1992; Yanai et al. 1992). In this case, however, the upper-level apparent source of the moist static energy is significantly small as compared with low-level apparent sink. Some errors in evaluation will cause this discrepancy. Here we note only qualitatively that, within the intense rainfall zone, the apparent sink and source of the moist static energy are seen in the lower and upper troposphere, respectively.

For the discussion in Section 6, the following remarks are added. The low-level apparent sink of the moist static energy and upper level apparent source of the moist static energy correspond to the low-level apparent sink of $\theta e$ (negative $d\theta e/dt$) and upper level source of $\theta e$ (positive $\theta e/dt$), respectively. As noted by Ninomiya and Akiyama (1992), $\theta e/dt$ is caused by $-\partial[(\omega^2 \theta e^2)]/\partial p$, which means the vertical divergence/convergence of convective transport of $\theta e$.

6. Maintenance of the Meiyu/Baiu front

Many authors (e.g., Matsumoto et al. 1971; Akiyama 1973; Ninomiya 1984; Ninomiya and Akiyama 1992) pointed out that the Meiyu/Baiu frontal zone is characterized by the strong merid-
ional gradient of the mixing ratio of the water vapor $q$ and equivalent potential temperature $\theta_e$, and also by the nearly moist neutral stratification. In this section, we will study the process by which the strong gradient of $\theta_e$ and the nearly moist neutral stratification are sustained within the frontal zone.

Figure 13A presents the 10-day averaged gradient of $\theta_e$ ($\nabla \theta_e$) at 700 hPa. The very strong gradient of $\sim 30$ K/(1000 km) is found between the heavy rainfall zone and the mid-latitude air mass. The large sink of $(Q_S - Q_M)/c_p$ in the lower troposphere in the intense rainfall zone means the decrease of the low-level $\theta_e$ and the decrease of $\theta_e$ gradient. Some large-scale process is, therefore, required to sustain the large gradient of $\theta_e$ against the aforementioned decrease of $\theta_e$ gradient.

In a previous analysis (Ninomiya 1984), the effects of the large-scale circulation onto the front was studied in terms of the frontogenesis by evaluating $d(\nabla \theta_e)/dt$. In the present study, however, we evaluate $\partial(\nabla \theta_e)/\partial t$ at 700 hPa due to the large-scale process, because the Meiyu/Baiu front remains quasi-stationary at the almost same place for the 10-day period under the consideration.

The local time change of $\nabla \theta_e$ is expressed by

$$\partial[(\partial \theta_e/\partial x)^2 + (\partial \theta_e/\partial y)^2]^{1/2}/\partial t$$

$$= 1/[(\partial \theta_e/\partial x)^2 + (\partial \theta_e/\partial y)^2]^{1/2}$$

$$\times \left\{ [\partial \theta_e/\partial x \cdot \partial (d\theta_e/dt)/\partial x + \partial \theta_e/\partial y \cdot \partial (d\theta_e/dt)/\partial y] - (\partial \theta_e/\partial x \cdot \partial (u\theta_e/\partial x) + u\partial \theta_e/\partial y + \partial \theta_e/\partial y \cdot \partial (u\theta_e/\partial y + v\partial \theta_e/\partial p)/\partial x + v\partial \theta_e/\partial y + \partial \theta_e/\partial x \cdot \partial (v\partial \theta_e/\partial y + \partial \theta_e/\partial p)/\partial y] \right\}$$

$$= 1/[(\partial \theta_e/\partial x)^2 + (\partial \theta_e/\partial y)^2]^{1/2}$$

$$\times \left\{ [\partial \theta_e/\partial x \cdot \partial (d\theta_e/dt)/\partial x + \partial \theta_e/\partial y \cdot \partial (d\theta_e/dt)/\partial y] - (\partial \theta_e/\partial x \cdot \partial (u\theta_e/\partial x) + u\partial \theta_e/\partial y + \partial \theta_e/\partial y \cdot \partial (u\theta_e/\partial y + v\partial \theta_e/\partial p)/\partial x + v\partial \theta_e/\partial y + \partial \theta_e/\partial x \cdot \partial (v\partial \theta_e/\partial y + \partial \theta_e/\partial p)/\partial y] \right\}$$

(4)
discuss the influence of large-scale circulation onto the Meiyu/Baiu front by evaluating

\[
\text{"local frontogenesis due to large-scale circulation"} = -1/\left(\left(\frac{\partial \theta e}{\partial x}\right)^2 + \left(\frac{\partial \theta e}{\partial y}\right)^2\right)^{1/2} \times \left\{ \frac{\partial (\theta u e/\partial x + \theta v e/\partial y)}{\partial x} + \frac{\partial (\theta u e/\partial y + \theta v e/\partial x)}{\partial y} \right\}.
\]

The 10-day averaged value of "local frontogenesis due to large-scale circulation" evaluated at 700 hPa is shown in Fig. 13B. Large value of \(\sim 30 K/(1000 \text{ km}) \text{ day}^{-1}\) is obtained over the northern rim of the frontal zone. This large-scale influence works to sustain the strong gradient of \(\theta e\) against the sink (decrease) of \(\theta e\) in the frontal zone, which is due to the active convection.

Next we examine features of the convective stability in the lower troposphere. Figure 14A shows the map of the 10-day averaged \(-\left(\frac{\partial \theta e}{\partial p}\right)\) evaluated for the layer between 850 and 500 hPa. In general, the oceanic area to the south of \(\sim 30^\circ\text{N}\) and the continental area south of \(45^\circ\text{N}\) are characterized by the convectively unstable stratification. It is an important feature that the Meiuy/Baiu frontal zone associated with strong convective precipitation is not characterized by the strong unstable condition but by the nearly moist neutral stratification. This indicates the stabilizing effect of active convective clouds.

For the sustenance of the strong convective precipitation for \(\sim 10\)-day, some large-scale process must generate the vertical instability against the stabilizing effect of the convective clouds. About this, we will use the following equation to study local time change of the vertical stability,

\[
\frac{\partial}{\partial t} (-\delta \frac{\partial \theta e}{\partial p}) = -\delta \frac{\partial (\delta \theta e/\partial t)}{\partial p} + \delta \theta u \frac{\partial (\theta u e/\partial x)}{\partial x} + \delta \theta v \frac{\partial (\theta v e/\partial y)}{\partial y} + \omega \frac{\partial (\theta e/\partial p)}{\partial p}.
\]

The first term in the right hand, i.e., \(-\delta (d\theta e/dt)/dp\), indicates the stabilization accompanied by the upper level source of \(\theta e\) and low level sink of \(\theta e\). The terms in second parentheses \([\ ]\) in Eq. (6) represent the change of the stratification due to the differential advection of \(\theta e\) in the large-scale circulation.

Figure 14B present the 10-day averaged value of the second term of Eq. (6) evaluated for the layer bounded by 850 and 500 hPa surface. The area of negative differential advection (generation of the convective instability) is seen in and around the Meiuy/Baiu frontal zone. This indicates that the differential advection generates successively convective instability against the release of the instability due to the convection. As the result of these two processes, the large precipitation and nearly moist neutral stratification are maintained within the heavy rainfall zone.

7. Circulation associated with the North Pacific subtropical anticyclone and Tibetan anticyclone

As already shown in Fig. 6, strong ascent motion appears within the Meiuy/Baiu frontal cloud zone, whereas strong subsidence appears in the north and south side of the zone of the ascending motion. In order to relate the distribution of the vertical motion to the large-scale circulation systems, the map of 10-day averaged divergence at 850 hPa and 250 hPa are presented in Fig. 15A and Fig. 15B, respectively.

Figure 6A (10-day averaged wind field at 850 hPa) and Fig. 15A indicate that a pair of strong convergence (northern side) and divergence (southern side) forms in association with the westward protruding North Pacific subtropical anticyclone. Although there will be several interpretations of the low-level convergence/divergence, here we like to relate the ageostrophic wind velocity \((u_a, v_a)\) to the acceleration along the periphery of the subtropical anticyclone by using the following equations,

\[
\frac{\partial u}{\partial t} + \omega^2 u / \partial x + v \omega u / \partial y + \omega u / \partial p = f v_a \quad (7)
\]
\[ \frac{\partial \nu}{\partial t} + u \frac{\partial \nu}{\partial x} + v \frac{\partial \nu}{\partial y} + \omega \frac{\partial \nu}{\partial p} = -fu. \quad (8) \]

The map of 10-day averaged ageostrophic wind velocity at 850 hPa is presented in Fig. 16A. It is clear that the large ageostrophic wind from the subtropical anticyclone toward the frontal zone is associated with the large acceleration along the periphery of the anticyclone where both the anticyclonic curvature and increase of velocity are large. Consequently, a pair of large convergence and divergence appears with the westward protruding anticyclone. Aforementioned ageostrophic wind velocity and convergence/divergence are seen not only at 850 hPa but in the deep tropospheric layer from 925 hPa up to 500 hPa.

Next, the 10-day averaged divergence field at 250 hPa (Fig. 15B) is inspected. Strong high level divergence and convergence appear over the low-level convergence and divergence zone, respectively. The 10-day averaged ageostrophic wind velocity at 250 hPa evaluated from Eq. (7) and Eq. (8) is presented in Fig. 16B. The strong divergence of ageostrophic wind over \(\sim 35^\circ N/115^\circ E\) is associated with the acceleration in the anticyclonic flows with strong curvature along the periphery of the eastward protruding Tibetan anticyclone. The ageostrophic flow from the anticyclone become smaller over the southern coast of China, and consequently, the strong upper-level convergence appears over the low-level divergence area to the south side of the intense rain zone.

The strong condensation heating in the heavy rainfall zone will enhance upper-level anticyclone and produce the strong upper-level divergence over the strong ascending motion. Figure 7C also indicates that significant compensating subsidence occurs in north and south side of the zone of maximum condensation heating. Aforementioned features are consistent with the appearance of zone with very small precipitation in south and north to the intense rainfall zone (Fig. 1).

8. Influence of the mid-latitude circulation systems on the Meiyu/Baiu front

The influence of mid-latitude circulation systems on the Meiyu/Baiu frontal zone has been discussed in several papers (Akiyama 1990; Ding 1991; Ninomiya and Akiyama 1992; Ninomiya and Muraki 1984).

A typical case of such influence is seen on TBB map at 00 UTC 6 July 1991 (Fig. 17A). When a mid-latitude disturbance moves toward the Meiyu/Baiu frontal zone, a large \(\lambda\)-shaped cloud pattern has been accomplished over China and convective clouds are significantly enhanced over \(\sim 32.5^\circ N, 115^\circ E\), where the mid-latitude disturbance is approaching to the frontal zone. The map of potential vorticity

\[ \nabla \times \mathbf{\Omega} = -\frac{1}{\rho_0} \frac{\partial \mathbf{u}}{\partial z}. \]
on isentropic surface of 324 K and the map of div $Q$ at 500 hPa for 00 UTC 6 July 1991 is presented in Fig. 17B and Fig. 17C, respectively. The Q-vector $(Q_1, Q_2)$ is defined by

$$Q_1 = \frac{R}{p} (\partial u_g/\partial x \cdot \partial T/\partial x + \partial v_g/\partial x \cdot \partial T/\partial y) \quad (9)$$

and

$$Q_2 = -\frac{R}{p} (\partial u_g/\partial y \cdot \partial T/\partial x + \partial v_g/\partial y \cdot \partial T/\partial y). \quad (10)$$

These maps indicate clearly that the convective clouds in the frontal zone are significantly enhanced in front of a mid-latitude trough with large potential vorticity, where significant ascent motion is induced.

During the 10-day period, a few mid-latitude disturbances approach successively toward the frontal zone with a period of $\sim$3-day (wave length $\sim$3000 km) and enhance frontal activity. Therefore, the influence of mid-latitude circulation systems is seen not only at the certain map time but also in the mean field. The 10-day averaged Q-vector and that of div $Q$ are presented in Fig. 18 A and B, respectively. The southward Q-vector from the middle latitudes to the intense rainfall zone indicates the influence of the mid-latitude circulation.

9. Meso-$\alpha$-scale circulation systems in the Meiyu/Baiu frontal zone

In the maps of 10-day averaged precipitation (Fig. 1) and that of TBB (Fig. 3), the Meiyu/Baiu front is identified as a quasi-stationary cloud band associated with intense rainfall. It is also seen in the TBB map at 00 UTC 6 July (Fig. 17A), that the cloud/rain band consists of, in actual, several sub-synoptic and/or meso-$\alpha$-scale cloud systems. They has been identified as the Baiu frontal meso-$\alpha$-scale cloud clusters (Akiyama 1984 and 1989; Ninomiya and Akiyama 1992, Ninomiya et al. 1987; Takeda and Iwasaki 1987). In this regard, the Meiyu/Baiu front is considered as the chain of meso-$\alpha$-scale disturbances aligned with 500~1000 km interval. In the present paper we will elucidate the relation between these circulation systems and the Meiyu/Baiu front.
front as a whole.

Figure 19 shows the longitude-time section of the TBB along 32.5°N. In Fig. 19, the diurnal variation in TBB with a maximum in afternoon is evident around 95°E. These maximum areas shift eastward up to ~105°E with time. The feature of the diurnal variation in this period is consistent with the report by Murakami (1983), Akiyama (1989), Kato et al. (1995) and Asai et al. (1998). However, these diurnally generated cloud systems do not necessarily develop into the migrating cloud clusters. Only some of the cloud systems develop into the Baiu frontal cloud cluster selectively. They propagate east-northeastward along the Meiyu/Baiu front with a speed of ~1000 km/day and develop significantly over the intense rainfall area around 115°E. It should be noted that these disturbances do not necessarily remain in the time section at certain latitude, since they move east-northeastward. In 7–11 July, disturbances disappear around 125E from Fig. 19, since they move to ~35°N.

Figure 20 presents the longitude-time section of 850 hPa relative vorticity at 35°N. Eastward propagation of the meso-a-scale circulation systems and their development around 115°E is evident. However, the correspondence of the cloud cluster in Fig. 19 to the respective circulation system in Fig. 20 is not always clear. This is due to the insufficient time and horizontal resolution (12-hour interval and 2.5° lat. and lon. grid) of the data utilized.

The horizontal distribution of the standard deviation (STD) of V at 850 hPa, ω at 500 hPa and T at 850 hPa for the analyzed period were already presented in Fig. 8. The maximum STD of ω at 500 hPa appears over the intense rainfall zone in China. The magnitude of the STD of ω significantly amplified around 110°E. On the other hand, the large value of the STD of V, which means the large amplitude of the circulation, appears along the north side of the Baiu cloud zone, in which small positive value of \((\partial \theta / \partial p)\) is found. This suggests that active convective clouds develop in the convectively unstable area, while the meso-a-scale frontal depressions tend to develop in the zone of weak convectively stable area north to the Baiu cloud zone.

According to the Eddy’s type baroclinic instability model (Gill 1982), the maximum growth rate of the baroclinic wave, \(\sigma_{\text{max}}\), is given by

\[
\sigma_{\text{max}} = 0.310(f/N)(\partial U_g/\partial z)
\]

at the wave number defined by

\[
Nh = 0.803f,
\]

where \(N\), \(k\), \(H\), \(f\) and \((\partial U_g/\partial z)\) are the Brunt-Vaisalla frequency \((N^2 = gd\ln \theta/\partial z)\), wave number in the direction of the thermal wind vector, half-depth of the layer in which the disturbance exists, Coriolis parameter and the vertical shear of the geostrophic wind speed, respectively.

We estimate the wave-length of the most unstable baroclinic wave and its growth rate, from the 10-day averaged quantities in the lower troposphere for the area in 125–140°E. If the potential temperature \(\theta\) is used to evaluate \(N\), the obtained wave-length is ~3100 km and the maximum growth rate is ~5.4 × 10^{-6} \text{ s}^{-1}, which correspond to the e-holding time of ~48 hour. Thus obtained wave-length of the most unstable wave coincides with the wave-length of the mid-latitude disturbances discussed for Fig. 17.

For the discussion of the meso-a-scale Meiyu/Baiu frontal disturbance, the equivalent potential temperature \(\theta e\) be used for the evaluation, since the frontal zone is very moist. In this case, \(N\) in Eq. (11) and Eq. (12) is replaced by \(N*\) \((N*^2 = gd\ln \theta e/\partial z)\). By this, the obtained wave-length of the most unstable wave is ~600 km and its growth rate is
April 2000 K. Ninomiya 155

\[ \sim 2.8 \times 10^{-5} \text{ s}^{-1} \], which correspond to the \( e \)-holding time of \( \sim 10 \) hour. This estimation suggests that necessary condition for formation of the meso-\( \alpha \)-scale Baiu frontal disturbance is the nearly moist neutral stratification along the north side of the Baiu frontal zone. The fact that these circulation systems begin to develop over the area of maximum rainfalls around 115°E, also suggests the important role of condensation heating for the generation of these circulation systems. This inference is consistent with the conclusion of the theoretical discussion on the Baiu frontal wave disturbance (Tokioka 1973).

10. Concluding remarks

The large- and meso-\( \alpha \)-scale features of the active Meiyu/Baiu front in 1–10 July 1991, which brings intense rainfalls over the Yangtze River Basin, are studied by utilizing mainly the Geostationary Meteorological Satellite (GMS) IR data and ECMWF re-analysis data. Important features found in the present study are:

(1) The intense Meiyu/Baiu frontal rainfall occurs in association with the westward protrusion of the North Pacific subtropical anticyclone. This intense rainfall event is characterized by distinct boundary between the intense rainfall area and the area with very small precipitation.

(2) The intense rainfall zone of \( \sim 500 \) km width extends from the eastern foot of Tibetan Plateau to the western North Pacific. The frontal zone is characterized by the low-level southwesterly jet stream, nearly moist neutral stratification, large gradient of equivalent potential temperature \( (\theta e) \).

(3) The large-scale confluence/convergence in the frontal zone sustains strong gradient of \( \theta e \) against the low-level sink of \( \theta e \) due to the convective transport. The differential advection of \( \theta e \) generates the convective instability against the stabilizing effect of the convection, and thus, the nearly moist neutral stratification is sustained during the period of intense convective rainfalls. The strong condensation heating is one of the factors to sustain the ascent motion in the frontal zone and the compensating subsidence outside of the frontal zone.

(4) The intrusion of the mid latitude disturbances into the frontal zone, which occurs in the vicinity of the cut off vortex, enhances the rainfall in the frontal zone by inducing the ascent motion and the cold-dry advection.

(5) The pair of low-level strong convergence (in the intense rainfall zone) and strong divergence (to the south of the intense rainfall zone) is owing to the northward strong ageostrophic wind, which is associated with the strong acceleration along the northwestern periphery of the westward protruding North Pacific subtropical anticyclone.

(6) The strong upper-level divergence over the intense rainfall zone is owing to strong southward ageostrophic wind, which is associated with the acceleration along the southeastern rim of the upper level Tibetan anticyclone enhanced by the strong condensation heating.

(7) The largest diurnal variation of convective cloud is found in 90–100°E, whereas such significant diurnal variation is not seen within the heavy rainfall zone. The eastward passage of meso-\( \alpha \)-scale cloud systems is seen along the Baiu front in 120–160°E. They are intensified in the heavy rainfall area, and develop into meso-\( \alpha \)-scale frontal depressions during propagation along the frontal zone, where the significant baroclinicity is seen within the lower troposphere of nearly moist neutral stratification.

Some of the large-scale features found in the present study are accordant with the results in previous case studies on the parts of Meiyu or Baiu front. The meaning of the present study is in that the meso-\( \alpha \)-scale features of whole Meiyu/Baiu front, influence of high-mid latitude circulation systems and that of the tropical-subtropical circulations are studied synthetically and quantitatively.

Acknowledgments

The GMS data is prepared by “Cooperative research on the tropical convective activity and variability of large-scale circulation” between the Meteorological Research Institute and the Center for Climate System Research. About this, the present author would like to express his sincere thanks to Dr. M. Yamazaki and Dr. T. Nakazawa of MRI. The author extends sincere thanks to the members of HUBEX, especially Prof. J. Matsumoto of the University of Tokyo, Prof. K. Kato of Okayama University and Prof. K. Tsuboki of Nagoya University for providing the valuable observation data.

References


**Journal of the Meteorological Society of Japan**

156

Vol. 78, No. 2


Si, G.-W., K. Kato and T. Takeda, 1995: The early summer seasonal change of large-scale circulation over East Asia and its relation to change of the frontal features and frontal rainfall environment during 1991 summer. Advance in Atmospheric Sciences, Beijing, **12**, 151–176.


1991年7月1〜10日の強い降雨を伴う梅雨前線の大規模及びメソ-α-規模の様相

二宮洗三
(東京大学 気候システム研究センター)

1991年7月1〜10日に揚子江流域を中心とする東アジアに豪雨をもたらした梅雨前線の大規模およびメソ-α-規模の様相と、その維持に寄与する周辺循環系の作用を解析した。

この期間、大西洋亜熱帯高気圧の東西伸張に伴って、梅雨前線は著しく強化された。前線帯下層における水蒸気流束の収束は、大西洋高気圧の北西縁で極大となり、特に南北収束が大きい。これに対し、南シナ海の高気圧圏内では大きな東西収束と南北発散が見られる。前線帯の大きな潜熱放出による熱源は、同時に前線帯の鉛直循環の維持に寄与する。

梅雨前線帯下層の相当温位のシンクは相当温位傾度を弱めるが、大規模場の合流収束場の移流過程は相当温位傾度を強め、両者がほぼ均衡を維持する。また、対流活動は前線帯の鉛直不安定を解消するが、二次元的デファレンシャルアドベクションは鉛直不安定を増加させ、両者がほぼ均衡し豪雨域で湿潤中立に近い成層を維持する。

梅雨前線帯下層の強い収束とその南側の強い発散は、大西洋高気圧西北縁の大きな曲率を持つ流れの加速度に対応する強い非地衡風によってもたらされ、多降水量と寡降水量の著しいコントラストを生じる。

この期間、〜50N、〜110Eに切離低気圧があり、その後面では中高緯度から擾乱が南下し梅雨前線に接近して、梅雨前線帯の対流活動を活発化した。

〜30Nゾーンの90〜100Eでは積雲対流の日変化が大きいが、〜105E以東では東進するメソ-α-規模雲システムが顕著である。それらは下層の低気圧性循環を伴い豪雨域で強化され、梅雨前線の中立に近い湿潤安定成層の傾斜ゾーンを東進しつつ小低気圧に発達する。