The Role of Offshore Convergence on Coastal Rainfall during TAMEX IOP 3

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ABSTRACT

The role of offshore convergence on the coastal rainfall maximum along the northwestern coast of Taiwan is analyzed based on TAMEX (Taiwan Area Mesoscale Experiment) data and numerical experiments using the fifth-generation Pennsylvania State University–National Center for Atmospheric Research (PSU–NCAR) Mesoscale Model (MM5).

From a case study during TAMEX IOP 3, moderate rainfalls (~40 mm h⁻¹) are observed along the northwestern coast of Taiwan associated with the arrival of three rainbands in succession during 1000–1600 LST. These rainbands form over the Taiwan Strait with northeast–southwest or north–south orientation. They intensify off the northwestern coast. There is a tendency for the most intense echoes to align in a northeast–southwest orientation off the northwestern coast. An orographically enhanced convergence zone with a northeast–southwest orientation occurs in the area where the deflected southerly flow converges with the prevailing southwesterly flow that is modified by the storm-induced westerlies immediately behind the convective line.

From the numerical experiment with synoptic-scale forcing but without the island topography over Taiwan, a large-scale cloud band is simulated within the Taiwan Strait without the localized rainfall maximum along the northwestern coast. The coastal rainfall maximum is simulated when the influences of the orographic effects are included. The results from the radar analyses and numerical simulations indicate that the convergence off the northwestern coast is a combination of the synoptic-scale forcing, orographic effects, and the feedback of the convection. The offshore convergence is important for the production of the rainfall maximum along the northwestern coast of Taiwan.

1. Introduction

Flash floods are major meteorological disasters that occur around the globe. They pose a challenging problem for both scientific research and operational forecasts. A quasi-stationary belt of monsoon rainfall occurs over East Asia in the early summer. The onset of monsoon rainfall in the northwestern part of the South China Sea is in early May and advances to the southern China coast in the middle of May (Tao and Chen 1987). It migrates to the Yangtze River valley over central China in mid-June. In most cases, heavy rain episodes are associated with the passage of a Mei-yu front. In the past, the analysis of monsoon circulations has focused on the peak phase of the east Asian summer monsoon during June–July over the Yangtze River valley (Tao and Chen 1987). Over Taiwan, the early-summer rainy season is typically from 15 May to 15 June (Kuo and Chen 1990), one month earlier than the peak phase. The early-summer rainy season over Taiwan is an ideal situation for studying heavy precipitation events because of its steep topography (Fig. 1), abundant moisture, and the frequent passage of surface frontal systems. Taiwan Area Mesoscale Experiment (TAMEX) was jointly conducted by Taiwan and the United States during May–June 1987. Its goal was to better understand the causes of severe floods. TAMEX was the first field experiment to focus on heavy rain problems in a distinct flow regime: subtropical flow and under orographic influences. In this study, we will study orographic effects on localized heavy rainfall that occurred near the northwestern coast of Taiwan during TAMEX IOP 3 (22 May 1987).

In regions with a complex terrain, large-scale circulations regulate whether or not widespread precipitation will occur, whereas local circulations and orographic effects govern the details of precisely where and when the convection will occur. The pressure and wind patterns that occur during a steady flow over a...
mountain have received considerable attention in the past (Smith 1979; Blumen 1990 and others). The mountains also disrupt the basic airflow to force ascending currents and initiate clouds (Banta 1990). The kinds of lifting that an isolated mountainous island can provide are direct, such as mechanically forced ascent as the air moves over the slope (Giambelluca et al. 1986; Chen et al. 1991; Chen and Feng 2001), or more indirect, such as when the airflow is blocked by the island obstacle (Smolarkiewicz et al. 1988; Rasmussen et al. 1989; Akaeda et al. 1995) or when mountains act as a high-level heat source (sink) during the day (night) and produce land–sea (mountain–valley) breeze (Leopold 1949; Garrett 1980; Chen and Nash 1994; Yeh and Chen 1998).

In the past, synoptic conditions favorable for the development of heavy rainfall over the Taiwan area were studied extensively (Kuo and Chen 1990; Chen and Yu 1988; Chen and Li 1995b and others). In this study, we will focus our efforts on the interaction between the island-induced circulations and the environment and on the role of this interaction for the development of localized heavy rainfall when synoptic conditions are favorable.

Prior to TAMEX, it was well known that a shallow mesoscale low pressure center often forms to the southeast of Taiwan in the prefrontal southwest monsoon flow with a windward ridge/leeside trough pressure

FIG. 1. Daily rainfall distribution 22 May 1987. Rainfall contours (heavy solid) are 10-mm intervals starting from 10 mm. Terrain contours (dashed) are 1500-m intervals starting from 1500 m.

FIG. 2. Winds (m s$^{-1}$) at 1-km level superimposed with CP4 radar reflectivity. The full barb and half barb represent 5 and 2.5 m s$^{-1}$. The interval of terrain contour is 1500 m. Locations of conventional radar at Kaoshiung (KS), and CP4 and TOGA Doppler radars.

FIG. 3. Synoptic-scale chart for 0000 UTC (0800 LST) 22 May 1987. Here, (a) 300 hPa: geopotential heights (solid) every 120 m, isotherms (dashed) every 4°C; (b) 850 hPa: geopotential heights (solid) every 30 m, equivalent potential temperature (dashed) every 10°C. Winds (m s$^{-1}$) with one pennant, full barb, and half barb represent 25, 5, and 2.5 m s$^{-1}$. (After Chen and Li 1995b.)
Fig. 4. GMS (geostationary meteorological satellite) infrared images 22 May 1987: (a) 0700, (b) 1000, (c) 1100, and (d) 1300 LST.

pattern (Wang 1986). TAMEX studies (Chen et al. 1989; Trier et al. 1990; Chen and Hui 1990, 1992) confirmed that the windward ridge/leeside trough pressure pattern is typical under the prefrontal southwesterly flow regime. This pressure pattern was also reproduced in numerical results (Sun et al. 1991). Taking the horizontal dimension of the central mountain range as approximately 100 km and $U \approx 10$ m s$^{-1}$, the Rossby number for the prefrontal southwesterly flow was approximately 1. Smith (1982) suggested that the windward ridge/leeside trough and upstream turning of the winds down the large-scale pressure gradient are features inherent to steady inviscid, initially barotropic flow if the Rossby number is on the order of 1 or greater. The windward ridge is caused by lifting and positive density anomalies aloft as the stable, low-level air is pushed up the windward slope (Smith 1982). The relatively higher pressure next to the slope decelerates the flow coming up the slope. The wind aloft traverses the mountain barrier and descends on the lee side, resulting in a leeside trough. Chen and Li (1995a) studied the evolution of a sea level pressure pattern and the surface airflow related to the changes in the wind speed and direction. As the synoptic-scale trough/low-level jet approached, the wind aloft turned SSW to WSW and strengthened. The windward ridge/leeside trough pressure pattern was most significant when the maximum axis of the southwesterly flow (or low-level jet) was over Taiwan.

Based on the scale analysis, Overland and Bond (1993, 1995) showed that for orographic blocking and strong low-level wind along the coast or channel, the alongshore momentum balance is determined by the Burger number ($B = (h/L)(N/f)$) or dynamically scaled mountain slope, where $h$ is the mountain height, $L$ is the mountain half-width, $N$ is the static stability, and $f$ is the Coriolis parameter. For the island of Taiwan, $h \approx 2$ km, $L \approx 50$ km, $N \approx 10^{-2}$ s$^{-1}$, $f \approx 10^{-4}$ s$^{-1}$, $B$ is greater than 1. Under this situation, a mountain Froude number, $Fr = U/hN$, where $U$ is the onshore component of wind, can be used to characterize the appropriate hydrodynamic regime for scaling the length and magnitude of response for the region of blocked flow (Overland and Bond 1995). Under the prevailing southwest monsoon flow over the Taiwan area, $U \approx 10$ m s$^{-1}$, $h \approx 2$ km, $N \approx 10^{-2}$ s$^{-1}$, Fr is about 0.5. From the scale analysis, for $Fr < 1$, the initial disturbance will grow seaward to a limit given by a Rossby radius $L_R = U/f$.
and the enhancement in the alongshore wind component \((V)\) is on the order of \(U\) (Overland and Bond 1995). For a small Froude number (<0.5) flow regime, most of the flow goes around the island (Drazin 1961; Smolarkiewicz et al. 1988; Sun et al. 1991). For the southwesterly flow region with a small Fr (0.2–0.5), the surface airflow moved around the island over the southwestern coast of Taiwan with acceleration of the airflow around the southern tip of the island (Trier et al. 1990; Li et al. 1997). The northern branch of the splitting airflow also accelerated downstream (Chen and Hui 1992). Chen and Li (1995b) analyzed the airflow within the Taiwan Strait and on the windward side using aircraft flight level data, rawinsonde, and pibal soundings during IOP 3. They found that below the 2000-m level, winds were weak over southwestern Taiwan as well as upstream off the coast because of flow deceleration. In the lowest levels, winds moved around the topography. Along the western coast, the northern branch of the splitting airflow had a cross-contour wind component down the windward ridge and accelerated downstream resulting in a barrier jet along the northwestern coast. Based on limited aircraft data, Li and Chen (1998) showed that the force balance along the coastal jet over northwestern Taiwan was dominated by the inertial advection term and pressure gradient force term. The barrier jet had a local wind speed maximum (∼10 m s⁻¹) between 0.5 and 1.5 km. It reached its maximum intensity when the windward ridge–leeside trough pressure pattern was most significant.

Li et al. (1997) showed that during IOP 13 (24–25 June 1987), a long-lived, convective rainband formed in a prefrontal localized convergence zone between an orographically induced barrier jet and the westerly flow behind the trough off the northwestern Taiwan coast. It occurred in the region beneath the upper-level divergence and along the axis of the warm, moist tongue.
TABLE 1. Thermodynamic variables and vertical wind shear between 0.5 and 1.5 km from the Makung soundings (Station 46734) shown in Fig. 6.

<table>
<thead>
<tr>
<th>Variable</th>
<th>0000 UTC 22 May</th>
<th>0600 UTC 22 May</th>
</tr>
</thead>
<tbody>
<tr>
<td>LCL</td>
<td>970 mb</td>
<td>960 mb</td>
</tr>
<tr>
<td>LFC</td>
<td>885 mb</td>
<td>920 mb</td>
</tr>
<tr>
<td>EL</td>
<td>300 mb</td>
<td>320 mb</td>
</tr>
<tr>
<td>CAPE</td>
<td>711 m² s⁻²</td>
<td>688 m² s⁻²</td>
</tr>
<tr>
<td>Vertical wind shear</td>
<td>$7 \times 10^{-3} \text{ s}^{-1}$</td>
<td>$9 \times 10^{-3} \text{ s}^{-1}$</td>
</tr>
</tbody>
</table>

During IOP 3 (22 May 1987), heavy rainfalls occurred along the southeastern China coast. The 12-h rainfall accumulation for 0000–1200 UTC (0800–2000 LST) 22 May exceeded 150 mm (Chen and Li 1995b). During the same period, the maximum rainfall occurred along the northwestern coast of Taiwan. As will be shown later, the rainfall there was produced by the arrival of rainbands in succession ahead of the 850-hPa trough.

The infrared (IR) satellite imagery (Fig. 4) showed that the precipitation systems developed and persisted along the southeastern China coast ahead of the troughs during 22 May 1987. Within the Taiwan Strait, the convection developed around 0700 LST (Fig. 4a), intensified off the northwestern coast of Taiwan, and moved onshore (Figs. 4b and 4c). During daytime hours, the convective clouds covered most of the northwestern coast. Most of the daily rainfall along the northwestern coast was produced during 1000–1600 LST (Fig. 5). The Makung (23.55°N, 119.62°E) soundings at 0800 LST (Fig. 6a) and 1400 LST (Fig. 6b) were characterized by a weakly unstable environment (Table 1) with low convective available potential energy (CAPE) ~ 711 m² s⁻² and 688 m² s⁻², respectively. The CAPE is calculated by vertically integrating the positive area on the skew $T$–
log $p$ diagram between the temperature profile and the moist-adiabatic curve from the level of free convection (LFC) to the equilibrium level (EL) for a surface air parcel. The vertical wind shear in low levels between 0.5 and 1.5 km was moderately strong, $\sim 7 \times 10^{-3}$ s$^{-1}$ and $9 \times 10^{-3}$ s$^{-1}$ at 0800 and 1400 LST, respectively. The low-level shear could be an important contributor, along with the offshore convergence,
Fig. 9. The tracks of long-lived (>1 h) reflectivity maxima: (a) B1, B2, B3 (rainband 2) during 0750–1212 LST; (b) C1, C2, C3 (rainband 3) during 1235–1556 LST. The numbers indicate the time at the beginning and at the end of the reflectivity maxima with reflectivity ≥30 dBZ.

in influencing the linear organization of deep convection (Byers and Braham 1949; Rotunno et al. 1988).

3. Data and analyses

The data from 808 daily rainfall stations (Yeh and Chen 1998) were used to construct the daily rainfall distribution over Taiwan during IOP 3. There were 85 surface stations with hourly rainfall observations and 125 IOP rain gauge stations with observations every half hour (Kuo and Chen 1990). Data from these stations were used to construct hourly rainfall maps.

The radar echoes collected by the Kaoshiung conventional radar (KS; southwest Taiwan) (Fig. 2) during 0507–1546 LST on 22 May 1987 were used to describe the evolution of convective activities for this coastal rainfall event. The coverage of the radar over the ocean is quite good, but the steep topography of Taiwan precludes accurate measurements of radar reflectivity over the island due to ground clutter and blockage (Parsons and Trier 1990). The maximum range of this radar is 250 km. The evolution of the entire coastal rainfall event is well monitored by the Kaoshiung conventional radar. The CP4 radar data are used to track reflectivity maxima associated with the rainbands off the northwestern coast. Both the CP4 and TOGA radar have relatively limited range (~100 km). In addition, with a 5-cm wavelength, the observed radar reflectivities from both radars were also affected by rainfall attenuation.

The mesoscale airflow off the northwestern coast is studied by performing the dual-Doppler radar (CP4 and TOGA) analyses from radial winds collected during 1114:50–1119:09 LST and 1420:05–1425:15 LST. During these periods, the convection is very active with a well-defined northeast–southwest-oriented convective line off the northwestern coast. The procedures in reconstructing three-dimensional wind fields from two Doppler radars included 1) data editing, 2) the interpolation of observational data on the spherical coordinate to grid points, and 3) synthesis of two Doppler winds on Cartesian coordinates with the help of the anelastic mass-continuity equation and a reflectivity–terminal velocity relation (Atlas et al. 1973).

The ambiguous Doppler velocities are unfolded using the RDSS (Research Data Support System) software package (Oye and Carbone 1981) developed by the Field Observing Facilities (FOF) at the National Center for Atmospheric Research (NCAR). Ground-clutter-contaminated data are discarded. The interpolation and synthesis are performed by using NCAR’s CEDRIC (custom editing and display of reduced information in Cartesian space) software package (Mohr et al. 1986) under the steady-state assumption during the sampling period (~5 min). Vertical velocities are calculated by upward integration of the anelastic mass-continuity equation from the lower to upper boundaries where vertical velocities are constrained to vanish by the O’Brien (1970) method.

The derived wind fields from dual-Doppler radar analysis contain numerous sources of errors. With limited collection times, there are temporal errors associated with the advection and the evolution of the storm. The errors caused by storm evolution during the data collection time are irrecoverable. The advection errors are corrected by introducing storm motion into the dual-Doppler analysis. The statistical uncertainties are usu-
Fig. 10. Horizontal distributions of (a) winds in m s\(^{-1}\) and the radar reflectivity (dBZ) (shaded), (b) divergence (10\(^{-3}\) s\(^{-1}\)), (c) vertical motion (m s\(^{-1}\)), (d) vorticity (10\(^{-3}\) s\(^{-1}\)) at 1 km during 1114:50±1119:09 LST. In (a), the shading interval for radar reflectivity is 10 dBZ and the solid line is the northwestern coastline of Taiwan. The contour interval for convergence (dashed) and divergence (solid) is 1.0 \times 10^{-3} \text{s}^{-1}. The contour interval of the vertical motion and vorticity is 0.5 m s\(^{-1}\) and 1.0 \times 10^{-3} \text{s}^{-1}, respectively.

ally less than 2 m s\(^{-1}\) for the horizontal winds and a few meters per second for the vertical velocities related to height-dependent mass density, station separation (Doviak et al. 1976; Wilson et al. 1984; Lee et al. 1992), and the boundary conditions.

4. Model description and initial conditions

a. Model description

The Penn State–NCAR nonhydrostatic Mesoscale Model version 5 is used. The prognostic equations (Dudhia 1993) are solved by finite differences and a time-splitting scheme on an Arakawa type-B staggered grid (Arakawa 1972). The vertical sigma coordinate, similar to a terrain-following height coordinate, is defined by a reference-state pressure that is a function of height (Dudhia 1993). There are 36 sigma levels from the surface (\(\sigma = 1.0\)) to the 100-hPa level (\(\sigma = 0.0\)) in our model domain. To better simulate the orographic effects in low levels, high vertical resolutions are used below the 625-hPa levels. The vertical resolution between the surface and 950 hPa is 6 hPa (60 m), and it is 10 hPa (100 m) between the 950- and 900-hPa levels. Between 900 hPa and 625 hPa, the vertical resolution is 25 hPa. Above 625 hPa, the model has a vertical resolution of 50 hPa. A fine domain covering 966 \times 903 \text{km}^2 with 21-km spacing is nested in a coarse domain covering 3465 \times 3339 \text{km}^2 with 63-km spacing (Fig. 7). The model topography in the fine domain is derived from National Central University (NCU) 1-km Taiwan terrain data and interpolated onto the fine grids. For the coarse grids, the model topography is derived from the NCAR 30-min (~56 km) terrain data.

The model physics include the planetary boundary
layer (PBL) processes, shortwave and longwave radiation, and precipitation physics. A revised version of the Blackadar PBL model (Blackadar 1979; Zhang and Anthes 1982) is used to forecast the vertical fluxes of heat, moisture, and momentum at each vertical layer within the PBL. The Blackadar PBL scheme includes the nocturnal regime and the free convective regime based on the bulk Richardson number (Grell et al. 1994). The calculations of the radiation include the net radiative fluxes at the surface and the radiation transfer in the earth’s atmosphere. At the surface, net radiative fluxes includes both net shortwave and longwave irradiances under clear or cloudy skies (Grell et al. 1994). The atmospheric longwave radiation is calculated by the broadband emissivity method (Stephens 1984).

The precipitation is calculated from both the explicit and implicit schemes. The resolvable-scale precipitation is directly calculated by explicit prognostic equations for three water phases: water vapor, cloud water, and rainwater (Dudhia 1989). The implicit schemes (cumulus parameterization) treat the unresolved subgrid-scale precipitation. In this study, the subgrid-scale precipitation in the control run is calculated from Grell’s cumulus parameterization scheme (Grell 1993).

b. Initial conditions

The National Centers for Environmental Prediction (NCEP) global analyses with 2.5° latitude–longitude resolution are interpolated to the grid points of the model domains. The first-guess fields are then enhanced by blending the observational data using the Cressman objective analysis technique to incorporate mesoscale features. The model is initialized at 0200 LST 22 May using the linear time interpolation of the analyses between 2000 LST 21 May and 0800 LST 22 May. For the lateral boundary conditions, the relaxation method is used for the coarse domain and the time-dependent boundary conditions are used for the nested domains to provide the forcing from the outside domain. The radiative boundary condition is used in the upper boundary to allow wave energy to pass without reflection (Klemp and Durran 1983; Bougeault 1983).

For the control (CTRL) case, MM5 was run with full model physics and island topography to simulate the convergence off the northwestern coast of Taiwan. For the experiment NTRN (no terrain), the island topography within the model domains is removed. A comparison between CTRL and NTRN will provide us insight on the orographic effects on the mesoscale airflow and the role of offshore convergence on localized rainfall along the northwestern coast of Taiwan. To examine the effects of horizontal resolution on the simulated rainfall, a fine nested domain covering 511 × 385 km² with 7-km spacing is nested in the fine domain with 21-km spacing (Fig. 7).

5. Radar analyses

a. Evolution of radar echoes

Based on the time series of radar echoes from the Kaoshiung conventional radar, three rainbands are identified. The assigned number (Fig. 8) for each rainband is based on the time sequence of the arrival of each rainband onshore. The Kaoshiung radar is a 10-cm conventional radar with a 2.25° beam width. At a distance ~150 km from the radar site, the resolution is about 6–7 km in both the horizontal and vertical directions. Thus the radar beam may not be able to resolve the echo structure of rain cells or reflectivity maxima embedded in the rainbands. Nevertheless, because of its range, it
will be used to describe the entire evolution of this coastal rainfall event. Detailed information on the structure of radar echoes and the motion of reflectivity maxima embedded within the rainbands will be described using NCAR CP4 radar data when the rainbands moved toward the coast.

At 0507 LST (Fig. 8a), rainband 2 formed southeast of Kin-men (KM) Island off the southeastern China coast. It was a northeast–southwest-oriented narrow convective line consisting of isolated rain cells. The echoes over the southwestern coastal plain of Taiwan were ground clutter. After it formed, rainband 2 moved eastward. Two hours later (Fig. 8b), rainband 1 with a north–south orientation formed off the northwestern coast of Taiwan ahead of rainband 2. In addition, precipitation echoes were also observed along the southeastern China coast. At 0836 LST (Fig. 8c), rainband 1 moved inland and decayed quickly, producing light rainfall (~1 mm) along the northwestern coast of Taiwan. At this time, rainband 3 formed to the west of rainband 2 and later merged with rainband 2 (Fig. 8d).

Rainband 2 intensified off the northwestern coast of Taiwan and continued to move eastward (Fig. 8e). The
satellite imagery also showed that the convection intensified off the northwestern coast of Taiwan (Figs. 4b and 4c) consistent with the evolution of rainband 2 observed by the Kaoshiung radar. At 1107 LST, weak echoes were observed behind the northern portion of the leading convective line (Fig. 8e). At this time, rainband 3 was a narrow line stretching southwestward. The precipitation system along the southeastern China coast was still active and moved eastward. At 1306 LST (Fig. 8f), rainband 2 moved inland producing appreciable rainfall along the coast and dissipated. Rainband 3 reached the northwestern coast of Taiwan by 1437 LST (Figs. 8g and 8h) and deposited considerable rainfall there. In the meantime, the precipitation echoes within the Taiwan Strait behind rainband 3 became weaker and dissipated later.

Each rainband contained several individual reflectivity maxima during its lifetime. Using a time series of plan-position indicator (PPI) from CP4 radar, reflectivity maxima with lifetimes more than 1 hr were tracked. The average time interval for the tracks was about 15 min. Unlike 10-cm conventional radar, attenuation by raindrops may affect the radar reflectivity measurements causing the measured radar reflectivity to be lower than the actual value. This problem would in turn affect the radar-derived rainfall estimate. In this study, the main focus of our radar analyses is on the structure and airflow associated with the rainbands. Thus, this shortcoming would not affect our results significantly.

For rainband 2, the reflectivity maxima (B1, B2, B3; Fig. 9a) formed upstream within the Taiwan Strait. They moved from southwest to northeast under the prevailing southwesterlies with lifetimes of ~1.2, 1.4, and 2.2 h, respectively. For rainband 3, the reflectivity maxima (C1, C2, C3; Fig. 9b) formed 10–30 km off the northwestern coast of Taiwan. Except for C2, they moved parallel to the northwestern coast from southwest to northeast. The lifetimes for these reflectivity maxima were ~2.5, 1.1, and 1.1 h, respectively.

As these reflectivity maxima approached the northwestern coast of Taiwan, they were advected northward by the southerly winds. During 1100–1600 LST, there was a tendency for the most active rain cells to align in a northeast–southwest orientation (Fig. 10), suggesting that the orographically enhanced low-level convergence zone offshore (Fig. 2) may be important for the intensification of rainbands. In the next section, the mesoscale airflow off the northwestern coast of Taiwan will be investigated from the dual-Doppler analyses using CP4 and TOGA radar data.

b. Dual-Doppler analyses

To investigate the mesoscale airflow off the northwestern coast of Taiwan, dual-Doppler analyses were performed during 1114:50–1119:09 LST for rainband 2 and during 1420:05–1425:15 LST for rainband 3. These are the periods that a long-lived reflectivity maximum within each rainband reached its maximum intensity off the northwestern coast, respectively.

1) Rainband 2
   (i) **Horizontal views**

Figure 10a shows the derived horizontal winds (related to ground) superimposed with the radar reflectivity at the 1-km level. The leading convective line of the rainband (Fig. 10a) and the associated convergence zone with a northeast–southwest orientation (Fig. 10b) occurred in the area where the deflected southerly flow converged with the southwesterlies off the northwestern coast of Taiwan. The flow immediately behind the convective line is from the west rather than from the southwest. Note that the low-level prevailing flow is from the southwest (Figs. 2 and 3b). The westerly flow behind the line between y = -20 km and y = -40 km were storm induced. Area-averaged radar-derived wind profiles over box 1 and box 2 (shown in Fig. 10a) were computed. Box 1 is over the area ahead of the maximum radar reflectivity near the northwestern coast of Taiwan. Box 2 is over the area immediately to the west of the maximum radar reflectivity behind the leading convective line. The wind profile over box 1 shows a strong southerly wind component (>12.5 m s⁻¹) and relatively...
weak westerly wind component between 1 and 2 km (Fig. 11). In contrast, the wind profile of the upstream sounding along the southeastern China coast has a weak southerly wind component and a strong westerly wind component in low levels. It is apparently that the strong southerly wind component off the northwestern coast (box 1) is orographically enhanced as a result of the prevailing southwesterlies impinging on the island topography of Taiwan (Li and Chen 1998). In addition, the wind profile over box 2 showed that the strongest westerly wind component occurred at 1 km (Fig. 11). But the southerly or northerly wind component was zero below 2 km, different from the wind profiles shown in the upstream sounding and box 1. Thus, the wind profile immediately behind the convective line suggests that the prevailing southwesterlies were modified by the storm-induced westerlies. As will be shown later from the vertical cross section, the low-level westerly flow behind the line is the descending rear inflow associated with the rainband.

The maximum convergence (Fig. 10b) at the 1-km level was \(3.0 \times 10^{-3} \text{s}^{-1}\) coinciding with the strongest upward motion (Fig. 10c). The vorticity field (Fig. 10d) showed cyclonic (positive) vorticity along the leading convective line of the rainband with the maximum vorticity located around the radar reflectivity maximum. The weak echoes northeast of the rainband (Fig. 10a) were associated with the dissipating rain cells that moved onshore earlier.

At 10 km, the divergent airflow (Fig. 12) representing the outflow from the deep convective clouds was observed over the reflectivity maximum. Winds were deflected westerly along the southern edge or south of these reflectivity maxima. Strong southerly winds occurred over the northern edge and immediately north of the third reflectivity maxima from the lower-left corner. As shown in Fig. 12, strong southerly winds occurred over the second reflectivity maximum from the lower-left corner.

(ii) Vertical cross sections

In addition to describing the horizontal mesoscale airflow in low levels, a vertical cross section along B1–B2 (Fig. 10a) normal to rainband 2 is constructed to show the internal structure of the rainband. The hori-
Fig. 16. Vertical cross section along line D1–D2 in Fig. 15a. (a) Convective line–related winds and radar reflectivity (shaded) with 10-dBZ interval, (b) convective line–related $u$ component winds with 5 m s$^{-1}$ interval, and (c) vertical motion with 0.5 m s$^{-1}$ interval.

Horizontal wind components in the vertical cross section are convective line–related winds determined by subtracting the mean speed of the rainband from the derived winds. The mean speed of each rainband is determined by the mean moving speed of the convective line using CP4 radar data. The mean speed for rainband 2 (rainband 3) is 12.0 m s$^{-1}$ (7.7 m s$^{-1}$).

The vertical cross section along B1–B2 (Fig. 13a) near the central part of the rainband shows that the rainband consists of a convective region with the strongest radar reflectivity of 44.6 dBZ at $x = -20$ km accompanied by a trailing stratiform rain region. A relatively weak reflectivity region occurs in the transition zone (Smull and Houze 1987; Braun and Houze 1994) between the convective and the stratiform rain regions. In the convective region, a developing cell occurs at $x = -20$ km along the outflow boundary ahead of an old cell. Two strong rising motion maxima (Fig. 13c) are located at the 5-km level (5 m s$^{-1}$) and the 8.5-km level (5.3 m s$^{-1}$) associated with the developing cell and the
old cell, respectively. This pattern is similar to the multicellular updraft cores across the mature squall line noted by Smull and Houze (1987). Within the transition zone, the sinking motion dominates. The interface between the mesoscale updraft and downdraft tilts upward from the leading edge. Behind the transition zone, a trailing stratiform region (Fig. 13a) is characterized by a bright band with radar reflectivity greater than 30 dBZ. There are extensive stratiform echoes ahead of the rainband associated with dissipating rain cells that moved onshore earlier.

The convective line–related winds normal to the rainband (Fig. 13b) show also that the rear inflow jet enters...
Vertically integrated cloud water (kg m$^{-2}$) for all model layers in the coarse domain of the control run after (a) 9-h (1100 LST), and (b) 12-h (1400 LST) simulation. The contour interval of the cloud water is 0.3 kg m$^{-2}$ starting from 0.3 kg m$^{-2}$.

The stratiform rain region at the 3-km level. It descends and converges with the front inflow along the outflow boundary. The front inflow tilts upward within the convective region and extends rearward in the upper levels associated with the outflow of the convection. In addition, the convective line–related winds ahead of the developing cell flow forward (southeastward) in the middle and the upper levels. Overall, the structure near the central part of this rainband is similar to a mature squall line found elsewhere around the world (Zipser 1977; Leary and Houze 1979; Smull and Houze 1987; Houze et al. 1989).

A schematic diagram showing the rainband evolution is given in Fig. 14. Rainband 2 formed within Taiwan Strait at 0507 LST. It intensified and reached the maximum intensity off the northwestern coast of Taiwan around 1107 LST and produced considerable rainfall along the northwestern coast as rain cells associated with it moved inland. In contrast to TAMEX IOP 13 (Li et al. 1997), the offshore convergence zone occurred within the prevailing southeasterly flow regime ahead of the 850-hPa trough in the absence of a wind shift line. Under the influence of island topography over Taiwan, the orographically induced strong southerly flow in low levels plays an important role for the presence of this offshore convergence zone.

(i) Horizontal views

The derived horizontal winds and the reflectivity at the 1-km level during 1420:05–1425:15 LST are shown in Fig. 15a. The maximum reflectivity for rainband 3 reached 43 dBZ. The airflow along the coast has a large southerly wind component whereas within the Taiwan Strait, winds are southwesterlies consistent with the larger mesoscale airflow shown in Fig. 2. The maximum convergence at this time (Fig. 15b) is $2.0 \times 10^{-3}$ s$^{-1}$. At the 10-km level (Fig. 15c), the radar reflectivity is weak. The airflow is dominated by southwesterlies.

(ii) Vertical cross sections

A vertical cross section across the central part of the rainband D1–D2 (Fig. 16a) shows an intense cell with a reflectivity maximum (42.9 dBZ) at $x = -15$ km ahead of the decaying cells. The strongest upward motion associated with this cell occurs along the leading edge of the rainband with a rear inflow layer between 2 and 4 km. The rear inflow jet reaches 12.5 m s$^{-1}$ at the 3-km level around $x = -20$ km (Fig. 16b). The vertical motions associated with the decaying cells are weak. The echo top of the developing cell reaches the 10-km level, lower than in rainband 2. The front inflow with a maximum of 10 m s$^{-1}$ occurs near the surface and converges with the rear inflow along the leading edge of the rainband. The maximum radar reflectivity and the strongest vertical motion (Fig. 16c) occur where the southerly flow along the coast converges with the southwesterlies from the Taiwan Strait (Fig. 15a). An anvil overhang (Fig. 16a) tilts southeastward with height consistent with the storm-relative flow (Fig. 16b) in the middle and the upper levels.

Similar to rainband 2, the most intense cells of rainband 3 (Figs. 2 and 9g) occur off the northwestern coast within the convergence zone (Fig. 15b). In addition to the observational evidence presented, we also conducted numerical experiments to study the orographic effect on the mesoscale airflow. The role of the convergence off
the northwestern coast of Taiwan on the coastal rainfall maximum is also simulated. The model results will be compared with satellite imagery, radar analyses, and the observed rainfall.

6. Model results and verifications

Before examining the mesoscale features related to the convergence zone off the northwestern coast of Taiwan and its role on the production of coastal rainfall, it is important to ensure that the Mesoscale Model MM5 has captured the synoptic-scale weather systems reasonably well as compared with observations.

a. Synoptic-scale flow

In the coarse domain, the 12-h (1400 LST) results of the CTRL run reproduce the 300-hPa (Fig. 17a) and the 850-hPa troughs (Fig. 17b) and compare well with observations (Fig. 18). The vertically integrated cloud water (kg m$^{-2}$) at 1100 and 1400 LST (Fig. 19) shows that convective activities are simulated ahead of the troughs over the southeastern China coast, Taiwan Strait, East China Sea, and Korean peninsula. The presence of simulated cloud water in these areas is consistent with the cloud distributions from the IR satellite imagery (Fig. 4). Within the Taiwan Strait, the convergence associated with subsynoptic-scale secondary circulation across the Mei-yu trough is about 3 × 10$^{-5}$ s$^{-1}$ (not shown). However, the convective clouds over western Taiwan as shown in Fig. 4 are not simulated in the coarse domain. This deficiency is primarily due to the fact that the orographic effects are not adequately resolved by the coarse grid. This problem will be examined in the nested domain.

b. Offshore convergence and rainfall distributions

Based on the results of the CTRL run at the 12-h (1400 LST) forecast time within the nested domain (21-
km horizontal resolution), the convergence off the northwestern coast of Taiwan and its role in localized rainfall along the northwestern coast are investigated in this section.

1) OFFSHORE CONVERGENCE

At 1400 LST, the airflow pattern at 903 hPa (~1 km) (Fig. 20a) shows that the incoming prevailing southwesterly monsoon flow decelerates and splits over the southwestern coast. The orographically induced southerly flow converges with the prevailing southwesterly flow off the northwestern coast. The maximum convergence in the nested domain is about $1.2 \times 10^{-4}$ s$^{-1}$ (Fig. 20b) where the maximum of vertically integrated cloud water (1.5 kg m$^{-2}$) (Fig. 20c) is also simulated. Note that the coarse domain also shows the convergence within the Taiwan Strait associated with the subsynoptic-scale forcing. Therefore, the large-scale convergence is enhanced orographically off the northwestern coast of Taiwan. However, the simulated convergence off the northwestern coast of Taiwan is weaker than observed from the dual-Doppler analyses (Fig. 15b). The storm-induced westerly flow is not well resolved in numerical simulations. The deficiency is possibly related to model resolution and physics, which may not be adequate to capture the small-scale airflow related to the convective system. In addition, the distribution of the vertically integrated cloud water (Fig. 20c) is consistent with the CP4 radar echoes (Fig. 2) with the maximum reflectivity (>30 dBZ) off the northwestern coast associated with the rainband. This mesoscale feature is not simulated in the coarse domain. Moreover, a convergence zone is also simulated ahead of the 850-hPa trough (Fig. 20b) within the Taiwan Strait and off the southeastern China coast consistent with the observed convective clouds there (Fig. 4d).

2) THE ROLE OF OFFSHORE CONVERGENCE ON COASTAL RAINFALL

The simulated 12-h (0800–2000 LST) accumulated rainfall (Fig. 21a) in the CTRL run shows a rainfall

![Simulated rainfalls (mm) in the 21-km nested domain of the control run during 0800–2000 LST: (a) total rainfall, (b) resolvable-scale rainfalls, (c) unresolvable subgrid-scale rainfalls. The contour interval is 10 mm.](image-url)
maximum (~70 mm) off the northwestern coast. It coincides with the strongest radar echoes (Fig. 2) off the coast. Considerable rainfall (~50 mm) is simulated along the northwestern coast of Taiwan consistent with the observed rainfall over Taiwan (Fig. 1) and the IR satellite imagery (Fig. 4). Note that the rainbands during IOP 3 contributed to most of the rainfall over Taiwan during 1000–1600 LST (Fig. 5). Along the northwestern coast of Taiwan, the simulated rainfall has a maximum of about 40 mm (not shown) during this period that is comparable to the observations. The simulated rainfall also occurs along the southeastern China coast and over the northeastern coast of Taiwan (Fig. 21a). In addition, the rainfall simulated by MM5 consists of resolvable-scale (Fig. 21b) and unresolvable subgrid-scale components (Fig. 21c). The resolvable-scale rainfall contributes to most of the simulated rainfall. Chen et al. (1998) showed that for a high-moisture and low-CAPE environment, the subgrid-scale convection did not contribute as much as grid-scale convection toward the total rainfall as compared to a much drier and higher-CAPE environment. The TAMEX IOP 3 rain case also occurs in a low-CAPE and high-moisture environment. Under this situation, most of the rainfall during TAMEX IOP 3 is produced by resolvable-scale vertical motion and is less sensitive to cumulus parameterization, consistent with the results of Chen et al. (1998).

To examine the role of orographic effects on the enhancement of the convergence off the northwestern coast and the resulting coastal rainfall maximum, the results from the NTRN experiment is compared with the CTRL run. In the NTRN experiment, the low-level deceleration and the split of the incoming prevailing southwesterly flow over the southwestern coast are absent (Fig. 22a). The orographically induced southerly flow along the northwestern coast is also absent. There-
fore, the enhanced low-level convergence (Fig. 22b) off the northwestern coast is not simulated and the simulated vertically integrated cloud water (Fig. 22c) there is significantly less. Within the Taiwan Strait, the large-scale convergence and the associated cloud water are still simulated. Nevertheless, the simulated 12-h rainfall accumulation during 0800–2000 LST (Fig. 23) over the northwestern coast of Taiwan is much less (<10 mm) than in the CTRL run. These results indicate that the convergence zone off the northwestern coast of Taiwan is enhanced orographically in the CTRL and resulted in the coastal rainfall maximum.

3) SENSITIVITY TO PRECIPITATION PARAMETERIZATION AND HORIZONTAL RESOLUTION

To further assess whether the production of the rainfall in the low-CAPE and high-moisture environment is sensitive to the cumulus parameterization used or not, the simulation using the Kain and Fritsch (1993) scheme instead of Grell's scheme used in the CTRL run is performed. Similar to the CTRL run, the simulated rainfall (Fig. 24a) during 0800–2000 LST primarily occurs over the northwestern coast of Taiwan and within the Taiwan Strait. The resolvable-scale rainfall also contributes to most of the simulated rainfall (Figs. 24b and 24c).

To assess the sensitivity of the model results to higher horizontal resolution, the results of the 12-h simulation in the fine nested domain with a 7-km resolution are compared with the 21-km nested domain. The airflow pattern at 903 hPa (~1 km) (Fig. 25a) reproduces the orographically induced southerly flow along the western coast of Taiwan and converges with the prevailing southwesterlies (Fig. 25b) over the northwestern coast. The orographic blocking on the prevailing southwesterly flow is also simulated. But the storm-induced westerly flow within the Taiwan Strait as shown in dual-Doppler radar analyses is not simulated. These results are similar to the results shown in the 21-km nested domain (Fig. 20). For the simulated rainfall (Fig. 25c) during 0800–2000 LST, the significant rainfall over the northwestern coast of Taiwan and within the Taiwan Strait is also reproduced. This is similar to the CTRL run and consistent with the observed rainfall. However, the orographic precipitation over the mountainous region in northwestern Taiwan is overpredicted.

7. Conclusions

During TAMEX IOP 3 (22 May 1987), moderate rainfall (≥50 mm day⁻¹) are observed along the northwestern coast of Taiwan because of the arrival of three rainbands in succession. This heavy rain event occurs within the southwesterly monsoon flow regime ahead of the 850-hPa trough. These rainbands form over the Taiwan Strait with a northeast–southwest or north–south orientation. In the formative stage, these rainbands consist of isolated convective rain cells. They intensify off the northwestern coast. There is a tendency for the most intense echoes to align in a northeast–southwest orientation off the northwestern coast. An orographically enhanced convergence zone with a northeast–southwest orientation occurs in the area where the deflected southerly flow converges with the prevailing southwesterly flow that is modified by the storm-induced westerlies immediately behind the convective line.

In numerical studies, the coarse domain in the CTRL run reproduces the synoptic-scale weather conditions well, including the upper-level and low-level troughs over southern China, and low-level strong southwesterly flow ahead of the low-level trough. The vertically integrated cloud water and large-scale convergence are also simulated over the southeastern China coast, Taiwan Strait, East China Sea, and Korean peninsula. These results are consistent with the convective activities observed by the IR satellite imageries. However, the active convective clouds over the northwestern coast of Taiwan during the period with significantly localized rainfall are not simulated because the orographic effects are not properly represented in the coarse horizontal resolution.

In the nested domain, the CTRL run simulates the deceleration and splitting of the incoming southwesterly flow off the southwestern coast of Taiwan due to orographic blocking. The orographically induced southerly flow along the western coast converges with the prevailing southwesterly flow off the northwestern coast and enhances the large-scale con-
vergence offshore. The simulated rainfall along the northwestern coast of Taiwan is in good agreement with the observed rainfall pattern. The resolvable-scale rainfall contributes to most of the simulated rainfall because the environment has a high moisture content and low CAPE. However, the storm-induced westerly flow immediately behind the convective line is not well resolved in the numerical experiment. Thus, the magnitude of the simulated offshore convergence is smaller than that in the dual-Doppler analysis. This deficiency may be related to model resolution and precipitation physics, which may not be adequate to capture the convectively driven small-scale airflow associated with the convective systems.

Without the island topography of Taiwan, the southerly flow along the northwestern coast is absent in the model simulation. The orographically enhanced convergence off the northwestern coast of Taiwan is not simulated. The simulated rainfall along the northwestern coast is much less than in the CTRL run. The results from the radar analyses and numerical simulations indicate that the convergence off the northwestern coast is a combination of the synoptic-scale forcing, orographic effects, and the feedback of the convection. The offshore convergence is important for the production of the coastal rainfall maximum.

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Fig. 25. The 12-h (1400 LST) CTRL simulation in the 7-km nested domain. (a) Winds (m s$^{-1}$) with full barb and half barb represent 5 and 2.5 m s$^{-1}$. The contours of wind speed (solid) are 2.5 m s$^{-1}$ interval starting from 12.5 m s$^{-1}$. (b) Divergence field with an interval $3 \times 10^{-2}$ s$^{-1}$: Dashed (solid) lines represent convergence (divergence). (c) Simulated rainfalls (mm) during 0800–2000 LST. The contour interval is 10 mm.

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