The Structure of a Subtropical Prefrontal Convective Rainband. Part II: Dynamic and Thermodynamic Structures and Momentum Budgets

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ABSTRACT

A thermodynamic retrieval method was employed to investigate the dynamic and thermodynamic structures of a subtropical prefrontal convective rainband associated with the Mei-Yu front on 25 June 1987 over northwestern Taiwan. Three-dimensional wind fields were derived from the dual-Doppler data based on the methodology developed at Saint Louis University. Subsequently, fields of perturbation pressure and temperature were retrieved from the detailed wind field using the three momentum equations.

Results show that the maintenance of this long-lived rainband at the times of dual-Doppler analysis is caused by the combined effects of a gust front arising from the convective downdrafts ahead of the front and developments of new convection along the gust front. In the lowest layer, high pressure occurs behind the cold front with low pressure to its southeast. A buoyancy-induced low pressure area lies beneath the frontal updraft corresponding to the rising warm environmental air. The precipitation core associated with the frontal updraft is elongated toward the southeast side with the environmental shear in the mid- and upper troposphere forming the convective downdraft on the warm side of the surface front. This precipitation-induced downdraft transports cooler air downward producing a high pressure area underneath the convective downdraft. This high is accompanied by a temperature deficit resulting in cold horizontal outflows in the boundary layer. Part of these cold outflows interacts with the high-$
\delta$-southwest monsoonal flow to form a gust front ahead of the surface front. New convection develops along a gust front and then merges with the old convection, thereby prolonging the lifetime of the rainband. The vertical flux convergences and divergences of horizontal momentum by organized convection are largely responsible for forming a midlevel jet and weakening a low-level jet. The momentum budget calculation shows that the horizontal and vertical flux convergences and divergences of horizontal momentum by the mean and eddy motions are the major contributor to maintain the mean momentum.

1. Introduction

In Lin et al. (1992), referred to as Part I hereafter, the kinematic structure of a subtropical prefrontal convective rainband, which occurred on 25 June 1987 over northwestern Taiwan, was investigated. The dual-Doppler data employed were obtained from NCAR (National Center for Atmospheric Research) and TOGA (Tropical Ocean Global Atmosphere) radars during the Taiwan Area Mesoscale Experiment (TAMEX). The dynamic and thermodynamic structures as well as horizontal momentum fluxes and budgets of the same rainband are reported in this paper.

The structures of a fast-moving subtropical squall line over the Taiwan Strait during TAMEX intensive observing period 2 (IOP 2) were reported in studies by Wang et al. (1990) and Lin et al. (1990). They found that the overall structural features of the squall line were quite similar to those of a tropical squall line. In the lowest layer a mesohigh formed behind the gust front due to the saturated convective downdraft. To the east of a mesohigh, a mesolow developed below the convective updraft due to its buoyancy. In the mid- and upper layers, the updraft–shear interaction resulted in high pressure on the upshear side with low pressure downshear. The orientation of horizontal pressure gradients was in the approximate direction of the average shear vector in the domain. The mesoscale pressure field accelerated the updrafts front-to-rear resulting in a westward tilt of the updrafts in the layer below the low-level jet (LLJ). The rear-to-front air entering from the back of the line was negatively buoyant producing a sloping downdraft behind the main cell. As the cool descending midtropospheric air approached the ground, it diverged horizontally to form a cold outflow.

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behind the gust front. Part of the descending air moved forward, interacting with the advancing environmental high-$\theta_e$ air at the leading edge to form new cells ahead of the old (main) cells. The interplay between a cell's cold surface outflow and the low-level shear contributed to the maintenance of the squall line. The vertical transport of horizontal momentum normal to the line was downward corresponding to the west-tilted convective updrafts, while the line-parallel component produced upward momentum transport. Convergences of horizontal and vertical momentum fluxes by the mean and eddy motions were mainly responsible for mean momentum generation.

Unlike the fast-moving squall line in TAMEX IOP 2, the prefrontal convective rainbow in IOP 13, detailed in Part I of this study, developed on the warm side of the Mei-Yu front over northwestern Taiwan as the front was approaching the northwest coast. The southwesterly monsoon flow in the warm sector was accompanied by an LLJ near the 900-mb level. The Mei-Yu front provided large-scale lifting in the planetary boundary layer (PBL) to initiate organized convection quite close to the front where the front was located far north of the island. As the system moved down the coast, the reflectivity cores associated with the frontal updraft were elongated toward the southeast in the direction of the shear in the mid- and upper troposphere. These reflectivity cores induced the convective downdrafts ahead of the surface cold front due largely to precipitation loading. The descending air of these downdrafts resulted in cold diverging outflows in the PBL, interacting with the incoming high-$\theta_e$ air to form a gust front on the southeast side of the front. As a result, new cells formed along the gust front and then merged with the old cells, thereby prolonging the life span of the squall line. Such mechanisms are different from those of the frontal rainbow in northern California studied by Carbone (1982). The prefrontal squall line traveled at a slow speed south-southeastward at 2–3 m s$^{-1}$, and produced heavy rainfall on the west coast of Taiwan. It consisted of many cells in a narrow band. Each cell was accompanied by the moderate convective updraft (6–8 m s$^{-1}$) and the weak downdraft (2–4 m s$^{-1}$). These updraft–downdraft speeds are likely to be underestimated due to the 1-km grid spacing used and the filtering inherent in the objective analysis procedure (see Part I for details). This point must be kept in mind when the results are interpreted.

The purpose of this paper is to study the dynamic and thermodynamic structures of the prefrontal rainbow reported in Part I. The thermodynamic retrieval method of Gal-Chen (1978) was employed to recover thermodynamic variables from dual-Doppler-derived winds at two analysis times. Subsequently, the budget and vertical transports of horizontal momentum, averaged over the horizontal domain of interest, were computed at each analysis level. Results were then compared to those for a fast-moving subtropical squall line in IOP 2. The goal of the research was to gain a better understanding of the mesoscale structure of a Mei-Yu prefrontal convective system that produced heavy precipitation in northwestern and central Taiwan.

A brief description of the weather situation is presented in section 2. Section 3 summarizes dual-Doppler analysis and computational procedures, and section 4 shows some squall-line characteristics determined from the surface observations and dual-Doppler analysis. The retrieval results are discussed in section 5, followed by the summary and conclusions in section 6.

2. Weather situation

The synoptic weather situation and environmental condition were described in detail in Part I. Briefly, a low pressure center was located to the southwest of Japan at 1200 UTC 24 June 1987. [Note that the local standard time (LST) is 8 h later than UTC.] A shear line extended from the low center southwestward to southeast China. Ahead of the shear line, an LLJ from the southwest at 15 m s$^{-1}$ prevailed over the northern tip of Taiwan. Winds veered with height at all levels, exhibiting large-scale warm advection. In the upper troposphere, an area of diffusiveness occurred over northeastern Taiwan with large-scale confluence at low levels. These synoptic weather patterns contributed to unstable environmental conditions in the warm sector favorable for thunderstorm development. Calculation of vertical velocity ($p$ velocity) for the same case by Jou and Deng (1990) showed that large-scale upward motion dominated at the times before and during the frontal passage with large-scale convergence below and divergence aloft.

The cold front passed through Panchiao near Taipei at 2300 LST 24 June and Makung by 1900 LST 25 June. Up to 200 mm of rainfall in 24 h was reported on the west coast of central Taiwan in association with the prefrontal rainbands.

The prefrontal and postfrontal environmental conditions extracted from the rawinsonde data from Panchiao and Makung were detailed in Part I. Values of CAPE (convective available potential energy) were relatively small, ranging from 1000 to 1400. These values are comparable in magnitude to those in GATE [GARP (Global Atmospheric Research Program) Atlantic Tropical Experiment] and TAMEX IOP 2. Note that CAPE is a measure of the amount of buoyant energy available in the environment to drive updrafts. A smaller value of CAPE implies that convective updrafts within the rainbands are likely to be weak or moderate, but not intense (see Part I for details).

3. Data analysis and computational procedures

As described in Part I, two volume scans obtained from CP-4 and TOGA at 0653 and 0700 LST 25 June were considered in this study. We employed the data
analysis and reduction procedures similar to those reported in studies by Lin et al. (1986) and Wang et al. (1990) to process dual-Doppler data. The derived three-dimensional winds were subjected to variational adjustment to render them internally consistent (see Lin et al. 1986 for details).

Once the detailed wind field was obtained, the thermodynamic retrieval method of Gal-Chen (1978) was employed to recover fields of deviation perturbation pressure and virtual temperature using the three-momentum equations. Note that the retrieved virtual temperature contains a contribution from cloud water. For details, see studies by Lin et al. (1986), Parsons et al. (1987), etc. The retrieved fields are subjected to momentum checking (E_r) to determine the level of confidence. Relatively low E_r values indicate that the agreement is good between the retrieved (best fit) horizontal pressure gradients and the pressure gradients derived from the individual momentum equations (including all terms). Values of E_r obtained in this study range from 0.29 to 0.35 with a volume mean of 0.33.

4. Squall-line characteristics

The surface observations in 30-min intervals collected during TAMEX IOP 13 were used to derive certain important properties of the prefrontal squall line in relation to the Mei-Yu front. Some of these surface analyses have already been presented in Part I (see Figs. 7 and 8 and section 2c). Results obtained show that the frontal position, indicated by the vertical dashed line in the figures, trailed the leading edge of the heavy rainfall by 2.5 h at Hsinchu (Fig. 7b) and by 14 h at Ching-Chung-Kang (CCK) (Fig. 8a). Recall that the TOGA radar was positioned at the CCK Air Force Base. These surface observations further reveal that the leading edge of the squall line passed through Hsinchu and CCK at 0400 and 0800 LST, respectively. At the time of squall passage, the wind changed in direction from the southwest to west, gusting to 7–10 m s^{-1} and pressure rose rapidly (~1 mb). Heavy rainfall with a maximum amount of 40–50 mm in one hour occurred after the squall passage. In addition, the surface observations taken at the Civil Aeronautic Administration (CAA) and CP-4 radar sites also exhibit the similar features. The surface traces all show that the passage of the prefrontal squall line was accompanied by signs of a gust front with temporary cooling and temporary wind shifts.

Table 1 presents some highlights extracted from the surface observations in the period from 0100 to 2300 LST 25 June. Among those seven stations listed, the CCK and Wuchi traces are very instructive since they are located near the southern edge of the dual-Doppler coverage area (see a circle in Fig. 1). In Fig. 1, the approximate positions of the Mei-Yu front (solid line) in 3-h intervals, determined from the surface analysis, are displayed. In addition, centers of the precipitation maxima at intervals of every hour, and the positions of the squall-line gust front (dashed line) at some selective times are also displayed. From Table 1 and Fig. 1, it is apparent that the squall line was located very close to the Mei-Yu front at the CAA radar site around 0330 LST. As the frontal system moved slowly toward the south-southeast along the northwest coast of Taiwan, the squall line quickly moved away from the cold front. The leading edge of the squall line reached Taoyuan around 0400 LST, approximately 1 h ahead of the frontal passage. As the system continued to move down the coast, the squall line moved more rapidly away from the cold front. As a result, the frontal position trailed the squall-line gust front by 3.5 h at Hsinchu, by 7.5 h at the CP-4 radar site, and by 15 h at CCK. The precipitation maxima depicted in Fig. 1 also exhibited similar results. The surface analyses further reveal that the Mei-Yu front became quasi-stationary in the dual-Doppler coverage area after 0900 LST (see the successive frontal positions in Fig. 1 for comparison). It traveled at only 0.5–1 m s^{-1} toward the TOGA radar. As a result, the frontal position trailed the leading

<table>
<thead>
<tr>
<th>Station Name</th>
<th>Station Code</th>
<th>Passage time (LST)</th>
<th>Surface traces during squall passage</th>
</tr>
</thead>
<tbody>
<tr>
<td>CAA (CKS Airport)</td>
<td>686</td>
<td>0330 0330</td>
<td>SW→N Yes Rise Strong Heavy</td>
</tr>
<tr>
<td>Taoyuan</td>
<td>697</td>
<td>0400 0500</td>
<td>SW→N Yes Rise Moderate Heavy</td>
</tr>
<tr>
<td>Hsinchu</td>
<td>756</td>
<td>0400 0730</td>
<td>SW→W Yes Rise Moderate Heavy</td>
</tr>
<tr>
<td>CP-4*</td>
<td>—</td>
<td>0530 1300</td>
<td>SW→W Yes — — Heavy</td>
</tr>
<tr>
<td>TOGA (CCK)</td>
<td>770</td>
<td>0800 2300</td>
<td>SW→W Yes Rise Strong Very heavy</td>
</tr>
<tr>
<td>Wuchi</td>
<td>777</td>
<td>0730 2200</td>
<td>S→SW Yes Rise Strong Very heavy</td>
</tr>
<tr>
<td>Makung</td>
<td>734</td>
<td>1330 1900</td>
<td>SSW→SW Yes Rise Moderate Moderate</td>
</tr>
</tbody>
</table>

* Based on the report at the CP-4 radar site.
The successive positions of the Mei-Yu front from 0200 to 2000 LST 25 June 1987 in 3-h intervals. The hatched area shows topographical heights greater than 1 km. The surface station code numbers are defined in Table 1. A circle signifies the dual-Doppler coverage area of CP-4 and TOGA. The squall-line gust front at selective times as listed in Table 1 is indicated by the dashed line. An open circle represents the center of precipitation maximum over the period from 0400 to 0900 LST in intervals of every hour.

The dual-Doppler analysis at 0653 and 0700 LST 25 June presented in Part I shows that the wind-shift line, corresponding to the position of the Mei-Yu front, is located about 20–30 km northwest of TOGA. The leading edge of the squall line, on the other hand, is located about 8–10 km northwest of TOGA. Figure 3 displays the reflectivity and radial velocity fields along the 315° radial (see line AB in Fig. 2) at 0653 LST as seen from TOGA. Unfiltered reflectivity data are used in the display. It is seen that the Mei-Yu front (heavy dashed line) is located 23 km northwest of TOGA. It is accompanied by high reflectivity (Fig. 3a) and low-level convergence (Fig. 3b). The prefrontal rainband associated with the squall line is evident in a broad area southeast of the Mei-Yu front with a reflectivity maximum of 40 dBZ. The leading edge of the squall line (GF) is signified by the low-level weak convergence at 9 km on the northwest side of TOGA. The reflectivity core (shaded) ahead of the gust front is associated with the new cell. This new cell eventually merges with the old cells behind the gust front, thereby prolonging the lifetime of the prefrontal rainband (see Part I for details).

The observational evidence presented above leads us to hypothesize that the squall line being investigated forms quite close to the Mei-Yu front over the far north of Taiwan. As the system moves slowly down the central west coast, the squall line quickly moves away from the cold front as gust fronts are formed by convective cells at the leading edge of the system. The convective cells generate cold outflows and these outflows are the likely source of the more rapid movement of the squall line in comparison with the slow movement of the Mei-Yu front. The results are more applicable to those of Trier et al. (1991) from PRE-STORM (Preliminary Regional Experiment for STORM-Central) in their study of a prefrontal squall line that developed in an unstable environment with low vertical shear. It is less applicable to the case of a severe frontal rainband studied by Carbone (1982) where strong lifting was maintained along the leading edge of a cold front in very high vertical shear and near-neutral stability.

5. Discussion of thermodynamic retrieval results

In Part I of this study, fields of the horizontal system-relative wind, convergence and divergence, and vertical velocity were presented in a north–south rectangle with horizontal dimensions of 40 km × 36 km using a 1-km grid spacing. The orientation of the rainband was almost in a northeast–southwest direction within the domain of interest, and the line moved slowly toward the south–southeast at 2.5 m s⁻¹. This line motion was
Fig. 2. Fields of the (a) horizontal system-relative wind with reflectivity contours superimposed, and (b) horizontal convergence and divergence for 0.75 km at 0700 LST 25 June 1987. Reflectivities larger than 30 dBZ are shaded. The heavy dashed line shows the approximate position of the Mei-Yu front, while the dotted line indicates the squall-line gust front (GF). The dash-dot line shows the position of GF at 0653 LST. Distances are in kilometers from TOGA (T).

determined from the reflectivity displays obtained from the Kaohsiung radar at 20-min intervals. The same method was employed by Lin et al. (1990) for a subtropical squall line during TAMEX IOP 2 with success. The mean line motion obtained is in agreement with the leading edge of the rainband as seen from CP-4 over the period from 0634 to 0706 LST 25 June (not shown). It is much slower than cell motions (~6–7 m s\(^{-1}\) from the west-southwest) determined from the CP-4 radar information.

As noted earlier, the momentum budget and vertical transports of horizontal momentum, averaged over the domain of interest, are stressed in Part II. For this reason, it is necessary to rotate the horizontal coordinates having the x axis toward the southeast in a direction normal to the line, and the y axis toward the northeast parallel to the line (see Fig. 4 for illustration). This approach is similar to that employed by LeMone (1983) for studying a tropical convective line. In this rotated coordinate system, a value of the normal wind \(u\) is positive in the direction of line motion, while the line-parallel component \(v\) is positive toward the northeast. The origin of the rotated coordinates is located 45 km west and 17 km north of the TOGA radar (Fig. 4).

a. Pressure and temperature perturbations in deep convection

The pressure field is retrieved from Doppler-derived winds using the horizontal momentum equations at a given level with the Neumann boundary condition. For details, see studies by Hane and Ray (1985), Lin et al. (1986), Parsons et al. (1987), etc. Subsequently,
the virtual potential temperature deviation (θ'_{cd}) is recovered from the buoyancy equation in the form of

\[ \theta'_{cd} = \frac{\theta_{o}}{g} \left[ c_{p} \theta_{o} \frac{\partial \pi}{\partial z} + \left( \frac{Dw}{Dt} - \frac{Dw}{Dt} \right) \right] \]

\[ VPG \quad VAC \]

\[ + g(q_{r} - \langle q_{r} \rangle) + (f_{1} - \langle f_{1} \rangle) \] \( \text{(1)} \)

\[ RWL \quad VFF \]

where \( \pi = (P'_{d}/P_{o})^{R/\gamma} \) is a perturbation Exner function deviation; \( P'_{d} = P_{o} \pi^{R/\gamma} \) is the pressure deviation; \( f_{1} \) is the forces other than pressure gradients; \( \theta_{o} \) is the environmental mean virtual potential temperature; \( \langle \cdot \rangle \) is the horizontal area mean; \( \langle \cdot \rangle' \) is the deviation from the layer mean; and \( q_{r} \) is the rainwater mixing ratio. The other symbols have their usual meanings. A value of \( q_{r} \) was estimated from radar reflectivity, and the environmental mean (subscript \( o \)) was determined from the prefrontal mean sounding reported in Part I.

Terms VPG, VAC, RWL, and VFF in (1) represent the contributions of vertical perturbation pressure gradient, vertical acceleration, rainwater loading, and friction, respectively, to \( \theta'_{cd} \). With the aid of the perturbed potential temperature equation, we obtain \( T'_{cd}/T_{o} \approx \theta'_{cd}/\theta_{o} \) since the \( P'_{d}/P_{o} \) term is much smaller than the other two terms based on the results obtained. The retrieved temperature represents a virtual cloud temperature since it accounts for both deviation of virtual temperature perturbation and cloud-water content (Roux et al. 1984). It also contains errors in winds, pressure, and rainwater mixing ratio estimates.

These must be kept in mind when the results are interpreted. Since the grid spacing used is 1 km in all three directions, only the thermodynamic field for large-scale features and some large cells within the prefrontal rainband with wavelength larger than 6–7 km can be adequately resolved. Additionally, the updraft–downdraft speeds are likely to be underestimated due to the filtering inherent in the objective analysis scheme that is used (see Part I for details).

\[ b. \text{Horizontal view at 0.75 km} \]

Fields of the horizontal system-relative wind with radar reflectivity contours superimposed, the vertical velocity, the deviation perturbation pressure, and the deviation perturbation virtual temperature at 0653 LST are shown in Fig. 5. Recall that the \( x \) axis of the rotated coordinates is directed toward the southeast, positive in the direction of line motion (Fig. 4). Distances are in kilometers from the origin \((0, 0)\). Note that the low-level wind-shift line (heavy dashed line) is evident at this level. Its position corresponds well to the Mei-Yu front (see Part I for details). On the left side of the wind-shift line, the winds are predominantly from the northwest carrying cooler, drier air from northern China to the Taiwan Strait. Note that the northwesterly wind becomes northeasterly at the surface due to the “channel effect” in the Taiwan Strait. Conversely, the southwest monsoon flow prevails in the warm sector (right side of the wind-shift line). This strong low-level flow is associated with the LLJ on the warm side of the front as depicted in Part I. It transports high-\( \theta_{o} \) air from the South China Sea all the way to the wind-shift line. Such air feeds the convective updrafts, resulting in a broad area of upward motion (see the hatched areas in Fig. 5b) in the vicinity of the wind-shift line. The maximum updraft speed at this level is about 4 m s\(^{-1}\). Conversely, downward motion dominates in a broad area southeast of the wind-shift line. As noted in Part I, such downward motion is associated with the precipitation-induced convective downdrafts on the warm side of the Mei-Yu front. These convective downdrafts generate cold outflows in the PBL. Part of these cold outflows move toward the southeast, interacting with incoming high-\( \theta_{o} \) air from the southwest to form a gust front (dotted line). This gust front indicates the leading edge of the squall line. It is characterized by low-level convergence, upward motion, high reflectivity, and slight wind shift.

It must be pointed out that the cold air associated with the Mei-Yu front is relatively shallow (\( \sim 1 \) km) as compared to that for a midlatitude front. For example, see TAMEX studies by Chen et al. (1989), Trier et al. (1990), Ray et al. (1991), etc. Additionally, Chen (1993) showed that the Mei-Yu front during TAMEX IOP 13 was characterized by weak temperature gradients and large moisture contrasts as it moved across the island. The effect of warm water in the Taiwan
Strait is largely responsible for modifying the characteristics of a Mei-Yu front, and decreasing the temperature contrast as the front moves slowly from the northern tip to the southern tip of Taiwan (Trier et al. 1990).

Based on the aforementioned studies and the observed data presented in Part I, we believe that the cold air associated with the Mei-Yu front in IOP 13 was only 0.5–1 km in depth at the times of analysis. The depth of the cold air decreased gradually from the west.
The coast of Taiwan toward the Taiwan Strait due to the effect of warm water over the strait. This must be kept in mind when the retrieved thermodynamic fields are interpreted. At the 0.75-km level, the northeast part of the domain (upper half of Fig. 5) is quite close to the coast, and is below the top of the cold pool. Hence, this region is considerably influenced by the cold air behind the front. On the other hand, the southwest portion of the domain (lower half of the figure), located 30–40 km west of the coast, appears to be above the top of the cold pool, where the air behind the surface front is relatively warm and moist. The postfrontal sounding at Panchiao (see Fig. 12b in Part I) reveals that the air immediately above the front is moist and unstable. The $\theta_e$ at the surface is only 346 K, indicative of the low-$\theta_e$ air from northern China. The value of $\theta_e$ increases to 353 K at 0.8 km and remains high ($\sim 350$ K) up to 4.5 km, showing the presence of high-$\theta_e$ air above the front. This high-$\theta_e$ air blows from the west in the layer between 900 and 600 mb, and provides inflow for the convective rainband located on the warm side (southeast) of the front. The postfrontal Makung sounding also exhibits the same features, except the moist layer above the front is somewhat shallower (900–700 mb).

The retrieved $P_d$ field at 0.75 km is presented in Fig. 5c. The magnitude of $P_d$ varies from $-90$ Pa ($-0.9$ mb) to $60$ Pa (0.6 mb) resulting in strong horizontal pressure gradients within the rainband. Such pressure gradients are of the order of 0.02 m s$^{-2}$ and are nearly balanced by the horizontal accelerations along the same directions derived from Doppler winds. The alongfront pressure variations are mainly caused by the accelerations and decelerations of the southwest flow on the southeast side of the windshift line. Because the vertical acceleration is small compared with the pressure and buoyancy terms in the lower layer, the perturbation pressure and buoyancy are nearly in hydrostatic balance. Notice that the high pressure centers behind the cold front (heavy dashed line) and the squall-line gust front (dotted line) over the northeast portion of the domain ($y > 18$ km) are attributed to the cooler air over those regions that are denser than the surrounding air. The low pressure centers ahead of the cold front appear to be due to buoyancy, since the low centers lie beneath more buoyant high-$\theta_e$ environmental air. On the other hand, the low pressure centers behind the surface front over the southwest part of the domain ($y < 18$ km) are associated with the warmer air from the west above the cold pool mentioned previously. The presence of this warm air over the southwest domain is also supported by the retrieved temperature field at the same level.

Figure 5d displays the retrieved $T'_{ed}$ field. It is generally in agreement with the updraft–downdraft structure (Fig. 5b) with relative warming in updrafts and cooling in downdrafts. The temperature deficits over a broad area on the southeast side of the front are largely caused by the evaporation-cooled downflow associated with convective downdrafts over that area (see Part I for details). The density of such downflow air is further increased by precipitation loading in the elongated precipitation cores mentioned previously. As a result, values of $T'_{ed}$ are generally negative with a maximum of $-3^\circ$C in a widespread area southeast of the front. Such negatively buoyant air is partially responsible for developing and sustaining the squall-line gust front located on the southeast side of the prefrontal rainband (dotted line). As explained earlier, such cold outflows are the likely source of the more rapid movement of the squall line as compared to the slow movement of the Mei-Yu front. Over the southwest portion of the domain (lower half of Fig. 5d), a temperature excess with a maximum of 2°C is found in a broad area behind the surface cold front (heavy dashed line). As explained before, this warm region is associated with the warm moist air from the west above the front due to the shallowness of the cold pool. Inspection of Fig. 5 further reveals that centers of maximum warming and cooling do not exactly match with centers of updrafts and downdrafts. Instead they are slightly tilted toward the southwest side of the updraft and downdraft centers, respectively, as observed previously by Lin et al. (1990) and others.

Calculation of the individual terms in (1) reveals that terms VPG and BUO have much larger values than terms VAC, RWL, and VFF at this level and levels above (at 1.75 and 2.75 km). Since buoyancy is derived from VPG and VAC [see Eq. (1)], it therefore must be of the order of VPG if VAC is very small. This pressure–buoyancy relation results from a near-hydrostatic balance between the perturbation pressure and temperature. Similar behavior has been reported in studies by LeMone (1983), Lin et al. (1990), etc.

c. Horizontal view at 3.75 km

Horizontal distributions of the system-relative wind (with reflectivity contours superimposed), the vertical velocity, the deviation pressure, and the deviation temperature at 3.75 km are presented in Fig. 6. At this level, the wind-shift line is no longer evident. Winds are predominantly from the west and southwest at speeds larger than 10 m s$^{-1}$. On the southeast side of the front, reflectivities are not homogeneous with values ranging from 10 to 35 dBZ. Conversely, reflectivities behind the front are generally weak and homogeneous exhibiting stratiform precipitation. Notice that many cells are embedded within the rainband. The high-reflectivity region with $Z > 30$ dBZ (shaded) corresponds to each precipitation core. Between cells, the reflectivity gap is apparent.

Upward motion (Fig. 6b) prevails in a broad area along the surface cold front. The maximum updraft at this level is about 8–9 m s$^{-1}$ and is located to the southeast of the front near $x = 22$ km and $y = 30$ km.
Another zone of upward motion is located in the vicinity of the squall-line gust front over the southeastern edge of the domain. Downward motion dominates in a broad area behind the gust front with a maximum of $-4 \text{ m s}^{-1}$. Such downward motion is related to the convective downdrafts that occur in the aforementioned warm sector. Most of these convective downdrafts are located in high reflectivity regions, indicative of the effect of precipitation loading on downdrafts.

The pressure field (Fig. 6c) as a whole displays high pressure on the northwest side of the surface front with low pressure on the southeast side, resulting in strong horizontal pressure gradients in the direction from the west-northwest to east-southeast. A maximum hori-
zontal pressure gradient in the vicinity of some updrafts is oriented approximately in the direction parallel to the environmental shear vector $V_x$. This shear vector is from $300^\circ$ at this level. High pressure also occurs in a broad area behind the gust front. It is associated with the downdraft in the high reflectivity region noted earlier. Magnitudes of $P_d'$ vary from 30 to $\sim 60$ Pa.

As stated in Lin et al. (1990), the pressure field in a subtropical squall line was quite similar to that for a tropical convective line, for example, LeMone et al. (1984). This field was caused by both the hydrostatic and dynamic effects. In the lower layer ($z < 2$ km), the buoyancy effect was more important than the dynamic effect. In the mid- and upper layers, both the dynamic and buoyancy effects contributed significantly to the pressure perturbation. In a similar manner, this study also shows that the pressure–buoyancy relation, due to a near-hydrostatic balance between perturbation pressure and temperature, prevails in the lower layer. Warmer air to the southeast of the wind-shift line can produce low pressure below the frontal updraft, while cooler air to the northwest of the wind-shift line results in a high pressure area below the downdraft (Fig. 5).

In the mid- and upper layers, the shear–updraft interaction term appears to become the major contributor to the pressure perturbation. Calculation of the individual terms in (1) shows that terms $VAC$, $VPG$, and $BUO$ are the three dominant terms. Unlike 0.75 km, the vertical accelerations at 3.75 km and levels above have magnitudes comparable to the vertical pressure gradient and buoyancy terms. Consequently, the perturbation pressure–buoyancy relation associated with the hydrostatic effect mentioned before is not as evident at this level and levels above.

In these mid- and upper layers, the pressure perturbation is largely determined by both the dynamic and buoyancy effects, but not the buoyancy alone. These findings are to some extent similar to those presented in the study by Lin et al. (1990) for TAMEX IOP 2.

The retrieved temperature field (Fig. 6d) is closely related to vertical velocity (Fig. 6b). In general, warming occurs in convective updrafts with cooling in downdrafts. A temperature excess is largely attributed to the release of latent heat by condensation, while a temperature deficit is caused by evaporative cooling. As in Fig. 5, centers of the maximum and minimum temperature deviations occur slightly to the west of updrafts and downdrafts, respectively. The magnitude of $T_{vd}'$ ranges from $-2$ to $2^\circ$C in the convective region and becomes smaller in the stratiform region.

d. Horizontal view at 6.75 km

Figure 7 displays the features of the prefrontal rainband at 6.75 km. System-relative winds (Fig. 7a) are predominantly from the northwest and west at speeds greater than 12 m s$^{-1}$. The rainband is still visible on the southeast side of the surface front showing several cells with reflectivity maxima of 30–35 dBZ (shaded). The precipitation cores of these convective cells are elongated southeastward following the prevailing flow in the direction nearly normal to the line. Upward motion prevails in the narrow zones along the surface front and the gust front with maximum speeds of 4–5 m s$^{-1}$ (Fig. 7b). Between cells, downward motion occurs. Like the precipitation cores, the convective updrafts and downdrafts are also elongated toward the southeast.

The $P_d'$ field (Fig. 7c) exhibits a pattern of high pressure on the east side of updrafts with low pressure on the west side. Such a pattern is opposite to that presented earlier for 3.75 km. This is attributed to the fact that the wind decreases its speed with height above the 5.4-km level resulting in negative vertical shear in the layer between 5.4 and 7.5 km (see Fig. 13 in Part I). The environmental shear vector at this level is from the east, in contrast to that from the west-northwest at 3.75 km.

Calculation of the individual terms in the horizontal pressure equation shows that the updraft–shear interaction term is dominant over the other terms at this level and those above and below. This result indicates the importance of the dynamical effect on the perturbation pressure in the upper troposphere. It is similar to that found in TAMEX IOP 2 (Lin et al. 1990) for a subtropical squall line. The overall temperature pattern (Fig. 7d) matches well with the updraft–downdraft structure with warming in updrafts and cooling in downdrafts. The magnitude of $T_{vd}'$ ranges from $-2$ to $3^\circ$C.

e. Vertical cross section

Figure 8 displays fields of the system-relative wind (with reflectivity contours superimposed), vertical velocity $w$, deviation pressure $P_d'$, and deviation temperature $T_{vd}'$ along line $AB$ in Fig. 5a. Notice that the front (heavy dashed line) is located near 20 km from the origin. Ahead of the front, reflectivity is deep but not intense. The height of 20-dBZ contour reaches 10 km and higher.

As explained in Part I, the reflectivity cores (shaded) on the warm side of the front shift toward the southeast (right side of the figure) with the environmental shear. Downward motion prevails in these high-reflectivity regions (Fig. 8b), indicative of the close relationship between precipitation loading and downward motion. The descending air of each convective downdraft carries much cooler air from the mid- and upper layers, resulting in a high pressure area near the surface (Fig. 8c). Such cooler air is evident in the temperature field (Fig. 8d) with a temperature deficit of up to $-3^\circ$C in a widespread area underneath the downdraft. As the descending air approaches the surface, it spreads out horizontally, forming cold outflows in the PBL. Part of these cold outflows moves southeastward, interacting with the southwest monsoon flow to form a gust front (GF) at 32 km. As noted earlier, this gust front indicates the leading edge of the
A new cell forms at the GF with a reflectivity maximum of 35–40 dBZ. This new cell eventually merges with the old cells behind the GF, thereby prolonging the life span of the prefrontal rainband. High pressure forms behind the GF with low pressure to its southeast ahead of the GF. Such horizontal variations in pressure are consistent with those for tropical and subtropical squall lines reported in the literature. Upward motion at the GF is accompanied by a temperature excess due to the latent heat release by condensation.

As explained in Part I, the Mei-Yu front lifted the low-level, warm, moist southwest flow to initiate organized convection along the front. The frontal updraft seen at 21 km is caused by the combined influence of frontal forcing in the PBL and positive buoyancy above.
These processes continue over the Taiwan Strait as the front moves slowly toward the south-southeast. A northwestward branch of the surface cold outflow (20–24 km from the origin) enhances low-level convergence at the leading edge of the front. As a result, upward motion continually prevails at the front with a maximum speed of 6 m s⁻¹. The frontal updraft is tilted toward the southeast with the shear at heights greater than 3 km. Low pressure underneath the frontal updraft is hydrostatic in character since the updraft air, fed by the low-level high-θₑ environmental inflow from the southwest, is warmer and lighter than the surroundings. On the other hand, the low pressure center (~70 Pa) at $x = 23$ km and $z = 3$ km appears to be caused by the combined effects of the updraft–shear interaction and buoyancy. This interaction results in high pressure on the upshear side (northwest) of the frontal updraft with low pressure downshear (southeast) as depicted earlier.

Behind the front, high pressure (20 Pa) is observed near the surface in the area of downward motion. A horizontal pressure gradient develops across the front with high to its left and low to its right. This high pressure center also generates an upward perturbation pressure gradient at the leading edge of the front. Notice a rapid pressure decrease with height above the high pressure center. It is worth mentioning that the vertical pressure gradient displayed in Fig. 8c is less reliable than the horizontal one.

**f. Discussion**

As described in Part I, the prefrontal environment during TAMEX IOP 13 was characterized by moist air in the lower troposphere with relatively dry air in the midtroposphere. The sounding exhibited conditional instability with the lifting condensation level (LCL), level of free convection (LFC), and freezing level at 0.9, 2.1, and 5.5 km, respectively. The cold pool behind the Mei-Yu front mechanically lifted the warm, moist strong southwest monsoon flow in the PBL as illustrated in Fig. 14 in Part I. In that figure, the system-relative mean wind normal to the line ($V_s$) had a negative value of 4–5 m s⁻¹ in the PBL, showing the cross-frontal inflow from the warm side of the front. The lifted parcel became saturated near 1 km. In the absence of mixing with the environment, this parcel (after about 100 mb of lifting) became positively buoyant and ascended freely above 200 mb. Hence, organized convection was initiated and maintained by frontal lifting of an LLJ in the PBL, and the main moisture supply came from the high-θₑ monsoon air at low levels. The squall line formed in the vicinity of the Mei-Yu front when the system was located far north of the island. The observational evidence presented in this study shows that the frontal lifting did play an important role in the initiation and maintenance of the squall line in the earlier stage of the squall-line life cycle. However, the Mei-Yu front gradually lost its direct influence on the squall line as the system moved down the central west coast of Taiwan. During that period, the squall line began to move away from the front as gust fronts were formed by cells at the leading edge of the system. The convective cells generated low-level cold outflows in the warm sector to the southeast of the front. Part of these cold outflows moved toward the southeast, interacting with the strong high-θₑ southwest monsoon flow to form a gust front in the warm sector ahead of the front. As a result, new cells developed along the gust front. These new cells then merged with the old cells, thereby prolonging the lifetime of the rainband. The density-current mechanism resulting from the cold outflows of convective cells in the lower layer was likely to be responsible for the more rapid movement of the squall line than the slow move-
ment of the Mei-Yu front. This result implies that the line propagated by the growth of new cells, as well as some southeastward motion of the cells, which is consistent with reports from GATE and TAMEX IOP 2. However, the squall line in IOP 13 traveled at a speed far slower than the fast-moving squall line in IOP 2. The difference in line motion between the two cases is as much as 14 m s$^{-1}$. Why is the squall line moving so slow in general and in regard to IOP 2 in particular? Why doesn’t the squall line in IOP 13 contain extensive stratiform precipitation as do tropical squall lines during GATE and a subtropical squall line during TAMEX IOP 2? These problems will be further studied in our future research.

g. Momentum flux

Observational evidence has shown that deep convection organized into lines is responsible for mixing and redistribution of horizontal momentum. Further, the wind shear may be increased or decreased depending on the structure of the convective line. For example, see observational studies by LeMone (1983), Lin et al. (1990), LeMone and Jorgensen (1991), etc. In the case study by Lin et al. (1990), the effects of a subtropical squall line during TAMEX IOP 2 on the vertical distribution of horizontal momentum normal and parallel to the line were investigated using the dual-Doppler-derived winds. LeMone and Jorgensen (1991) studied momentum generation and redistributions in a TAMEX convective band based on the aircraft in situ measurements. These studies all show that the horizontal and vertical flux convergences and divergences of horizontal momentum by convective-scale eddies are the major contributor to momentum generation.

Figure 9 shows profiles of the vertical transports of horizontal momentum, averaged over the domain of interest (35 × 40 km$^2$), for 0653 LST 25 June. For the normal (U) component, the eddy transport term (dotted line) is positive throughout the troposphere. This implies that positive u momentum is continually transferred upward due to convective eddies in the rainband. It is consistent with the southeastward-tilted updraft depicted in Fig. 8. Like the eddy momentum flux, the total transport term (solid line) also exhibits positive values at all levels, showing that the horizontal momentum is being transported upward by organized convection at the time of analysis. Examination of the mean winds (Fig. 10) reveals that the normal component (U) has positive shear from the PBL to 7.5 km. Hence, both the total and eddy momentum transports displayed in Fig. 9a are against the vertical momentum gradient, contrary to mixing-length theory predictions. This finding is in agreement with that reported in LeMone (1983) for a line of cumulonimbus during GATE. It further shows that the parameterization of cumulus friction using a simple entrainment model for cloud momentum will fail to predict the cumulus-induced acceleration of the upper-level flow in squall lines (Wu and Yanai 1991).

The vertical fluxes of u momentum presented in Fig. 9a are quite similar to those for a subtropical squall line in TAMEX IOP 2 (Lin et al. 1990), except the sign is opposite. However, the sign will be the same (negative) if the x axis is reversed in orientation, positive toward the northwest and negative in the direction of line motion.

For the line-parallel component (Fig. 9b), the vertical flux of v momentum by convective eddies (dotted line) is downward from the PBL to 6.5 km and is upward higher up. The V profile (Fig. 10) reveals that dV/dz has positive values at heights below 4 km with negative values aloft. As a result, the eddy term (dotted line) is downgradient in the layers below 4 km and above 6.5 km. This finding is also consistent with that reported in LeMone (1983) for the line-parallel component.
Profiles of the vertical flux convergences and divergences of $u$ momentum, $-\frac{\partial (\rho_0 \bar{w})}{\partial z}$, are plotted in Fig. 11a. Our sign convention is positive (negative) for flux convergence (divergence). The total term (solid) is negative (flux divergence) at levels below 3.5 km and is positive (flux convergence) higher up. This result shows that the combined effects of the mean and eddy momentum fluxes, averaged over the rainband, contribute to a decrease in $u$ momentum at levels below 3.5 km and an increase above. Similar to the normal component, the line-parallel component, $-\frac{\partial (\rho_0 \bar{w})}{\partial z}$ (see the solid line in Fig. 11b), exhibits flux divergence below 3.5 km with flux convergence aloft.

The aforementioned result is significant since it can explain the formation of a midlevel jet (MLJ) and the demise of an LLJ at the times of organized convection described in Part I. It strongly suggests that the interaction between the mesoscale convective system (MCS) and its environment can lead to the generation and redistribution of mean horizontal momentum. For example, the vertical flux convergence of horizontal momentum in the 5-6-km layer (Fig. 11) is of the order of $10^{-3}$ N m$^{-3}$. This process would contribute to the growth of mean momentum at the rate of approximately 3.5 m s$^{-1}$ h$^{-1}$. In other words, the contribution from the vertical flux convergence of horizontal momentum alone is adequate to form an MLJ at the times of convection. Recall that the observed environmental winds at 500 mb increased from 10 to 20 m s$^{-1}$ over a period of 3 h during convection (see Fig. 10a in Part 1). A net increase of 3.4 m s$^{-1}$ h$^{-1}$ is comparable to that estimated from the vertical flux convergence term at the same height (3.5 m s$^{-1}$ h$^{-1}$) mentioned previously. This finding demonstrates that organized convection associated with the convective rainband is capable of producing an MLJ through the vertical transports of horizontal momentum by the mean and eddy motions.

At low levels the vertical flux divergence term is of the order of $-10^{-3}$ N m$^{-3}$ (Fig. 11). This term could contribute to a decrease in mean momentum at the rate of 3.5 m s$^{-1}$ h$^{-1}$. Notice that the decrease in mean momentum at low levels is offset by the increase at midlevels. It shows that the strong low-level momentum associated with an LLJ is being transported upward to develop an MLJ due to organized convection.

h. Momentum budget

The momentum budget for a subtropical squall line during TAMEX IOP 2 was reported in the study by Lin et al. (1990). The overall result is quite encouraging. It provides a possibility of investigating the dynamical interaction between the LLJ and the MCS over Taiwan and its vicinity. Furthermore, a possible feedback mechanism due to the convergence and divergence of the vertical flux of horizontal momentum to the mean flow normal or parallel to the convective band can likely be studied using the Doppler-derived winds. Thus, a momentum budget is required to determine the influence of organized convection on its environment.

According to Lin et al. (1990), the $u$ momentum equation, averaged over the horizontal domain of interest, can be written as:

\[
\frac{\partial (\rho_0 \bar{u})}{\partial t} = -\frac{\partial (\rho_0 \bar{u} \bar{v})}{\partial x} - \frac{\partial (\rho_0 \bar{u} \bar{v})}{\partial y} - \frac{\partial (\rho_0 \bar{w})}{\partial z} - \frac{\partial P}{\partial x} + \bar{j}_1, \quad (2)
\]

where $u = \bar{u} + u'$; $v = \bar{v} + v'$; $w = \bar{w} + w'$; the overbar represents the horizontal area mean; and the prime denotes the deviation from that mean. Terms A and B in (2) represent the horizontal flux convergences and divergences of $u$ momentum; term C, the vertical flux convergences and divergences of $u$ momentum; term D, the perturbation pressure gradient force; and term E, the forces other than pressure gradients.

In a similar manner, the $v$ momentum budget can be calculated using the formula similar to (2). Note that the $P'$ in (2) can be replaced by $P'_0$ whenever it appears in horizontally differentiated form.

The $u$ momentum budget is computed from (2) based on the Doppler-derived winds and retrieved pressure fields. Figure 12a plots vertical profiles of all five of the terms in (2). Note that terms A and C are dominant over the other terms in (2) in most levels. These two terms have the opposite sign at heights below 3 km. The dominance of term A at low levels is largely due to the presence of the southwest monsoon winds on the warm side of the front with the westerlies and northwesterlies behind the front. Consequently, the $u$ component has relatively small values along the south-
eastern edge (front) of the domain with larger positive values along the northwestern boundary (rear); see Fig. 5a for comparison. As a result, the horizontal flux convergence of $\vec{u}$ momentum dominates the lower troposphere. On the other hand, term B is much smaller than term A at all levels, similar to that found in TAMEX IOP 2 (Lin et al. 1990). This finding is also in agreement with that of numerical simulation studies reported in the literature. Recall that the significance of term C has already been discussed earlier (see section 5g and Fig. 11 for details).

The contribution from the pressure gradient force (term D) to the $\vec{U}$ momentum generation is quite small at low levels. It becomes larger at heights above 4 km. As in IOP 2, the pressure term is at least one order of magnitude smaller than the flux convergence terms in the lower and midlayers. It must be pointed out that we employed the relatively large size of the domain in the $x$ direction (35 km) in comparison with the convective band (5--10 km). Additional calculations (not shown) reveal that the magnitude of term D will be significantly increased, especially at low levels, if the smaller domain ($15 \times 40 \text{ km}^2$, see LeMone and Jorgensen 1991), is used instead for budget calculation. However, we prefer to use the larger domain ($35 \times 40 \text{ km}^2$) since it generates the mean variables ($\vec{U}, \vec{V}, \vec{W}$), which are more representative of the mean conditions than those obtained from the smaller domain. Note that term D has the sign opposite term A at all levels. It provides a negative contribution to the time change of $\vec{U}$ momentum.

In a similar manner, the $\vec{V}$ momentum is mainly maintained by terms A and C (Fig. 12b). The $\partial / \partial y$ term (B) is much smaller than the $\partial / \partial x$ term (A) as in Fig. 12a. Like term B, terms D and E are also small compared to terms A and C at most levels.

Figure 13 displays vertical profiles of the total advection term (AD), the pressure gradient force term (PG), and the time tendency term (TT) for both components. Term AD is reproduced from the sum of terms A, B, and C in Fig. 12. The tendency term is obtained as a residual of the momentum equation. It represents the part of air acceleration that is not balanced by the retrieved pressure field. It also contains indeterminate contributions of radar and numerical errors and actual storm evolutions. Therefore, it must be interpreted with caution. For the normal component (Fig. 13a), the total advection term is positive at all levels, contributing to an increase in $\vec{U}$ momentum. Conversely, the pressure term (PG) is negative throughout the troposphere, resulting in northward pressure gradients, especially in the layers above 4 km. Such gradients are required to balance the horizontal accelerations in the same layers. As pointed out before, the retrieval of pressure and temperature is made using the 1-km grid spacing in all three directions. This grid spacing can only resolve large-scale features and some mesoscale features with wavelength greater than 6--7 km. As a result, the pressure term in the convective rainband is considerably underestimated, especially in the lower troposphere where the convection dominates, due in part to the smoothing inherent in the objective analysis scheme that is used (see Part I). This would have a significant impact on the estimate of the tendency term (TT) at low levels. The tendency term has positive values at all levels showing an increase in $\vec{U}$ momentum with time during the convection. Once again, caution must be exercised in interpreting the tendency term because of the uncertainties mentioned previously.

The line-parallel component (Fig. 13b) reveals that the total advection is negative at levels below 2 km and above 7 km, and is positive in between. It shows that the $\vec{V}$ momentum is increased in the midlayer and is decreased in the lower and upper layers as a result of three-dimensional advections. The pressure term (PG) has negative values at heights above 4 km with positive values below. The time tendency shows an increase in $\vec{V}$ momentum in the midlayer and a decrease in the lower and upper layers. Note that the decrease in the
lowest layer is corresponding to the demise of the southwest LLJ at the times of convection described previously.

The computed tendencies presented in Fig. 13 are probably calculated on too short a time interval compared to the observed tendencies determined from a 7-min time interval between the scans. For this reason, no attempt has been made in this study to compare the computed tendency to the observed one. In the future, we plan to average the tendency terms deduced for the three successive scans and then to compare them with an independent measurement of evolution such as the wind profiles deduced from prefrontal and postfrontal soundings.

The momentum budget for the $u$ component normal to a tropical squall line was investigated by LaFore et al. (1988) using a three-dimensional cloud model. Results showed that the contributions from the three advection terms (total advection) are dominant over the other terms in the budget equation. The horizontal advection terms have the same order of magnitude as the vertical term, but the sign is opposite. These findings are in good qualitative agreement with those of a subtropical squall line during TAMEX IOP 2 (Lin et al. 1990).

The momentum budget of an oceanic MCS during TAMEX was investigated by LeMone and Jorgensen (1991) using airborne Doppler and in situ (flight level) P-3 aircraft data. Results showed that some behavior of the system is different from that of the earlier system studied by LaFore et al. (1988). The momentum flux and pressure gradient reinforce to produce eastward (front) acceleration of air exiting the band at levels above 7 km. This eastward acceleration is consistent with low pressure on the east side of the updraft due to the updraft–shear interaction (Rotunno and Klemp 1982). Such a rear-to-front acceleration was not observed by LaFore et al. (1988) because of the lack of strong $\bar{U}$ shear aloft. It was opposite in direction to that for a subtropical squall line in IOP 2 (Lin et al. 1990) and a subtropical prefrontal rainbow in IOP 13.

6. Summary and conclusions

The dynamic and thermodynamic structures of a subtropical prefrontal convective rainbow over northern Taiwan were studied using the thermodynamic retrieval method of Gal-Chen (1978). This rainbow was associated with the Mei-Yu front that occurred on 25 June 1987 during TAMEX IOP 13. A detailed wind field was derived from dual-Doppler measurements using the procedures outlined in Part I of this study. Fields of the deviation perturbation pressure and temperature were retrieved from the Doppler-derived winds at each analysis level. These fields together with a detailed wind field were then used to compute the vertical momentum fluxes and budgets for both the normal and line-parallel components.

Results show that the overall structural features of the rainbow are different from those of a subtropical squall line during TAMEX IOP 2 because of the significant differences in environmental conditions. The maintenance of this long-lived rainbow at the times of dual-Doppler analysis is caused by the combined effects of a gust front arising from the convective downdrafts ahead of the cold front and developments of new convection along the gust front. Such mechanisms are quite different from those found in a severe frontal rainbow in the midlatitudes. In the lowest layer, high pressure forms behind the front with low pressure to its southeast. A buoyancy-induced low pressure area lies beneath the frontal updraft. The cold pool behind the front is very shallow (0.5–1 km). On top of the cold pool, high-$\theta_v$ air from the west prevails in the 1–4-km layer. This warm and moist air provides additional inflow for the frontal updraft. The precipitation core associated with the frontal updraft is elongated toward the southeast with the environmental shear in the mid- and upper troposphere, forming the convective downdraft on the southeast side of the surface front. This precipitation-induced downdraft carries cooler air with it from the mid- and upper troposphere to the lower troposphere, forming a high pressure area underneath the downdraft. This mesohigh results in a negatively buoyant outflow, interacting with the high-$\theta_v$ monsoon flow in the boundary layer. As a result, new convection develops along a gust front on the southeast side of the surface front. It merges with the old convection, thereby prolonging the lifetime of this prefrontal rainbow. The vertical flux convergences and divergences of horizontal momentum by organized convection are responsible for initiating an MLJ and weakening an LLJ. The increase (or decrease) in mean momentum is largely caused by the horizontal and vertical flux convergence terms.

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