A Diagnostic Study of a Retreating Mei-Yu Front and the Accompanying Low-Level Jet Formation and Intensification

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ABSTRACT

During 7–8 June 1998, an organized mesoscale convective system (MCS) formed within the mei-yu frontal cloud band and moved northeastward to produce heavy rain over the island of Taiwan. During this period, the section of the mei-yu front east of Taiwan moved northward, most significantly for about 300 km over 12 h. Meanwhile, a low-level jet (LLJ) developed within the environmental southwesterly flow to the south of the mei-yu front and the MCS.

Observations revealed that the front retreated as low-level meridional wind components over the post-frontal region shifted from northerly to southerly. Using European Centre for Medium-Range Weather Forecasts (ECMWF) analyses with piecewise potential vorticity (PV) inversion technique and other methods, a diagnostic study was carried out to investigate the northward frontal movement and the formation of the LLJ.

Results indicated that diabatic latent heating from the MCS, large enough in scale, generated positive PV and height fall at low levels. The enhanced height gradient induced northwestward-directed ageostrophic winds and the LLJ formed southeast of the MCS through Coriolis torque. The southwesterly flow associated with this diabatic PV perturbation led to rapid retreat of the frontal segment east of Taiwan at a speed of about 25 m s\(^{-1}\), while the movement was dominated by horizontal advection in the present case. During this process of readjustment toward geostrophy, a thermally indirect circulation also appeared over and south of the front, and the LLJ formed within its lower branch at 850 hPa. The enhanced southwesterly winds reached LLJ strength because they were superimposed upon a background monsoon flow at the same direction. To the lee of Taiwan, the topography also played the role in enhancing local wind speed at lower levels and contributed toward the frontal retreat at nearby regions.

1. Introduction

Climatologically, the seasonal rainfall maximum over Taiwan and southern China from mid-May to mid-June, known as the mei-yu season, was largely caused by the mei-yu frontal cloud band (Ding 1992; Chen 2004). During this rainy season, organized mesoscale convective systems (MCSs) embedded within the cloud band often develop and produce heavy rainfall and flash floods. South of the front, also frequently observed is the low-level jet (LLJ), identified to be closely related to heavy rainfall events (Chen and Yu 1988; Li et al. 1997; Chen et al. 2005). During 7–8 June 1998, a section of the mei-yu front near Taiwan moved northward rapidly, for a distance of about 300 km over 12 h. Although such a frontal retreat is not a rare event on weather maps, it is interesting since mei-yu fronts in the subtropics usually move slowly southward/southeast-
ward or remain quasi-stationary (e.g., Chen and Chi 1980; Chen 1988, 1992). The mechanism of the northward retreat of a mei-yu front, however, has never been investigated. This case was therefore selected in the present study as a first effort to diagnose the physical processes responsible for the retreating movement of a mei-yu front. As will be shown later (section 3), during this period, organized MCSs developed, intensified, and moved east-northeastward along the front to produce heavy rainfall over southwestern Taiwan, with the largest amount well exceeding 200 mm (24 h)^{-1} in 7 June 1998.

The theoretical study of mei-yu frontogenesis by Cho and Chen (1995) suggested that the feedback mechanism similar to the conditional instability of the second kind (CISK), in which the front helps organize the convection while latent heat release of cumuli enhances low-level potential vorticity (PV) and the frontogenetic process, could lead to rapid growth in shear vorticity along the front. Chen et al. (2003) selected a mei-yu front case over southeastern China, and preformed PV diagnosis to investigate the mechanism responsible for the intensification and maintenance of the frontal vorticity. It was found that the apparent frontogenesis was driven almost entirely by an outbreak of deep convection along the front, through the CISK mechanism proposed by Cho and Chen (1995). Kinematically, frontal movement is controlled by horizontal advection process if the front is considered a material surface. For a slow-moving mei-yu front, Chen and Chang (1980) found that the horizontal advection at either side of the front, as expected, contributed very little toward frontal vorticity. Meanwhile, without advection a front could also move in response to the frontogenetic process, which could occur on one or both sides of the front. The present case appears to be also ideal for studying the role of these different physical processes, advective or frontogenetical, in causing the northward retreat of a frontal segment.

In this case, an LLJ at 850 hPa formed and intensified to the south of the mei-yu front. Observational (Lin and Chiu 1985; Chen and Yu 1988), theoretical (e.g., Chen 1982), and numerical studies (Chou et al. 1990; Nagata and Ogura 1991; Chen et al. 1997; Chen et al. 1998, 2000) in the literature all suggested that an LLJ could form to the south of the heavy convective rainfall area associated with the mei-yu front, presumably through Coriolis acceleration of the northward-moving lower branch of the induced secondary circulation. Geostrophic forcing and lee cyclogenesis processes (Chen and Pu 1985; Pu and Chen 1988; Chen and Chen 1995; Chen and Tseng 2000), as well as upper-level jet streak forcing (Tsay and Kau 1989), were also suggested as possible mechanisms for the formation of LLJs in the mei-yu season. The LLJ in the present case also appeared to be a good subject for studying the role of various physical processes on its formation and intensification, as well as its relationship with the frontal retreat.

2. Data and analysis method

Six-hourly (at 0000, 0600, 1200, and 1800 UTC) gridded operational analyses from the European Centre for Medium-Range Weather Forecasts [ECMWF; Tropical Ocean Global Atmosphere (TOGA) advanced] were used as the primary data source for illustrating synoptic situations and for calculation and diagnosis. The dataset has a horizontal resolution of 0.5° latitude × 0.5° longitude and provides geopotential height, temperature (T), u and v components of horizontal wind, vertical velocity (w), mixing ratio (r), and relative humidity (RH) at nine pressure (p) levels (1000, 925, 850, 700, 500, 400, 300, 250, and 200 hPa). Besides ECMWF data, the Japan Meteorological Agency (JMA) surface maps were used to locate the surface mei-yu front. Collected during the South China Sea Monsoon Experiment (SCSMEX), 6-hourly sounding data at Ishigakijima, Tungsha, and several other stations near Taiwan (cf. Figs. 1a and 1c) were used to identify the LLJ, to verify gridded data, and to carry out further diagnosis. Data at most of these stations are routinely assimilated into the ECMWF operational analysis. Infrared (IR) imageries of the Geostationary Meteorological Satellite (GMS) were employed to depict frontal cloud band and convection, while gauge network data over Taiwan were also used to analyze rainfall distribution.

In this paper, Ertel’s PV (EPV; e.g., Hoskins et al. 1985) inversion technique (PVIT) and the nonlinear balanced diagnostic system developed by Davis and Emanuel (1991) were used to study the frontal retreat and the formation and intensification of the LLJ. Principles of this method are described in section 5. Vorticity budget, PV budget on isentropic surfaces, and ageostrophic winds were analyzed, and vertical cross sections (and their composites) were constructed to examine frontal movement and the origin of the LLJ. These results are presented in section 4, as well as sections 6 through 8.

3. Synoptic situations

At 1200 UTC 7 June, the JMA surface map (Fig. 1a) showed that the mei-yu front extended from western Pacific, passing through the southern tip of Taiwan, to the coastal region of southern China, and was oriented
in an east–west direction. Prior to this time, the mei-yu front was nearly stationary (not shown). A weak frontal disturbance, meanwhile, formed over the southern Taiwan Strait (near 22°N, 117°E; Fig. 1a). This low developed slightly and moved near the southwestern coast of Taiwan at 1800 UTC along the front (not shown), and farther to the northeast of Taiwan at 0000 UTC 8 June (Fig. 1b). During this 12-h period, the front migrated from southern to northern Taiwan, most rapidly over the section east of Taiwan, for about 300 km. Ishigakijima station, located at about 250 km east of Taiwan, experienced a change in wind direction and air temperature due to this rapid northward frontal movement (Figs. 1a and 1b). As the disturbance continued to move northeastward along the mei-yu front in the next 6 h, the frontal segment east of Taiwan further retreated while that over and west of Taiwan remained quasi-stationary (Fig. 1c).

The GMS IR imageries showed a frontal cloud band with an organized MCS over the frontal disturbance to the southwest of Taiwan at 1200 UTC 7 June (Fig. 2a). In the following 12 h, the MCS moved northeastward along with the frontal disturbance, to the east of Taiwan at 0000 UTC 8 June (Fig. 2b). The frontal low and the accompanied MCS continued to move along the front in the next 6 h, to a location northeast or east of Taiwan at 0600 UTC 8 June (Figs. 1c and 2c). On 7 June, upon moving across Taiwan the MCS produced heavy rainfall over the island, up to more than 200 mm (24 h)^{-1}, particularly along the southwestern coast and over the western slope of the southern Central Mountain Range (CMR; Fig. 3). Peak values of 6-h rainfall reached 196 mm from 1200 to 1800 UTC 7 June, and 357 mm from 1800 UTC 7 June to 0000 UTC 8 June (not shown). Figure 4 presents 850-hPa synoptic maps between 1200 UTC 7 June and 0600 UTC 8 June 1998. During this period, a monsoon trough extended east-northeastward from the low west of 105°E and strengthened markedly, such that southwesterly flow prevailed south of the trough over the coastal region of southern China, northern South China Sea, Taiwan, and nearby regions over northwestern Pacific. As defined by the axis of maximum relative vorticity (ζ), the segment of the mei-yu front east of Taiwan at 1200 UTC 7 June (dotted line, Fig. 4a) was collocated with the wind shift (shear) line and about 75–150 km north of the surface front (cf. Fig. 1). Although the associated wind shift was less apparent at later times, the frontal vorticity remained comparable (to be discussed later). During this period, the eastern section of the front at 850 hPa came to coincide with the surface front, and retreated to pass Ishigakijima by 0000 UTC 8 June as well (Figs. 4b–d). Within the southwesterly monsoon flow, a trough in the height field existed over southern Taiwan Strait (to the southwest of Taiwan) at 1200 UTC 7 June, and was associated with the organized MCS as presented in Fig. 2a. This area was also where the surface frontal disturbance formed in Fig. 1. This 850-hPa trough moved eastward, and then deepened to the east of Taiwan in association with the organized MCS over the same area in the following 18 h (Figs. 4b–d). In response to the deepening of the trough, the...
geopotential height gradient and the accompanied winds strengthened, and southwesterly flow intensified to the south of the organized MCS. The maximum wind speed at 850 hPa increased from about 11 m s\(^{-1}\) at 1200 UTC 7 June to 17.5 m s\(^{-1}\) over the area southeast of Taiwan at 0000 UTC 8 June, and further to almost 20 m s\(^{-1}\) 6 h later (Fig. 4). Thus, the LLJ formed and strengthened to the south of the organized MCS, which was associated with the low-level frontal disturbance (Figs. 1 and 2).

Synoptic maps at 500 hPa are presented in Fig. 5. In the middle troposphere a cutoff low was located over central China, with westerly flows prevailing to its south over southern China, Taiwan, and the surrounding area in northwestern Pacific. Identified through wind shift, a weak short-wave trough was over the Bashi Channel (to the south of Taiwan) at 1200 UTC 7 June, although winds in its vicinity were relatively weak (Fig. 5a). Twelve hours later (Fig. 5b), winds north of Taiwan strengthened as the geopotential height gradient increased mainly due to the eastward movement of the cutoff low. Meanwhile, the short-wave trough over the Bashi Channel intensified (as identified by the associated height wave) and moved to the southeast of Taiwan, such that winds east of it not only increased in speed but also changed from westerly to southwesterly (Fig. 5b). At 300 hPa, a weak trough was located over central China above the 500-hPa cutoff low at 1200 UTC 7 June, and winds from the west-northwest prevailed in the vicinity of Taiwan (Fig. 6a). As the trough moved eastward at 0000 UTC 8 June, winds surrounding the Taiwan area intensified (Fig. 6b). East of Tai-
Fig. 4. Synoptic maps at 850 hPa analyzed using ECMWF gridded data every 6 h from (a) 1200 UTC 7 Jun to (d) 0600 UTC 8 Jun 1998. Geopotential heights (gpm, contour) are analyzed at 10-gpm intervals. Full and half barbs represent 5 and 2.5 m s\(^{-1}\), respectively, and areas with wind speed \(\geq 10\) m s\(^{-1}\) are shaded (grayscale shown at lower-right corner). Heavy solid (dotted) lines indicate a mei-yu front section as defined by maximum \(\varphi\) gradient (relative vorticity), and the symbol \(\star\) marks the location of the Ishigakijima station.
wan, wind direction shifted from west-northwest to west-southwest, similar to the conditions at 500 hPa. Figure 7 presents the time series of temperature and horizontal wind at mandatory levels observed at Ishigakijima located about 250 km east of Taiwan (cf. Fig. 1a). At and below 850 hPa, temperatures increased during 1200 UTC 7 June–0000 UTC 8 June (note that the vertical scale is reversed), indicating the northward movement of the mei-yu front. During the same period, winds below 700 hPa veered with time and also suggested that the (low level) mei-yu frontal wind shift line retreated to pass Ishigakijima between 1800 UTC 7 June and 0000 UTC 8 June, as clearly shown by Fig. 4. Moreover, wind speed at 700–925 hPa increased dramatically over a 6-h period prior to 0000 UTC 8 June, from about 5–7.5 to 12.5–15 m s$^{-1}$ (Fig. 7b). This indicates the intensification (or passage) of the LLJ near Ishigakijima in observational data and supports the ECMWF analysis. At Lutao station just off southeastern Taiwan (cf. Fig. 1c) low-level winds also strengthened from about 7.5 to 20 m s$^{-1}$ from 0600 UTC 7 June to 0000 UTC 8 June (not shown). Since the LLJ intensified to the southeast of Taiwan, one wonders whether the leeside effect of topography (e.g., Smolarkiewicz and Rotunno 1989; Rotunno et al. 1999) was the major mechanism for its intensification in the present case. Nevertheless, since the wind speed maximum at 850 hPa first appeared upstream from Taiwan (near 19°N, 117°E at 1200 UTC 7 June) and exhibited a clear tendency to propagate downstream (to 23°N, 124°E at 0600 UTC 8 June) in Fig. 4, there must exist some other mechanism in play, which was crucial at least to the initial formation of the jet. In later sections, this mechanism will be identified and its relative importance to the topographical effect will be further discussed.

4. Frontal structure and vorticity budgets

Figure 8 presents the averaged vorticity and divergence at both 925 and 850 hPa, across the (850 hPa) mei-yu front over the period of 1200 UTC 7 June–0600 UTC 8 June 1998. Here, the front was determined through relative vorticity and its position is indicated by dotted lines in Fig. 4. Values along lines 0.25° apart and normal to the front were first interpolated from 5.5° south (or southwest) to 5.5° north (or northeast) of the front (also at 0.25° intervals), then averages were made among lines within the period. Thus, the front was constantly placed at “0” along the abscissa during the averaging. As measured by high positive vorticity, the width of the mei-yu front east of Taiwan in the present case was about 200 km. At 925 and 850 hPa, the cyclonic vorticity was of comparable values (peaking at 5–6 × 10$^{-5}$ s$^{-1}$) and exhibited nearly no vertical tilt, consistent with earlier discussion on the northward frontal retreat. The vorticity was also in phase with horizontal convergence over the frontal zone (Fig. 8), in agreement with Chen and Chang (1980) and suggesting the importance of vortex stretching process in maintaining the frontal vorticity. As the convergence was stronger at 925 hPa (about −3 × 10$^{-5}$ s$^{-1}$) than at 850 hPa, the frontogenetical process (and thus the vortex stretching) was apparently dominated by boundary layer forcing. The 850-hPa convergence was also strong at about 250 km north of the front (Fig. 8b), and a comparison with Fig. 4 suggests that this second convergence zone was at the leading edge of the LLJ and associated with maximum $\theta_e$ gradient (solid lines in Fig. 4). Since the $\theta_e$ gradient and baroclinity over this segment of mei-yu front were weak (not shown), as typically found at such low latitudes (Chen and Chang 1980), and also since a similar convergence zone was absent at 925 hPa, the relative vorticity was selected as...
the indicator for the mei-yu front throughout the present study.

Using the same method as in Fig. 8, the composite NW-SE vertical cross sections of vertical velocity $\omega$, and anomalies of temperature ($T'$) and mixing ratio ($r'$) across the front are constructed to illustrate the averaged frontal structure (Fig. 9). Here, the anomaly was the deviation from the time mean of 15 May–15 June 1998. Strong upward motion, maximized at 850 hPa (about $-0.3$ Pa s$^{-1}$), was observed over the front (at “0” along the abscissa), consistent with Fig. 8. Another center of ascent was located about 400 km south of the front in the upper troposphere, and was apparently related to the short-wave trough at 500 hPa (cf. Fig. 5). A thermally direct circulation (TDC) across the front, with rising warm air ($\omega < 0$ and $T' > 0$) to the south and sinking cold air ($\omega > 0$ and $T' < 0$) to the north, is clearly shown by the negative correlation between $\omega$ and $T'$ (Figs. 9a and 9b). Near the surface at 1000 and 925 hPa, also consistent with Fig. 4, the relatively weak frontal baroclinic zone (peaking at about 1 K per 100 km) appeared at 100–200 km north of the front, which was defined by and collocated with vorticity maximum. The shallowness of the front is revealed by the fact that the baroclinic zone was confined below 700 hPa. The distribution of mixing ratio anomaly shows a moist tongue extending from the south to the north of the front in the lower troposphere, and the largest $r'$ gradient appeared to its immediate north, at about 250 km north of the front at 850 hPa (Fig. 9c).
The moist tongue was apparently related to the existence of the LLJ, and the distributions of $T'$ and $r'$ were also consistent with the northward displacement of the maximum $\theta_e$ gradient during the frontal retreat as discussed earlier (Figs. 4 and 8).

Time variations of the vorticity budget terms at 850 hPa are presented in Fig. 10. Because the twisting term was relatively small compared to other terms, it is therefore not shown. The stationary nature of the front before 1200 UTC 7 June is clearly illustrated by the near-zero local tendency along the front, while its rapid northward movement afterward is indicated by the large positive tendency north of the front (Fig. 10a). The positive tendency reached maximum during 1800 UTC 7 June–0000 UTC 8 June, in agreement with the period of most rapid frontal retreat as well as the significant intensification of the LLJ from upstream to the lee of Taiwan (Figs. 4 and 7). Thus, a close relationship apparently existed between the LLJ and the retreat of the mei-yu front, an argument supported by the concurrent strengthening of the jet and the horizontal vorticity advection process (Fig. 10b). The effect of horizontal vorticity advection was in phase with the tendency term after 1200 UTC 7 June with positive value north of the front (maximized at a distance of about 50–100 km), and was the primary contributor to the latter (Figs. 10a and 10b). All these facts together suggest the vital role of the vorticity advection process associated with the LLJ in causing the northward retreat of the front in the present case. In contrast, both divergence and vertical advection terms contributed toward local vorticity generation along the front (mainly prior to about 1800 UTC 7 June), rather than to the north of it, and therefore the mei-yu frontal intensity at 850 hPa was indeed maintained by convergence and the upward vorticity transport from the boundary layer as discussed earlier.

5. The method of piecewise PV inversion and diagnosis

The ECMWF-TOGA data were used to calculate EPV ($q$), which is defined as

$$q = \frac{1}{\rho} \eta \cdot \nabla \theta,$$

where $\eta$ is the three-dimensional absolute vorticity vector, $\nabla$ is the gradient (del) operator, $\theta$ is potential temperature, and $\rho$ is air density. In practice, using data in $p$ coordinates, EPV can be computed from (the vertical component of) relative vorticity in isentropic coordinates ($\zeta_o$), expressed as

$$\zeta_o = \left[ \zeta + \frac{R}{\sigma p} \left( \frac{\partial v}{\partial p} \frac{\partial T}{\partial x} - \frac{\partial u}{\partial p} \frac{\partial T}{\partial y} \right) \right]_p,$$

where $\zeta$ is the total vorticity, $R$ is the gas constant, $\sigma$ is the lapse rate, $u$ and $v$ are the zonal and meridional wind components, and $T$ is temperature.
where \( \sigma = \frac{T}{C_p} - \frac{(p/R)(\partial T/\partial p)}{C_p} \), \( \zeta \) is relative vorticity in \( p \) coordinates as mentioned, \( R \) the gas constant, and \( C_p \) the specific heat at constant pressure.

Based on the conserved property and invertibility of EPV for adiabatic and frictionless flow (Hoskins et al. 1985), Davis and Emanuel (1991) and Davis (1992a,b) developed the method of piecewise PVIT that satisfies the nonlinear balanced dynamics (Charney 1955). Because of a more accurate dynamical constraint and the ability to quantify effects among individual components in the atmosphere, the piecewise PVIT has been widely applied by researchers as a useful tool for diagnosis (e.g., Wu and Emanuel 1995; Davis et al. 1996; Shapiro 1996; Chang et al. 1998; Morgan 1999; Chen et al. 2003).

**Fig. 8.** Composite vorticity (\( \zeta \)) and divergence (\( D \), both in \( 10^{-5} \text{s}^{-1} \); ordinate) at (a) 925 and (b) 850 hPa normal to and across the mei-yu front (positioned at zero in abscissa) from 5.5° south/southwest (negative) to 5.5° north/northeast (positive) of the front over the period of 1200 UTC 7 Jun–0600 UTC 8 Jun 1998. The section and position of the front, as defined by maximum relative vorticity, used for composite are indicated by dotted lines in Fig. 4.
Following Davis and Emanuel (1991) and Davis (1992a), a mean EPV distribution must be chosen to obtain a balanced mean field first. Here, the time mean of the period from 15 May to 15 June 1998 was used, and the EPV perturbation \( q' \) was defined as the departure from this mean. Then, by partitioning total \( q' \) in the calculation domain into several anomaly components, the balanced mass and wind fields that correspond to each of the components can be obtained through piecewise inversion. Thus, with proper partitioning, the contribution of individual physical processes, to the northward movement of the mei-yu front and to the formation/intensification of the LLJ in our case, can be isolated and quantified. During the inversion, the mathematical treatment to linearize the balanced equation set was the same as in Davis and Emanuel (1991), that is, by prescribing a “pseudo-mean” that equals the mean plus one-half of the total perturbation at all points. For a complete description of the nonlinear balanced piecewise PVIT, readers
are referred to Davis and Emanuel (1991) and Davis (1992a).

In our study, the method of partitioning PV perturbations followed the strategy of Davis (1992b) closely, as illustrated schematically in Fig. 11. The total $q'$ is divided into four parts associated with (a) diabatic heating, represented by $q'$ of nearly saturated (RH $\geq 70\%$) high EPV ($q' \geq 0$) tropospheric (925–500 hPa) air, and denoted by LLh; (b) adiabatic effect, associated with $q'$ of unsaturated tropospheric air (i.e., 925–500-hPa $q'$ other than LLh) and denoted by LLd; (c) lower-boundary effect, represented by the averaged 1000–925-hPa $\theta$ perturbation ($\theta'$, denoted by LB); and finally (d) upper-level effect due to $q'$ and $\theta'$ at 400–200 hPa (denoted by UL). The criterion of $q' \geq 0$ in defining LLh is to identify only the condensational heating effect that produces positive vorticity. A threshold that RH $\geq 70\%$ was enforced (in addition to $q' \geq 0$) for the perturbation at a grid point to be considered as related to latent heating. This 70% threshold, as chosen by Davis (1992b), was low enough to include PV that might be advected out from a precipitation area but high enough to exclude PV of upper-level origin. The

![Fig. 10. Time variations of vorticity budget terms (a) local rate of change ($\partial \zeta / \partial t$), (b) horizontal advection [−$\nabla \cdot (\zeta + f)$], (c) divergence [−$(\zeta + f) \nabla \cdot \nabla$], and (d) vertical advection [−$\omega (\partial \zeta / \partial p)$], all in $10^{-5}$ s$^{-1}$ (6 h)$^{-1}$ across the front at 850 hPa during 0000 UTC 7 Jun–0600 UTC 8 Jun 1998 (abscissa). Values were averaged along the front from 5.5° south (negative, ordinate) to 5.5° north (positive). Isolines are analyzed at 1 $\times 10^{-5}$ s$^{-1}$ (6 h)$^{-1}$, and heavy dashed lines indicate frontal position.](image)

![Fig. 11. Scheme used in this study for partitioning the total Ertel's PV perturbation ($q'$) for piecewise inversion. Data pressure levels are indicated on the ordinate. See text for details.](image)
value also accounted, to some extent, for saturation with respect to ice in subfreezing environments.

When the piecewise PVIT is employed as a diagnostic tool, it is necessary to validate the nonlinear balanced fields obtained from the inversion. Figure 12 compares 850-hPa geopotential height and wind fields in the ECMWF analysis with the corresponding non-linear balanced fields at 0000 UTC 8 June. The inverted fields exhibit patterns very close to the analysis, but height values are systematically raised by about 10 gpm due to the imposed constraint of nonlinear balanced relationship. Because only the rotational part of the flow is retained after the inversion (Davis 1992a), such differences are both expected and considered acceptable. The inverted fields are also somewhat smoother, another unavoidable outcome of the inversion. Both the mei-yu front as defined by the axis of vorticity maximum and the LLJ to the southeast of Taiwan are well preserved in the balanced fields. Figure 13 presents the mean geopotential height and wind fields over the period of 15 May–15 June 1998 used for the inversion. It is clear that the retreating movement of the mei-yu front and the formation/intensification of the LLJ occurred in a background of prevailing southwesterly monsoonal flow.

6. Formation and intensification of LLJ

To examine the physical processes involved in the formation and intensification of the LLJ, ageostrophic wind components perpendicular, and local tendency of winds parallel, to the geopotential height contours at 850 hPa are presented in Fig. 14. At 1200 and 1800 UTC 7 June when the newly formed LLJ (≈10 and 12 m s$^{-1}$, respectively) was still to the southwest and south of Taiwan, significant southeasterly ageostrophic winds appeared across the jet axis toward the organized MCS (Figs. 14a and 14b; cf. Fig. 2a). Recall that there existed a trough at 850 hPa over the area of the MCS (Figs. 4a and 4b), and the accompanied height fall appeared to be responsible for the generation of ageostrophic winds during this period. South of the MCS, the LLJ formed
at the immediate north of strong ageostrophic flow, with an axis oriented normal to the latter. In the following 12 h as the LLJ intensified to the southeast of Taiwan (Figs. 14c and 14d), ageostrophic winds again prevailed across the jet toward the MCS, where the trough also enhanced (Figs. 14c,d and 4c,d). Although the core of the LLJ at 0000 UTC 8 June was very close to the prominent leeside trough (Figs. 14c and 14c,d), the region of maximum wind acceleration (i.e., positive local tendency) along geopotential height contours was constantly located downstream from the jet center, rather than fixed to the southeast of Taiwan, during the entire period (Fig. 14). This confirms that the LLJ was migratory in nature, but the possible role played by topography in strengthening the jet as it passed through the lee, obviously, cannot be ruled out. In our case, the ageostrophic wind across the jet peaked at about 7 m s\(^{-1}\) (in Fig. 14c) while the Coriolis parameter at 20°N is roughly 5 \times 10^{-5} s^{-1}. Therefore, Coriolis torque of the ageostrophic flow would produce a wind speed increase of about 7.5 m s\(^{-1}\) in 6 h, in close agreement with the intensification rate of the LLJ. This suggests that the LLJ formed and intensified largely, if not entirely, through the process of Coriolis acceleration of ageostrophic winds to the south of the MCS, which propagated northeastward together with the surface frontal disturbance. Prior to 0000 UTC 8 June, large ageostrophic winds at 850 hPa also existed near Taiwan and the coast of southeastern China (Figs. 14a and 14b), most likely in response to the eastward extension of the monsoon trough and the associated height fall (cf. Fig. 4). As a result, stronger southwesterly flow (of about 8–13 m s\(^{-1}\)) also appeared over these regions during 1800 UTC 7 June and 0600 UTC 8 June (Figs. 4b–d).
Vertical cross sections across the LLJ axis from northwest to southeast, along AA’ in Fig. 4a, at the time of the jet formation (1200 UTC 7 June) are presented in Fig. 15 to reveal its dynamic and thermodynamic environments. The line AA’ intersected with the mei-yu front (defined by $\zeta$ at 850 hPa) west of Taiwan (position marked), which had a different orientation from that east of Taiwan (cf. Figs. 1a and 4a). At this time, the front exhibited stronger gradient in equivalent potential temperature ($\theta_e$) than in $\theta$, but also only below 700 hPa (Figs. 15a and 15b). On the section plane, the LLJ formed to the southeast of the front in a potentially unstable environment ($-\partial \theta_e / \partial p < 0$). The maximum values of $\theta_e$ and $r$ (in the lower to middle troposphere existed over the area of organized MCS near the front (Figs. 15b and 15c), to the southwest of Taiwan (cf. Fig. 2a), and were mainly due to upward transport of moisture by convection. At about 118.5°–122.5°E, a thermally indirect circulation (TIC) also clearly existed with rising motion ($\omega < 0$) over the front and sinking farther to the southeast (Fig. 15d), as also indicated in Fig. 14a. The LLJ (>10 m s$^{-1}$ at this time) formed within the lower branch of this secondary circulation, at 850 hPa where the transverse ageostrophic flow was the strongest. Northwest of the front, weak descent also appeared, forming a TDC over and north of the front, in agreement with Fig. 9. In the lower troposphere, the front was accompanied by maximum $\zeta$ reaching $7 \times 10^{-5}$ s$^{-1}$ and high EPV values of about 0.55 PVU (1 PVU = $10^{-6}$ K m$^2$ kg$^{-1}$ s$^{-1}$; Figs. 15e and 15f). The centers of cyclonic $\zeta$ and $q$ in the middle-to-upper troposphere south of the front, on the other hand, were associated with the short-wave trough over the Bashi Channel (cf. Fig. 5a). Thus, the analysis here shows that the LLJ formed at 850 hPa within the lower branch of a TIC near 1200 UTC 7 June (Figs. 14 and 15), before it reached the leeside of Taiwan.

Figure 16 presents the isentropic PV (IPV) at the 305-K $\theta$ surface (near 850 hPa; cf. Fig. 15a) during 1200 UTC 7 June–0600 UTC 8 June. It is clear that the band structure of maximum PV east of Taiwan was associated with the mei-yu front through the retreat, and its magnitude (and thus $\zeta$) was reasonably maintained during this period. At 1200 UTC 7 June (Fig. 16a), regions of elevated PV (>0.8 PVU) to the west and southwest of Taiwan (near 23°N, 118°E and 21°N, 117°E) were collocated with the organized MCS (cf. Fig. 2a), suggesting the role of latent heat release in PV generation in the lower troposphere there. At the same time, LLJ formed to the southeast of the PV maximum. The high PV area to the southwest moved closer to Taiwan at 1800 UTC 7 June, and the values increased during the next 12 h when it reached the leeside (along with the MCS; Figs. 16b–d). Meanwhile, the LLJ also moved northeastward and intensified to the southeast of the PV center. Apparently, the formation and movement of the LLJ were closely related to convective latent heating, which generated positive PV (and produced height fall) in the lower troposphere. Except the moving PV maximum, another quasi-stationary PV center also appeared to the immediate southeast of Taiwan by 1800 UTC 7 June (Fig. 16b). The two centers merged at 0000 UTC 8 June (Fig. 16c), and grew stronger (>1.2 PVU) as low-level flow surrounding southern Taiwan also intensified (cf. Fig. 4). The existence and evolution of this stationary PV center suggested that the lee vortex (or PV) generation mechanism (e.g., Smolarkiewicz and Rotunno 1989; Rotunno et al. 1999; Epifanio and Durran 2002) also contributed toward the intensification of the LLJ when it moved through the leeside of Taiwan (Fig. 16). Similarly, the PV minimum over Taiwan was likely linked to vortex contraction as the airflow ascended the topography (e.g., Zehrnder 1991; Holton 1992). This feature was also more prominent on 8 June when the low-level flow was stronger, and was associated with the development of a windward-side ridge (cf. Figs. 4c and 4d).

For adiabatic and frictionless flow, air parcels move along isentropic surfaces and the PV is conserved ($dq/dt = 0$). The local PV tendency ($dq/dt$), therefore, is only and totally due to advection on the isentrope ($-V \cdot \nabla q$), and the net change in $q$ can be obtained by integrating the PV advection with respect to time. The difference between observed and integrated $q$ is hence the (nonconserved) PV generation or destruction from diabatic processes and friction (provided that the computational error is sufficiently small). As discussed earlier, because the 305-K isentrope was about 1.5 km above the surface (except over the Taiwan interior), PV generation by friction is assumed negligible. Figure 17 presents the local change in PV on the 305-K $\theta$ surface due to diabatic processes over 6-h periods from 1200 UTC 7 June to 1200 UTC 8 June, computed using the above-mentioned method. It is clear that positive PV generation by diabatic processes occurred persistently to the northwest of the LLJ, over areas of organized MCS and high PV, as the jet moved from upstream to the lee of Taiwan (cf. Figs. 2 and 16). The effect was also quite strong after 1800 UTC 7 June (Figs. 17c and 17d), suggesting that convective latent heating was an important source of PV generation throughout the period, including the period when the jet moved to the leeside of Taiwan. Although smaller in magnitude, positive PV generation also occurred along most of the mei-yu front during its northward retreat (Fig. 17), also
Fig. 15. Vertical cross section along AA’ in Fig. 4a at 1200 UTC 7 Jun 1998 for (a) $\theta$ (K), (b) equivalent potential temperature ($\theta_e$, solid, K) and $-\partial \theta_e/\partial p$ (dashed, K hPa$^{-1}$) over negative regions, (c) mixing ratio ($r$, g kg$^{-1}$), (d) $\omega$ (10$^{-5}$ Pa s$^{-1}$) and ageostrophic secondary circulation (vectors), (e) $\zeta$ (10$^{-3}$ s$^{-1}$), and (f) $q$ (10$^{-2}$ PVU). Contour intervals are (a) 3 K, (b) 3 K and 0.02 K hPa$^{-1}$, (c) 2 g kg$^{-1}$, (d) 2 $\times$ 10$^{-5}$ Pa s$^{-1}$, (e) 1 $\times$ 10$^{-3}$ s$^{-1}$, and (f) 10 $\times$ 10$^{-2}$ PVU, respectively, while solid (dashed) lines represent positive (negative) values in (d) and (e). Areas of wind speed $\geq$10 m s$^{-1}$ are gray shaded at 2 m s$^{-1}$ intervals. Arrows along abscissa indicate the position of the mei-yu front at 850 hPa.
indicating the role played by latent heat release in maintaining the front.

7. Piecewise PV diagnosis

To reveal the role of diabatic latent heating on the northward retreat of the mei-yu front and on the formation and intensification of the LLJ, the EPV perturbation \( \left( q' \right) \) due to diabatic heating effects (LLh) and the associated (inverted) balanced winds at 850 hPa are presented in Fig. 18 for 1200 UTC 7 June–0600 UTC 8 June. It is clear that the area of largest \( q' \) (from LLh) collocated, moved, and intensified with the organized MCS over the mei-yu frontal area (cf. Figs. 1 and 2). East of Taiwan, winds to the north of the mei-yu front were weak northeasterly at 1200 UTC 7 June (Fig. 18a, note that the use of barbs is not conventional). During the next 18 h, balanced winds over the same region shifted into much stronger southerly or southwesterly flow of about 3–5 m s\(^{-1}\) due to latent heating (Figs. 18b and 18c), and this shift of wind within the postfrontal cold air caused the front to move northward. It is also obvious that the southwesterly wind component due to LLh over the area of the LLJ contributed significantly to the generation and intensification of the jet, in agreement with Fig. 17. The balanced geopotential height field due to the LLh component at 850 hPa (Fig. 19) reveals that the latent heat release, being positive in \( q' \), corresponds to negative height values throughout the region, with a trough over the organized MCS (cf. Fig. 2). From 1200 UTC 7 June to 0000 UTC 8 June (Figs. 19a and 19b), due to continuous cumulus convection the balanced height values over the MCS fell further, and the height gradient and the accompanied wind speed over the jet area, immediately southeast of the MCS, increased significantly (cf. Figs. 2 and 4). Thus, the 850-hPa trough and the weak low at the surface were both the result of latent heating (Figs. 1 and 4). As the 500-hPa upper-level forcing was weak and farther to the north (Fig. 5), without the MCS and moist processes it was unlikely for the LLJ to form through the typical process of baroclinic cyclogenesis.

Based on results of the piecewise PV inversion presented in Figs. 18 and 19, the latent heating effects over the mei-yu frontal area generated positive EPV perturbation and a negative height anomaly (trough) at

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Fig. 16. Distribution of 305-K IPV (10\(^{-1}\) PVU) every 6 h from (a) 1200 UTC 7 Jun to (d) 0600 UTC 8 Jun 1998. Contour intervals are \( 1 \times 10^{-3} \) PVU, and negative regions are shaded. Heavy solid and dotted lines indicate LLJ axis and the frontal position at 850 hPa, respectively.
850 hPa. The balanced wind associated with this diabatic heating was the primarily cause for the formation and intensification of the LLJ and the subsequent northward retreat of the mei-yu front. Figure 20 presents the temporal variation of averaged relative vorticity over the region of maximum frontal area at 850 hPa during 0600 UTC 7 June–0600 UTC 8 June, and the partition between different components from the piecewise PV inversion. Here, the average was made over a $4 \times 4$ gridpoint box, centered at the 850-hPa frontal maximum (along dotted lines in Fig. 4) and moving with it. Over the period when the mei-yu front retreated most rapidly (1200 UTC 7 June–0000 UTC 8 June), the averaged value of total balanced vorticity increased significantly (rather than weakened), from about 1.3 to $1.9 \times 10^{-5}$ s$^{-1}$, while an increase by $0.5 \times 10^{-5}$ s$^{-1}$ in the contribution from latent heating effect (LLh) also occurred (Fig. 20). The contribution from the “mean” component, although it exceeded that of LLh both prior to and at 1800 UTC 7 June, experienced rapid decrease mainly because the front was moving into an environment with weaker background vorticity (cf. Fig. 13). The adiabatic effect in the lower troposphere (LLd) had tendency quite close to that of LLh over the entire period, but its contribution to the total vorticity was negative most of the time except at 0000 UTC 8 June. Also negative was the lower boundary effect (LB) due to friction, while effects from upper levels (UL) were negligible as expected (Fig. 20). Therefore, results from the piecewise PV inversion suggest that latent heating effect associated with the organized MCS was the primary mechanism for the maintenance and strengthening of the low-level mei-yu front as it moved northward in the present case.

Likewise, through the piecewise PV inversion, contributions from different components to the LLJ at 850 hPa can be quantified, as presented in Fig. 21 for the time at 0000 UTC 8 June 1998 (cf. Fig. 4c). For the LLJ, the average was performed inside a hexagonal area aligned along the jet axis shown in Fig. 12 and containing a total of 19 grid points. The (balanced) total wind vector after the PV inversion was from the west-southwest at about 11.5 m s$^{-1}$, slightly weaker than the observed wind in ECMWF analysis, as also revealed by Fig. 12. About 6 m s$^{-1}$, or one-half, of the total wind was from the mean field, as shown in Fig. 13, while the...
8. Discussion

In previous sections, a variety of diagnostic methods were used to examine the mechanisms responsible for the northward retreat of a segment of the mei-yu front east of Taiwan, as well as for the formation and intensification of the accompanied LLJ in the present case over 7–8 June 1998 during SCSMEX. Although the contribution from topographical effects also existed as the LLJ moved pass the leeside of Taiwan, these methods yielded consistent results among themselves that are in agreement with Cho and Chen (1995), and a sequence of events that eventually led to the formation of the LLJ and the retreat of the front in the following order could be deduced. The organized MCS developed and propagated northeastward across Taiwan along the front (Fig. 2), and the latent heat release produced (nonconserved) PV (and $\zeta$) growth and height fall at low levels (Figs. 16–18 and 20). The increase in height gradient induced cross-contour, north-westward-directed, ageostrophic flow (Figs. 4 and 14), and the LLJ formed and further intensified to the southeast of the MCS through Coriolis acceleration in the midst of a favorable southwesterly background monsoon flow (Figs. 13, 14, 19, and 21). This shift in winds replaced the previously weak easterly flow north of the front (near Ishigakijima) with west-southwesterly flow (Figs. 7, 14, 18, and 19), and caused the section of the front nearby to move northward significantly prior to 0000 UTC 8 June, mainly through horizontal advection (Figs. 1, 4, and 10). During this process, the MCS brought heavy rainfall to Taiwan (Fig. 3), and the associated height fall from latent heating manifested on weather maps as a trough at 850 hPa and a weak frontal disturbance at the surface (Figs. 1 and 4). The above sequence of events was also revealed in vertical cross sections (Fig. 15) and could be understood in the context of transverse circulation associated with the front. As the MCS developed, latent heating of cumuli enhanced the frontal convergence and the TIC over and south of the front, and the LLJ formed inside the lower (northwestward moving) branch of this secondary circulation near 850 hPa (Figs. 14a and 15d). Finally, the appearance of the LLJ enhanced the northeastward advection and led to frontal retreat near Ishigakijima.
Fig. 19. Nonlinearly balanced fields of geopotential height (gpm) and wind (m s$^{-1}$) associated with LLh component at 850 hPa for (a) 1200 UTC 7 Jun, (b) 0000 UTC 8 Jun, and (c) 0600 UTC 8 Jun 1998. Geopotential heights are analyzed at 5-gpm intervals. The pennant, full barb, and half barb represent 5, 1, and 0.5 m s$^{-1}$, respectively, and areas with winds $\geq$ 5 m s$^{-1}$ are gray shaded (scale shown at bottom). The dotted lines indicate frontal position.
In the present study, the question of whether latent heating associated with the organized MCS was strong or effective enough in warming the atmosphere needs to be addressed. Here we present the vertical profile of apparent heat source \( Q_1 \), as defined by Yanai et al. (1973), over the area of the MCS at 1200 and 1800 UTC 7 June in Fig. 22. At both times, the heating occurred throughout the depth of the troposphere and reached the maximum of 2.2°–2.6°C (6 h)\(^{-1}\) at 400 hPa. The heating was stronger at 1200 UTC at and below this level, but stronger at 1800 UTC above 400 hPa, suggesting a slow ascent of the heating with time. From diabatic processes, the generation of nonconserved PV can be approximated as (Hoskins et al. 1985)

\[
\frac{dq}{dt} = -g \eta \frac{\partial \theta}{\partial p},
\]

where \( g \) is the gravitational acceleration and \( \theta = d\theta/dt \) is the heating rate. Thus, in a qualitative sense, maximum PV generation would occur at where the product of absolute vorticity \( \eta \) and vertical gradient of the heating rate was the largest, that is, in the lower troposphere near (or slightly below) 850 hPa (Figs. 15e and 22). Using the largest \( \eta \) value of about 1.2 \( \times 10^{-5} \) s\(^{-1}\) and \( \partial \theta/\partial p = -1 \times 10^{-4} \) K Pa\(^{-1}\) (6 h)\(^{-1}\) obtained from Fig. 22 (at 925–850 hPa) at 1200 UTC 7 June, the PV generation can be estimated to be about 0.12 PVU (6 h)\(^{-1}\). This value is comparable to the areal average over the MCS in Fig. 17a, and exceeds 1/5 of the 0.55 PVU at the PV center in Fig. 15f. Therefore, the latent heat released inside the organized MCS was sufficiently large to affect the frontal PV, and thus the frontal intensity, a conclusion also reached through the results of piece-

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wise PV inversion and IPV budget analysis (Figs. 17, 18, and 20).

The efficiency of latent heating in warming the atmosphere is linked to the scale of the convection in relation to the Rossby radius of deformation \( L_R \) (\( L_R = Nh/\eta \) for continuously stratified fluid, while \( N \) is the Brunt–Väisälä frequency and \( h \) is the system depth). As pointed out by Holton (1992), if the horizontal scale of the system (\( L \)) is larger than \( L_R \), the wind field will adjust toward the mass field once an imbalance from geostrophy occurs. However, if \( L \approx L_R \), the mass field will adjust toward the wind field, and the heating is inefficient since \( L \) is too small and its effects are dispersed rapidly by inertia-gravity waves. In our case, with the same \( \eta \approx 1.2 \times 10^{-4} \text{ s}^{-1} \), \( h \approx 7.5 \text{ km} \) estimated from the height of maximum heating (at 400 hPa; Fig. 22), and \( N \approx 1.15 \times 10^{-2} \text{ s}^{-1} \) at 1–2 km obtained from Tsungsha sounding at 1200 UTC 7 June (cf. Fig. 1a), the calculation yields an \( L_R \) of roughly 720 km. Judging from Fig. 2, the MCS was of a scale comparable to \( L_R \), and hence the embedded cumulus heating could warm the atmosphere effectively and lead to observable responses in the wind field, as shown in Figs. 4, 14, 18, 19, and 21.

From the definition of \( L_R \), the role played by the mei-yu front can also be stressed, at least qualitatively, in determining the heating efficiency. This efficiency, in cases where a mei-yu front is present, would be highest along the front (which is associated with maximum \( \zeta \)), and similar latent heating elsewhere is unlikely to cause a response at the same level in the wind field or in the growth in \( \zeta \). The ineffectiveness of convective heating at regions other than the front was exemplified in the case studied by Chen et al. (2003), while the nonlinear interaction between convection and the front, which provides a mean to organize the convection, is also consistent with Cho and Chen (1995). For LLJs, many of them over southern China and regions near Taiwan in mei-yu season are, as was the present case, related to deep convection and diabatic heating along the front (e.g., Chou et al. 1990; Hsu and Sun 1994; Chen et al. 1998). Under these circumstances, similarly, an LLJ would form only when the heating was effective and persistent enough, and its appearance could be viewed as a product during the process of readjustment between mass and wind fields toward a new state of geostrophic equilibrium (e.g., Chen 1982; Chou et al. 1990; Chen et al. 2003). Even so, a favorable background of moderate southwesterly flow (of about 5–6 m s\(^{-1}\)) nevertheless remained essential for the enhanced winds to reach LLJ intensity, as in many previous studies (e.g., Chen and Yu 1988; Zhong et al. 1996; Igau and Nielsen-Gammon 1998; Chen et al. 2003). In the present case, the additional contribution from topographical effect apparently also helped the LLJ’s intensity and the accompanied frontal retreat.

Chen and Chang (1980) performed vorticity budget analysis to diagnose a mei-yu front case that moved slowly southward at 9 km h\(^{-1}\). Their results showed that horizontal advection at both sides of the front contributed negatively to local tendency of \( \zeta \), and suggested minimal effects of advection in driving the frontal movement, as one would expect for stationary and slow-moving mei-yu fronts. In our study, however, the strengthening of the southwesterly flow and the formation of the LLJ in response to the diabatic latent heating of the MCS caused the low-level wind near Ishigakijima to shift into strong southwesterly. This shift in wind, over the area then behind the front, led to large positive contribution to \( \zeta \) tendency from horizontal advection to the north of the front and caused its rapid retreat at about 25 m s\(^{-1}\) over the region. Apparently, these changes in wind field also caused the maximum \( \theta_e \) gradient to separate from the \( \zeta \) maximum and the seemingly collapse of the frontal wind shift zone (Figs. 4 and 9). This dramatic transition in frontal characteristics during the retreat, into those more similar to a synoptic warm front (cf. Fig. 1c), is also an interesting aspect but not investigated in detail here. Nevertheless, the underlying mechanism leading to the northward retreat of a mei-yu front, to our knowledge, has not been studied, and the present work represents the first of such efforts.

9. Conclusions

During SCSMEX in 1998, a mei-yu front with MCS development near Taiwan exhibited rapid northward movement for about 300 km over a 12-h period from 1200 UTC 7 June to 0000 UTC 8 June. Gridded ECMWF-TOGA advanced analysis at resolutions of 0.5° × 0.5° and 6 h were employed as the primary data source, and various methods including the piecewise PV inversion were used to diagnose the frontal movement as well as the formation and intensification of an LLJ south of the front in this case. Results yielded using different methods were consistent, and major findings can be summarized as follows:

1) The organized MCS, while propagating northeastward along the mei-yu front, released diabatic latent heat and generated nonconserved PV and height fall at low levels. The enhanced height gradient induced northward-directed ageostrophic flow, and the LLJ formed/intensified southeast of the MCS through Coriolis acceleration. Then, the appearance of southwesterly flow east of Taiwan, with addi-
tional contribution from topographical effect, led to rapid retreat of the front over the region near Ishigakijima, for about 300 km at a speed of 25 m s\(^{-1}\).

2) Enhanced by the deep convection along the mei-yu front, a thermally indirect secondary circulation appeared over and south of the front in the NW–SE-oriented cross section. The southwesterly LLJ, maximized at 850 hPa, formed inside the lower branch of this transverse TIC in a potentially unstable environment during the process of readjustment toward geostrophic balance.

3) Results of piecewise PV inversion suggest that the latent heating effects were associated with a cyclonic flow pattern surrounding the MCS and were the major contributor toward frontal intensity and the increase in wind speed over the jet area. Nevertheless, the enhanced southwesterly wind was unlikely to reach LLJ strength had it not been superimposed upon a favorable background current of about 6 m s\(^{-1}\) in the same direction. Topographical effects, in addition, also enhanced local wind speed at lower levels as the jet migrated to the leeside of Taiwan.

4) In the present case, the horizontal scale of the MCS along the front was found to be comparable to the Rossby radius of deformation, and the latent heating was strong and efficient enough to produce observable response in the wind field.

5) This mei-yu front case was shallow (below 700 hPa) with little vertical tilt, and exhibited typical weak \(\theta\) gradient but significant horizontal shear vorticity. The vorticity budget analysis indicates that its northward movement was dominated by horizontal advection, as winds turned to southwesterly over regions initially north of the front.

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