A Case Study of Trade-Wind Rainbands and Their Interaction with the Island-Induced Airflow

JIAN-JIAN WANG* AND YI-LENG CHEN

Department of Meteorology, School of Ocean and Earth Science and Technology, University of Hawaii at Manoa, Honolulu, Hawaii

(Manuscript received 19 December 1995, in final form 21 April 1997)

ABSTRACT

A case study of trade-wind rainbands observed on 22 August 1990 during the Hawaiian Rainband Project is presented. It shows that the interaction between the morning rainbands and the island-induced airflow is important for the evolution of the rainbands. In the early morning of 22 August, there are two convective periods, 0400–0600 and 0700–0900 HST (Hawaii standard time), on the windward side of the island of Hawaii. For both periods, preexisting rain cells are observed in the trade-wind flow at least 40 km upstream of the island and move westward toward the island.

At night and in the early morning, the offshore flow opposes the trade winds resulting in a convergent region over the area immediately upstream of the island. As the first group of rain cells (0400–0600 HST) moves toward the island, the low-level convergent airflow provides a favorable kinematic background for the enhancement of the incoming rain cells. These rain cells merge in the convergent zone and become a well-defined rainband. However, after the first rainband meets the offshore flow, the cool air feeds into the lowest levels of the rainband. This is an unfavorable thermal condition for the rainband and is thus partly responsible for the decay of the first rainband over the windward lowlands. After the arrival of the first rainband, the depth of the offshore flow at Hilo increases from about 250 m to over 500 m. Its horizontal extent also extends from approximately 10 km to more than 20 km offshore.

The second group of rain cells (0700–0900 HST) also becomes a well-defined rainband as it moves over the convergent zone. Interacting with a deep and extensive offshore flow resulting from precipitation effects from the first rainband, the rain cells associated with the second rainband are much deeper and stronger than the first rainband. The second rainband moves toward the island during the morning transition, during which the offshore flow retreats and onshore flow begins. After the onset of the onshore flow, the low-level airflow in the Hilo Bay region diverges and splits around the island. This provides an unfavorable dynamic condition for the maintenance of the rainband. Therefore, the second rainband weakens. It dissipates and only reaches the eastern tip of the island.

1. Introduction

The island of Hawaii, the largest mountainous island in the Hawaiian island chain with peaks exceeding 4100 m in elevation, is a natural laboratory for investigating the interaction between atmosphere, ocean, and land processes. Climatologically, trade wind conditions are persistent (~70%) in Hawaii throughout the year, especially during the summer months, with the maximum occurrence of 92% of the time in August (Schroeder 1993).

Leopold (1949) first described the interactions be-
formation offshore. They suggested that for a low-Froude number regime, such as the island of Hawaii, the nighttime surface offshore flow is mainly driven by dynamic blocking; band clouds are frequently generated in the low-level convergence zone along the separation line between the surface return flow and the incoming trade winds. The nocturnal cooling only slightly modifies the strength and location of the dynamically driven flow. In their later discussion, Rasmussen and Smolarkiewicz (1993) suggested that dynamic blocking and nocturnal cooling are equally important in causing the nighttime downslope flow; however, the location of the convergence zone is still thought to be primarily determined by dynamic blocking.

The aircraft observations during the Joint Hawaii Warm Rain Project (JHWRP) in 1985 enabled studies to be done on the structure and evolution of the rainband offshore (Takahashi 1986, 1988). Takahashi et al. (1989) presented a schematic model for the rainfall accumulation. The rainband has a sloped updraft in the low-level convergence zone. But, different from other storm systems, Hawaiian rainbands are long lasting without a typical cold dome. Raga et al. (1990) suggested that the convergence could generate clouds by lifting the air to the level of free convection and that the generated buoyancy is sufficient to maintain the clouds once they leave the zone of convergence. LeMone and Jensen (1990) provided some evidence for those hypotheses of rainband maintenance and speculated that the clouds persist until evaporative cooling and water loading from the rainfall leads to their dissipation. Rasmussen et al. (1989) studied a band cloud during JHWRP that was quasi-stationary in the convergence zone. They suggested that the occurrence of this band cloud was linked to the convergence zone.

Hawaiian Rainband Project (HaRP) was conducted from 11 July to 24 August 1990 to better understand the island-induced airflow and the evolution of the trade-wind rainband frequently observed upstream from the island. Using the HaRP dataset, Chen and Nash (1994), Chen and Wang (1994), and Chen and Feng (1995) documented the diurnal variation of surface airflow, rainfall, and the thermodynamic fields over the entire island. In addition to the diurnal heating cycle, the variations of surface thermodynamic fields are also related to orography, airflow, and distributions of orographic clouds and rain showers. Chen and Wang (1994) found that the onset of upslope (downslope) flow is governed by the thermal contrast between the island surface and the adjacent air at the same elevation. To study the effects of precipitation and clouds on surface thermodynamic fields and airflow, Chen and Wang (1995) and Wang and Chen (1995a) performed both composite and individual case studies for dry and rain cases during HaRP. It is found that clouds and rainfall can modify the surface thermal fields and result in changes in the intensity of the diurnal circulations and the timing of the wind shift from downslope (upslope) to upslope (downslope) flow at the surface in the early morning (late afternoon). On the windward lowlands, most rain cases show earlier downslope flow onset in the afternoon, deeper downslope flow at night, and later upslope flow onset in the morning than dry cases. Carbone et al. (1995) also suggested that evaporation cooling of rainfall plays an important role for the flow reversal on the windward side. On the upper slope, Chen and Wang (1995) showed that for most rain cases, orographic clouds delay the turning from upslope to downslope flow. The delay of the downslope flow onset on the upper slope may be related to the reduction of outgoing longwave radiation by an extensive cloud cover as evidenced by a slower temperature decrease in the afternoon hours as compared to dry cases. Different from studies before HaRP, preliminary HaRP radar data analyses (Wang and Chen 1995b; Ochs et al. 1995) indicate that some rainbands offshore may originate upstream rather than over the convergence zone between the offshore flow and the coming trade winds.

Despite recent HaRP studies that provide a better understanding of local airflow, rainfall, and their interaction over the island, considerable gaps between our knowledge and the reality of the development of the trade-wind rainbands upstream from the island still exist. In particular, the interaction between the island-induced airflow and the rainband is not well understood. In this paper, we will use the HaRP dataset, mainly dual-Doppler radar and aircraft data, to present a detailed case study of 22 August 1990 focusing on the evolution of the two propagating morning rainbands and their interaction with the island-induced airflow. Section 2 describes the data we used. Section 3 presents the evolution of the first (earlier) rainband and the influences of downslope/offshore flow. The feedback of the rain showers associated with the first rainband on the low-level airflow, particularly the depth and horizontal extent of the downslope/offshore flow, is described in section 4. The formation and evolution of the second (later) rainband and interactions with the island-induced airflow are discussed in detail in section 5. A summary is given in the last section.

2. Data analysis

The primary data used for this study are radar and aircraft data collected during HaRP. The additional sources of information include synoptic charts, Hilo raingauges, the Portable Automatic Mesonet (PAM; Chen and Nash 1994) on the windward side, and tethered sondes (Wang and Chen 1995a) at Kaumana Elementary School.

The National Center of Atmospheric Research (NCAR) Electra is the primary source of airborne measurements obtained during HaRP. For the measurement of height at lower altitudes (<500 m), the output from the Stewart–Warner radar may not be valid; therefore, we use the output from the Sperry radio altimeter. The
wind data are derived from measurements taken from the radarome wind package and the inertial navigation system (INS). INS performance is good during HaRP. However, as suggested in the Hawaiian Rainband Project Data Catalog (1991), wind data reported during turns, and in particular climbing turns, should be used with caution because these conditions may severely affect the ambient wind calculation.

An important issue for aircraft data analysis is the difficulty of measuring temperature in cloud because of the sensor wetting on the surface of the Rosemount thermometer (LeMone 1980). We noticed that for several cloud penetrations, the temperature readings are 0.5°–1.0°C lower than the dewpoint. An alternative way to determine in-cloud temperature is to use the temperature readings from the side-looking radiometer, which does not expose an electrical element to the airstream (Jorgenson and LeMone 1989), and to correct side-look radiometer temperature readings with the systematic difference between side-looking radiometer and Rosemount thermometer temperature readings in cloud-free areas. However, such a temperature sensor is not available in HaRP. For the best guess in this study, we corrected both temperature and dewpoint values with one-half of the difference whenever the dewpoint exceeds the temperature. In general, the wetting problem of these sensors in our case study results in dewpoints less than 1°C higher than temperature. Therefore, the error of equivalent potential temperature calculated from corrected temperature and dewpoint should be within ±1 K.

During HaRP, two C-band Doppler radars were operated at Paradise Park (CP3) and General Lyman Field at Hilo (CP4), respectively (Fig. 1). The basic principles in deriving wind fields from the combined use of separated Doppler radars are given by Armijo (1969). Davies-Jones (1979) notes that a minimum beam-intersection angle (difference in azimuth of the two radars when they are scanning identical volumes) of 30° is usually necessary for qualitatively reasonable dual-Doppler analyses. Then, if the data are detected with this angle of intersection or greater, the three-dimensional wind field computed using the CEDRIC (custom editing and display of reduced information in Cartesian space) analysis package, developed at NCAR (Mohr et al. 1986).

Dual-Doppler radar analyses are prone to numerous sources of error. Doviak et al. (1976) predicted the variance of wind components synthesized from dual-Doppler velocity estimates combined with the terminal velocity and the continuity equation. In this study, vertical velocity is constrained to vanish at the ground and at the top of cloud bands. Combining the results of Gal-Chen and Wyngaard (1982), Carbone et al. (1985) suggested that 10 independent samples ensure approximately 90% recovery of flow amplitude, and 5 independent samples ensure approximately 75% amplitude recovery. For our case, based on a similar calculation, we believe that we can resolve more than 80% of the amplitude of features that have a spatial extent of 2 km (6–7 samples) or greater. To monitor the evolution of low-level winds at the radar sites, the velocity–azimuth display (VAD) method (Srivastava et al. 1986; Matejka and Srivastava 1991) is also applied in radar data analysis.

3. Environmental conditions
The surface chart at 0800 HST (Hawaii standard time) 22 August 1990 (not shown) exhibits a typical trade-wind situation on the Hawaiian Islands. The island chain is under the influence of the subtropical high centered at 47°N, 150°W (not shown). Based on the Electra sounding starting at 0607 HST approximately 140 km upwind from the island, the trade-wind inversion is at a height of about 2900 m (720 hPa) with a strength of 3.6 K (Fig. 2a). The surface mixing ratio q is 14.5 g kg⁻¹ and the potential temperature θ is 297 K. The convective available potential energy is only about 60 J kg⁻¹, with a small negative area (10 J kg⁻¹) below the level of free convection (890 hPa). The lifting condensation level is at 935 hPa (700 m). The equivalent potential temperature θₑ profile shows a moist, well-mixed surface layer in the lowest 500 m with θₑ of 336 K and a near-constant θₑ (~330 K) trade-wind layer from 500 m to the base of the trade-wind inversion. The trade winds are quite uniform below the trade-wind inversion, almost due easterly with an average speed of 8 m s⁻¹. The inversion height shown in Fig. 2a is higher than the height (2500 m) detected around 0645 HST by the later sounding about 70 km upstream of the island when Electra flew toward the island (not shown). This may
be because part of the earlier aircraft sounding (770 hPa–720 hPa) is taken in clouds.

Based on the Hilo sounding at 0700 HST, the height of the trade-wind inversion base is about 2500 m (760 hPa) (Fig. 2b). Also evident is the dry downslope flow in the lowest levels (~500 m), and a nocturnal inversion at 240 m with a strength of 2.4 K. The $\theta_e$ profile has a minimum (328 K) at the surface and increases to 334 K at about 110 m because of the temperature inversion. Another local minimum $\theta_e$ is located at 240 m corresponding to the minimum dewpoint. The $\theta_e$ increases to 335 K at 800 m and decreases gradually to 330 K at 1300 m. From there to the trade-wind inversion, $\theta_e$ is uniform, almost the same as the upstream value.

Figure 3 shows the surface winds from PAM data over the island and the aircraft winds at approximately 200 m over the ocean. In the early morning, the downslope flow prevails over the windward side of the island and extends approximately 10 km offshore of Hilo. A low-level convergent zone is observed in the region.
where the offshore flow encounters the easterly trade winds. Before reaching the convergent zone, nocturnal offshore flow decelerates and turns southward and northward in the upstream region south and north of the island, respectively.

4. The first rainband

a. Development of the rainband

The radar operations begin at 0406 HST at Hilo Airport (CP4) and 0532 HST at Paradise Park (CP3), respectively (Fig. 1). Radar echoes are already observed about 30–60 km upstream of Hilo by CP4 radar at 0406 HST (Fig. 4). These echoes merge later as they move close to the island and are enhanced by the offshore convergence zone. A convective cell is usually defined as the maximum radar reflectivity in excess of 40 dBZ and having a diameter of 1–5 km (e.g., Cheng and Houze 1979). In this study of trade-wind rainbands, we define a rain cell as 1) having a lifetime greater than 1 h, 2) with a maximum reflectivity above 30 dBZ, and 3) the maximum east–west extent of its reflectivity core (≥30 dBZ) during the lifetime is greater than 3 km. These rain cells may contain several convective elements with a maximum radar reflectivity over 40 dBZ and a much shorter lifetime of 15–20 min. Without the radar data during its earlier stage, the exact location of formation of these rain cells is unknown. However, in contrast to the observational and numerical studies prior to HaRP, these rain cells must have formed at least 40 km upwind rather than just over the region where the offshore flow interacts with the trade winds.

At 0406 HST, this convective line primarily consists of a few rain cells about 10–35 km south of Hilo and a northern portion including several individual rain cells with maximum reflectivities greater than 30 dBZ. Before the operation of CP3 radar, most CP4 scans are surveillance mode starting from 0.5° elevation angle at 2° intervals. For a rain cell 50 km from Hilo, the surveillance scans of 0.5°, 2.5°, and 4.5° provide data at the heights of 400, 2200, and 3900 m, respectively. This vertical resolution is not sufficient for the detailed analysis of rain cell structure and evolution. From 0406 to 0457 HST, these rain cells move toward the island at an average speed of 5–6 m s\(^{-1}\) (Fig. 4). In general, the strength of the northern portion of the line is fairly stable during this period with a maximum reflectivity, based on the 2.5° scans, around 35–40 dBZ from 0406 to 0457 HST. The variation of the maximum reflectivity may be partially related to the different levels detected by the radar scans at different times. For the northern portion of the line, the 2.5° scans correspond to approximately the 2300-, 1900-, and 1600-m levels at 0406, 0427, and 0457 HST, respectively. Around 0500 HST, a new rain cell forms between the early formed southern and northern rain cells about 25 km east of Hilo and ahead of the convergence zone (Fig. 4). This rain cell becomes a part of the well-formed rainband later along with the other rain cells as they move over the convergence zone. With good radar data coverage through its life cycle, we choose this rain cell as the target rain cell for the discussion that follows.

The target rain cell forms around 0500 HST, about 25 km from Hilo (Fig. 4). It takes about 10 min for the target rain cell to widen its 10-dBZ contour to 4 km and increase the maximum reflectivity to above 40 dBZ (Fig. 5). After that, the rain cell moves toward the island at an average speed of 5.5 m s\(^{-1}\). Figure 6 displays a CP4 RHI (range–height indicator) scan across the rain cell along 86° azimuth (from north) from Hilo at 0526 HST. At this time, the rain cell is 16 km upstream of Hilo, that is, 8–10 km from the leading edge of the offshore flow (Fig. 6). The maximum reflectivity is 44 dBZ. The 10-dBZ contour has an east–west extent of 5 km and extends to the 2700-m level, about 200 m above the trade-wind inversion. For the first rainband, all rain cells are below or slightly above the inversion base.

From 0505 to 0540 HST, the aircraft penetrates the rainband at six different levels from 150 m to 2400 m along an east-to-west axis before the rainband interacts with the offshore flow (Fig. 7a). For the lowest flight leg, the aircraft encounters the rainband at point C. Composite fields of relative \(u\) component (subtracting the rainband motion of 6.0 m s\(^{-1}\) westward from aircraft data) and \(\theta_e\) are constructed from the aircraft flight-level data based on the space–time conversion to account for the rainband motion (Figs. 7b, c). Point C is the reference point for the space–time conversion to adjust the
horizontal coordinate. Since it takes 35 min for the aircraft to complete its flight pattern, the mesoscale features associated with the rainband with a timescale much larger than 35 min, such as the mesoscale airflow relative to the rainband, are well sampled by the aircraft observations. The relative $u$ component (Fig. 7b) shows that the low-level flow into the band is primarily from the downwind side. The maximum westerly flow relative to the rainband (>4 m s$^{-1}$) is in the midlevel (900–1500 m) except close to the convection core. The only region with a negative relative $u$ component is found in the upper level (~2500 m) on the downwind side of the convection representing outflow from the rainband. It is important to note that before the rainband reaches the low-level confluence zone, the low-level inflow (<1 km) from the environment is the warm, moist maritime air from the western side of the rainband. The airflow pattern from this analysis represents airflow during the early stages of the rainband before it interacts with the offshore flow.

At the lowest level that the aircraft sampled (~200 m), $\theta_e$ is 334 K at 4 km upstream of Hilo (Fig. 7d). The $\theta_e$ of the offshore flow is higher as it extends farther over the ocean. At its leading edge, the offshore flow has $\theta_e$ over 336 K, equal to the far upstream value measured by the aircraft. The increase in $\theta_e$ of the offshore flow from land to ocean may be related to 1) the heat and moisture fluxes from the ocean and 2) the horizontal mixing between the air from the island and the air over the ocean. A high-$\theta_e$ tongue between 10 and 20 km from Hilo is also related to the lifting of air from the low levels. The highest $\theta_e$ (>336 K) is observed in the lowest levels on the western side of the rainband. A high-$\theta_e$ (>334 K) region extending vertically upward is apparently associated with rising motion in the rainband, which brings the warm and moist air from the lowest levels to the upper level (~2500 m) with very little mixing. The $\theta_e$ distribution is consistent with the mesoscale relative airflow presented in Fig. 7b in which the low-level inflow is from the western side of the rainband. In Fig. 7b, although the relative flow from the west is strongest at the 1-km level, this may not be the primary inflow level for the rainband. Note that above the 700-m level west of the rainband, the $\theta_e$ values are less than that within the $\theta_e$ core in the vicinity of the rainband. Since $\theta_e$ is a conserved quantity during moist adiabatic processes, it is unlikely that the strong westerly flow above the 700-m level represents the inflow entering the rainband and accounts for the high-$\theta_e$ core above the 1-km level. The $\theta_e$ distributions suggest that the inflow from the western side is mainly confined in the lowest 600–800 m. A large portion of the westerly flow at the 1-km level may move around the rainband rather than entering the rainband. Two regions of low $\theta_e$ (<330 K) are apparent on either side of the rainband. Comparing with the upstream $\theta_e$ profile (Fig. 2a), the low $\theta_e$ (326–330 K) in the far field of the rainband suggests the presence of mesoscale subsidence in the environment.

b. Peak and decay of the rainband

Around 0530 HST, the rainband is strengthened by the formation and the development of the new rain cells (Fig. 4). After that, these rain cells merge gradually and form a continuous band. About 0540 HST, the target rain cell (Fig. 5) meets the leading edge of the offshore flow (Fig. 3). The peak reflectivity (50 dBZ, Fig. 5) of the rain cell is located at the altitude of 1100 m at 0550 HST (not shown). The interaction of the offshore flow with the rainband also results in an increase of the width of all reflectivity contours from 0540 to 0550 HST (Fig. 5). The similar intensification of other rain cells in the rainband is also observed when they are close to the leading edge of the offshore flow (not shown). At that time, the target rain cell is too close to the baseline between CP3 and CP4, and, thus, not in a good position for dual-Doppler analysis. At 0552 HST, the CP4 surveillance scan (2.5$^\circ$), which covers the whole rainband, shows that the rainband is best developed and organized at that time. The radar reflectivity, especially the high reflectivity of the rain cell, weakens after the rainband passes the low-level confluence zone between the offshore flow and the trade winds (Fig. 5). When the rainband moves over the island at around 0610 HST, there are still several cores showing reflectivity about 40–45 dBZ (not shown, refer to Fig. 5 for the target rain cell). After that, the rainband weakens rapidly. At 0627 HST, most rain cells disappear (Fig. 4). The rainband dissipates over the windward lowlands before 0700 HST.

In the early morning, the low-level air over the island is very stable, with no convective available potential

---

**FIG. 5.** The east–west extent of 10-dBZ (solid line), 30-dBZ (dashed line), 40-dBZ (dotted line), and 50-dBZ (dashed–dotted line) reflectivity in the target rain cell as it moved toward the island. Time (HST) is marked for selected points.
energy or conditional instability (Fig. 2b). In other words, there is no positive buoyancy available for the island air to rise continuously even if it is forced to the lifting condensation level. Therefore, after the first rainband moves over the island, the cool and dry island air feeding into the lowest levels of the rainband would produce an unfavorable thermal condition for the maintenance of the rainband. This is the main reason responsible for the dissipation of the first rainband over the windward lowlands.

c. Influences of the first rainband

The first rainband produces small amounts of precipitation on the windward lowlands after 0600 HST (Fig. 8). Before the rainfall starts, the characteristics of near-surface thermal fields and winds at Kaumana School (asterisk in Fig. 1) are very close to that for clear cases, for example, 20 August described in Wang and Chen (1995a). At 0550 HST, a strong nocturnal inversion is found at about the 970-hPa level, with a strength of 3.2
K (Fig. 9a). Because of the katabatic winds in the lowest levels (Fig. 9b), the mixing ratio is higher at the top of the downslope flow (13.6 g kg\(^{-1}\)) than at the surface (11.2 g kg\(^{-1}\)) (Fig. 9a). The downslope flow extends above the nocturnal inversion with a depth of 15–20 hPa (Fig. 9b). A downslope wind maxima of 7.4 m s\(^{-1}\) is observed at the 970-hPa level just beneath the nocturnal inversion.

After the rainfall starts around 0615 HST, mixing ratio, temperature, and moist static energy all increase in the lowest levels from 0550 to 0650 HST with decreasing moist static energy aloft (Fig. 9a). These features indicate the existence of vertical mixing associated with rains as found by Wang and Chen (1995a, their Fig. 15b). It appears that the potentially warm, moist air aloft mixes with the potentially cold, dry air beneath the nocturnal inversion. The temperature decrease of 1–2 K in the layer of 970–965 hPa is primarily due to the evaporation of raindrops since the moist static energy remains about the same. Above the 970-hPa level, a near 2-K temperature drop occurs with only a slight decrease in moist static energy suggests that evaporative cooling is present. Part of the decrease in the temperature readings aloft may also be related to sensor wetting. From 0615 to 0700 HST, both the nocturnal inversion (~0.8 K) and downslope wind maxima (~4.5 m s\(^{-1}\)) weaken.
Fig. 10. The vertical cross section showing the boundary of downslope/offshore flow (heavy solid line) at (a) 0600, (b) 0700, and (c) 0830 HST. The data used are tethersonde \((X = -9 \text{ km})\), VAD \((X = 0 \text{ km})\), and aircraft (over the ocean) in (a) and (c), and Hilo sounding \((X = 0 \text{ km})\) and CP4 RHI scans (over the ocean) in (b). Thin lines indicate the island topography.

Fig. 11. Same as Fig. 3 but using the aircraft data at about the 100-m level over the ocean.

(Fig. 9b) because of the vertical mixing associated with rains as well as surface heating after sunrise \((\sim 0600 \text{ HST})\). Before 0700 HST, about an hour after sunrise, the cooling aloft due to rain evaporation results in a deep stable downslope flow with a depth of more than 34 hPa, which is more than twice as deep as before rainfall begins.

The vertical cross sections of horizontal winds constructed from VAD winds from the CP4 radar, Hilo sounding, and the aircraft data reveal that after rains start around 0600 HST, the downslope flow deepens at Hilo and extends farther offshore. Before rains start, the downslope/offshore flow is only 200 m deep at Hilo and reaches only 10 km offshore with a speed of 2–3 m s\(^{-1}\) over the ocean (Figs. 10a and 3). After the precipitation on the windward side, the downslope/offshore flow deepens to above 500 m and extends to 15 km offshore (Fig. 10b). Note that the deepening of the offshore flow occurs after sunrise \((\sim 0600 \text{ HST})\) as a result of evaporative cooling aloft. At 0830 HST, compared to the earlier time, the offshore flow becomes much stronger \((5–8 \text{ m s}^{-1})\) and reaches farther \((>20 \text{ km})\) over the ocean (Figs. 11 and 10c).

5. The second rainband

a. The development of the second rainband

At 0607 HST, two rain cells (B and C in Fig. 12) move into the radar range about 40–50 km upstream of the eastern tip of the island. A series of Geostationary Operational Environmental Satellite images (not shown) shows that these clouds form in the far upstream open ocean as trade-wind cumuli and move westward toward the island. Half an hour later, the third rain cell (A in Fig. 12) forms to the south of rain cell B. The maximum reflectivities of these rain cells are fairly constant \((40–45 \text{ dBZ})\) in the following hour. Around 0730 HST, rain cells D and G (Fig. 12) form 40–50 km upstream of Hilo and north of rain cell C. Rain cell E forms around 0800 HST between rain cells D and G. It is evident that new rain cells except cells B and C form in the upstream region near the island, and all rain cells merged into a well-defined rainband after they met the offshore flow (Fig. 13). Among these individual rain cells, rain cell C has the maximum east–west extent \((\sim 12 \text{ km})\) and the longest duration \((>3.5 \text{ h})\), whereas rain cell E is the strongest having an area of 1–2 km\(^2\) with the radar.
reflectivity in excess of 55 dBZ. Therefore, in this study, we focus our discussions on the evolution of rain cells C and E and their interactions with the island-induced airflow.

b. Evolution of rain cell C

From 0600 HST to 0730 HST, rain cell C moves toward the island at a speed of 5.8 m s\(^{-1}\). At 0730 HST, the 10-dBZ contour of rain cell C has an east–west extent of 5 km and reaches the 3000-m level (Fig. 14a), approximately 500 m above the trade-wind inversion (Fig. 2). The maximum reflectivity is about 36 dBZ. After 0730 HST, rain cell C moves at a slower speed of 2.7 m s\(^{-1}\). The slowing down of the rain cell C may be related to the fact that it moves toward the deep and far-extending offshore flow. At 0800 HST, the maximum reflectivity increases to 45 dBZ with the top of the 10-dBZ contour extending to the 3300-m level (Fig. 14b).

Because of the interaction between the deep and strong offshore flow and the onshore trade winds, the rainband is enhanced as it moves into this deep convergence zone. The areal averaged low-level (~100 m) convergence computed from the integration of the hor-

---

**Fig. 12.** The evolution of the second rainband based on CP4 surveillance scans. Dash lines indicate discontinuities among the rain cells of the rainband.

**Fig. 13.** Same as Fig. 4 but for the second rainband. Line DE indicates the track of Electra at the lowest level from 0825 to 0830 HST. Point F is where the aircraft enters the rainband.

**Fig. 14.** Vertical cross section of reflectivity for rain cell C at (a) 0730, (b) 0800, (c) 0818, (d) 0837, and (e) 0900 HST.
horizontal winds following the track of Electra (shown in Fig. 11) is about $3 \times 10^{-4} \text{ s}^{-1}$. Comparing to the upstream region where the low-level easterly winds are quite uniform, the presence of low-level convergence there apparently provides a favorable condition for the enhancement of the rainband. As rain cell C moves closer to the island, it exhibits a westward tilt with height because of flow deceleration in lower levels accompanied by stronger winds in upper levels. After 0818 HST, the rain cell becomes much wider, deeper, and stronger (Fig. 14c). The 10-dBZ contour is nearly 10 km wide and extends to a height of 3400 m. Rain cell C reaches its peak reflectivity ($>50 \text{ dBZ}$) around 0837 HST (Fig. 14d) after the rainband moves into the strong low-level convergence zone (Fig. 11). The 10-dBZ contour enlarges to 12 km and extends to above 3500 m (Fig. 14d). At that time, the rain cell is located to the north of and very close to the eastern tip of the island (Fig. 12). Around 0900 HST, the southern portion of the second rainband collapses over the eastern tip of the island (Figs. 12 and 14e).

From 0824 to 0850 HST, the Electra penetrated the rainband between the cores of rain cell C and D at five different levels (Fig. 15a). As in Fig. 7, the space–time conversion is used to derive vertical cross sections with point $F$ in Fig. 15a as the reference point. The speed of the rainband motion is about 2.7 m s$^{-1}$ at this time. At 0836 HST along the aircraft flight track, the vertical cross section of radar reflectivity shows a maximum reflectivity of 37 dBZ (the cross in Fig. 15a). At that time, the offshore flow extends more than 20 km offshore of Hilo (Fig. 11). It meets the rainband in the lowest 500 m (Fig. 15b). The vertical cross sections constructed from the aircraft data allow us to study the interaction between the incoming trade winds and the offshore flow as these rain cells arrive.

The mesoscale airflow relative to the morning rainband shows that in the lowest levels, the warm air on the eastern side of the rainband meets with the cool offshore flow (Fig. 15b). Above the offshore flow, weak relative westerly flow from the western side is present. In the upper levels, a well-defined relative westerly flow maximum is evident (Fig. 15b) downstream of the rainband suggesting that the outflow blows to the downwind side. The 10-dBZ contour of radar reflectivity extends slightly above the trade-wind inversion (Fig. 2).

In the lowest 500 m, the warm and moist incoming trade-wind flow is lifted as it encounters the deep offshore flow. A tongue of high $\theta_e (>336 \text{ K})$ tilts westward toward the island (Fig. 15c). A core of high $\theta_e (>334 \text{ K})$ is observed above the convergence zone coinciding with the reflectivity maximum in Fig. 15a. It is apparent that this high-$\theta_e$ air originates from the low level east of the convergence zone. The lifting of the warm, moist air by the deep offshore flow is important for the strengthening of the second rainband as it encounters the leading edge of the deep offshore flow. Minimum $\theta_e$ is observed for the cold and dry downslope flow from the island (Fig. 15c). The $\theta_e$ increases gradually offshore with the maximum gradient in the region where the offshore flow converges with the incoming trade winds; $\theta_e$ is less than 332 K on the western side and more than 336 K east of it. Away from the convergence zone, low $\theta_e (<330 \text{ K})$ is observed on either side of the convergence zone above the 1-km level. The low-$\theta_e$ air does not extend to below the 1-km level, suggesting that the compensating subsidence in the environment of weak echoes, if any, may be rather weak.

c. Evolution of rain cell E

Cell E forms after 0750 HST, about 32 km upstream of Hilo (Fig. 12). It takes about 10 min for the maximum reflectivity to increase to 30 dBZ (Fig. 16). At 0818 HST as the rain cell moves into the range of dual-Doppler analysis, the maximum reflectivity is about 35 dBZ (Fig. 17a, upper). The 10-dBZ contour is about 3 km wide and extends slightly above the trade-wind inver-
Fig. 16. Same as Fig. 5 but for the rain cell E of the second rainband.

The low-level cell-relative inflow is from the upwind side (east) of the rain cell with convergence dominating the layer below the 1000-m level (Fig. 17a, lower). In the core of the rain cell with reflectivity exceeding 30 dBZ, the convergence extends to near 2000 m with a maximum ($>2 \times 10^{-3}$ m s$^{-1}$) below the 1000-m level. Above the convergence, the flow becomes increasingly divergent, especially near 3000 m above the core of maximum reflectivity. The upper-level outflow mainly blows to the downwind side. Vertical velocities are obtained by downward integration of the anelastic mass-continuity equation from the upper to lower boundaries where vertical velocities are constrained to vanish. The upward motion is moderate, with a maximum value of 2 m s$^{-1}$ in the midlevel.

From 0817 to 0837 HST, the maximum reflectivity of rain cell E is around 35 dBZ (Fig. 16). At 0837 HST, the rain cell is just upstream from the leading edge of the deep and extensive offshore flow. The 10-dBZ contour enlarges to 6 km wide (Fig. 17b, upper). The maximum reflectivity (36 dBZ) is in the eastern part of the rain cell with strong low-level convergence ($>4 \times 10^{-3}$ m s$^{-1}$) and upward motion (3 m s$^{-1}$) (Fig. 17b, lower). In this portion, the low-level inflow is from the upwind...
side, ascending and flowing to the downwind side at the upper level. In addition, another reflectivity maxima (>20 dBZ) is observed in the downwind (west) side of the original rain cell. This new center forms just after the western part of the rain cell meets the leading edge of the offshore flow. The low-level convergence (>2 × 10^{-3} s^{-1}) and upward motion (~3 m s^{-1}) in that area may be related to the lifting along the leading edge of the deep offshore flow and enhanced by condensation.

The two reflectivity maxima of rain cell E merge at about 0840 HST (not shown). The rain cell intensifies rapidly after 0837 HST (Figs. 16 and 17c) in the low-level convergence zone as a result of interaction between the deep offshore flow and the trade winds. At 0848 HST, the 10-dBZ contour increases to 7 km wide and deepens to the 3400-m level (Fig. 17c, upper). A reflectivity maximum of 58 dBZ is located at the 1500-m level. This is the maximum reflectivity observed during the life cycle of rain cell E. The high reflectivity region of the rain cell has the strongest upward motion, up to 3.5 m s^{-1} (Fig. 17c, lower). Below the echo core, convergence exceeding 6 × 10^{-3} s^{-1} is observed in the lowest 1000 m. In contrast to the target rain cell of the first rainband that moves over the island, rain cell E of the second rainband weakens gradually in the offshore region (10–20 km from Hilo, Fig. 16) after 0848 HST and dissipates. The weakening of rain cell E occurs during the morning transition with a rapid change in the airflow on the windward side and the adjacent ocean. At about 0835 HST, the maximum horizontal extent of the offshore flow is more than 20 km offshore. After that, as a result of increasing solar heating, the offshore flow retreats toward the island. Its leading edge is 17 km offshore at 0845 HST, and 10 km offshore at 0900 HST (not shown). At 0848 HST, the core of rain cell E is about 20 km upstream of Hilo, which is on the upwind side of the offshore flow as the offshore flow retreats. Along with the weakening of the offshore flow, the rain cell moves faster at a speed of 4.1 m s^{-1}. As will be shown in section 5d, the offshore flow shifts to the onshore flow after 0900 HST. At 0932 HST, the maximum radar reflectivity for rain cell E decreases to about 34 dBZ (Fig. 17d, upper). Weak upward motion and low-level convergence are observed on the upwind side accompanied with divergence on the downwind side (Fig. 17d, lower) as the cell dissipates.

Takahashi (1986, 1988) suggests that the parabolic wind profile within the trade-wind layer is necessary for the formation of a strong and long-lived rainband. On 22 August, two consecutive rainbands form under the same trade-wind conditions, but the second rainband is much stronger than the first rainband (Fig. 18) because it interacts with a deeper and farther extending offshore flow. Most rain cells of the first rainband have a maximum radar reflectivity of 40–50 dBZ with the 10-dBZ contour top below the 3000-m level, whereas the second rainband records a peak radar reflectivity of 50–60 dBZ with the 10-dBZ contour top above the 3000-m level.

d. Decay of the rainband

In general, the surface airflow in the Hilo region shows convergence throughout the night, with a maximum value around sunrise (solid line in Fig. 19). After sunrise, the transition from downslope to upslope flow occurs earlier in the lowlands than along the coast (Chen and Wang 1994). As a result, the magnitude of surface convergence decreases. After the onset of upslope flow, surface divergence is observed in the Hilo region because of the splitting airflow (Chen and Nash 1994).

On 22 August, the arrival of the first rainband results in the maximum convergence in the Hilo region at 0600 HST (dashed line in Fig. 19). The surface winds turn to a divergent flow after 0630 HST with a peak immediately after rain showers because downdrafts associated with rainfall strengthen the downslope flow in the coastal area. Similar to the rain cases (e.g., 24 July) discussed in Wang and Chen (1995a), the effects of rainfall deepen the downslope flow. Therefore, the morning transition on 22 August starts later than on clear days (e.g., 20 August). The second rainband reaches its
peak from 0830 to 0845 HST after it meets the deep and farther-extended offshore flow (not shown). After that, the offshore flow retreats toward the island as a result of increasing solar heating. The rainband is now on the upwind side of the offshore flow and weakens. At 0900 HST, the offshore component only has a 10-km horizontal extent from Hilo (Fig. 20a). After the upslope flow onset at 0930 HST (Fig. 20b), the airflow over the Hilo area becomes divergent (Fig. 19) with splitting airflow in the Hilo Bay area (Fig. 20b). All rain cells of the second rainband dissipate around 1000 HST and do not result in any precipitation in the Hilo region (Fig. 13).

6. Summary and conclusions

In the early morning, there are two convective periods, 0400–0600 and 0700–0900 HST, on the windward side. For both periods, preexisting rain cells are observed in the trade-wind flow at least 40 km upstream of the island and move westward toward the island. Though the formation mechanism of these rain cells in the far upstream region remains unclear due to the lack of data, we think that it may be related to the trade-wind cumuli in the open ocean or other systems with scales larger than the island.

We found that the island-induced airflow affects the evolution of these two rainbands. At night, the low-level airflow of the windward lowlands and the upstream area near the island is convergent as a result of interaction between the downslope/offshore flow and the trade winds. As the first group of rain cells (0400–0600 HST) moves toward the island, this low-level convergent airflow enhances the existing rain cells. These rain cells merge in the convergent zone and become a well-defined rainband. However, after the first rainband meets the cool offshore flow, especially over the land, the island cool air feeds into the lowest levels of the rainband. That is an unfavorable thermal condition for the rainband and responsible for the decay of the first rainband over the windward lowlands.

After the morning rainfall along the windward coast, the cooling aloft due to rain evaporation results in a much deeper offshore flow (>500 m) with a larger horizontal extent (>20 km) over the ocean than before the arrival of the first rainband. The second group of rain cells (0700–0900 HST) also becomes a well-defined rainband as it moves over the convergent zone. Interacting with this deep and extensive offshore flow, the rain cells associated with the second rainband are much deeper and stronger than the first rainband. The second rainband moves toward the island during the morning transition, during which the offshore flow retreats followed by the onset of the onshore flow. After the onset of the onshore flow, the low-level airflow in the Hilo Bay region diverges due to a splitting airflow around the island. This provides an unfavorable dynamic condition for the maintenance of the rainband. Therefore, the second rainband weakens and dissipates offshore of Hilo. It only reaches the eastern tip of the island.

Acknowledgments. We would like to thank those who participated in HaRP, especially the Electra and radar crew whose dedication made this work possible. This work is supported by the National Science Foundation under Grants ATM-9301227 and ATM-9629886. Acknowledgment is made to the National Center for Atmospheric Research, which is sponsored by the National Science Foundation for the computing time used in this research. We would like to thank G. Barnes, T. Schroeder, and R. Carbone for their suggestions and J. Li for his assistance. Constructive reviews from R. Rasmussen and an anonymous reviewer helped in the presentation of this paper.

REFERENCES


