Retrieved Thermodynamic Structure of a Subtropical, Orographically Influenced, Quasi-Stationary Convective Line

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ABSTRACT
Airborne Doppler radar and flight-level measurements were used to document the three-dimensional thermodynamic structure of a convective line that developed off the southeastern coast of Taiwan on 16 June 1987 during the Taiwan Area Mesoscale Experiment. During the period of aircraft observation the convective line appeared to be quasi-stationary in a region ~90 km away from the coast. Calculations of perturbation pressure and buoyancy retrieved from the Doppler-radar-derived wind fields show that a high pressure perturbation associated with negative buoyancy existed in the region of heavy precipitation in the lowest level. In low levels, an elongated zone of low pressure was documented along the line, which was closely related to the distributions of buoyancy. These retrieved features are basically consistent with those calculated from in situ observations. In mid- to upper levels, the high (low) pressure perturbation lay around the updraft cores on the upshear (downshear) of environmental shear. The quantitative diagnosis for the perturbation pressure indicates that the positive buoyancy associated with the in-cloud latent heat release as well as evaporative cooling occurring near cloud base and subcloud regions is crucial to determine the low-level perturbation pressure distributions. However, dynamic forcing contributing to the pressure perturbation cannot be ignored in middle levels, because of updraft–shear interaction due to the presence of pronounced cross-line wind shear and corresponding updrafts reaching maximum values near midtroposphere.

In distinct contrast to squall lines, which are rapidly moving, the near-surface temperature gradient across this quasi-stationary line was very weak, and at low levels no gust front, rear inflow, or dynamically induced high pressure was evident near the leading edge of the line. Analyses presented here suggest that the persistent, stationary convergence, in association with the deceleration of low-level southeasterly flow as it approached the nearshore blocked flow, played an important role in maintaining and organizing the deep convection in this case.

1. Introduction

Deep moist convection within mesoscale convective systems (MCSs) frequently organizes into a line and it is usually referred to as a squall-line system when the leading convective line exhibits a significant propagation speed (i.e., fast-moving convective line) (Zipser 1977; Frank 1978; Martin 1975; Payne and McGarry 1977). The precipitation patterns of mature squall lines are usually characterized by a leading convective line trailed by a large region of stratiform precipitation, particularly when the lines organize perpendicular to the low-level wind shear. The general structure of midlatitude systems has been extensively examined and well documented (Houze et al. 1989). In particular, with the observations of dual-Doppler radars and application of dynamic retrieval techniques (Gal-Chen 1978), the detailed kinematic and thermodynamic structures of these systems have been revealed (Smull and Houze 1987; Chong et al. 1987; Lin et al. 1986; Roux et al. 1984; Roux 1988; Lin et al. 1990; Wang et al. 1990; Fankhauser et al. 1992; Jorgensen et al. 1997), which has led to a better understanding of the physical processes resulting in long-lived squall lines. However, there still are considerable gaps in our understanding of the physical mechanisms responsible for convection within slow-moving or quasi-stationary lines. In fact, only very limited aspects of slow-moving convective lines have been studied. Barnes and Sieckman (1984) used rawinsonde data from the Global Atmospheric Research Program’s Atlantic Tropical Experiment (GATE) to examine the environmental conditions in which fast-mov-
ing and slow-moving convective lines were embedded. Their results indicated that slow-moving lines are generally oriented parallel to the low-level vertical shear of the horizontal wind, whereas the fast-moving lines are oriented normal to the shear. Moreover, fast-moving lines appear to have a more pronounced minimum in equivalent potential temperature at ~700 mb. Zipser et al. (1981) studied a slow-moving case during GATE. They found that both the low-level wind and the vertical shear are parallel to the alignment of the line, and that the observed gust front and cross-line temperature contrasts are much weaker compared to that of squall lines. LeMone et al. (1984b) used aircraft observations to describe the composite structure of perturbation pressure and momentum fluxes within slow-moving (as well as fast moving) lines observed during GATE. Recently, Lewis et al. (1998) used airborne Doppler radar data to document a slow-moving convective band that formed in the wake of a squall line. Their analysis showed that the band is parallel to the low-level environmental shear and that the source of inflow air is found to be primarily from the squall-line stratiform region and from altitudes above 2 km.

An important but less certain distinction is that the deep moist convection within the squall line could maintain itself through propagation of the gust front (Rotunno et al. 1988), while development of the quasi-stationary convective line appears highly related to large-scale and/or orographic forcing. For example, Dudhia and Moncrieff (1987) used a three-dimensional numerical model to examine the effect of large-scale ascent on the development of quasi-stationary convective bands observed in GATE. They showed that the imposition of large-scale ascent is crucial to maintain deep moist convection within the bands. In another study utilizing a numerical cloud model, Crook and Moncrieff (1988) demonstrated that in the presence of large-scale convergence, deep moist convection could be maintained without the additional lifting produced by evaporative cooling, which is in a manner significantly different from the squall-line dynamics discussed comprehensively in Rotunno et al. (1988). Caracena et al. (1979) used surface and upper-air observations and radar data to study the Big Thompson storm occurring adjacent to the Rocky Mountains. Based on the mesoscale analyses, they suggested that the orographic uplift of the low-level inflow air was important for the release of the environmental convective instability, and the light winds at steering levels allowed the storm to remain nearly stationary over the foothills.

The objective of this study is to use aircraft measurements from the P-3 aircraft operated by the National Oceanic and Atmospheric Administration (NOAA) to document the three-dimensional thermodynamic structure of a convective line that was observed off the southeastern coast of Taiwan on 16 June 1987 during the Taiwan Area Mesoscale Experiment (TAMEX; Kuo and Chen 1990). This convective line appeared to be quasi-stationary throughout the period of NOAA P-3 observation (~4 h) (see Fig. 6 in Jorgensen et al. 1991). The event provides an excellent opportunity to investigate the physical processes contributing to the maintenance of quasi-stationary, organized deep convection occurring adjacent to a coastal barrier. The precipitation and kinematic structure of the 16 June convective line has been studied by Jorgensen et al. (1991) using airborne Doppler radar and flight-level data. Their results indicated that this subtropical convective line exhibited many circulation features different from tropical convective lines documented previously. In particular, no distinct low-level rear inflow or strong convective-scale downdrafts were seen. A preliminary, limited aspect of pressure and buoyancy features within the line has also been described in Jorgensen et al. (1991); however, with respect to the maintenance and organization of the convective line, they are far from definitive. Particularly, processes relevant to the terrain forcing and their associated contributions to the line remain largely unexplored. The main focus of the present study is to extend the work of Jorgensen et al. to include more complete, quantitative analyses of the thermodynamic structure of the convective line and to further investigate the possible roles of orography in influencing this convective system.

Section 2 provides an overview of synoptic and mesoscale environments in the region where the quasi-stationary system developed. Section 3 introduces the analysis methods applied in this study. The thermodynamic structure of the convective line retrieved from dual-Doppler winds is presented in section 4. Low-level thermodynamic characteristics derived from in situ measurements and their comparisons with retrieved results are presented in section 5. In section 6, the retrieved pressure field is diagnosed via a time-independent pressure diagnosis equation derived from the three-dimensional momentum equation. Major results of the study are summarized in section 7.

2. Synoptic and mesoscale environment

The synoptic conditions accompanying this event have been reviewed in several previous studies (e.g., Chen and Liang 1992; Jorgensen et al. 1991). A brief description follows. During the period of aircraft observation (~1400–1800 UTC), a mei-yu1 front and its associated pressure trough were located ~200 km south of Taiwan, arcing northeastward from the South China Sea to a location just south of Japan. A tropical depression lay ~300 km southwest of Taiwan over the South China Sea with circulation extending up to 700

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1 Seasonal rainfall in east Asia reaches a maximum in the transition period from the winter northeast monsoon to the summer southwest monsoon. The rainfall occurring during this period is called “mei-yu” (plum rain) (Chen 1983).
mb so that the observational region was generally characterized by moist east-southeasterly flow in the lower troposphere (Fig. 1). The low-level winds to the east of the convective line generally veered with height. At the middle troposphere a weak short-wave trough was moving over the Taiwan Strait (to the west of the observational region), and basically uniform, weak westerly flow prevailed at upper levels (Fig. 1). The flight-level measurements (not shown) reveal that the troposphere was quite moist to the east of the convective line. Both the lifting condensation level (LCL) and the level of free convection (LFC) were low with heights of 500 and 900 m above mean sea level (MSL), respectively, and the corresponding convective available potential energy (CAPE) had a moderate magnitude of $\sim 1500$ J kg$^{-1}$ (Jorgensen et al. 1991).

As the mei-yu front advanced slowly southward and passed over southern Taiwan at about 0000 UTC on 16 June 1987, a strong north-northeasterly flow started to develop at low levels along the southeastern coast of Taiwan. The winds at $\sim 310$ m measured by the NOAA P-3 (Fig. 2) revealed some important aspects of this nearshore airflow. An obvious flow transition was observed from easterly/southeasterly flow farther offshore to north-northeasterly flow near the southeastern coast. Nearshore wind speeds generally exceeded 15–17 m s$^{-1}$ but decreased southward and seaward. The sounding launched at Lutao (Green Island) (Fig. 3) indicated that this mountain-parallel (or barrier) flow had maximum winds $>17$ m s$^{-1}$ in the lowest 700 m (all heights are MSL) and decreased rapidly with height above 1 km. Stronger stratification was found within this barrier flow but was only confined to the lowest 1 km, corresponding to the mean altitude of the coastal terrain in southern Taiwan (Fig. 2). Relatively weak stability (compared to that nearshore) was found in the lowest 1 km farther offshore, where the environment was dominated by pre-

**Fig. 1.** Hodograph of composite environmental wind on the east side of the convective line. The lower part of this sounding (below $\sim 5.4$ km) is taken from aircraft descent at a location $\sim 50$ km northeast of the line near 1434 UTC 16 Jun 1987. Upper portion of the sounding is taken from a rawinsonde released from Lutao (Green Island) (location indicated in Fig. 2) at 1200 UTC.

**Fig. 2.** Flight-level winds at $\sim 310$ m MSL measured by the NOAA P-3 during 1434–1501 and 1710–1733 UTC 16 Jun 1987. Winds observed at Lutao and Lanyu Islands are also indicated. Full wind barbs correspond to 5 m s$^{-1}$, half-barbs to 2.5 m s$^{-1}$. The inset square denotes the dual-Doppler analysis domain ($54 \times 54$ km$^2$). Terrain height (m MSL) in southern Taiwan is indicated by shading (key at top).

**Fig. 3.** Vertical profiles of potential temperature (K) and wind (full and half wind barbs represent 5 and 2.5 m s$^{-1}$, respectively) measured from a rawinsonde released at Lutao (Green Island), $\sim 35$ km offshore from the southeastern coast of Taiwan, and from the P-3 flight legs just to the east of the convective line, $\sim 120$ km offshore from the southeastern coast of Taiwan.
vailing easterly/southeasterly flow (Fig. 3). Above about 1 km, there was no significant difference in stratification between nearshore and farther offshore. These observed features strongly suggest that the low-level postfrontal northeasterly flow was blocked by the topography to form a strong barrier wind along the southeastern coast of Taiwan. As noted in Yu et al. (1999) and Chen and Liang (1992), the basic wind and pressure patterns associated with the barrier flow persisted until 1200 UTC 17 June.

It should be noted that the offshore extent of the observed barrier wind is consistently predicted by the upstream influence of topography as revealed by theory. For an idealized two-dimensional barrier, Pierrehumbert and Wyman (1985) showed that the Burger number2 ($B$) is an important parameter in determining the degree of upstream blocking. When $B$ exceeds unity, the flow is deflected by the barrier and the mountain is said to be “hydrodynamically steep” (Overland and Bond 1995). Pierrehumbert and Wyman also showed that, in the non-rotating limit (i.e., when the Coriolis effects are ignored), the terrain influence on the oncoming flow can propagate upstream infinitely far. However, when considering the effect of earth rotation, the Coriolis force is found to constrain the upstream extent of flow blocking that attains a maximum on the order of the Rossby radius of deformation ($L_R = NH/f$, where $N$ is the Brunt–Väisälä frequency, $H$ is the mountain height, and $f$ is Coriolis parameter). Since the region of interest for this case was generally filled with precipitation, static stability is approximated using the saturated Brunt–Väisälä frequency ($N_s$) derived by Durran and Klemp (1982) and is equal to $5.28 \times 10^{-3}$ s$^{-1}$. If we take $\sim 1000$ m and $\sim 30$ km representative of the mountain height and half-width (averaged along the coastal barrier adjacent to the region of the aircraft observation), respectively, and the latitude of 22° gives a value of $f = 5.46 \times 10^{-3}$ s$^{-1}$, we obtain $B \sim 3.2$ and $L_R \sim 96$ km. Given the large value of $B$ (i.e., $>1$), upstream blocking by the coastal terrain appears to be significant. The estimated magnitude of $L_R$ is consistent with the observed barrier wind extending offshore to a location $\sim 90$ km from the coastal barrier (Fig. 2).

3. Methodology

Since the 16 June convective line developed over the Pacific Ocean $\sim 90$ km away from the southeastern coast of Taiwan, the measurements (including airborne Doppler radar and flight-level data) from the NOAA P-3 aircraft provided a unique opportunity to document this system. In this study, the three-dimensional flow and precipitation structure of the convective line are primarily obtained by use of pseudo-dual-Doppler syntheses derived from multiple-view radial velocity/reflectivity data through L-shaped flight pairs (each leg $\sim 5$ min in duration) as described in detail by Jorgensen et al. (1991). Doppler data processing is performed to remove contamination like sea clutter and noise and to correct folded radial velocities. The location of the analysis domain for the dual-Doppler syntheses is shown by the inset square ($54 \times 54$ km$^2$) in Fig. 2, with the lowest analysis level 0.5–14.5 km MSL in the vertical. The horizontal and vertical grid spacing was set to 1 km within the analysis domain.

Once the three-dimensional flow and precipitation fields are obtained, the horizontal momentum equation is used to retrieve the pressure field within the convective line through the retrieval technique described by Gal-Chen (1978). In this technique, the least squares method is used to yield a two-dimensional Poisson equation for the perturbation pressure field, which can be solved by the successive overrelaxation method (SOR) with the Neumann boundary condition. According to Gal-Chen, a unique solution can exist only in $p' - \langle p' \rangle$, where $p'$ denotes departures from pressure values in a undisturbed environment and the symbol, $\langle \rangle$, represents horizontal area mean. In this study, the values of the momentum checking (Gal-Chen and Hane 1981) are smaller than 0.5 at all analysis levels. While it is possible to deduce the temperature field by substituting obtained pressure deviations to the vertical momentum equation directly (as described in Gal-Chen), we utilize an alternative procedure proposed by Roux et al. (1984), which solves a horizontal Poisson equation for the perturbation temperature field. The major advantage of applying this procedure is to avoid errors resulting from varied coverage of Doppler radar observations at different analysis levels. Similarly, a unique solution in this calculation, exists only in $\theta'_c - \langle \theta'_c \rangle$, where $\theta'_c$ is the perturbation virtual cloud temperature with $\theta'_c = \theta'_v - \theta_e q_c$; $\theta'_v$ is the perturbation virtual temperature, and $q_c$ is the cloud water mixing ratio.

In the above calculations, the steady-state assumption is applied because local time derivatives in the momentum equation cannot be accurately estimated due to a large gap in time between individual three-dimensional wind fields derived via the pseudo-dual-Doppler syntheses ($\sim 10$ min on average). Fortunately, our Doppler observations revealed no notable variations in both kinematic and precipitation fields within the convective line at different analysis times. This result suggests that the contribution of local tendency should be relatively small compared to other advection terms in the momentum equation. In order to evaluate the reliability of these retrieved results, we also infer the perturbation pressure and temperature from independent flight-level measurements. These comparisons will be discussed in section 5.

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2 The Burger number is defined as $B = \frac{HN}{Nf}$, where $H$ is the mountain height, $N$ is the Brunt–Väisälä frequency, $L$ is the mountain half-width, and $f$ is the Coriolis parameter.
4. Retrieval results

In this section, the retrieved perturbation thermodynamics from two analysis periods of airborne Doppler observations (1546–1556 and 1616–1626 UTC) are presented, in which deep moist convection was highly organized and best sampled by the airborne Doppler radar. While the thermodynamic field is our focus, we also illustrate its corresponding kinematic fields, which can help interpret the retrieved results.

The 2.5 km MSL radar reflectivity and wind, vertical velocity, perturbation pressure, and perturbation temperature at 1546–1556 UTC are shown in Fig. 4. Within the analysis domain, the precipitation was oriented approximately north–south with a maximum reflectivity value of ~40 dBZ. The horizontal structure of airflow is quite simple. The low-level southeasterly flow entered the region of heavy precipitation from the east while weaker southerly flow prevailed to the west of the line. In the lower troposphere, the horizontal structure of dual-Doppler wind at different analysis levels remains similar except at the lowest level (0.5 km MSL), where the orographically blocked north-northeasterly flow was found to the west of the line as described in section 2. Generally, the regions of highest reflectivity were also characterized by positive vertical velocities with a magnitude of ~1–3 m s\(^{-1}\). The main feature of the perturbation pressure field is an elongated zone of low pressure (with a perturbation of ~0.1–0.3 mb) found along but slightly to the east of the line. This zone was characterized by several low pressure cores and was coincided with the regions of positive buoyancy (~0.3–0.8 K) and upward motion. Qualitatively, the retrieved structures are two dimensional.

The 5.5 km MSL kinematic and thermodynamic fields are shown in Fig. 5. Southerly flow prevailed at this level, with evidence of rear-to-front flow\(^3\) (i.e., component of westerly flow) near the northern portion of the line (Fig. 5a). At this level, the vertical velocity was stronger than at lower levels. Two pronounced updraft cores were observed along the line at \(X = 36, Y = 37\) km and \(X = 32, Y = 17\) km, with maximum magnitudes of 4–5 m s\(^{-1}\). The high–low pressure couplets lay around the updraft cores (Fig. 5c), and their associated pressure gradients were maximized roughly along the direction of the midlevel environmental shear vector (Fig. 1). The positive buoyancy reached a maximum (greater than 1 K) at this level and coincided with the updraft cores. The regions of positive buoyancy did not exactly coincide with those of low pressure, suggesting that the pressure distributions were not solely determined by the buoyancy effect. The amplitudes of the pressure and temperature perturbations at the upper troposphere were much weaker than those at midlevels, but their structures were quite similar (not shown).

About 30 min later, the basic airflow and precipitation structure of the convective line was similar (Fig. 6); however, some evolving aspects could be found at this time. A new precipitation core (\(X = 36, Y = 23\) km) was observed immediately to the east of the line and was characterized by moderate updrafts (+2 m s\(^{-1}\)) and positive perturbation temperatures (+0.2 K). Examination of the high temporal resolution of P-3’s lower fuselage radar during this period indicates that this precipitation core developed rapidly and moved westward, then merged into the region of main precipitation along the convective line. While the low-level thermodynamic fields became less two-dimensional at this time (which is due partly to the thermodynamic modifications in association with the formation of a convective cell to the east of the line), the perturbation pressure, temperature, and vertical velocity fields were still closely related to each other (Figs. 6b–d), similar to those observed at 1546–1556 UTC. At midlevels (Fig. 7), a main feature in the pressure field was still the high–low couplets around the updraft cores (\(X = 27, Y = 15\) km and \(X = 28, Y = 34\) km). While upward motion regions were generally associated with positive values of perturbation temperature, the regions of maximum positive perturbation temperature were not collocated with the updraft cores (Fig. 7d).

In pursuit of a more complete description of vertical thermodynamic structure, vertical sections are selected to pass through the updraft cores and to be parallel to the midtropospheric shear vector of the environment. The locations of vertical sections chosen for 1546–1556 and 1616–1626 UTC are shown in Fig. 5a (AB) and Fig. 7a (CD), respectively. The vertical structure at 1546–1556 UTC (Fig. 8) indicates that low-level southeasterly flow entered the major precipitating regions and was lifted rearward (toward the west). High pressure existed in the lowest 1.5 km near the leading edge and was associated with a deficit in temperature field (~1 K), which likely resulted from evaporation cooling due to rainwater falling into the unsaturated environment near or below the cloud base. A low perturbation pressure region extended at low levels rearward from the line, indicative of warm air aloft from detraining convective elements (LeMone et al. 1984b). The magnitude of this low is not as pronounced as seen in squall-line systems (e.g., Jorgensen et al. 1997) probably because the tilt of the leading edge (LeMone et al. 1984a) was not as pronounced. Such characteristics have also been found in tropical convective lines as documented by LeMone et al. 1984c). Their results indicate that the slow-moving (fast moving) lines exhibit a weaker (stronger) amplitude of mesolows located underneath sloping updrafts. At midlevels, the predominant feature of the pressure field is that of a high–low couplet on

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\(^3\) The terminology “front” and “rear” are somewhat ambiguous for this line since the line was not moving, so we have adapted it for this case: front refers to the east side of the line since the low-level inflow had an easterly component; similarly, rear refers to the west side of the line.
the flanks of the updraft core, with low (high) pressure on the downshear (upshear) side of the environmental shear vector. A notable positive perturbation temperature was evident within the updraft core, indicating the warming effect due to the release of latent heat. A relatively pronounced cooling was found at altitudes higher than 10 km MSL, which was probably caused by overshooting convective towers near the cloud top. Owing to the reduced horizontal coverage of dual-Doppler data at higher levels, the retrieved results above about 10 km MSL are less reliable. The vertical structure at 1616–1626 UTC (Fig. 9) shows many basic features similar to those at 1546–1556 UTC, although some of the low-level inflow was crossing the region of heavy precipi-
Fig. 5. As in Fig. 4 except at 5.5 km (MSL). In (a) thick line segments AB and EF mark locations of vertical cross sections shown in Fig. 8 and Fig. 10a, respectively. In (c) and (d) shading indicates the regions of vertical velocity greater than 3 m s\(^{-1}\). The H marks the location of maxima in the pressure perturbation field.

As described in section 2, the orographic effect might have extended offshore to a region ~90 km from the coast. Our dual-Doppler observations further demonstrate that the convective line was located near the flow transition from the low-level southeasterly flow well offshore to north-northeasterly flow near the coast. This relationship can be best seen from the vertical sections shown in Fig. 10. The locations of vertical sections chosen are indicated in Fig. 5a (EF) and Fig. 7a (GH). In order to depict the region occupied by nearshore north-northeasterly flow, those low-level winds having a northerly component are highlighted with hollow circles along the vertical section. It is clear that the barrier flow existed in the lowest level, generally to the west of heaviest precipitation within the convective line. The low-level moist southeasterly inflow entered the system
from the east, and started to be lifted westward immediately ahead of the eastern edge of the low-level barrier flow. Recall that the LFC in the inflow region had a height of $\sim 1$ km, which implies an additional lifting was still needed to allow air parcels to overcome negative buoyancy in the lowest 1 km. Airflow and precipitation features shown in Fig. 10 strongly suggest that the deceleration of warm southeasterly flow as it encountered the nearshore barrier flow provided the necessary low-level convergence for triggering and maintaining convection. We will further elaborate on this subject in section 6.

5. Low-level thermodynamic characteristics

In this section, the low-level pressure and temperature fields accompanying the convective line are further examined using flight-level data. The perturbation pressure ($p'$) is defined as deviations between the measured pressure and environmental pressure, which can be written as

$$p' = p - p_0(Z),$$  \hspace{1cm} (1)$$

where $p$ is the observed pressure, subscript $0$ denotes the environment, and $Z$ is the altitude of the aircraft. The perturbation pressure defined above can also be
inferred from the $D$ value (LeMone and Tarleton 1986) through $p' = \rho g D$, where $\rho$ is the air density and $g$ is the gravity acceleration. Similar to the perturbation pressure, the perturbation temperature is defined in terms of deviations between the measured temperature and environmental temperature as follows:

$$\theta' = (\theta_v - \theta_v^0) - \frac{L_v}{\rho} \theta_v$$

where $\theta_v$ is the virtual potential temperature, $L_v$ is the cloud liquid water content, and $\rho$ is the air density. The definition of perturbation temperature (i.e., the perturbation virtual cloud temperature, $\theta'_v$) in (2), which includes the effect of cloud liquid water, is identical to that of the retrieved temperature defined in section 3.

The perturbation pressure and temperature along a flight track as the aircraft passed through the convective line at an altitude of $\sim 3064$ m MSL from 1557 to 1602 UTC are shown in Fig. 11. This flight leg was approximately normal to the convective line and its position was close to the northern edge of the dual-Doppler synthesis domain. Owing to the Rosemount and dewpoint
sensor wetting problems on this pass that have been associated with these probes, we apply a procedure suggested by Jorgensen and LeMone (1989) to mitigate the influences of the wetting problem. The in-cloud region was basically characterized by negative perturbation pressure and positive perturbation temperature. The perturbation pressure (temperature) reached a minimum (maximum) (with $-1.4$ mb and $+3$ K, respectively) in a region having considerable liquid water and convective upward motion. The pressure and temperature fields were generally out of phase and their peak values existed slightly to the east of strongest convective updraft, which are qualitatively consistent with retrieved features described in section 4.

However, the absolute values of the perturbations derived from in situ data appear much larger than those retrieved from dual-Doppler winds. Such a result is expected because the in situ perturbation is the difference...
Fig. 11. Flight-level measurements and perturbation fields derived from the flight track during the interval 1557–1602 UTC at ~3064 m MSL. (a) Temperature (T, °C) and dewpoint temperature (Td, °C), cloud liquid water content (LWC, g m⁻³) and vertical air motion (w; m s⁻¹). (b) Perturbation pressure (mb). (c) Perturbation temperature (K). An approximate horizontal scale is indicated in (b) based on the ground speed of aircraft.

between the cloud and environment, whereas the perturbation pressure and temperature retrieved from dual-Doppler winds only represent the departure from the horizontal area mean within the analysis domain. The magnitude of the horizontal area mean of the perturbation fields can be roughly estimated by subtracting retrieval values from those deduced from in situ data. At this altitude, the horizontal area mean of the perturbation pressure and temperature are equal to ~−1.2 mb and ~+2.2 K, respectively.

A quantitative evaluation of the retrieved results can be made through comparisons of the horizontal gradient of the perturbation fields. The in situ derived perturbation fields shown in Fig. 11 exhibited the pattern of a single wave, with a half-wave length of ~20 km. This gives a horizontal gradient (i.e., in the west–east direction) for perturbation pressure and temperature equal to ~0.05 mb km⁻¹ and ~0.38 K km⁻¹, respectively. The mean west–east perturbation pressure and temperature gradient estimated from Figs. 4c,d are equal to ~0.06 mb km⁻¹ and ~0.63 K km⁻¹, respectively. It appears that the horizontal gradient of perturbation pressure calculated from both datasets is in good agreement, while it is noted that a larger gradient found in the retrieved temperature may imply relatively less accuracy for the retrieval of the temperature field. However, we cannot rule out temperature sensor wetting as a potential error that would reduce the in situ calculated gradient.

The flight track from 1722 to 1727 UTC as the aircraft crossed the line at an altitude of ~308 m MSL (close to the cloud base) is used to assess near-surface pressure and temperature features. This flight leg was located near the northern portion of the dual-Doppler synthesis domain, with an orientation approximately normal to the line. In order to examine the actual temperature contrast across the convective line, the potential temperature [instead of perturbation temperature defined in (2)] is calculated as shown in Fig. 12. An obvious deceleration of easterly flow component (from ~11 to ~3 m s⁻¹) accompanying higher liquid water content was found in the eastern portion of the line. The precipitating region was characterized by relatively cool temperature and weak downward motion (less than ~−1 m s⁻¹). The weak temperature deficit (~−0.5 K) within the line suggests that the subcloud evaporative cooling is not significant, which is probably due to a moist environment at low levels. Note that the along-track pressure and temperature fields were generally out of phase, similar to low-level retrieved structures shown in Figs. 4c,d. In addition, the high pressure perturbation (~−0.2 mb) observed in the region of precipitation is consistent with that retrieved from dual-Doppler winds shown in Figs. 8 and 9, which depict a shallow high confined to the lowest 1.5 km MSL in a region of cooler air within the line.

6. Diagnosis of the perturbation pressure

Three major features of the pressure field have been documented within the line, including the low-level low
pressure zone, the midlevel high–low pressure couplets around the updraft cores, and the lowest-level high pressure. In this section the pressure diagnosis equation is used to examine physical processes responsible for these distributions of pressure perturbation. Ignoring Coriolis force and friction, the momentum equation can be written as

$$\frac{dv}{dt} = -\frac{1}{\rho} \nabla p' - \frac{C_p}{C_p} \rho g \hat{k} + \frac{\theta_c'}{\theta_o} g \hat{k} - q_r g \hat{k},$$  \hspace{1cm} (3)

where $v$ is the vector of air parcel motion, $\rho$ is the air density, $p'$ is the perturbation pressure, $\rho_o$ is the reference pressure, $g$ is the gravity acceleration, and $q_r$ is the rainwater mixing ratio. The term on the left-hand side of (3) is the acceleration following an air parcel and the four terms on the right-hand side represent the pressure gradient force, pressure buoyancy, thermal buoyancy, and rainwater drag force, respectively. Combination of pressure and thermal buoyancy terms gives a definition for the buoyancy term $-\rho g p'/\rho_o$, where $\rho'_o = \rho' + L_s$, which represents an air parcel accelerating upward due to horizontal variation in air density. Taking the differential operator $\nabla \cdot$ on (3) and applying the anelastic approximation ($\nabla \cdot \rho_o v = 0$), a time-independent pressure diagnosis equation can be obtained:

$$\nabla^2 p' = -\rho_o \left[ \frac{\partial^2 u}{\partial x^2} + \left( \frac{\partial u}{\partial y} \right)^2 + \left( \frac{\partial u}{\partial z} \right)^2 + 2 \frac{\partial v}{\partial x} \frac{\partial u}{\partial y} + 2 \frac{\partial w}{\partial x} \frac{\partial u}{\partial z} + 2 \frac{\partial w}{\partial y} \frac{\partial u}{\partial z} - w^2 \frac{d^2 \ln \rho_o}{dz^2} \right]$$

$$+ \left[ \frac{\partial}{\partial z} \left( -\rho_o \frac{C_p}{C_p} p' g + \rho_o \frac{\theta_c'}{\theta_o} g \right) + \frac{\partial}{\partial z} (-\rho_o q_r g) \right],$$  \hspace{1cm} (4)

where $\rho_o$ is the reference density, $C_p$ is the specific heat at constant pressure, $\theta_o$ is the reference potential temperature, and $L_s$ is the saturation latent heat of vaporization.
It is clear from (4) that the physical processes determining distributions of perturbation pressure include dynamic forcing terms (A, B1–B3, C) and buoyancy forcing terms (D, E). The term A is proportional to the Laplace of the kinetic energy and can be considered the spatial distribution of fluid kinetic energy. Physically, it represents the Bernoulli effect (Yau 1979). Mathematically, this term always contributes positively to the perturbation pressure. The term B represents the contributions of fluid shear. For a situation having pure rotation, it is proportional to the square of vorticity (Rotunno and Klemp 1982). The term C represents the contribution of the density variation in the vertical. In general, its magnitude is very small in the actual atmosphere and is thus ignored. The terms D and E represent the contributions of the vertical variations of the thermal buoyancy and rainwater drag. If a wavelike structure for the perturbation pressure is assumed, (4) can be expressed in a simplified form as

\[-p' \propto \nabla^2 p' = -D_y + \frac{\partial B}{\partial z}, \tag{5}\]

where \(D_y\) denotes the dynamic forcing terms and \(\frac{\partial B}{\partial z}\) denotes the buoyancy forcing terms. Since the retrieved fields are a deviation from their horizontal area mean, (5) cannot be applied directly to the diagnosis of retrieved fields. Hence, we take a horizontal average on (5) and subtract it from (5), yielding

\[p^* \propto D^*_y - \frac{\partial B^*}{\partial z}, \tag{6}\]

where \(p^* = p' - \langle p' \rangle\), \(D^*_y = D_y - \langle D_y \rangle\), and \(B^* = B - \langle B \rangle\). In (6), the retrieved perturbation pressure is proportional to the horizontal deviations of the dynamic forcing term and buoyancy forcing term. The major advantage of employing (6) instead of (5) for diagnosis is that the uncertainty due to the unknown value of horizontal area mean is removed and all forcing terms in (6) can be entirely determined by the dual-Doppler winds and retrieved pressure and temperature. The rainwater mixing ratio in (4) is estimated from precipitation reflectivity by using the \(q_r-Z\) empirical formula (Hauser and Amayenc 1986). It is important to note that the only assumption we made from (5) to (6) is that the horizontal area mean of the vertical differential of buoyancy is equal to the vertical differential of the horizontal area mean of buoyancy (i.e., \(\frac{\partial B}{\partial z} = \frac{\partial (B)}{\partial z}\)). Strictly speaking, this relationship may not remain valid when the data coverage of Doppler observations is varied with height. Fortunately, the data coverage of our Doppler observations between two adjacent analysis levels is similar except at the lowest analysis level, and hence the errors should not be significant for analysis levels above 1.5 km MSL.

Because the diagnostic results from two analysis periods of Doppler observations are similar, we only present the results from 1546–1556 UTC. The horizontal distribution of the total forcing term, dynamic forcing term, and buoyancy forcing term at 2.5 km are shown in Fig. 13. Basically, the patterns of total forcing (Fig. 13a) are in good agreement with the horizontal distribution of perturbation pressure (Fig. 4c). The negative regions of total forcing coincide with the low pressure centers. It is apparent that the buoyancy forcing is a primary contributor to the formation of the low-level low pressure zone (Fig. 13c), although the dynamic forcing term also contributes positively to the negative perturbation pressure (Fig. 13b). Examination of the horizontal plots for individual dynamic forcing terms and buoyancy forcing terms as expressed in (4) indicates that the vertical variation of the thermal buoyancy (term D) is the primary driver for the buoyancy forcing, and in contrast, the contribution of vertical variation of rainwater drag (term E) is quite weak. Recall that positive buoyancy associated with release of latent heat generally increased with altitude and reached a maximum near midtroposphere (Fig. 8); hence the vertical variation of thermal buoyancy is an important contributor at low levels. The dominating effect for dynamic forcing is the interaction between the updraft and cross-line wind shear (term B2), but with a relatively weak contribution to the low pressure zone. The remaining dynamic forcing terms indicate a near-zero contribution (not shown).

Figure 14 illustrates the horizontal distribution of the total forcing term, the dynamic forcing term, and the buoyancy forcing term at 5.5-km altitude. In general, at this level the buoyancy forcing also plays a predominant role in determining distributions of the pressure field over most of the analysis area. However, in contrast to the low levels, in which the buoyancy forcing was dominated by negative values (Fig. 13c), the buoyancy forcing at this level exhibited a considerable positive contribution along the line. An obvious, negative dynamic forcing is found near the location of the existing low pressure on the downshear side of the updraft core (\(X = 34, Y = 17\) km in Fig. 14b) and its associated magnitude is comparable to that of the buoyancy forcing (\(-9 \times 10^{-6} \text{ s}^{-2}\)). The plots of individual dynamic forcing terms and buoyancy forcing terms (not shown) reveal that the dynamic forcing originated from the interaction between cross-line wind shear and updraft [term B2 in (4)]. Thermal buoyancy forcing [term D in (4)] is also a contributor to this midlevel downshear low. The environmental shear exhibited pronounced cross-line shear (cf. Fig. 1) and in-cloud vertical velocities also reached a maximum (cf. Fig. 8) at the midlevel, both of which may explain the increased importance of the dynamic forcing on the distribution of the pressure perturbation.

It is noteworthy that the pressure pattern (i.e., high–low couplet) induced by the dynamical effect of the updraft–shear interaction has been previously observed in cumulus/cumulonimbus clouds and squall lines (e.g., LeMone et al. 1988; Lin et al. 1986). By utilizing a numerical cloud model, Rotunno and Klemp (1982)
demonstrated that the vertical pressure gradient force associated with the downshear low can explain how a veering environmental wind shear can cause an initially symmetric updraft to grow preferentially to the right of the shear vector. In this view, it is possible that the existence of the midlevel downshear low could favor the acceleration of low-level inflow air to higher levels. However, given that the dynamic effect of updraft–shear interaction was notable only at midlevels, such a process should be less important for the lifting of boundary layer air.

Both retrieval results and in situ observations show that a high pressure perturbation existed near and underneath the cloud base within the line. As mentioned earlier, owing to a relatively limited coverage of Doppler observations in the lowest analysis level (0.5 km MSL), (5) cannot be used for diagnosing the generation of this pressure perturbation. Owing to the lack of rotational flow and strong vertical velocities at low levels, the contribution of fluid shear [terms B1–B3 in (4)] to the pressure field is expected to be minor. The forcing of rainwater drag (term E) should not be pronounced at low levels as the foregoing. Within the region occupied by the high pressure perturbation, an obvious deceleration of warm, southeasterly inflow was observed as it entered the precipitation region from the east of the line (cf. Fig. 12).
This reveals that the Bernoulli effect (i.e., term A) might be a contributor. Given the two-dimensional characteristic of the line, the magnitude of horizontal pressure gradient (in the cross-line direction) caused by the Bernoulli effect can be approximated by (4) via

$$\frac{\Delta p'}{\Delta x} = -\rho \left( \frac{\partial u}{\partial x} \right)^2.$$  (7)

where $\Delta x$ is the horizontal distance valid for the estimate of the pressure gradient. The horizontal gradient of the $u$ component of the wind in Fig. 12 near the eastern edge of the line is equal to about $10^{-3}$ s$^{-1}$, which gives a west–east pressure gradient of $\sim 10^{-3}$ Pa km$^{-1}$. However, this Bernoulli-induced pressure gradient is far smaller than the in situ pressure gradient calculated from Fig. 12 ($\sim 4.4$ Pa km$^{-1}$). This implies that the Bernoulli effect is not the dominant forcing. As illustrated earlier, the thermal buoyancy (term D) is an important contributor to the distribution of pressure perturbation within the line. It also may have been important to the lowest-level high pressure as well, although its significance is difficult to validate quantitatively through (4). An alternative approach to quantifying the importance of thermal buoyancy to the lowest-level pressure field is taking a differential (in the west–east direction) of the hydrostatic equation,$^4$ approximating the pressure gradient in the cross-line direction as

$$\frac{\Delta p'}{\Delta x} = -\rho \left( \frac{\partial u}{\partial x} \right)^2.$$  (7)

$^4$Given general near-zero vertical velocities in the lowest 1 km along the line indicated by both flight-level data (cf. Fig. 12) and dual-Doppler synthesis results (cf. Fig. 8), the hydrostatic approximation is assumed near ground level.
where $\delta z$ is the depth of relatively colder air associated with the high pressure perturbation and is assumed to be $\sim 1$ km (Fig. 8). The temperature decreased by $\sim 0.5$ K within 5 km across the eastern edge of the line (Fig. 12). Substituting these values into (8) yields a value of cross-line pressure gradient of $\sim 3.8$ Pa km$^{-1}$, which is comparable to the magnitude of the in situ observed pressure gradient. Hence, it is suggested that the thermal buoyancy would also play a dominant role in the generation of the lowest-level pressure perturbation.

It is noteworthy that the lack of appreciable dynamic forcing at low levels is in contrast to many previously documented squall lines. High pressure perturbations are frequently found in their leading edge, caused by the dynamic effect of cold low-level frontward flow behind the line colliding with warm inflow air in their front (Roux et al. 1984; Roux 1988). The vertical acceleration caused by the nonhydrostatic pressure gradient associated with the dynamic high has been shown to be important for the initiation/maintenance of intense convective updrafts along the leading edge of squall lines (e.g., Roux 1988; Weisman et al. 1988). For the case studied herein, given such weak dynamic forcing rooted in the heavy precipitation region, the orographically induced low-level convergence in association with the deceleration of warm southeasterly flow as it encountered the nearshore barrier flow may have been closely linked to the maintenance of deep convection. Further support for this is that the position of the observed convective line remained quasi-stationary and collocated with the flow boundary between nearshore nearly steady barrier flow and warm southeasterly flow well offshore, as described in section 4. Moreover, the fact that the observed line organized in a direction approximately parallel to the coastal barrier is also highly suggestive of a causal relationship between the orographic forcing and maintenance of the convective line.

7. Conclusions

Using airborne Doppler radar and flight-level measurements, the detailed thermodynamic structure of a convective line that developed off the southeastern coast of Taiwan on 16 June 1987 during the TAMEX experiment was examined. The convective line appeared to be quasi-stationary during the period of airborne observations ($\sim 4$ h). The major features of the in-cloud perturbation pressure distribution within the line and accompanying the airflow in the region where it formed are summarized schematically in Fig. 15. Analyses of retrieval results and in situ data show that a shallow high pressure (H1) was found below 1.5 km, an elongated zone of low pressure (L1) was organized along the line at low to midlevels, and the high–low couplets (H2, L2) lay around the updraft cores at mid to high levels. Quantitative diagnosis of the pressure perturbation indicates that at low levels the thermal buoyancy (i.e., subcloud evaporation cooling and in-cloud latent heat release) is a dominant factor determining the pressure distribution. However, in the midtroposphere, in addition to the contribution of thermal buoyancy, the dynamic impact on the generation of pressure perturbation cannot be ignored in the vicinity of principal updraft regions. This is a result of an enhanced dynamic effect from the interaction of updrafts with pronounced environmental shear.

In distinct contrast to squall lines, the near-surface temperature gradient across this quasi-stationary line was very weak, and at low levels no gust front, rear inflow, or dynamically induced high pressure were evident. Given such weak dynamic forcing rooted in the convective region, it is suggested that the orographic forcing may have played an important role in maintaining deep convection for this event. In particular, a stably stratified north-northeasterly flow was observed adjacent to the steep terrain along the southeastern coast of Taiwan (i.e., to the west of the convective line), with an offshore extent of $\sim 90$ km consistent with the upstream influence of orography predicted by theory (i.e., Rossby radius of deformation, $\sim L_R$). Dual-Doppler observations further show that the convective line developed and remained quasi-stationary near the flow boundary between nearshore orographically blocked north-northeasterly flow and southeasterly flow farther offshore (Fig. 15). The existence of the persistent, stationary convergence in association with the deceleration of low-level southeasterly flow as it encountered the nearshore nearly steady barrier flow may explain the quasi-stationary nature of the convective line, its ob-
served location relative to the landmass of Taiwan, as well as its organized alignment approximately parallel to the coastal barrier. The ability of barrier jets to influence atmospheric moist convection occurring adjacent to steep coastal terrain has received increased attention of late over different geographical locations such as that found along the western coast of the United States (Bond et al. 1997) and the northwestern coast of Taiwan (Li et al. 1997; Li and Chen 1998). The study presented contributes to this issue.

Although our analysis suggests the importance of the terrain-induced flow boundary/convergence on the maintenance of the observed quasi-stationary line, we are unable to definitively address the extent to which it relates to the forcing induced by the convective processes. The presence of the convectively generated high pressure perturbation (i.e., H1) would probably provide additional convergence at low levels as the warm, moist inflow approaches the leading edge of the line and decelerates due to the opposite horizontal pressure gradient, which in turn favors the subsequent development of new convection ahead of the line. Moreover, as described in section 6, the midlevel low (i.e., L2) partially produced by the dynamic effect implies the existence of an upward nonhydrostatic pressure gradient on the inflow side, which is favorable for the acceleration of low-level air to higher levels. The impact of these convectively generated pressure perturbations on influencing this quasi-stationary event is difficult to quantify owing to a relatively coarse temporal resolution and limited data coverage of the available observations. The specific contribution of these factors to convection maintenance deserves future examination using a cloud/mesoscale numerical model. Moreover, future detailed observations of slow-moving and quasi-stationary lines occurring in a variety of environments will be required for a more complete understanding of their dynamical processes which appear to be different from that of moving squall lines.

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