A Numerical Study of a Low-Level Jet and Its Accompanying Secondary Circulation in a Mei-Yu System

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(Manuscript received 17 June 1992, in final form 30 August 1993)

ABSTRACT

A primitive equation numerical model has been used to study the Mei-Yu system that occurred in the middle of May 1987. Although cumulus heating was not included in one of the experiments, all of the major features of a Mei-Yu system other than deep convection are reproduced in that experiment. These features are a cloud band along the coastline in southern China; a weak temperature gradient, a strong moisture gradient, and a strong wind shear across the cloud band; and a supergeostrophic low-level jet (LLJ) with an accompanying secondary circulation.

When the moist air coming from the south encounters the migrating high pressure system in the north, it turns to the east-northeast and becomes saturated. The resulting stratus cloud covers all of southern China and the pressure trough at the 850-hPa level deepens by 1.5 hPa due to the latent heat released. It induces convergent motion in the lower level and generates a direct, cross-frontal secondary circulation that helps keep the low-level wind supergeostrophic. Without such large-scale latent heating, the low-level wind would be weaker; the amount of moisture that can be transported to the frontal zone would also be drastically reduced. Once the LLJ forms, it may trigger heavy precipitation in the later stages of the Mei-Yu system. Our study suggests that cumulus heating may not have played a critical role in the formation of the LLJ, at least for the particular Mei-Yu system under consideration.

1. Introduction

The Mei-Yu (plum rain) is a climatological phenomenon affecting a large area of east Asia. It occurs in late spring and early summer and lasts one to two months. The Mei-Yu season is characterized by the presence of a quasi-stationary cloud band extending from southern Japan to southern China. On surface weather maps, the cloud band is associated with a surface front (the Mei-Yu front, or Baiu front) separating the Pacific sub-tropical high to the south and the migratory high over central China. The nature of the "front," however, is quite different from that of a typical polar front. The temperature gradient across the front is usually weak, while the moisture gradient and horizontal wind shear are very large (Akiyama 1973a; Kato 1985). In addition, the surface pressure depression associated with the front is very weak compared with a polar front. When viewed at a particular instant in time, a mature Mei-Yu cloud band is (during peak Mei-Yu season around June) actually composed of many deep convective systems. The rain produced by the convection can sometimes be very heavy (over 100 mm day$^{-1}$) and persists for several days (Chen 1983; Matsumoto et al. 1971). Thus, many people are affected by, and interested in, this phenomenon. The detailed climatological aspects of the system are documented in Ninomiya and Murakami (1987) and Tao and Chen (1987).

One may wonder how the Mei-Yu system maintains its strength over such a long period of time. It is generally believed (Ninomiya 1984; Matsumoto et al. 1971) that a southwesterly low-level jet (LLJ) observed just to the south of the front between 700 and 850 hPa is responsible for the system's duration. Ninomiya (1984) calculated the vertically integrated water vapor transport of the Northern Hemisphere over a seven-day period in June 1975. He showed that a Baiu front is predominant due to the LLJ. The LLJ not only feeds a lot of moisture into the frontal zone but the advective process also contributes to greater potential instability, triggering heavy rainfall in the downstream region of the LLJ (Ninomiya 1980). A statistical study by Chen and Yu (1988) found that 84% of all 1965–84 Mei-Yu heavy rainfall events in Taiwan were ac-
accompanied by an LLJ (defined as wind speed over 12.5 m s^{-1}). They also showed a 91% likelihood of a heavy rainfall event 24 h after the formation of an LLJ over Taiwan.

There have been quite a few studies of the mechanisms involved in the evolution of the LLJ. Early works (Matsumoto and Ninomiya 1971; Matsumoto 1972, 1973; Akiyama 1973b; Ninomiya and Akiyama 1974) hypothesized that the LLJ forms as a result of the cumulus-scale vertical transport of larger horizontal momentum from the upper level to the lower level. The upper-level wind over the Mei-Yu front, though, is usually from the west or northwest, rather than the southwest. Uccellini and Johnson (1979) pointed out that simple momentum exchange cannot account for the direction change between the upper and lower levels.

Uccellini and Johnson (1979) demonstrated that, through a mass--momentum adjustment process, a supergeostrophic LLJ can be generated by coupling with an upper-level jet streak. This theory does not seem to apply to the Mei-Yu LLJ, though, because, in most cases the upper-level wind is not very strong. A distinct jet streak over a Mei-Yu front is rarely observed (Nagata and Ogura 1991), due to the weak temperature gradient across the frontal zone.

A number of studies have concentrated on the effect of convective heating on the formation of the LLJ. Ninomiya (1980) showed that the Baiu front is significantly enhanced when a convective adjustment process is included in his four-level primitive equation model. Without subgrid-scale (∆x = 381 km) convection, the 700-hPa wind speed to the south of the front is much weaker in his simulation. Kuo and Anthes (1982) also showed the importance of the latent heat associated with cumulus convection in maintaining a Mei-Yu cloud band and the low-level cyclonic vorticity over southeast Asia. Chou et al. (1990) simulated the formation of the LLJ with a two-dimensional numerical model. By imposing a deformation wind field (constant with time) as background large-scale forcing (frontogenetical forcing), they obtained a thermally direct secondary circulation to the south of the frontal zone. The lower branch of this secondary circulation turns to its right and becomes a westerly jet (LLJ) due to the Coriolis force. They also stressed the importance of deep cumulus convection in maintaining a secondary circulation and suggested that symmetric instability may be at work in organizing the deep convection.

This paper also investigates the formation of the Mei-Yu LLJ but from a much different perspective. Emphasis is placed on the role of the stratiform cloud over the area. The fact that the occurrence of the LLJ often proceeds the onset of a heavy precipitation event leads us to suspect that cumulus heating may not have played a critical role in producing the LLJ. As we will show later, we are able to reproduce the LLJ observed at 0000 UTC 16 May 1987 over southern China without including cumulus heating in one of the experiments. The latent heating of the large-scale stratiform clouds instead provides the forcing necessary for maintaining a secondary circulation and other features of the Mei-Yu front in the developmental stage of its life cycle.

2. Fundamental equations

The primitive equations in the σ coordinate are applied in this model. Here, σ is defined as

\[
\sigma = \frac{p - p_i}{p^* - p_i} \equiv \frac{p - p_i}{p - p_i},
\]

where \( p_i \) and \( p_i \) are the pressures at the top and bottom of the domain. The equations in σ coordinate are

\[
\frac{\partial u}{\partial t} = \text{adv}(u) + f\nu
\]

\[
- \left[ \frac{1}{\rho} \frac{\partial p^*}{\partial x} + \frac{1}{p^*} \frac{\partial p^*}{\partial \sigma} \left( \frac{\partial \phi}{\partial x} \right) \right] + \text{diff}(u),
\]

\[
\frac{\partial v}{\partial t} = \text{adv}(v) - f\nu
\]

\[
- \left[ \frac{1}{\rho} \frac{\partial p^*}{\partial y} + \frac{1}{p^*} \frac{\partial p^*}{\partial \sigma} \left( \frac{\partial \phi}{\partial y} \right) \right] + \text{diff}(v),
\]

\[
\frac{\partial \theta_c}{\partial t} = \text{adv}(\theta_c) + \frac{L q_w}{c_p} \frac{d}{dt} \left( \frac{p_0}{p} \right)^* + \text{diff}(\theta_c),
\]

\[
\frac{\partial q_w}{\partial t} = \text{adv}(q_w) + \text{diff}(q_w),
\]

\[
\frac{\partial \phi}{\partial (\ln p)} = -R_d f (1 + 0.61 q_w - q_i),
\]

\[
\frac{\partial p^*}{\partial t} = -\int_0^\sigma \nabla \cdot (p^* \mathbf{V}) d\sigma,
\]

\[
\dot{\sigma} = -\frac{1}{p^*} \int_0^\sigma \nabla \cdot (p^* \mathbf{V}) d\sigma + \frac{\sigma}{p^*} \int_0^\sigma \nabla \cdot (p^* \mathbf{V}) d\sigma,
\]

where \( \theta_c = \theta + (L/c_p) (\theta/T) q_i \) is the equivalent potential temperature, \( q_w = q_i + q_1 \) (where \( q_1 \) is specific humidity and \( q_i \) is the liquid water content), and \( p^* \) is the perturbation pressure field calculated from a reference atmosphere at the same height. The advection operator \( \text{adv}(\cdot) \) is defined as

\[
\text{adv}(\cdot) = u \frac{\partial (\cdot)}{\partial x} + v \frac{\partial (\cdot)}{\partial y} + \dot{\sigma} \frac{\partial (\cdot)}{\partial \sigma}.
\]

The mass divergence in (2.7) and (2.8) is calculated by

\[
\nabla \cdot (p^* \mathbf{V}) = \left[ \frac{\partial (p^* u)}{\partial x} + \frac{\partial (p^* v)}{\partial y} \right] \sigma.
\]
The symbols are conventional. We use $\theta_c$ and $q_v$ instead of $\theta$ and $q_v$ in the prognostic equations because they are semi-conservative quantities in the absence of precipitation, radiation, or diffusion. The liquid water content, $q_v$, can be calculated diagnostically from the two prognostic variables. This method has been proved to be quite accurate in resolving stratiform clouds in the authors’ other papers (Sun and Hsu 1988; Hsu and Sun 1991). Subgrid-scale convection is parameterized with a modified Kuo scheme (Molnar 1982). Radiation is not considered in the present study. Detailed discussion of the original equations and the calculation of the (explicit) diffusion terms $[\text{diff}(\cdot)]$ are presented in Sun and Hsu (1988) with the following modifications.

1) In the momentum equations, the pressure gradients are calculated by using the pressure perturbation $p'$ instead of the total pressure $p$.

2) The variation of $d(p_0/p)'/dt$ is also included in the heat equation.

3) Different lateral boundary conditions and horizontal smoothing are applied here and will be discussed briefly in the next section.

Sun and Chern (1993), using this model, have successfully reproduced the observed mesoscale circulation and lee vortices during the Taiwan Area Mesoscale Experiment (TAMEX) of 1987.

3. Numerical methods and grid points

The integration of the governing equations is split into two different time steps; a smaller time interval with a forward–backward scheme (Sun 1980, 1984) is applied to the terms involving inertia–gravity waves. This scheme not only allows twice the time interval used in the leapfrog scheme but also avoids the $2\Delta t$ waves that usually exist in a central-difference scheme in time. A larger time interval, limited by the horizontal advection, is applied to the Coriolis, the horizontal advection, and the diffusion terms.

The domain in the vertical direction, from the surface up to 100 hPa, is divided into 25 levels. Unevenly distributed $\Delta \sigma$ is used in the vertical direction in order to have better resolution in the lower atmosphere (about 300 m in the lowest 3 km).

The Arakawa C grid is used. The interior area of the domain is about 3000 km $\times$ 3000 km with 40 grid points in each direction; hence, $\Delta x = \Delta y = 75$ km. It is surrounded by a buffer zone with a larger space interval ($3\Delta x$) to avoid serious reflection from the lateral boundary (Sun et al. 1991). It also allows the change of wind field in the buffer zone corresponding to the change of the pressure gradient. All values at the outer boundary are imposed as functions of time. They are updated every time step and interpolated linearly from 12-h European Centre for Medium-Range Weather Forecasts (ECMWF) analysis.

The computation domain and terrain elevation used in the study is shown in Fig. 1. The area covers a small portion of Tibet to the west and the Korean peninsula and Pacific Ocean to the east. Since only limited smoothing is imposed on terrain elevation, Tibet is as high as 5 km above the mean sea level. Such high topography and steep slope around the plateau poses a serious challenge to a numerical model with a terrain-following $\sigma$ coordinate. Fortunately, our model handles the problem very well by using pressure perturbation instead of total pressure in calculating the pressure gradient terms in the horizontal momentum equations, as mentioned earlier.

4. Overview of weather

At 0000 UTC 15 May 1987, there are two low-level migratory high pressure systems in the area of interest (Fig. 2). One is located just off the east coast of China and another in the eastern Tibetan Plateau. At the time, there is little convective activity, except for some clouds in the area over central China (Fig. 3). The convection in central China is apparently associated with a weak trough between the two high pressure systems. Although three sounding stations just to the south of the convective clouds reported wind speed over 15 m s$^{-1}$ at the 850-hPa level (not shown), the ECMWF analysis shows that the maximum wind speed is less than 12 m s$^{-1}$ in Fig. 4. The averaged wind speed for nine sounding stations near the coastline in southern China is only about 6 m s$^{-1}$ at this hour. There is no well-defined LLJ in southern China.

![Terrain height](image.png)

Fig. 1. The model domain and terrain elevation with contour interval of 500 m. The heavy solid line indicates the position of the cross sections displayed in Figs. 14 and 15.
One day later (at 0000 UTC 16 May), the two high pressure systems moved quickly toward the east (not shown in figure). A Mei-Yu front developed just to the north of the coastline in southern China. The front seems to show the typical characteristics discussed in Ninomiya (1984) and Tao and Chen (1987). These characteristics are the following.

1) A mesoscale convective system developed within a cloud band (Fig. 5). The strength and location of the deep convection can be illustrated by the outgoing longwave radiation (OLR) flux [Fig. 6, taken from Kau and Wu (1991)] derived from a satellite picture. Small OLR flux implies low cloud-top temperature and strong convection. The minimum OLR flux in Fig. 6 is 120 W m$^{-2}$, which is smaller than 160 W m$^{-2}$, the minimum for the convective cloud in central China 24 h earlier.
2) A supergeostrophic LLJ to the south of the Mei-Yu front (Fig. 7). The geostrophic wind speed to the west of Taiwan, along the coastline at height $z = 1500$ m, is about 11 m s$^{-1}$ (estimated from both observations and ECMWF analysis in Fig. 8a). But the averaged wind speed for nine sounding stations near the coastline reaches 15 m s$^{-1}$ (not shown in figure). The LLJ also shows up clearly from ECMWF analysis (Fig. 7).

3) Large low-level cyclonic vorticity to the north of the jet stream (Fig. 8b). Wind shear at 850 hPa across the Mei-Yu system is usually very large. The Mei-Yu front can be defined as the line of maximum cyclonic vorticity, as suggested by Kuo and Anthes (1982). Vorticity along the front is as large as $5 \times 10^{-5}$ s$^{-1}$.

4) Large moisture gradient (Fig. 8c). Since the air to the north of the Mei-Yu system is from the north around the Tibetan Plateau, it is drier than the tropical air brought by the LLJ. The moisture gradient can be further enhanced by the deforming wind field in southern China.

5) Weak temperature gradient (Fig. 8d). The virtual potential temperature gradient across the front in southern China is much weaker than that of a polar front to the north.

Another 12 h later (at 1200 UTC 16 May), the LLJ still maintains much of its strength and the mesoscale organized convection cells spread to the eastern part of the coastline in southern China (not shown). A squall line passes through the Taiwan Strait between 1200 and 1800 UTC (Wang et al. 1990). At the time, the TAMEX was in progress; the second intensive observing period (IOP 2) started at 0600 UTC 16 May. Major events are summarized in Fig. 9.

It is interesting to note that deep convection in the east occurred after the LLJ had already become very strong over the area. As for the area to the west, it is difficult to tell which event took place first because the wind is observed only at 12-h intervals.

5. Results of the without-cumulus experiment

Data from the ECMWF analysis (at a resolution of $2.5^\circ \times 2.5^\circ$, at standard pressure levels) at 0000 UTC 15 May 1987 were fitted into the model grid points as initial conditions. Three numerical experiments were carried out. We will first examine the results of the without-cumulus experiment (no subgrid-scale cumulus heating) in this section. It will be shown that the Mei-Yu system, along with a strong LLJ, can be adequately simulated in this experiment with heating coming only from stratiform clouds. The second experiment does not include any type of latent heating (neither resolvable nor cumulus heating) in the model. Finally, we will present the third experiment, which examines the role of cumulus clouds. All results are taken from model outputs at the end of the 24-h period simulation and are compared with the ECMWF analyses at 0000 UTC 16 May (see Fig. 9).

a. Migrating high and trough

The pressure trough, initially located in northern China (Fig. 4), moved eastward along with the migratory high over Tibet. This can be seen by examining...
the wind field in Fig. 10a. The cold air behind the trough was initially blocked from extending toward the south by the mountains. Once the system moved eastward, it plunged southward and downward from the upper level around the northeastern tip of the mountain. (see Fig. 1 for terrain elevation.) The downward motion will be shown later in a vertical cross section. This dry (and cold) advection process is clearly evident by examining the moisture distribution in Fig. 10b. The specific humidity is very low behind the trough. The leading edge of the cold and dry air reaches central China and the result is comparable to the observations (Fig. 8c).

At the same time, a Mei-Yu front develops in southern China. The simulated LLJ (Fig. 10a) associated with the Mei-Yu front agrees roughly with the observations (Fig. 7). The maximum wind speed, 16 m s\(^{-1}\), is slightly larger than the observed maximum, 15 m s\(^{-1}\).

The LLJ brings moist air from the tropics to southern China. As the moist air moves farther northward, it encounters the northerly wind behind the trough in central China. Thus, the specific humidity over southern China at \(z = 1500 \text{ m}\) increases from 10 g kg\(^{-1}\) to over 14 g kg\(^{-1}\) during the 24-h period (Fig. 10b). In the region slightly to the north, the air becomes drier. The strong deformation field, created by the confluence of the northerly wind and the southwesterly LLJ, generates a strong moisture gradient in southern China. Mei-Yu fronts are often

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**Fig. 8.** ECMWF analyses at \(z = 1500 \text{ m}\), 0000 UTC 16 May 1987: (a) pressure (contour interval 1 hPa); (b) vorticity \((10^{-2} \text{ s}^{-1})\), dashed contours represent negative values; (c) specific humidity \(q_v\) (1 g kg\(^{-1}\)); and (d) virtual potential temperature \(\theta_v\) (1 K).
of southern China with a cloud top at 2 km in the west, 5 km in the middle, and 8 km in the eastern section (cloud-top information not shown in diagram). In this without-cumulus experiment, the stratiform cloud in the western section of the Mei-Yu front is rather shallow, while the observed cloud top is very high over the area, as implied by the OLR flux in Fig. 6. The observed high cloud is associated with deep convection, which cannot be simulated in this experiment. The deeper cloud in the east is produced by the large moisture convergence in the LLJ exit area. The maximum liquid water content allowed in the model is 1 g kg⁻¹. Excessive liquid water over this artificial limit is treated as precipitation. Although it is difficult to verify the existence of these low and middle clouds through satellite pictures and surface weather maps, ECMWF analysis of relative humidity suggests the presence of stratiform clouds over the area (Fig. 13). The layered and shallow clouds appear to be very important during this time of the year in southern China (Kato 1985). Ninomiya (1989) found that such cloud type is the predominant cloud type over the area during May of 1979. The large coverage of layered clouds simulated in this case study may not be just coincidental. In section 7, we shall explain that the latent heating provided by these clouds in the middle and lower troposphere is responsible for the formation of the LLJ.

c. Vertical cross section

Figure 14 shows model results at a vertical cross section (heavy solid line in Fig. 1) perpendicular to the Mei-Yu cloud band. The vertical plane passes through the middle section of the cloud band where the low-level wind speed is the largest. Isentropic surfaces (Fig. 10).
14a) slope from the north-northwest (the right-hand side of the figure) in the upper level to southern China in the lower level. Downward motion in the right-hand side of Fig. 14b corresponds to the northerly wind (Fig. 14c) behind the trough in central China discussed earlier. The leading edge of a pool of very cold air coming around the plateau from the north is marked by the dashed arrow in the bottom of each panel. The dry advection process behind the polar front is clearly shown in Fig. 14d. The wind component perpendicular to the vertical cross section (Fig. 14e) is from the east-northeast (negative), as in a typical polar front.

The Mei-Yu cloud band (Fig. 14f) is located about 600 km to the south of the polar front. The solid arrow at the bottom indicates the maximum low-level vorticity center. The stratiform cloud is rather shallow with the cloud top at 5 km. The upward vertical motion (Fig. 14b) and the LLJ (Fig. 14e) are also confined to the bottom half of the troposphere. The simulated maximum vertical motion, 4 cm s⁻¹, is slightly smaller than the observed value, about 5 cm s⁻¹ in Fig. 15a, because the convective clouds are not included in this experiment. The simulated wind field and mixing ratio distribution are also quite close to the ECMWF analysis (Figs. 15b,c). The latent heating by the stratiform cloud generates a cross-frontal secondary circulation (Fig. 14c) to the south. Since the location of the upward motion matches the heat source, it is a direct circulation. It is interesting to note that even without cumulus parameterization, this experiment can still reproduce all of the major features of the Mei-Yu front, discussed in section 4, except convective clouds.

6. Without-latent-heat experiment

Without including latent heat in the simulation, the maximum wind speed near the vicinity of the LLJ in-
Fig. 14. Simulated (without-cumulus experiment) results at a vertical cross section (heavy solid line in Fig. 1): (a) virtual potential temperature $\theta_v$ (contour interval 2 K); (b) vertical motion $w$ (0.02 m s$^{-1}$); (c) wind vectors, $w$ is exaggerated by 100 times; (d) specific humidity $q_v$ (1 g kg$^{-1}$); (e) wind component perpendicular to the vertical cross section (3 m s$^{-1}$); and (f) liquid water content $q_l$ (0.2 g kg$^{-1}$). Dashed contours represent negative values. Arrows in the bottom of each panel denote positions of the Mei-Yu front (solid arrow) and the polar front (dashed arrow) discussed in the text.
increases slightly, from 11.85 m s\(^{-1}\) (Fig. 4) to 12.67 m s\(^{-1}\) (Fig. 16), due to migration of the weather system. However, the increment of 0.82 m s\(^{-1}\) is much smaller than the value of 4.17 m s\(^{-1}\) (16.02 – 11.85), calculated from the case with grid-scale condensation. It is also smaller than 2.82 m s\(^{-1}\) (14.67 – 11.85), the increment estimated from the ECMWF analysis. Without condensation, wind is nearly in geostrophic balance (not shown). The Mei-Yu low-level vorticity center does not exist. In addition, the axis of the maximum wind speed shifts northward by two grid points (150 km). To understand the role of latent heat release associated with the stratiform cloud, the difference between the two experiments (without-cumulus versus without-latent-heat) for many fields is shown in the next section. A Mei-Yu secondary circulation can be seen clearly through the practice.

7. Secondary circulation

The virtual potential temperature at a lower level in southern China is higher (positive in Fig. 17a) for the without-cumulus experiment, as expected. The area affected by the latent heating is wider than the cloud band shown in Fig. 12. This is due mainly to the horizontal diffusion applied to each of the prognostic variables as required for the stability of any numerical model using a finite-difference method. The heating, however, is primarily restricted near the vicinity of the Mei-Yu system, and the axis of the maximum heating coincides with the Mei-Yu cloud band. The temperature difference is very small (<0.2°) in other regions.

The pressure difference (Fig. 17b) further demonstrates that the numerical model is well behaved. It produces almost identical results for the two simulations over areas not affected by the cloud band. The contour interval is as small as 0.2 hPa. Any small difference in pressure would show up easily in the figure. Warming in a vertical column induces low pressure in a lower level. The maximum perturbation, over 1 hPa, is located in the eastern section of the cloud band due to a deeper cloud layer, as described earlier. The total wind field will also change with time due to the change of the mass field through geostrophic adjustment. If the pressure perturbation remains independent of time, we may estimate (though not very accurately) the geostrophic wind speed from \((\rho f)^{-1} \Delta p (\Delta y)^{-1}\), where \(\rho\) is approximately 1 kg m\(^{-3}\), \(f \sim 6 \times 10^{-5}\) s\(^{-1}\), and \(\Delta p\)

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**Fig. 15.** ECMWF analyses at 0000 UTC 16 May 1987 for the same vertical cross section as in Fig. 14: (a) vertical motion \(w\) (contour interval 0.02 m s\(^{-1}\)); (b) wind component perpendicular to the vertical cross section (3 m s\(^{-1}\)); and (c) specific humidity \(q_v\) (1 g kg\(^{-1}\)). Dashed contours represent negative values.
is 0.8 hPa over a distance of five grids (~400 km). The geostrophic wind speed to the south of the low pressure is enhanced by about 3 m s$^{-1}$ due to latent heating. This perturbation geostrophic wind speed can already explain about one-half of the difference in LLJ wind speeds between the two experiments (Fig. 17c). Another important contribution comes from the acceleration along the downstream direction.

Although the maximum low-level wind speeds over southern China for the two experiments are, respectively, 16 and 12 m s$^{-1}$ (Fig. 10a and Fig. 16), the difference in wind fields can be larger than 6 m s$^{-1}$ due to the location shift of the jet axis when latent heat is turned off. The difference circulation (Fig. 17c) correlates closely with the pressure perturbation shown in Fig. 17b. Unlike the pressure difference, however, it is not symmetrical, with 2–3 m s$^{-1}$ easterly wind in the north and a westerly wind of over 6 m s$^{-1}$ in the south. The resulting wind shear is very large, as found in a typical Mei-Yu system. The vorticity (Fig. 17d) calculated from this circulation clearly defines the Mei-Yu system. The maximum vorticity is larger than the geostrophic vorticity corresponding to the pressure difference induced by latent heating. In other words, this circulation is actually stronger than that implied by Fig. 17b, and the LLJ is supergeostrophic. The upward vertical motion inside the cloud band is able to maintain a direct, cross-frontal secondary circulation that helps keep the LLJ supergeostrophic. A momentum budget analysis will be shown later in the section.

Figure 18a shows the difference in vertical motion at the same vertical cross section as in section 5. It is quite consistent with liquid water content in Fig. 14f. The maximum upward motion, over 3 cm s$^{-1}$, is located slightly to the south of the cloud band. Compensating downward motion can be found both to the north and the south. This implies two branches of secondary circulation: one to the north and one to the south of the Mei-Yu front (Fig. 18b). We will focus our attention on the southern branch because it is associated with the LLJ, as will be demonstrated in the next paragraph. The arrow in the lower-right corner of the figure denotes a horizontal wind speed of 5 m s$^{-1}$, or a vertical motion of 5 cm s$^{-1}$, because the vertical motion is exaggerated by 100 times in the figure. The secondary circulation is 5 km deep and 400 km wide. The low-level convergent (or upper-level divergent) wind speed is about 2.5 m s$^{-1}$. This is consistent with a scale analysis of the continuity equation, $W/H \sim U/L$, with $W$ about 3 cm s$^{-1}$.

The horizontal wind component associated with the secondary circulation is deflected to its right due to the Coriolis force. Therefore, two maxima and two minima of the along-the-front wind component show up in Fig. 18c in the vicinity of the Mei-Yu front. The LLJ corresponds to the lower-left maximum with southwesterly wind (compare Fig. 14e). Large wind shear in the lower level results in large low-level vorticity in Fig. 18d. Vorticity in the upper layer (2.5–5 km) becomes negative due to the divergent motion.

We can examine the momentum budget for the without-cumulus experiment to see how this secondary circulation is maintained. Since the cross-frontal vertical plane is almost parallel to the $y$ direction of the model domain, a budget on the $v$ component of the wind is performed for convenience. Because of the difficulty of distinguishing the effects of horizontal and vertical advection with equations based on the $\sigma$ coordinate, the following equation (in $z$ coordinate) is used:

$$\frac{\partial w}{\partial t} = -\left( u \frac{\partial w}{\partial x} + v \frac{\partial w}{\partial y} \right) - w \frac{\partial w}{\partial z} - f(u - u_0) + \text{residual.} \quad (7.1)$$

To obtain more representative forcings, the following budgets were averaged over a 1-h period between the 24th and 25th hour of the simulation.

Figure 19 shows the magnitude of each term (except for the residual term) of (7.1) at $z = 1500$ m for the without-cumulus experiment. The residual term is contributed mainly from diffusion that is negligible in this study. The third term (Fig. 19d) on the right-hand side of this equation gives the acceleration due to the ageostrophic wind, $u - u_0$. From this figure, we can see that the LLJ is supergeostrophic. The supergeostrophic LLJ generates southward (negative) acceleration against the existing secondary circulation to the south of the Mei-Yu front (the left branch in Fig. 18b). The ageostrophic acceleration is offset, however, by the vertical advection term (Fig. 19c), which is contributed pri-
Fig. 17. Difference of two experimental results (without-cumulus minus without-latent-heat) at z = 1500 m: (a) virtual potential temperature \( \theta_v \) (contour interval 0.5 K); (b) pressure (0.2 hPa), numbers in pascals; (c) wind vectors and isotachs (2 m s\(^{-1}\)); and (d) vorticity (10\(^{-5}\) s\(^{-1}\)). Dashed contours represent negative values.

primarily by the upward vertical motion inside the cloud band. With all terms considered, Fig. 19a shows that the time rate of change of \( v \) is positive in the area to the south of the Mei-Yu front in favor of the existing secondary circulation. Thus, we can conclude that condensation warming produces both low pressure and a low-level convergence along the cloud band. The wind speed increases and also turns gradually to the right as the air parcel moves toward the cloud band. Hence, the LLJ can be sustained by the south branch of this secondary circulation. It also explains that the location of the LLJ simulated with latent heat should be farther south than the axis of the wind maximum generated without condensation, as discussed in section 6.

8. With-cumulus experiment

Finally, we included a modified Kuo's cumulus parameterization scheme (Molinari 1982) in the numerical model to look into the role of deep convection on the Mei-Yu system. The stratiform cloud and stable precipitation is still included as in the without-cumulus experiment, and the physics involved in the final experiment is quite complete except for the exclusion of
radiation. We call this experiment the “with-cumulus experiment.”

In addition to the stratiform cloud (not shown), similar to that we obtained in the without-cumulus experiment, deep convection develops in southern China. The accumulated amount of cumulus precipitation during the simulation time period is shown in Fig. 20. The area with cumulus precipitation matches quite well with the area with deep convection (both shown in the IR picture and OLR flux in Figs. 5 and 6). The maximum precipitation amount occurs to the south of the resolvable stratiform cloud band. The amount of the stable precipitation (Fig. 21) is about the same as the cumulus precipitation. But the stable precipitation is more concentrated, with a maximum reaching 22.15 mm, while the maximum convective precipitation is only 10.8 mm.

The total amount of water vapor that has been converted into liquid water in the area of interest during the 24-h period is summarized in Table 1. As we can see from this table, unresolvable cumulus heating accounts for about 40% of the total latent heat that has been released. The other 60% comes from the resolvable stable cloud, with some of the cloud water falling to the ground as precipitation during the course of the integration. The total latent heat release from this stable process is about the same for the two experiments listed in Table 1.

The additional heating from cumulus convection, however, changes the low-level wind field only slightly. Although Fig. 22 shows that the maximum wind speed of the LLJ is only 14 m s⁻¹, the area enclosed by the 12 m s⁻¹ contour line is about the same as in the without-cumulus experiment (Fig. 10a). There are no significant differences between the two diagrams. From Fig. 20, the cumulus heating seems to spread out in a large area and to high levels, and its effect is not very noticeable at this stage of the Mei-
Yu system. The most significant difference between the results of the two experiments is in vertical wind speed. The upward vertical motion (Fig. 23) extends to the higher level in the area where the LLJ is located, compared with Fig. 14b. The vertical motion and the secondary circulation (not shown) in the lower atmosphere, however, are similar to those in the without-cumulus experiment (Figs. 14b,c). The small difference in low-level wind (both in the horizontal and the vertical directions) between the two experiments may also support our conclusion in section 5. That is, the influence of cumulus convection on the formation of the LLJ may not be as important as the stratified cloud and the stable precipitation from it.

The role of the cumulus clouds must have become more and more important as time progressed, as the LLJ brings more and more water vapor into the region. A mature Mei-Yu front, as described earlier, comprises many deep convective clouds. Our study suggests, however, that stable cloud and stable precipitation in the early stage of the Mei-Yu system may help induce an LLJ that later triggers those deep convective clouds.
Table 1. Total amount of water in stratiform cloud, accumulated stable and cumulus precipitation in the area shown in Fig. 20 (dashed box) for without-cumulus and with-cumulus experiments (units in $10^{12}$ kg).

<table>
<thead>
<tr>
<th></th>
<th>Without-cumulus experiment</th>
<th>With-cumulus experiment</th>
</tr>
</thead>
<tbody>
<tr>
<td>Stratiform cloud water</td>
<td>2.56</td>
<td>2.43</td>
</tr>
<tr>
<td>Accumulated stable precipitation</td>
<td>4.23</td>
<td>4.56</td>
</tr>
<tr>
<td>Accumulated cumulus precipitation</td>
<td>0.00</td>
<td>4.57</td>
</tr>
<tr>
<td>Total</td>
<td>6.79</td>
<td>11.56</td>
</tr>
</tbody>
</table>

approximately 24 h before a very pronounced LLJ appeared along the coastline in southern China. In the first experiment presented in this paper, it is shown that the LLJ can be simulated even though cumulus heating was not included in the model, while in the with-cumulus experiment there is no appreciable difference in either the pattern or strength of the LLJ compared with that of the without-cumulus experiment. The results seem to suggest that cumulus heating may not have played a critical role in the formation of the LLJ in a Mei-Yu system. The latent heat released by the stratiform clouds over the area instead contributes to the development of the LLJ. When latent heat is completely shut off, artificially, in one of the experiments, the low-level wind is much weaker. The characteristics of the observed Mei-Yu front at 0000 UTC 16 May can then no longer be identified as in the other two experiments.

Our analyses, at least for the particular Mei-Yu front under consideration, suggest that an LLJ develops...
through the following sequence of processes. A large amount of moisture is transported to the north by a marine tropical air mass in the South China Sea and Indochina during the Mei-Yu season. When the moist air encounters a migrating high pressure system in the north, it is blocked from moving farther northward. It turns to the east-northeast and becomes saturated. The resulting stratiform cloud covers all of southern China, and the pressure trough at the 850-hPa level deepens by about 1.5 hPa due to the warming process. The low pressure induces convergent motion and generates a direct, cross-frontal secondary circulation. The southern branch of the secondary circulation is ultimately responsible for the formation of the LLJ. The northward convergent wind turns to its right due to the earth’s rotation. Together with the synoptic west-southwest mean wind, the total wind speed becomes very high. Since this secondary circulation is vital for the formation and duration of the LLJ, a momentum budget is calculated to understand how it is maintained for a sufficient period of time for the Coriolis force to take effect. Although the geostrophic adjustment process induced by the supergeostrophic LLJ tends to oppose and decelerate the secondary circulation, it is found that the vertical transport of momentum by the secondary circulation itself is enough to offset the adjustment process. Thus, the secondary circulation can be maintained for a long period of time. In summary, we have demonstrated the importance of large-scale latent heating by the stratiform cloud in relation to the LLJ and its accompanying secondary circulation. Without such large-scale latent heating, the low-level wind would be weaker; the amount of moisture that can be transported to the frontal zone would also be drastically reduced. Once the LLJ forms, it may trigger heavy precipitation in the later stages of the Mei-Yu system.

Acknowledgments. The authors would like to extend thanks to Dr. Daniel Keyser and reviewers for their comments. We would also like to thank Mr. Jian-Dar Chern for helping install a cumulus parameterization scheme to our numerical model, and Mr. J. Gardner for proofreading. This study is coprisoned by the National Science Council of ROC (NSC 80-0202-M002-13) and the National Science Foundation of the United States (ATMS-8611729 and ATMS-8907881). The model simulations were performed on the Convex C120 computer in the Department of Atmospheric Sciences at National Taiwan University.

REFERENCES


