The Heavy Rain Event of 29 October 2000 in Hana, Maui*

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ABSTRACT

On 29 October 2000, the Hana region of Maui received 700 mm of rain in 7 h. Radar analyses revealed that the storm consisted of seven cells that were initiated along the southeast slope of Haleakala volcano. One of these cells survived for nearly 4 h and was responsible for 80% of the volumetric rainout from the storm. The interaction of low-level flow distorted by the island of Hawaii located farther east, the passage of a trough, and the topographic forcing caused by Haleakala volcano were major factors responsible for the evolution of the storm.

1. Introduction

During seven hours on the afternoon of 29 October 2000 over 700 mm of rain fell over the eastern tip of the island of Maui, Hawaii. The principal population center in the area is Hana (Fig. 1) and we shall refer to the event as the Hana storm. The area is remote, with one traffic artery that was severed by the ensuing flash floods. No lives were lost but some evacuations were required and several rescues were necessary.

The event’s notoriety was short lived due to the occurrence on the island of Hawaii within the ensuing 72 h of a much larger event with greater maximum rainfall (940 mm in 24 h), greater spatial extent, and much more substantial economic impact. The meteorology of the latter event is better understood (Schroeder 1978; Kodama and Barnes 1997) but that of the Hana storm is not.

The State of Hawaii has always been vulnerable to heavy rain events. Six flash floods occur annually on average (Kodama and Barnes 1997). Extreme rain rates (Fullerton and Wilson 1975; Carbone et al. 1998), small watersheds with rapid response times, and a data-sparse synoptic environment have made forecasting of heavy rains difficult. As a consequence, determination of time and place of flooding has generally been an exercise in nowcasting. Hawaii lacked meteorological radar support prior to the installation of the first Weather Surveillance Radar-1988 Doppler (WSR-88D) in 1993. The WSR-88D has improved the surveillance of heavy rain situations but research quality archival data were unavailable until recently. These radars were also sited to provide coverage for the major airports, not necessarily the regions prone to heavy rains. The Hana event was one of the first for which the level II archive data are available.

Ramage (1971) reviewed studies of heavy tropical rainfalls and concluded that the storms 1) were associated with a synoptic-scale disturbance, 2) drew upon plentiful supplies of moisture, and 3) were usually anchored by some discontinuity in surface roughness.

He further subdivided events into those with strong surface winds (tropical cyclone, vigorous monsoon disturbances) and those with weak surface winds (characterized by quasi-stationary thunderstorms). Schroeder (1977, 1978) applied these concepts to heavy rain events in the Hawaiian Islands. He found that island topography was a dominant factor in most events, producing extreme gradients of rainfall. Schroeder (1978), Cram and Tatum (1979), and Dracup et al. (1991) noted that warm clouds have produced heavy rains. With the exception of Dracup et al. (1991), no radar information existed for any of the above studies.

A sequence of notorious flash floods that occurred in rapid succession in the mid-1970s inspired a number of studies that applied Geostationary Operational Environmental Satellite (GOES) data. These included Big
Thompson, Colorado, on 31 July 1976 (Maddox et al. 1978; Caracena et al. 1979); Johnstown, Pennsylvania, on 19–20 July 1977 (Hoxit et al. 1978; Bosart and Sanders 1981); and Kansas City, Missouri, on 12–13 September 1977 (National Weather Service 1977). Satellite data demonstrated that deep convection was ubiquitous, and that the echoes were either quasi-stationary or tracking repeatedly over the heavy rain area.

Radar studies of heavy rain events confirm the presence of a nearly stationary storm with cell regeneration on the rear flank resulting in a series of cells passing over the same region (e.g., Chappell 1986; Akaeda et al. 1995; Bauer-Messmer et al. 1997; Baeck and Smith 1998; Petersen et al. 1999). In his comprehensive review of heavy rain events, Davis (2001) emphasizes that propagation or new cell formation must essentially counter the translation of the storm cells. For this to occur, outflows from downdrafts either do not move away from the heavy rain, or are not present with the depth nor density difference to initiate cells in locations away from the affected region. High relative humidity in the midtroposphere has been argued to be an inhibitor to the production of downdrafts (Akaeda et al. 1995; Kodama and Barnes 1997) and continues to be recognized as an important ingredient for heavy precipitation (e.g., Doswell et al. 1996; Konrad 1997; Harnack et al. 1999).

Maddox et al. (1979) prepared a comprehensive study of 151 heavy rain events for the continental United States and identified eight common features:

1) Torrential rains were associated with convective storms.
2) Surface dewpoints were high.
3) High moisture content was found through a deep tropospheric layer.
4) Weak shear of the horizontal winds existed through the cloud-bearing layer.
5) Convective cells repeatedly form and move over the same area.
6) A weak 500-hPa trough helped trigger and focus the storms.
7) Storms coincided with the position of the midtropospheric large-scale ridge.
8) Storms tended to occur at night.

Most of the features listed above are traits of the synoptic-scale environment, corresponding to the first two characteristics identified by Ramage (1971). Feature 5 above is a result of one or more mesoscale processes.

Kodama and Barnes (1997) analyzed 44 heavy rain events that occurred on the southeast flank of Mauna Loa volcano on the island of Hawaii. The region in question lies between the east rift of Kilauea volcano (1300 m) and the south rift of Mauna Loa (4200 m) and the role of local topographic focusing is well known (Schroeder 1978). They found that the events coincided with one of four synoptic types previously identified by Haraguchi (1977): tropical cyclone or remnant, subtropical cyclone, cold front/shearline, or upper-level trough/low. They found that low-level winds possessed an upslope component that was well correlated with the rains. They also found that with each disturbance the associated vertical ascent eroded the prevailing trade wind inversion and the midlevel environment moistened. This feature was similar to the findings of Maddox et al. (1979). They suggested that the K index (George 1960) was an excellent stability parameter for identifying these cases. They proposed that the midlevel moisture reduces the entrainment of dry air into the cloud and suppresses the formation of cold downdrafts. The absence of downdrafts precludes development of outflows that would displace new cells farther down slope or even upstream away from the area.

The findings of Maddox et al. (1979) and those of Hawaiian investigators are generally consistent. Two exceptions are warm cloud (cloud tops below 0°C) heavy rain events, and episodes of strong shears with heavy rains (e.g., Dracup et al. 1991), which have both been observed in Hawaii. In each instance the topographic influences were significant, unlike in the continental sample.

The Hana event is distinguished from other Hawaiian heavy rains by its isolation, exceptional short-term precipitation totals, and the availability of new observing systems (WSR-88Ds). In the following discussion...
we shall appraise the observations available, the synoptic environment, and analyze cell structure, evolution, rainfall production, and motion to describe this event.

2. Data

a. East Maui, Hawaii—Terrain and data sources

The eastern half of Maui is dominated by Haleakala volcano, the summit of which exceeds 3000 m in elevation (Fig. 1). Hana lies at the eastern end of the east ridgeline that separates the north and south sides of the volcano. Very limited surface weather observations are taken at the Hana airport (HAP). There are three rain gauges located within 10 m of each other at HAP. One is a part of a statewide array of telemetered recording gauges, but this gauge only supplied a total rain estimate rather than the usual 15-min temporal resolution. Also present are a gauge in the National Climatic Data Center Hourly Precipitation Data network (2.5-mm resolution) and a standard gauge (0.25-mm resolution) within the State of Hawaii climatic rain gauge network. The latter is sampled at 0800 local time. Gauge readings agreed for this event. An unofficial gauge several kilometers to the west (254-mm capacity) was read irregularly by a retired National Weather Service (NWS) employee during the event. The next nearest observations are 50 km to the west at Kahului airport (OGG) in the central valley northwest of Haleakala.

Rawinsondes are launched at 0200 and 1400 Hawaiian standard time [Hawaiian standard time (HST) = UTC – 10 h] at Hilo and Lihue (Fig. 1). Additional synoptic-scale information can be gleaned from the standard visible and infrared imagery from the GOES-10 satellite. The water vapor channel for this day was intermittent and therefore neglected. National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis data are used to assess the synoptic-scale environment. The NCEP–NCAR reanalysis is available four times a day with horizontal resolution of 2.5°.

b. Radar

There are four WSR-88D radars in the state (Fig. 1), sited by the Federal Aviation Administration to cover the key airports. The archived reflectivity and radial velocity, a part of what is commonly known as level II data, are used. The base reflectivity has a spatial resolution of 1° by 1 km (a bin), a reflectivity resolution of 0.5 dBZ, and a maximum range of 460 km. The WSR-88Ds can be set into one of four volume coverage patterns (VCPs). For the first half of the period the VCP was set to precipitation mode, which consists of nine 360° azimuthal sweeps from 0.5° to 19.5° elevation. It takes 6 min to complete a full volume scan. During the latter half of the period the VCP was changed to severe weather mode. Fourteen 360° sweeps are made in 5 min over the same volume during the precipitation mode.

The velocity azimuth display (VAD) can be used to create mean vertical wind profiles for the region around the radar in question. Scatterers must fill the volume around the radar site and steadiness must be assumed for the time it takes to complete the scan. Errors associated with VADs are discussed by Browning and Wexler (1968). Only the radar on Molokai offers complete profiles from 1324 to 1635 HST.

3. Results

a. Synoptic situation

The weather on 29 October deviated from the typical trade wind showers in the vicinity of Hawaii. A large area of convection developed north and east of Oahu (Fig. 2a), spread over Oahu (Fig. 2b), and gradually moved eastward and dissipated (Fig. 2c). An isolated feature appeared over East Maui (Hana) at 1230 HST and persisted through 1830 HST. Additional convection erupted to the west of Maui. The Hana storm was a minor feature on the radar scope for most of the day.

Synoptic analyses from the NCEP–NCAR reanalysis reveal a trough aloft over the islands. The trough was evident at 250 and 500 hPa and was reflected at the surface by a split in the surface ridge north of the islands and a weak “inverted” trough at the surface (Fig. 3). The surface winds over Maui were easterly in contrast to the normal trade winds. Such southeast flow was a typical feature of the storms studied by Kodama and Barnes (1997). The upper trough gradually drifted eastward. This was documented by subtle shifts in winds aloft as well as temperatures (300 hPa; Fig. 4). This trend coincides with the drift of the radar features (Fig. 2).

The 1400 HST Lihue and Hilo soundings bracket the trough aloft. The Hilo air mass (Fig. 5a) is slightly warmer through the lower and middle troposphere and moister (43 mm of precipitable water versus 36 mm for Lihue; Fig. 5b). Neither sounding contains a significant trade inversion. The K index and lifted index at Hilo each indicate reduced hydrostatic stability east of the trough. This destabilization is typical of the environment associated with upper-level troughs over Hawaii. The ascent associated with the advancing trough eliminates the subsidence responsible for the prevailing trade wind inversion (Kodama and Barnes 1997). The soundings suggest stability distributions consistent with the synoptic pattern. The radar history of the event
Fig. 2. Molokai base reflectivity (0.5° elevation angle) every hour from 0830 to 1830 HST on 29 Oct 2000. Color shading represents 10-dBZ increments starting at 15 dBZ.
(Fig. 2) is consistent with the movement of the trough to the east, with new cells developing farther eastward and the older cells to the west decaying as they come under the influence of negative vorticity and subsequent subsidence.

**b. Rainfall**

The three rain gauges at the Hana airport collected the same amount of rain: 249 mm. An unofficial gauge mentioned earlier collected a remarkable 686 mm (27 in.). These amounts fell between approximately 1200 and 1900 HST on 29 October 2000, based on the individuals that tended the instruments, and an assessment of when radar echoes were over the region. The 686-mm reading was at first suspect, but investigations by the NWS senior service hydrologist of the high-water mark left at nearby Mokulehua Stream, damage and the damage gradient between Hana and the gauge in question, and other testimonials from long-time residents confirm that an extremely heavy rain occurred. There was also another unofficial “gauge,” actually an empty circular waste receptacle that was located between the Hana airport and the gauge that received 686 mm. This gauge collected 550 mm.

Radar estimates of rain offer the advantage of high spatial and temporal resolution, but suffer the major disadvantage of requiring the appropriate calibration. Causes of poor rain estimates from radar are detailed in Battan (1973), Austin (1987), Rinehart (1991), Rosenfeld et al. (1992), Bauer-Messmer et al. (1997), Baek and Smith (1998), and Fulton (1999). Key problems are the identification of the appropriate $Z-R$ relationship, the height and incomplete filling of the beam, attenuation of the radar energy, contamination by the bright band, and the identification of a hail cap. Some of these problems can be mitigated, while others are not easily fixed. We will use differences between the two radars to determine an adjustment for incomplete beam filling. Comparisons between the rain gauges and the radar-derived rain will be used to identify the best hail reflectivity cap, and to select the best of three $Z-R$ relationships commonly applied in the Tropics.

Two WSR-88D sites offer coverage over the Hana region. The radar located at Kamuela on the northern part of the island of Hawaii is 60–100 km from the region while the Molokai radar is 100–140 km away. The Kamuela radar is most desirable, but it fails at 1719 HST and remains inoperative for the rest of the period of interest. The Molokai radar offers coverage during the entire storm, but for best results it needs to be adjusted for range.
There is a difference in the width of the beam given the approximately 40 km greater range for the Molokai radar. Rosenfeld et al. (1992) demonstrated that incomplete beam filling is particularly serious when the radar is viewing intense convective cells with large reflectivity gradients. As the beam widens with increasing range, high-reflectivity cores are averaged with the surrounding lower reflectivity resulting in an underestimation of the precipitation. A correlation between the maximum reflectivity observed by each radar for a particular location shows that Molokai has a low bias of 3.5 dBZ. We will apply this correction to the Molokai data for an estimate of storm total precipitation (STP) given that this radar offers coverage throughout the event.

Both radars have their lowest elevation scan between 2 and 3 km above sea level over the Hana region. The closer Kamuela radar is at this height because it is sited about 1 km above sea level. A lower scan height is always more desirable, but in this case much of the rain falls on terrain that is 0.5 to 1.0 km above sea level. If the lowest elevation scan were another 1 km lower, we would be plagued by ground clutter and beam blocking by Haleakala.

To compare the ground measurement with the radar-derived estimate, we identify the best-fit bin, which is a 1° by 1 km portion of the reflectivity field. This is the standard high-resolution product from the WSR-88D processing. We sought the nearest bin that is best correlated with the hourly rain record at the Hana airport. The bin for both radars that yields the best correlation with the gauge is offset 4–5 km to the southeast. Recall that the beams are 2–3 km above sea level. With a mean southeasterly flow of 5–6 m s$^{-1}$ in the lowest 3 km, it is likely that the rain at the beam level will fall to the northwest several kilometers away. This best-fit bin will be used to determine the appropriate hail cap and the

![Fig. 3. Synoptic-scale surface winds (m s$^{-1}$), pressure (hPa), and lifted index (°C) at 1400 HST 29 Oct 2000. Unstable lifted index values are shaded.](image-url)
Z–R relationship that comes closest to matching the rain that was recorded hourly by the rain gauges.

This storm did produce hail 25 mm in diameter, which yields high reflectivity and, correspondingly, overestimates the precipitation (Fulton 1999). We must therefore contend with the application of a hail reflectivity cap. A hail cap constrains all reflectivity values above a chosen threshold to that threshold. Operationally the Honolulu Forecast Office uses 51 dBZ for its radar-derived rain estimates. This threshold is somewhat lower than the typical value selected for tropical sites; so we conducted a test using the best-fit bin and assumed hail caps of 51, 53, and 55 dBZ with the operational Z–R relationship. Differences between the Hana airport official gauge and the operational Z–R relationship are substantial with a 51-dBZ cap; the radar overestimates hourly rain by as much as 75% compared to the rain gauges. The other higher hail caps yield even worse biases, being 250% and 330% higher than the ground measurement. We conclude that 51 dBZ is the best choice, but it is also clear that the operational Z–R relationship is not accurately describing this particular storm.

There are an almost inexhaustible number of Z–R relationships (Battan 1973). For rain in the Tropics, but not associated with hurricanes, three of the more widely used relationships are from Hudlow and Patterson (1979), Rosenfeld et al. (1993), and Tokay and Short (1996). The Rosenfeld et al. (1993) equation, \( Z = 250R^{1.2} \), is in operational use at the Honolulu Weather Forecast Office and was found by Birchard (1999) to yield the best match with rain gauges for a heavy rain event on Oahu. The three relationships yield similar rain rates for reflectivities less than 43 dBZ; above that value the rain derived from a given dBZ diverges rapidly (Fig. 6). Note that by 50 dBZ the difference between the highest and lowest estimates is 50 mm h\(^{-1}\).

The three Z–R relationships are applied to the best-fit bin, all using a 51-dBZ hail cap and are compared to the Hana surface measurements. For this storm the operational Z–R overestimates the STP for Hana (Fig. 7) the most, the Hudlow and Patterson (1979) Z–R offers some improvement, and the Tokay and Short (1996) equation, \( Z = 139R^{1.43} \) fares the best. Note that the rain estimate for the Kamuela radar, using the Tokay and Short algorithm, virtually matches the ground measurement till 1700 HST, when the radar fails. The adjustments are less successful for the Molokai radar, especially for the high rain rates between 1700 and 1800 HST, but they are still superior to the other two relationships.

Application of the 51-dBZ hail cap, the Tokay and Short (1996) Z–R, and the addition of 3.5 dBZ to the Molokai radar to account for incomplete beam filling provide a best-estimate storm total precipitation from 1200 to 1900 HST (Fig. 8). The STP from the radar manifests a very compact area, about 2 km\(^2\), with rain greater than 700 mm. The spatial scale of the rain greater than 70 mm is roughly circular, with a diameter slightly less than 20 km. We have not altered the STP map to account for the likely southeast to northwest drift of the drops. The radar-derived rain amount, with the likely drift of the drops taken into consideration, agrees well with the gauge that collected 686 mm. Agreement over Hana is poorer, but much closer than before the tuning exercise.

World record rainfall (Fig. 9) shows that the Hana storm is close to, but below, the line that defines the maximum potential rainfall for a given duration. Receiving approximately 700 mm in 7 h does appear to separate heavy rain events for tropical locations versus midlatitude locales. The extreme amounts noted for the Tropics are usually under the influence of a quasi-stationary tropical cyclone or monsoon flow [e.g., La Reunion events; see Barcelo et al. (1997)]. Midlatitude locations record lower amounts as they depend more on the cooperation of several phenomena of varying scale (e.g., Caracena and Fritsch 1983; Petersen et al. 1999; Pontrelli et al. 1999; Junker et al. 1999).

c. Evolution of the storm

Like all storms, one can partition the life into growth, mature, and decay phases. The spatial extent of the 45–60 dBZ from the two radars (no adjustments to the reflectivity field are made to determine storm struc-
Fig. 5. Skew T–log\(p\) diagrams for (top) Hilo and (bottom) Lihue at 1400 HST. Wind barbs are 5 m s\(^{-1}\) and triangles are 25 m s\(^{-1}\). LCL is lifting condensation level, LFC is level of free convection, EQLV is equilibrium level, PW is precipitable water integrated between surface and 100 mb, LI is lifted index, KI is K index, CAPE is convective available potential energy, and BRN is bulk Richardson number.
ture) shows growth from 1200 to about 1500 HST, a mature phase till 1730 HST, then a rapid decay with high reflectivities disappearing by 1900 HST (Fig. 10a). Maximum reflectivity for the two radars shows the 3.5-dBZ low bias for Molokai (Fig. 10b), but with reassuringly similar trends. Storm top (Fig. 10c), estimated with infrared sensors mounted on the GOES-10 satellite, is between 10 and 12 km; the higher spatial resolution of the Molokai radar shows evidence of tops reaching over 14 km and above the tropopause. Heavy rains in Hawaii can often be from clouds that reach only to the midtroposphere (e.g., Schroeder 1978), but the Hana storm clearly has deep convection.

During the growth stage there are several cells aligned north–south and there is no reflectivity reaching 10 km (Fig. 11a, 1315 HST). New cells appear on the south side of the island and track north over the ridge of Haleakala. The north–south alignment becomes more SSW to NNE about an hour later (Fig. 11b, 1409 HST), and though the storm has reached 10 km briefly before, there are no echoes at this level at 1409 HST. By 1538 HST (Fig. 11c) strong echoes reach to 10 km and the storm appears to be dominated by a single cell. This single cell reverses direction and moves to the SSE (Fig. 11d, 1608 HST). The heaviest rain is just offshore by 1659 HST (Fig. 10). The horizontal reflectivity gradients weaken and the tops are descending by 1731 HST (Fig. 11e).

Sample cross sections reveal that the vertical profiles along a SE–NW plane and through the high-reflectivity regions are indicative of weak ordinary cells (Fig. 12, 1320:43 and 1344:18 HST), but by 1537 HST (Fig. 12) there is an overhang with much higher tops. The storm maintains an overhang for over 1.5 h (Fig. 12, 1658:02 HST). This overhang is to the SE and on the inflow side of the storm and is interpreted as a weak echo region. By 1730:18+ HST (Fig. 12) the storm is losing altitude, and its overhang.

A cell is defined as maxima of reflectivity greater than 45 dBZ that covers greater than 5 km$^2$, survives for at least 15 min, and is present in at least the two lowest-elevation scans. Seven cells are identified and their characteristics are summarized in Table 1. Cells labeled C1–C5 and C7 move from the south-southwest, have maximum reflectivities less than or equal to 60 dBZ, and an average top for the 45-dBZ contour of 6.5 km.
Lifetimes vary, but the mean duration of these six cells is just over an hour. The remaining cell, C6, stands out, given that it is much longer lived (>3.7 h), several kilometers deeper, more intense, and has a contrasting mean motion. The mean vertical profile of reflectivity (Fig. 13) for each cell reveals that all but C6 have a low echo centroid structure (Baeck and Smith 1998), typically found in tropical systems (e.g., Houze 1977; Szoke and Zipser 1986) with the greatest reflectivity at and below 2–3 km. Contrasting these profiles is the one for C6, which is more akin to a midlatitude hailstorm (e.g., Chisholm and Renick 1972; Foote and Wade 1982) with maximum reflectivity near the freezing level.

The tracks of the centers of the 45-dBZ area for these cells (Fig. 14) demonstrate that they generally meet the cell criteria just about on the crest of the eastern ridge of Haleakala. The cells move to the north side of the ridge and decay, but C6 eventually changes direction and moves to the SSE. Its path is best described as like a horseshoe in shape. The cells generally reach their maximum reflectivity on the north side of Haleakala. The echoes collapse on this side, too. Downdrafts and outflows in the subcloud layer are most prevalent when the reflectivity profiles undergo collapse (e.g., Fujita and Wakimoto 1981; Wakimoto and Bringer 1988) and when rain has reached into the subcloud layer (e.g., Barnes and Garstang 1982), especially for tropical environments. We view this as evidence that the majority of the downdrafts and outflows would occur on the north side of the volcano, and away from the inflow that is responsible for the continued regeneration of cells on the south side.

When the rain amounts from both radars are plotted together with the times when each of the cells was present, C6 again distinguishes itself. The long-lived C6 is responsible for nearly 80% of the rain from the storm (Fig. 15).

![Fig. 9. World record rainfall accumulations as function of duration (from Ramage 1995); appended with the Hana event of 29 Oct 2000.](image)

![Fig. 10. (top) Areal extent of four different reflectivity magnitudes with data prior to 1719 HST from the Kamuela radar and subsequently from Molokai, (middle) maximum reflectivity from both radars, and (bottom) storm top, again from both radars as in (top) and from satellite as functions of time.](image)
Fig. 11. PPIs at 0.5°, 2.4°, and 6.2° for the Kamuela radar, corresponding to approximately 2.5, 4.5, and 10 km above mean sea level, respectively. Color shading represents 10-dBZ intervals beginning at 15 dBZ and dashed lines representing 60 dBZ. Letters A and B denote locations of cross sections presented in Fig. 12. Time of scan is noted in each frame with each series starting with the lowest elevation scan at (a) 1315, (b) 1409, (c) 1538, (d) 1608, and (e) 1731 HST. Panel (e) is from the Molokai radar with scan elevations of 0.5°, 1.5°, and 4.3°.
d. Change in cell character with C6

Duration, motion, intensity, height, and the amount of rain all highlight that C6 contrasts with all the preceding cells as well as the final cell. In fact, C6 has many of the characteristics of a supercell including a nearly 4-h life, a weak echo region with a forward reflectivity overhang, hail with a diameter greater than 25 mm, overshooting tops, and a bulk Richardson number of 35. Of course one can have a bulk Richardson number that is simply a ratio of weak convective available potential energy (CAPE) and weak shear, which is what either the Hilo or Lihue soundings offer. Perhaps the hallmark of a supercell is the strength of the dynamic pressure perturbation, caused by the interaction of the updraft with the vertical shear of the horizontal wind (Schlesinger 1975; Klemp and Wilhelmson 1978). Doswell (2001) argues that a supercell possesses a deep, persistent mesocyclone. Unfortunately, we do not have Doppler measurements to verify the existence of the mesocyclone nor do we have aircraft available to measure the dynamic pressure perturbation (e.g., LeMone et al. 1988). The issue, however, is not what label we give it. Instead, can we identify causes as to why a cell with such different characteristics forms? We will discuss the evolution in chronological order, highlighting the genesis of the storm, the cause of the first five cells, the birth of C6, and the system’s demise.

e. Key events in the evolution of the storm

The first five cells develop far from the main activity in the trough, on the southeast slope of Haleakala volcano. Clues to genesis are few given the sparse observations. However, a high-resolution satellite picture (Fig. 16) reveals that the blocking effect of the big island of Hawaii may have focused low-level flow into the SE portion of Maui. This effect of the big island has been discussed by numerous researchers (e.g., Smolarkiewicz et al. 1988). The bow cloud is a manifestation of the enhanced convergence associated with the deceleration of the trade wind flow. The enhanced convergence of the bow cloud reinforces the upslope motion on Maui in a location that coincides with the genesis region for the storm.

The Hilo sounding from 1400 HST (Fig. 5a) is representative of initial conditions. The trade wind inversion is very weak, and there is CAPE near 950 J kg\(^{-1}\). The sounding does have a dry layer from above 600 to 375 hPa that would limit cloud growth via entrainment.

The first two cells do not produce much rain (Fig. 15). Perhaps the modest instability is struggling against the dry air that would be entrained (see above 600 hPa in the Hilo sounding). The bulk Richardson number is 35 but these early cells have no characteristics that would invite a “super” label. The tracks of the early cells (Fig. 14) are toward the NNE, and they reach maximum spatial extent and maximum intensity on the
north side of the ridge. If these cells produced a substantial downdraft, the resulting outflow would occur on the north side of the island, away from the genesis point. There is little chance of disruption by these downdrafts so the cell formation continues unabated.

These early cells probably do condition the atmosphere and lower the inhibiting effect of dry air entrainment. However, there appears to be another factor that assists in the formation of C6—the passage of the trough. First, the trough passage shifts the winds more to the west, altering the alignment and track of the cells. This will allow any downdrafts and ensuing outflows to be east of the island where it can spread across the sea and perhaps interact with the inflow. Second, there is a moistening between 550 and 400 hPa, which reduces the inhibiting effect of entrainment (Figs. 5a,b). Third, cooler air is advected over the Hana region in the upper levels (Fig. 4). The cooling, which extends through the thickness of the trough, or from 700 to 260 hPa, is approximately 2.5°C, based on a comparison between the Lihue and Hilo soundings. This increases CAPE by about 50%, and invigorates C6. Note that an increase of CAPE would raise the BRN and favor ordinary cell formation.

Cell C6 initially moves to the northeast at 4 m s⁻¹, which is about 70% of the mean winds from the surface to 10 km for the Hilo sounding. Based on the VAD at Molokai, winds with a northerly component did not reach the radar site, nearly 120 km to the west of Hana, until 1630 HST. Cell C6 moves to the SSE nearly 40 min earlier. This seems to preclude the chance that NW flow, indicative of conditions on the west side of the trough line, was affecting C6 at this time. What has been occurring is that cells have been slowly shifting their track to a more NNE direction. A little after 1530 HST, C6 tops increase by about 2 km, and maximum reflectivity reaches 65 dBZ. The cell contains hail and probably produces a stronger downdraft and outflow. At the time when C6 turns toward the SE (1542 HST) the high reflectivity associated with the cell is centered over the coast where the terrain is between 150 m and sea level. Here the outflow from C6 is no longer confined by the east rift of the volcano, nor does it acquire any significant acceleration downslope that would drive it toward the NNE. Instead at least some of the outflow spreads toward the SE. Hana airport did record flow from the NNW at 6 m s⁻¹ after 1600 HST, supporting the argument that downdrafts exist and produced out-

<table>
<thead>
<tr>
<th>Cell No.</th>
<th>Mean speed (m s⁻¹)</th>
<th>Mean direction (°)</th>
<th>Start time (HST)</th>
<th>End time (HST)</th>
<th>Duration (min)</th>
<th>Max Z (dBZ)</th>
<th>Mean height of max Z (km)</th>
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<td>1303</td>
<td>1338</td>
<td>35</td>
<td>54</td>
<td>2.2</td>
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<tr>
<td>C3</td>
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<td>196</td>
<td>1338</td>
<td>1427</td>
<td>48</td>
<td>56</td>
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<td>C4</td>
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<td>1643</td>
<td>71</td>
<td>59</td>
<td>2.7</td>
<td>6.2</td>
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</table>

FIG. 13. Vertical profiles of maximum reflectivity through each of the seven cells that compose the Hana storm. Cell C6 is singled out in boldface.

FIG. 14. Origins (solid dots) and tracks of the seven cells composing the Hana storm. Open circles indicate location at which echo achieved “cell” status. Cell C6 singled out in boldface. Contours are 500 m.
flows that went toward the SE. The result is that for a period the downdraft and the SE low-level flow, both of similar magnitude but nearly opposed to each other, collide and the ensuing updraft continues to feed C6. This contributes to the longevity of the cell. Cell C6 benefits from higher CAPE, and also maintains contact with the inflow, by virtue of its motion away from the ridge. The outflow of C6 also disrupts the flow into the ridge and eventually isolates C7 from its moisture supply.

Clues to the collapse of the storm can be seen with the behavior of the cells seen in the large-scale view from the Molokai radar (Fig. 2). Convective cells intensify, then fade, in a west to east pattern as a function of time. Convection over Oahu intensifies from 1130 to 1330 HST, then weakens thereafter. Convection north of Molokai intensifies from 1330 to 1530 HST, then weakens. The storm over Hana intensifies from 1430 to 1730 HST, but by 1830 HST it also fades. The trough initially provides cooler air aloft that invigorates the cells, but later large-scale subsidence, accompanying the negative vorticity advection, terminates the episode.

f. Performance of numerical guidance

The large-scale Aviation Model (AVN) runs did not provide clues about the event at Hana. It did recognize the presence of the upper-level trough but as is often the case for the lower latitudes the guidance struggled with the direction of motion and the AVN had the trough retrograde away from the state. Precipitation was predicted for a region 300–500 km north-northwest of Maui and the islands remained rain free.

Two smaller-domain, higher-resolution simulations were examined to see if the event could be forecast. The Regional Spectral Model (RSM) has 10-km resolution and includes the entire Hawaiian archipelago (Wang et al. 1998). The large volcanoes Mauna Loa, Mauna Kea, and Haleakala are present, but the gradient and height of the terrain are compromised. The RSM is initialized with the large-scale fields from the NCEP model (T62L28).

The RSM did have partial success in placing rain in the Hana region during the afternoon. At 1100 HST (Fig. 17a) there is an arc-shaped region of precipitation just east of the Big Island that is a manifestation of the barrier effect due to Mauna Kea and Mauna Loa. A convective-scale region of rain separated from the main body of precipitation appears by 1400 HST over Hana (Fig. 17b) and continues for several hours (Figs. 17c,d). Antecedent and during the early stages of the event the simulated surface winds impinging on eastern Maui (Fig. 18) are from the ESE in response to the aforementioned barrier effect of the upstream Big Island. The arc of precipitation is similar to the low-level clouds seen in Fig. 16 and the simulated surface flow matches the ESE flow of 6–7 m s⁻¹ recorded at Hana airport. The rain amounts are, unfortunately, a gross underestimation of the event, and other isolated regions of rain that appeared in the simulation did not

![Fig. 15. Volumetric rainout of the Hana storm. Derived from Kamuela (solid) and Molokai (dashed) radars using the Rosenfeld et al. (1993) Z–R relationship with a hail cap of 51 dBZ. Arrows denote life span of each cell. Again C6 is identified in boldface.](image1)

![Fig. 16. GOES-10 visible image of the region of the main Hawaiian Islands at 2000 UTC (0800 HST), revealing the presence of a bow–wave cloud reaching East Maui.](image2)
occur. The RSM also did poorly with respect to the motion of the main body of precipitation to the north of the islands, tending to retrograde the rain field rather than lifting it out to the northeast as observed (cf. Fig. 17 and Fig. 2). The simulation did not produce flow from the northwest that was observed at Hana after 1600 HST. The partial success of the RSM is due to three factors: recognition of the disturbed low-level flow by the big island located upstream, the lifting of this flow along the southeast slopes of Haleakala, and the presence of conditionally unstable air. Poorly resolved convective-scale motions and a smoothing of the height gradient along the southeast portion of Haleakala resulting in weaker lifting over a wider area are partially to blame for the gross underestimation of the precipitation.

A finer-scale (1-km resolution) simulation with a domain that includes only Maui was run with the mesoscale spectral model (MSM). This model is currently under development at Hawaii. It is a triply nested grid model with the outermost fields supplied by the aforementioned NCEP model, and the intermediate grid determined from the RSM, which does include the entire archipelago. The simulation was initiated at 0200 HST and run for 24 h. The total precipitation over the Hana region generated by this simulation (Fig. 19) was about a factor of 2 low, compared to the actual (Fig. 8), but the fact that the simulation produced heavy rain in the Hana region was gratifying. However, the model produced a significant amount of rain north of Maui and heavy rain on the northeast portion of west Maui, neither of which occurred. The hourly forecast maps manifest substantial problems with the timing and location of the rain. Precipitation is generated in many places where there never is any, and the rain over Hana starts too early and lasts 6 h longer than what was observed. The simulation does not produce any unusual cell behavior, contrary to the observations. A forecaster faced with the hourly guidance would quickly tend to disregard the entire simulation.

Nair et al. (1997), Roebber and Eise (2001), and Li et al. (2003) discuss the need for very fine resolution, detailed initial conditions, or inhomogeneous initializing of the simulation to capture flash flood events. The accuracy of the RSM or other mesoscale models offers clues as to forthcoming conditions, but just as often they mislead. A high-resolution model, tuned for the islands, and capable of ingesting data from new sources such as microwave sensors on satellites, is still in the developmental stages for Hawaii.

![Fig. 17. Precipitation fields generated from the RSM simulation at 1100, 1400, 1700, and 2000 HST. Color scale at bottom delineates mm h$^{-1}$.](image-url)
4. Conclusions

This first Hawaiian application of WSR-88D level II data reveals some intriguing aspects of the storm, but one is forced into a more speculative vein without accompanying surface and upper-air data on smaller spatial and shorter temporal scales.

The presence of an upper-level trough modifies the typical Hawaiian situation with large-scale lifting, the erosion of the trade wind inversion, and the enhancement of instability. The heavy rain episode does not occur near the main body of convective activity. Instead, the interaction of the big island barrier effect with the topography of eastern Maui creates a narrow zone with enhanced uplope and the development of a series of cells that track over the same area for 7 h. The eastern ridge of Haleakala volcano separates the low-level outflows of the mature cells from the genesis region, allowing the storm to have the observed longevity. No cells in the main precipitation region north of the islands achieved a similar life span. The passage of the trough enhances one of the cells mostly via cold-air advection in the middle and upper troposphere. This cell contrasts the others with higher tops, greater maximum reflectivity, and a track that mimics a horseshoe shape. Its outflow, no longer confined by the terrain, spreads to the southeast and collides with the environmental flow, creating a continual inflow that sustains the cell for nearly 4 h. This enables this cell to account for 80% of the rain associated with this storm. The storm collapses when the trough axis passes.

The mesoscale simulations produce a heavy rain event over the Hana region, but the timing and the total amount of rain depart significantly from reality. The simulations do supply supporting evidence for our speculation that orographic lifting and low-level flow modified by the upstream big island are important factors affecting the location of the event.

Nowcasting this episode would demand attention to subtle changes such as the shift in cell alignment, timely RHIs to identify the weak echo region, and keen awareness of the trough axis. The presence of an upper-level synoptic-scale feature alerts forecasters to a possible heavy rain episode, but the radar provides the information to nowcast the location of the event.

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