

Doppler radar investigation of Hawaiian rain¹

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ABSTRACT

Doppler radar measurements in a number of Hawaiian rain echoes indicated a variable amount of convection, with vertical air velocities ranging from unmeasurably small values in some cases to values of 7 m/sec in the more convective cases. The echoes exhibiting weak vertical motions usually had a stratiform appearance when plotted on coordinates of height versus time: contours of signal intensity and Doppler velocity tended to be approximately horizontal. The intensity decreased with altitude in these cases, and vertical gradients as strong as 20 db/km were not uncommon.

In the echoes exhibiting strong vertical air motions, signal intensities, echo depth, and surface rainfall rate tended to be higher. Radar reflectivity factors as high as 10^5 mm⁶/m³ were observed. Although height-time patterns revealed less order than in the echoes with weak updraft velocities, there was a tendency for the leading (upslope) portion of a convective echo to be more active than the trailing portion. As echoes progressed further upslope they tended to become more stratiform in character.

The fact that moderate rain could at times fall from shallow, stratiform clouds, taken in conjunction with the strong vertical reflectivity gradients observed in these cases, indicates a high efficiency of the precipitation process in Hawaiian orographic clouds. The observations in convective echoes clearly demonstrate that convective motions enhance the process.

Introduction

As part of the summer 1965 study of rainfall on Hawaii, Cornell Aeronautical Laboratory investigated the radar structure of rain on the island's windward slopes during a six-week period. The radar was a 3-cm wavelength Doppler set which has been used for several years by CAL in studies of precipitation around Buffalo, New York. Data obtained in a number of shower and uniform rain situations were reduced to give reflectivity patterns and Doppler velocity patterns of the radar echoes. A method was devised to compute the patterns of vertical air velocity on the basis of these measurements. Attempts were also made to infer the distributions of raindrop size in the echoes on the basis of measured Doppler spectra.

This paper includes representative data examples and a discussion of those aspects of the observations which are believed pertinent to the warm rain process. All of the examples presented here were included in a project final

report (Rogers & Jiusto, 1966), but the discussion in this paper has been modified somewhat from that given in the earlier version. In another project report (Rogers, 1966a), problems of radar operation and data interpretation were covered more extensively than elsewhere. An abbreviated version of selected parts of this earlier material was presented at a conference on radar meteorology (Rogers, 1966b). For detailed information about radar techniques and data reductions, the reader is referred to these earlier publications.

Equipment and measurement techniques

The CAL Doppler radar and associated instrumentation were designed for meteorological research and have been in use since 1963. Important characteristics of the radar, as used in Hawaii, are summarized in Table 1. For more details, the reader is referred to Tripp (1964) or Rogers & Tripp (1964). A recent survey of meteorological applications of Doppler radar is given by Lhermitte (1966).

In all the experiments a range gate of 0.5 μ sec duration was used to isolate the signal from a small portion of weather echo for analysis. An-

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Table 1. *Radar characteristics*

Wavelength	3.22 cm
Peak power output	2 kw nominal
Antenna	8-ft paraboloid
Pulse duration	0.5 μ sec
Pulse repetition frequency	5 kc
Beamwidth	0.9°
Receiver dynamic range	40 db

tenna azimuth and elevation angles and range gate position were controlled and indicated at the radar console. Measurements were taken either by fixing the gate in space and permitting echoes to drift through, or by moving the gate out in range at a constant speed. In either case, the echo was essentially scanned over the path traced out by the gate.

The range-gated signal from a weather echo may be characterized by its short-term time average intensity and its Doppler frequency content. A spread of Doppler frequencies is observed in the signal because the meteorological scatterers in the gated volume have velocities slightly different from one another. Each velocity gives rise to a Doppler frequency component with frequency f related to the radial component u of velocity by

$$f = 2u/\lambda,$$

with λ the radar wavelength. The Doppler frequency content of the signal is described by the Doppler spectrum, which indicates how the energy is distributed over Doppler frequency.

The CAL radar is equipped to measure the signal intensity and the Doppler spectrum. When desired, measurements may be performed more rapidly by omitting the measurement of the entire spectrum in favor of only its first and second moments. The first moment of the spectrum is called the mean Doppler velocity, or simply Doppler velocity, and the square-root of the second moment about the mean is called the Doppler standard deviation, or simply Doppler spread. Symbolically, the Doppler spectrum may be denoted by $S(u)$, where $S(u)du$ is proportional to the energy returned in the velocity interval between u and $u + du$. Then the Doppler velocity $\langle u \rangle$ is given by

$$\langle u \rangle = \int u S(u) du / \int S(u) du.$$

The Doppler spread σ_u may be defined by a similar expression involving the second moment of $S(u)$. For the special case of vertical viewing, the Doppler velocity is denoted by W . It is convenient to measure $S(u)$ on an ordinate scale such that the total signal energy, as indicated by the signal intensity I , is equal to the area beneath the spectrum function. That is,

$$I = \int S(u) du.$$

An AGC circuit is used in the receiver of the CAL radar to insure that all incoming signals lie within the dynamic ranges of the receiver and the analog signal processing instruments. The AGC voltage is recorded continuously and, by means of a calibration with a signal generator, provides a measure of signal intensity, I . Because of the wide variation of signal intensity, it is standard practice to refer measurements to a logarithmic scale. The measured quantity, denoted by P , is related to intensity by $P = 10 \log_{10} I$ with I in milliwatts. Doppler velocity is measured with a frequency tracker (Rogers & Tripp, 1964), a discriminator circuit which generates a voltage proportional to $\langle u \rangle$. A portion of the received signal is subjected to linear detection and thence passed through an R-meter (Rutkowski & Fleisher, 1955) to provide an indication of σ_u , the Doppler spread.

The quantities I , $\langle u \rangle$, σ_u , and the range gate position were automatically recorded as functions of time on a four-channel Sanborn strip chart recorder. An example is given in Fig. 1 of the Sanborn record resulting from a range scan with the antenna pointing vertically. Most of the data were obtained in this form, namely as height profiles. The profile in this example begins at a height of 600 m above the radar which was the minimum usable range, and extends to the echo top, in this case 2.3 km, where the return signal falls essentially to the noise level. The intensity decreases from a maximum value of about -70 dbm at the closest range to a value below -100 dbm at echo top. Part of this decrease of intensity with altitude arises from the r^{-2} fall-off of return power associated with the divergence of the beam. This fall-off was always compensated for in data reductions. The Doppler velocity is seen to be negative at all altitudes (corresponding to motion *toward* the radar or downward),

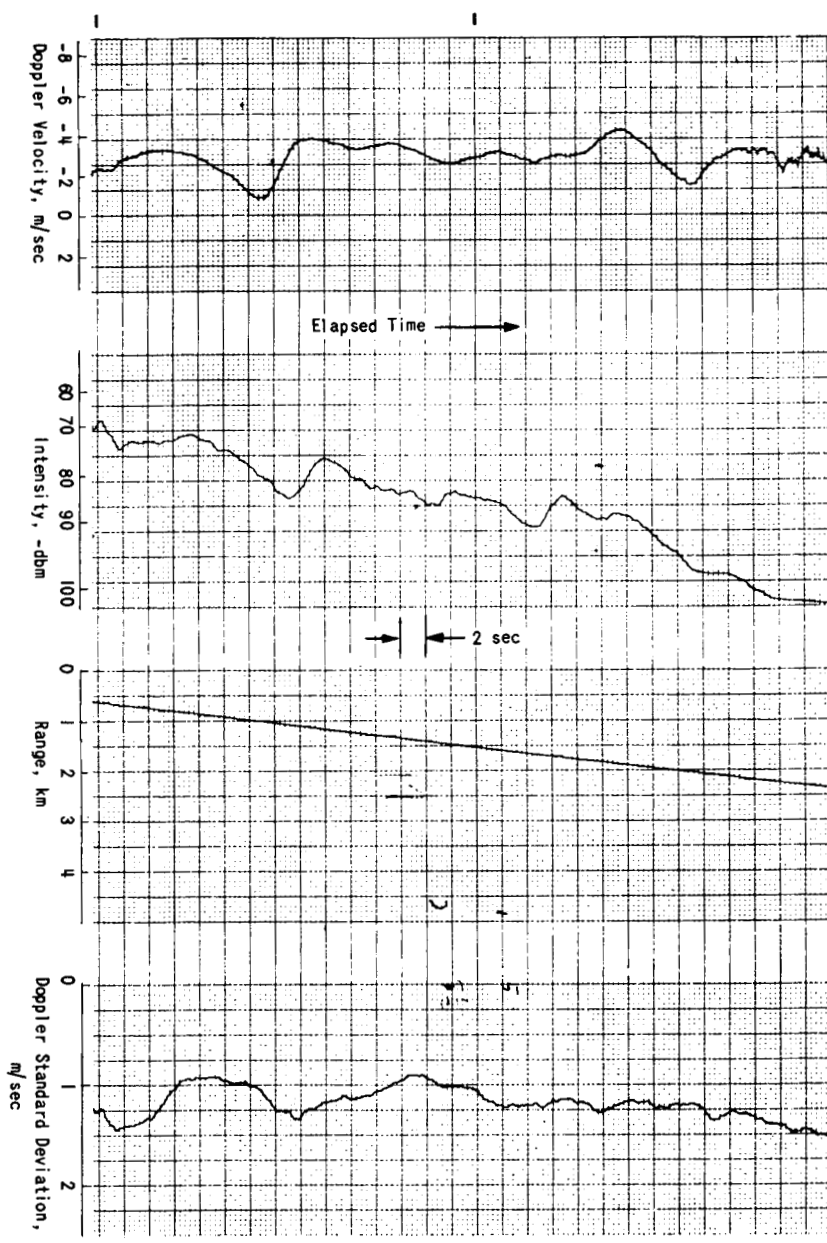


Fig. 1. Example chart record of a height profile.

varying between approximately -1 and -4 m/sec. Owing to the influence of receiver noise, the frequency tracker does not operate properly for signals weaker than about -95 dbm. It can be seen on the record that the velocity trace becomes noticeably noisy at altitudes

above about 2.0 km, where the intensity first fell below -95 dbm. To preclude the possibility of erroneous velocity data, Doppler velocity measurements were not used in the analysis unless the signal was stronger than -95 dbm. The fourth channel on the example record

shows that the Doppler spread varies between 0.9 and 1.4 m/sec in an irregular manner. As a consequence of equipment design, the R-meter is subject to the same noise limitations as the frequency tracker and Doppler standard deviations were not used in analysis when signal intensities were below -95 dbm. An inspection of the recorder time scale shows that the measurement of this profile required approximately 1 min and that the gate speed was 30 m/sec.¹ The time constant of each of the four channels was approximately 1 sec. Antenna directions and other remarks or observations were entered by the operator directly on the chart record.

In addition to the data recorded on strip charts, complete Doppler spectra of signals from selected locations were recorded when conditions were appropriate. A high-resolution spectrum analyzer, built to CAL specifications by Raytheon, Inc., was used for this purpose. This instrument consists of an array of 250 highly tuned, contiguous filters of 4 cps bandwidth, providing a total analysis band of 1000 cps or, in terms of Doppler velocity, of 16.1 m/sec. The output of each filter is passed through a linear detector and integrated for 3 sec. A commutator operating at a frequency of 3 cps samples the outputs of each detector consecutively, synchronizes the trigger on a Tektronix 585 oscilloscope, and provides a spectral display on the scope. The position of zero velocity within the 16.1 m/sec range is arbitrary, and was usually offset so that a wider range of negative (approaching) velocities could be accommodated than positive (receding) velocities. This analysis band was found to be sufficiently broad for almost all of the vertically-viewed spectra encountered in Hawaii. An example of a spectrum, for vertical viewing, is given as Fig. 2.

With reference to the example, it is noted that the detected and integrated output of each filter appears as a dot on the oscilloscope display. The spectrum function is thus displayed as a discrete function of velocity (or of Doppler frequency). Although the gain of each filter can be individually controlled so that, in principle, the filters can be adjusted to have the same response, it is not feasible, in practice, to keep the gains exactly uniform. Consequently,

¹ At other times, a speed of 100 m/sec was employed to reduce the time requirement for data acquisition.

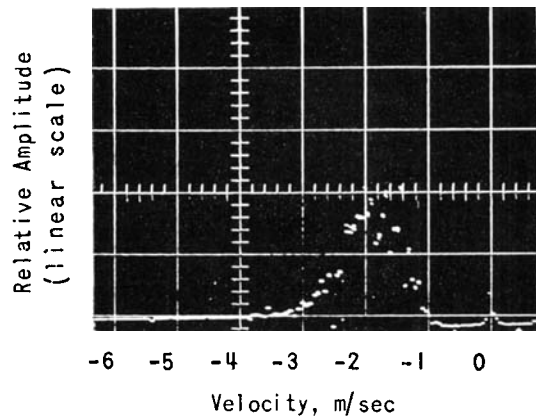


Fig. 2. Example of Doppler spectrum for vertical viewing.

some irregularity is introduced into the spectrum by the small differences in gains of the different filters. Furthermore, a few of the filters behaved erratically in Hawaii, always reading too high or too low, and could not be repaired for lack of parts and proper test equipment. Therefore, some smoothing of the displayed function was justified and indeed necessary before any spectral computations were made.

In the example shown, the return signal is seen to have spectral content at velocities ranging between approximately 1 and 3.5 m/sec toward the radar (downward). On the assumption that updraft velocities were small compared to the terminal fall velocities of the raindrops, the spectrum indicates the presence of drop sizes ranging from approximately 0.25 mm to 1 mm diameter. Relationships between the spectrum and the functional form of drop-size distribution are discussed in a later section.

The measurement accuracy of this radar has been discussed previously (Rogers & Tripp, 1964). Calibration procedures used in Buffalo enable the measurement of signal intensity to within ± 3 db and mean Doppler velocity to within ± 0.2 m/sec. This accuracy was not possible in Hawaii because the power source, a diesel generator, was not capable of supplying the steady and continuous service required. To minimize errors, calibrations were performed for the velocity-measuring equipment at least once every day that data were taken and often

before every series of runs. Additionally, relative calibrations of receiver AGC voltage were conducted on a daily basis. Absolute receiver calibrations were obtained by means of a standard target at the beginning, middle, and near the close of field operations. On the basis of the care taken in calibration procedures and the variations observed to occur in calibrations, it is estimated that the mean Doppler velocity was measured to within 0.5 m/sec and that signal intensity was measured to within 4 db. On a given day the relative intensity measurements were more accurate, probably to within 3 db.

Observational procedure

The radar was situated at the so-called nine-mile site, located several hundred yards north of Saddle Road about 9.5 mi from Hilo.¹ Elevation of the site was approximately 750 m (msl) from which a relatively unobstructed horizontal view was possible in directions ranging from north to southeast. Horizontal range was limited in other directions by Mauna Kea and Mauna Loa. The site was chosen because of this unobstructed view to windward and because, on the evidence of existing reports (e.g. Mordy, 1957), it was expected to be at the base of typical Hawaiian orographic clouds and in the area of maximum rainfall. It served also as the site for most of the Japanese equipment and observations. From July 10 through August 24, during which period a tipping-bucket raingage was operated at the site, nearly 16 inches of rain were recorded.

Data on weather targets were obtained on 19 days between July 15 and August 23. Data from 6 of these days were not reduced or analyzed because the echoes were either very weak or distant. Consequently, the radar analysis was based on 13 days' data. Weather types encountered ranged from steady orographic drizzle, which seemed to be fairly characteristic of precipitation at the site, to convective showers and associated heavy rain.

The equipment was kept on standby basis on days that showed promise of rain. When rain occurred at the site, the radar was used almost exclusively for vertical measurements,

as updrafts and drop sizes were the principal quantities of interest. Occasionally, measurements were taken at several azimuths at low elevation angles in order to determine the direction and speed of movement of the echoes and to look for indications of convective activity. Particularly in those cases when convective showers were moving upslope toward the radar, attempts were made to measure horizontal velocities and signal intensities in the echoes until they reached the radar site, at which time vertical measurements would commence.

Once rain reached the site, a series of several vertical profiles was usually taken as a preliminary investigation of the echo. From these could be determined the echo height, the general strength of the shower (as indicated by signal intensity) and, to some extent, the degree of convectivity (as indicated by the magnitude and the variability of the Doppler velocity). If the shower was found to be only mildly convective, the height profiles would be interrupted by occasional sequences of Doppler spectrum photographs at different altitudes. For the cases that appeared to be at least moderately convective, spectrum photographs were generally not taken and uninterrupted height profiles were continued. The procedure was different for the two cases because significant time variability in the measured quantities might occur in the convective case and be unobserved during the time required for a sequence of spectrum photographs; furthermore, Doppler spectra can not be used to obtain reliable estimates of drop sizes in convective situations, as will be discussed later in the section on data analysis. Most of the data that were reduced for analysis, therefore, were in the form of height profiles of signal intensity, Doppler velocity, and Doppler spread, occasionally with interspersed Doppler spectrum photographs.

Examples of data

(a) *Height-time structure of echoes*

To facilitate the examination of a series of height profiles, height-time ($h-t$) patterns were constructed for those series which were judged to be worth detailed analysis. Data were read off individual profiles (such as the example in Fig. 1) at height intervals of 100 m and plotted on coordinates of height versus time.² Isopleths

¹ A map of Hawaii showing the project site is given as Fig. 1 in the paper by Lavoie in this series.

² The vertical resolution of the radar was 150 m.

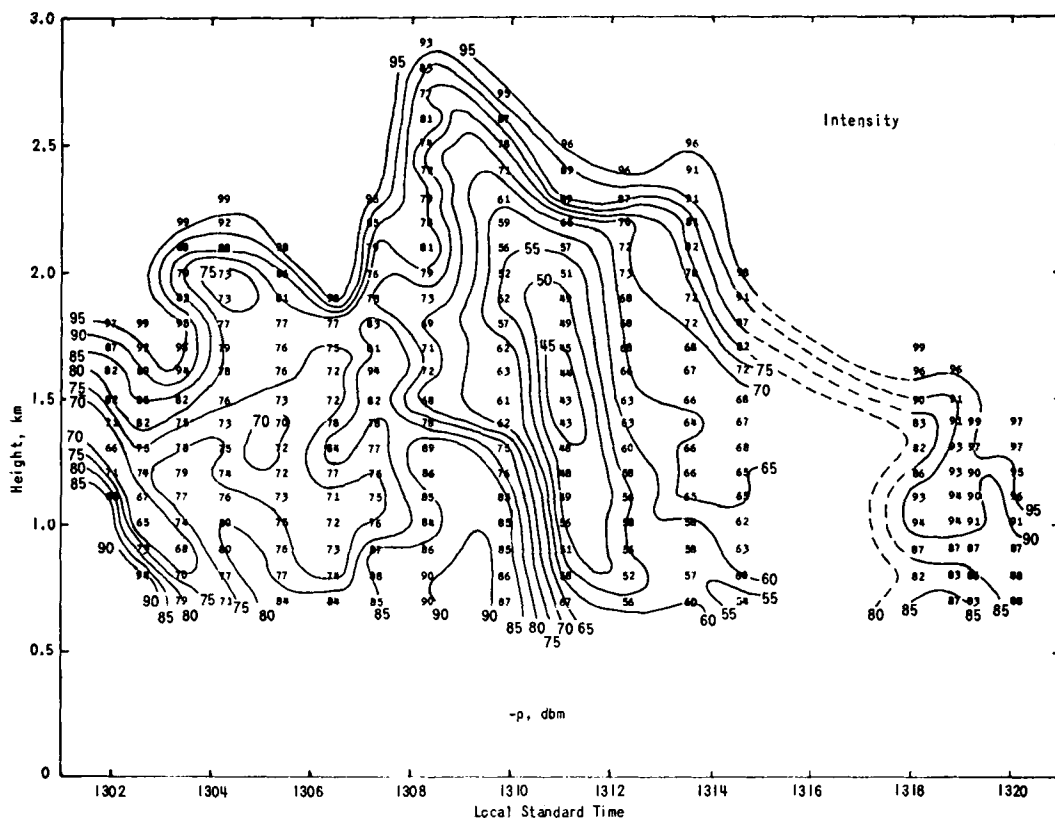


Fig. 3a. Height-time pattern of signal intensity, August 9. Values in $-dbm$, normalized to 1 km.

were then constructed to provide an $h-t$ pattern of the measured quantities for a given series of profiles. Such a pattern only indicates the structure of an echo in the measurement coordinates of height and time and cannot, in general, be interpreted as either a vertical space-section through the echo or as a time sequence of echo evolution. It reflects both time and space variations of echo structure. Interpretation is further complicated when cloud elements at different levels move with different directions. Problems of interpreting this kind of data presentation are considered in some detail by Miles (1956). Height-time patterns are used here as a convenient means of displaying the data; it is not necessary that they be interpreted exactly as vertical sections through the echoes. Several representative examples are given in this section. They are listed by date of observation.

(1) *August 9.* On this day the most intensely convective showers that were observed passed

over the radar site. In the early afternoon, individual convective cells began to appear windward and to move toward the site. The horizontal Doppler velocities measured in the approaching echoes were approximately 6 m/sec, a somewhat higher value than ordinarily observed. Just before 1300 (local standard time) an echo was observed to be very close and moving directly toward the site. Height profiles were started at 1302 and continued for 18 min while the echo passed overhead. The resulting $h-t$ pattern of signal intensity is shown in Fig. 3a. All of the data points—signal intensities in $-dbm$, normalized to a range of 1 km—are entered to give an idea of the measurement grid. Contours are drawn for 5 db intervals and smoothed to a degree which is compatible with the estimated data accuracy, namely ± 3 db in intensity and ± 50 m in height. The data are entered, for convenience, on vertical lines corresponding to the starting times of the height profiles, al-

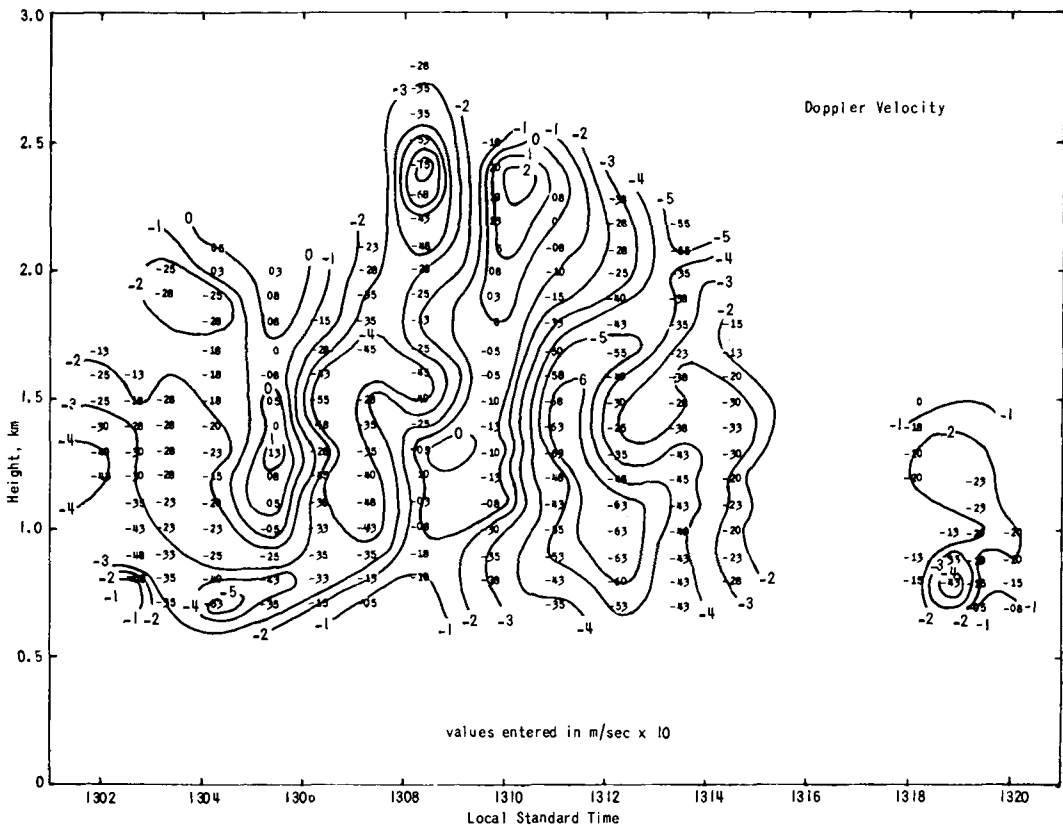


Fig. 3b. Height-time pattern of Doppler velocity, August 9. Contour values in m/sec, positive upward.

though in fact the data should be entered on slightly tilted lines to indicate the actual paths of the gate through space-time coordinates. The effect of this small tilt on the h - t patterns is insignificant, so that the use of vertical lines to represent the approximate paths of the gate is justified.

The pattern began with echoes aloft extending from 1.1 to 1.8 km. By the fourth profile (approximately 1304), the echo had descended to the minimum usable range, in this case 700 m.¹ By this time also the echo top had extended to 2.2 km. Maximum echo height of nearly 3.0 km occurred at 1308 and maximum intensities of about -45 dbm were observed three minutes later extending from about 1.4 to 1.7 km. Intensity and echo top diminished

gradually after this time, essentially disappearing by the last profile at 1320. The three-minute data gap beginning at 1315 was caused by malfunction of a recorder pen. The pen was repaired and profiles resumed, as shown, at 1318. When an appreciable gap occurred in any of the data, arising either from equipment problems or from the time required for a sequence of spectrum photographs or horizontal measurements, the gap is indicated by dashed lines on the h - t patterns.

With each value of normalized signal intensity in Fig. 3a can readily be associated a radar reflectivity factor, Z . For any (linear) radar system, the range-normalized intensity p is related to Z by

$$p = C + 10 \log_{10} Z.$$

For our radar $C \approx -95$ db when p is in dbm normalized to 1 km and Z is in mm⁶/m³. Thus, for example, an intensity value of -95 cor-

¹ Ranges closer than 600 or 700 m were not usable because the return from ground targets in the antenna pattern sidelobes became as strong as the direct return from weather echoes.

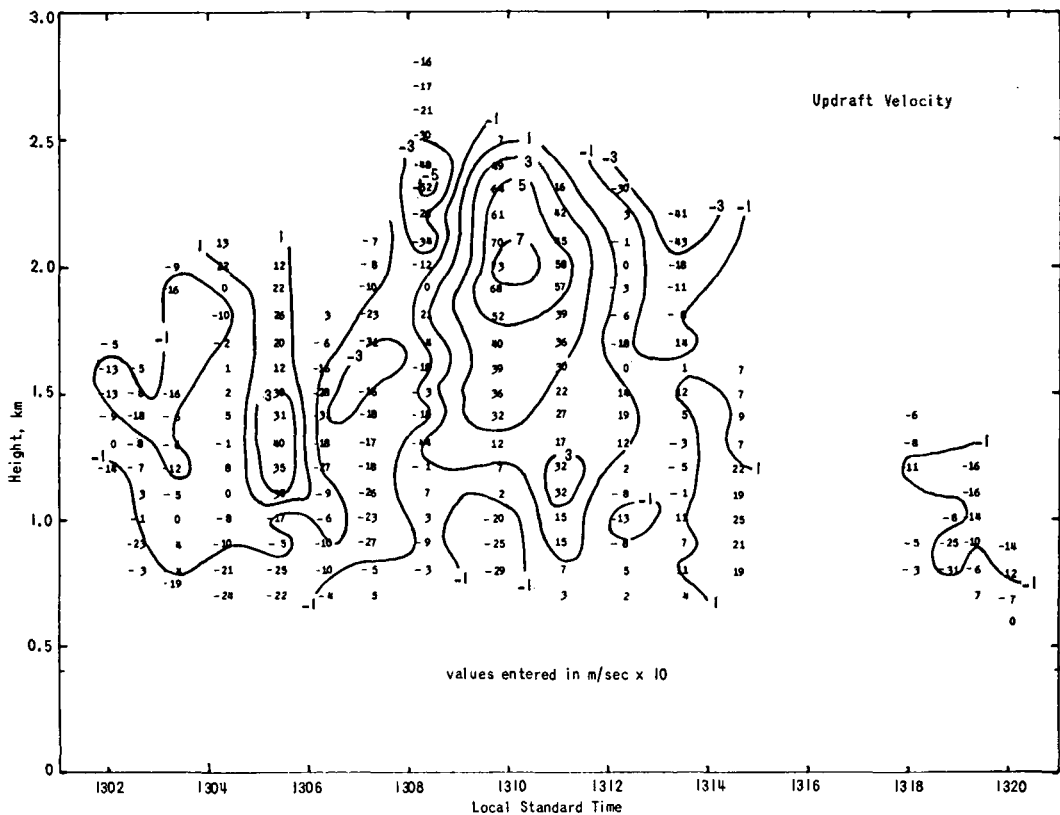


Fig. 3c. Height-time pattern of computed updraft velocity, August 9. Contour values in m/sec, positive upward.

responds to $Z = 10^4$; $p = -85$ corresponds to $Z = 10^4$, etc. The maximum intensities in this pattern thus correspond to Z values of about 10^5 which are typical of reflectivities for a continental cumulonimbus cloud. (A criterion for hail in New England storms has been found, for example, to be Z values of $10^{5.5}$ or higher (Geotis, 1963).) To one familiar with radar characteristics of mainland storms, the observation of such strong signals in an echo only 3 km deep is surprising. Although the explanation of this observation rests in the difference between mainland and Hawaiian droplet populations, as discussed later, it is worth emphasizing that the showers of August 9 were not typical of Hawaii; echo tops and signal strengths were both very high on this date.

There is evidence that the shower of Fig. 3a was in an early stage of development. No rain was observed at the ground until 1312, approximately 10 min after the profiles were

started. This suggests that updrafts or turbulence were responsible for preventing the highly reflective raindrops from reaching the ground, a point borne out by the Doppler data discussed next. During the 18 minutes of echo passage, the raingage bucket did not tip, indicating that less than 0.01 in of rain fell.

The Doppler velocity $h-t$ pattern for this echo is shown in Fig. 3b. As in Fig. 3a, the data points (in units of m/sec) are entered to indicate the measurement grid and the degree of smoothing employed in drawing contours. Notice here that it was not feasible to fill in the analysis during the 3-min gap in data, as was done for the intensity pattern. By the fourth profile (1304) a positive Doppler velocity had appeared at the echo top. On the next profile, positive velocities were present from 1.8 km to echo top and in the interval 1.1–1.5 km. The positive velocities disappeared by 1306 and the strongest negative (downward) velocity, 7.5

m/sec, occurred soon thereafter at 2.4 km. Positive values appeared again at high levels at about 1310, but were present only for two profiles. From then on, all Doppler velocities were negative. Obviously the Doppler velocity pattern is related to the updraft velocity pattern. In the absence of updrafts, drops fall at terminal velocity and the Doppler velocity is negative with a magnitude depending upon drop sizes. Positive Doppler velocities are a certain indication of updrafts, but a one-to-one association cannot be established between Doppler velocity and updraft velocity without additional information about terminal fall velocities, and hence drop sizes. The signal intensity provides such additional information; for, in the absence of updrafts, large signal intensities are associated with large negative velocities. The h - t pattern of updrafts can be inferred on the basis of measured intensity and Doppler velocity. In brief, a given radar reflectivity implies a hypothetical Doppler velocity, W_0 , which would be observed in the absence of updrafts. Updraft velocity w is estimated by determining W_0 from the intensity measurement and subtracting this value from the measured Doppler velocity:

$$w = W - W_0.$$

The relationship between W_0 and reflectivity depends upon drop-size distribution. The most suitable relationship for the case of Hawaiian rain was found to be the semiempirical formula¹

$$W_0 = -0.46 Z^{1/3},$$

where Z is the radar reflectivity factor in mm^6/m^3 , and W_0 is in m/sec. The derivation of this equation and several justifications for its use are discussed in the references cited in the in-

¹ This formula should be termed semiempirical because of the way it was derived. Basically, it is the relationship between Doppler velocity and reflectivity factor that would be observed if the raindrops were essentially all less than 1.2 mm in diameter, if their sizes were distributed according to an exponentially-decreasing law, and if the liquid water content were 1 gm/m³. To that extent it is a theoretical formula for this special population of raindrops. The empirical element enters because the formula was found to provide a good fit to the actual measurements of Z and W in those echoes having stratiform character, where updraft velocities were expected to be negligibly small.

roduction above. It is estimated that the updraft velocities computed by this procedure are accurate to within ± 1 m/sec most of the time.

The updraft estimates computed by this means from data in Figs. 3a and 3b are shown plotted in Fig. 3c. The two regions of high-level updrafts are the most prominent characteristics of this pattern. The first occurred at about 1306; the second and strongest at about 1310. Both of these narrow updraft regions were bordered by relatively broader, but weaker, downdraft regions, suggesting a system of updrafts and compensating downdrafts. It is noteworthy that the second updraft region corresponds very closely in position to the intensity maximum in Fig. 3a. This region of updrafts contained the highest updraft value observed in any of the Hawaii data, namely 7 m/sec. This is an appreciable updraft, even for continental cumulus clouds, although Battan & Theiss (1966) in Arizona & Rogers (1963) in New York have observed updrafts of 20 m/sec in tops of cumulonimbus with Doppler radar. Fig. 3c raises the question of where was the origin of the air in the updrafts. There is no indication in the data that this air came from below, for the only appreciable low level updrafts occurred at about 1314. This same pattern—stronger updrafts at high levels than at low levels—has been found to characterize the structure of continental convective echoes. Possible explanations are mid-level entrainment as proposed by Easterbrook (1967) or updrafts that are tilted with respect to the surface of measurement, as suggested by Battan and Theiss.

In summary, the shower analyzed in Figs. 3a-3c was a highly convective one, containing some of the strongest signals and largest updrafts observed in Hawaii. On the evidence of meager rainfall at the surface and vigorous updrafts in the echo, it is suggested that this shower was in the early stages of development.

There were numerous distinct shower echoes in the vicinity of the radar on the afternoon of August 9. About a half hour after the passage of the echo discussed above, a second sizable echo, located about 6 km to windward, was observed to be heading toward the radar site. Several range scans through this approaching shower were made at elevations between 6° and 30°. Horizontal Doppler velocities were approxi-

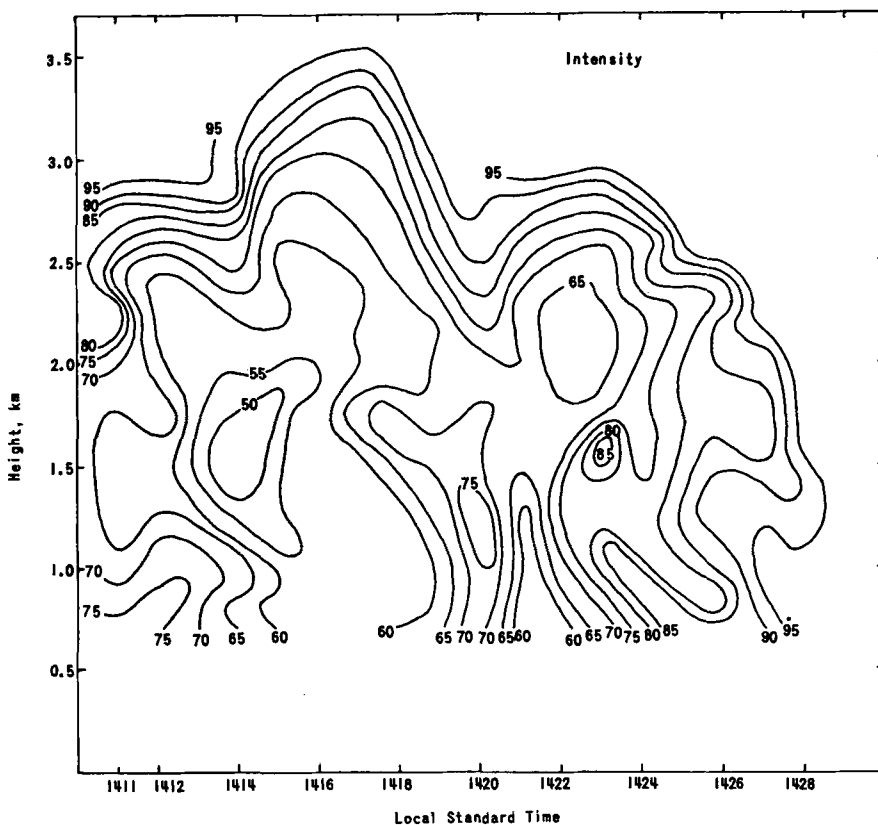


Fig. 4a. Height-time pattern of signal intensity, second series August 9. Values in -dbm, normalized to 1 km.

mately 4 m/sec, and somewhat variable throughout the echo, suggesting irregular motion of the horizontal wind. From the first horizontal probing until the appearance of the initial overhead echo for this shower, 28 minutes elapsed and the echo moved 5.7 km. The rate of movement was thus approximately equal to the average horizontal Doppler velocity. It is likely that the shower, as it passed overhead, was somewhat different from the one first scanned horizontally. The movement and general behavior of this and many other convective showers suggested that the echo, as a relatively distinctive entity, was not simply being carried along by the wind, but rather was continually changing in a manner consistent with generation on its leading edge and dissipation behind. However, because of the continuity of movement over the half hour required for the echo to reach the radar site after initial observation it seems evident that the organized convective

system responsible for the echo was moving more or less with the wind.

The $h-t$ pattern of intensity for this second shower of August 9 is given in Fig. 4a. The data consist of 13 height profiles, fairly regularly spaced on the time axis. For clarity the data points are not entered on this or subsequent $h-t$ patterns. The echo top, reaching about 3.6 km at 1417, was higher for this shower than for any other observed in Hawaii. The signal intensity maximum, occurring at 1414 at an altitude of about 1.5 km, corresponds to a Z value of about $10^{4.5}$ and was not quite as strong as the maximum signal in the first echo example. In Fig. 4b is shown the associated $h-t$ pattern of Doppler velocity. The data on these two figures were used to construct the pattern of updraft estimates shown in Fig. 4c. The strongest updrafts occurred at mid-levels on the leading edge of the shower, with a maximum value of about 5 m/sec at 1.8 km. The region of

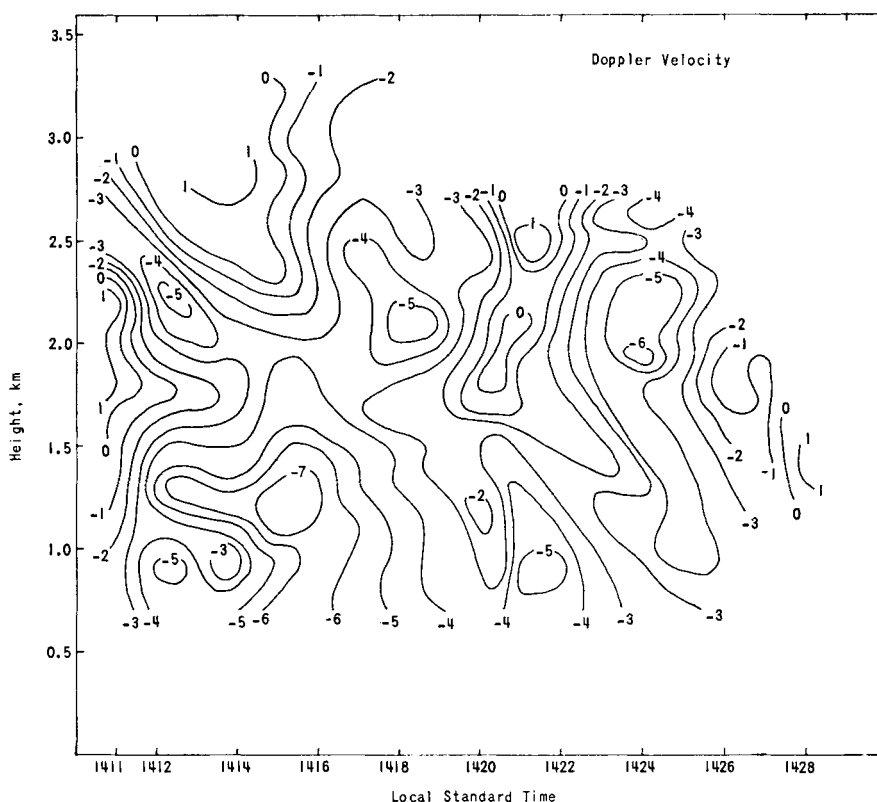


Fig. 4b. Height-time pattern of Doppler velocity, second series August 9. Values in m/sec, positive upward.

updraft maximum moved upwards over the next 4 min, being centered at 2.4 km at 1415. A small region of high level downdrafts followed the passage of the highest part of the echo and was soon replaced by a broader and stronger region of updrafts. Immediately after passage of another apparent convective center at 1423, downdrafts again appeared at all levels. These data clearly indicate that updrafts were associated with the leading edge of the shower and downdrafts with the trailing, dissipating edge; the shower apparently consisted of at least two of these simple convective systems. As in the previous example, there is a correspondence in locations of intensity maxima and updraft maxima.

Rainfall data at the surface suggest that this shower was in a further stage of development than the shower of the first example. Relatively large raindrops (some exceeding 1.5 mm diameter) were falling at the time of the first height profile and heavy rain continued during

shower passage. A total rainfall of 0.15 in occurred during the 18 min required for passage. Although remarkably steady, the rain was somewhat heavier at the front of the shower than at the rear. This shower was quite similar to the first one, but being taller and with weaker updrafts, it apparently permitted raindrops to grow to large size and to fall out rather than to be suspended by strong updrafts.

The Doppler spread data (not shown) suggest that turbulence as well as updrafts may have been connected with the propagation of this shower. All spreads of 1.6 m/sec or greater occurred in the first 6 minutes of data, while almost all spreads of 1.2 m/sec or less occurred later. Since small scale turbulence is one of the contributors to Doppler spread, this observation is consistent with the presence of a greater degree of turbulence in the generating portion of the shower than in the dissipating portion.

(2) August 5. In the early morning weather was clear over Hilo and the mountain slopes,

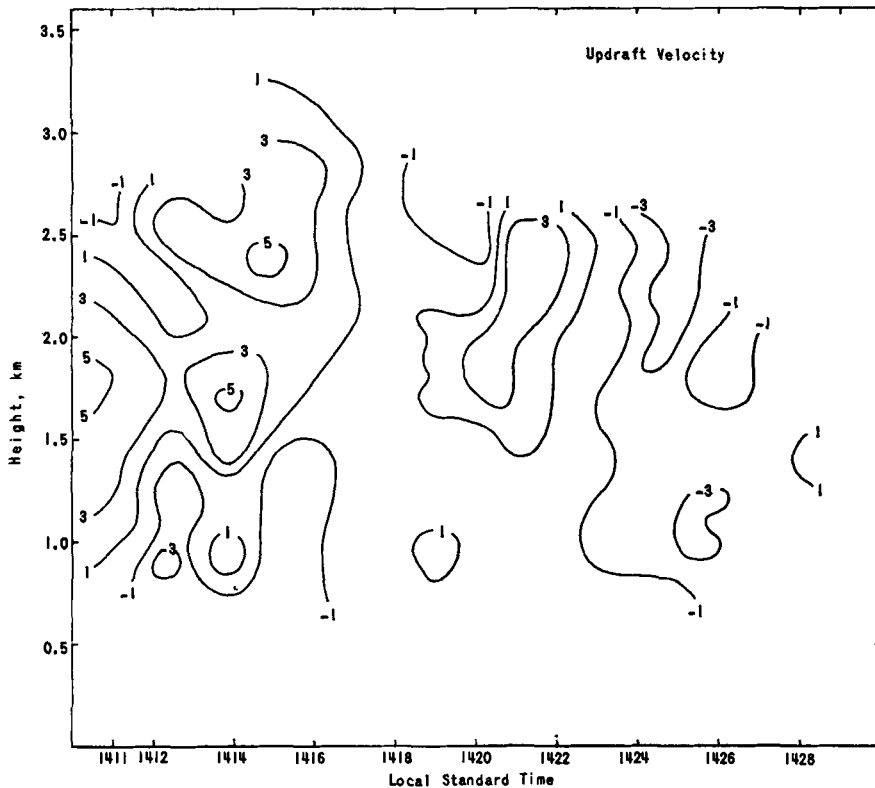


Fig. 4c. Height-time pattern of computed updraft velocity, second series August 9. Values in m/sec, positive upward.

but, as was often the case, it rapidly became cloudy by mid-morning and there was light to moderate rain all afternoon and evening at the radar site. During the late morning a variety of weak echoes was in the vicinity and spot measurements showed that horizontal Doppler velocities were 3 m/sec or less from the northeast. A very small echo, probably a single convective cell, was noted at about 1100 to be moving directly toward the radar. Vertical measurements were begun at 1114 and the resulting $h-t$ patterns of intensity and Doppler velocity are shown in Fig. 5. This shower was apparently dissipating, for it contained no measurable updrafts, i.e. no updrafts as large as 1 m/sec. Only a very light rain, consisting of many small drops, fell at the surface. Total accumulation was less than 0.01 in.

Initial signal maxima, located at about 1.6 km, corresponded to Z values of $10^{1.5}$. The intensity contours are seen to descend with time as the shower simultaneously dissipated

and moved out of the beam. The pattern here has the character of a stratiform echo although the shower was convective originally. It was frequently observed that echoes of an apparent convective nature downslope became progressively more spread out and stratified as they moved up to the nine-mile site and beyond. This example is believed to be a good case in point. The upslope dissipation might be associated with the squeezing effect created by the rising ground below and the quasi-horizontal inversion above, which restricts the vertical extent of clouds and thereby limits the time available for droplet growth in a convective trajectory. It may be significant that on August 9 the inversion amounted to only 1°C while on August 5 the inversion was 5°C . On both days the inversion base was about 3.5 km above Hilo, or about 2.8 km above the radar site. Thus the inversion was penetrated by the active showers on August 9 and drops were permitted to attain relatively large sizes, while

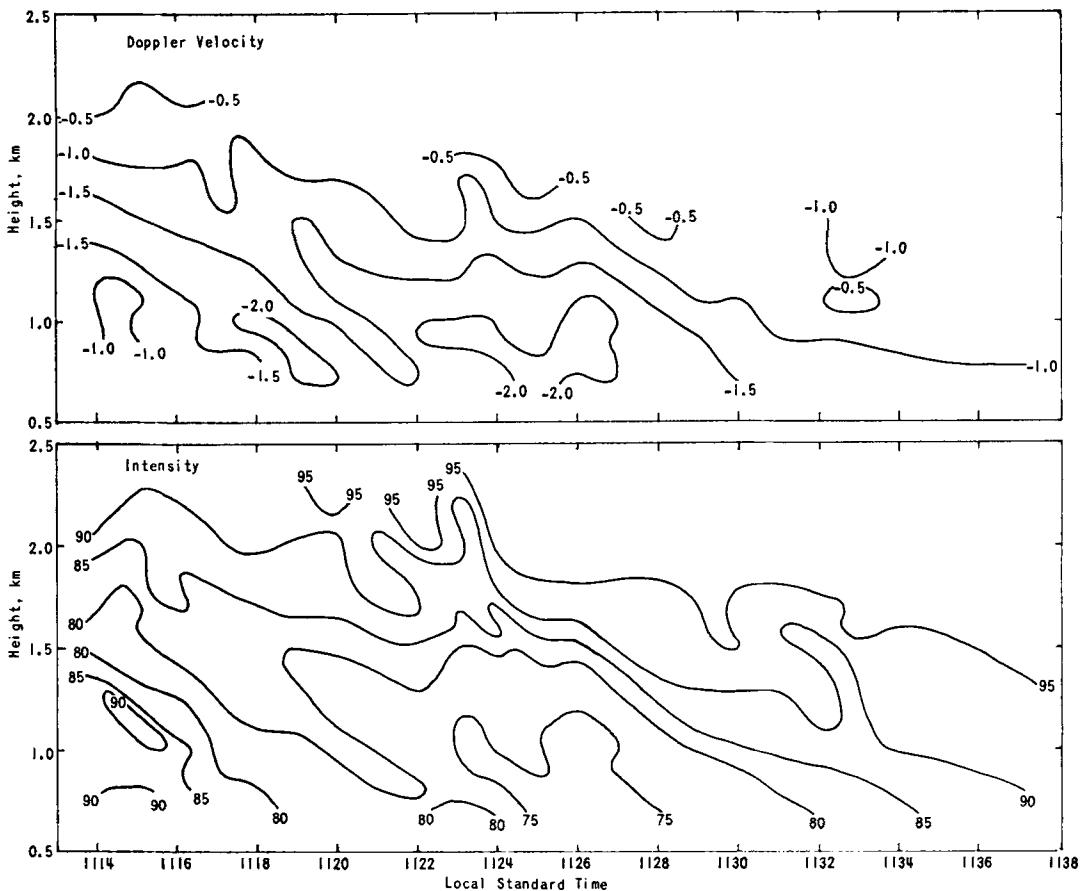


Fig. 5. Height-time pattern, August 5. Intensities in $-dbm$, normalized to 1 km. Velocities in m/sec , positive upward. The vertical air velocities inferred for this case were less than 1 m/sec in magnitude throughout the echo extent and have not been plotted.

the strong inversion of August 5 may have suppressed convection and contributed to the upslope dissipation of echoes.

In part because the inversion was usually strong, there was a tendency for most of the rain at the radar site to be in the form of rather small drops falling from stratiform clouds or from convective clouds that had been spread out. A fairly typical example of this kind of situation was obtained later in the day on August 5. Light rain and drizzle prevailed all afternoon and at about 1600 a system of strong echoes over the bay was observed to be moving inland. Horizontal profiles through the echoes indicated strong, irregular horizontal winds as they moved onshore. These radar observations were confirmed, incidentally, by the report of strong gusts at the University campus in Hilo.

This system of echoes underwent considerable modification in the half hour required to reach the nine-mile site. In Fig. 6 are displayed the $h-t$ patterns of signal intensity and Doppler velocity as the system moved overhead. Up until this time, there had been very light rain at the ground from weak echoes extending to about 1.5 km. These echoes are still present in the first few minutes of the intensity pattern. The arrival of the new system is indicated by the increase in echo height occurring at about 1626. Rainfall at the ground suddenly increased and 0.07 in fell between 1625 and 1640. From 1640 until the end of the pattern, only 0.01 in of additional rain fell. The maximum signal intensities of about $-65 dbm$ ($Z \approx 10^3$) occurred in the lowest part of the echo at 1633. This presence of maximum signal at the lowest altitudes

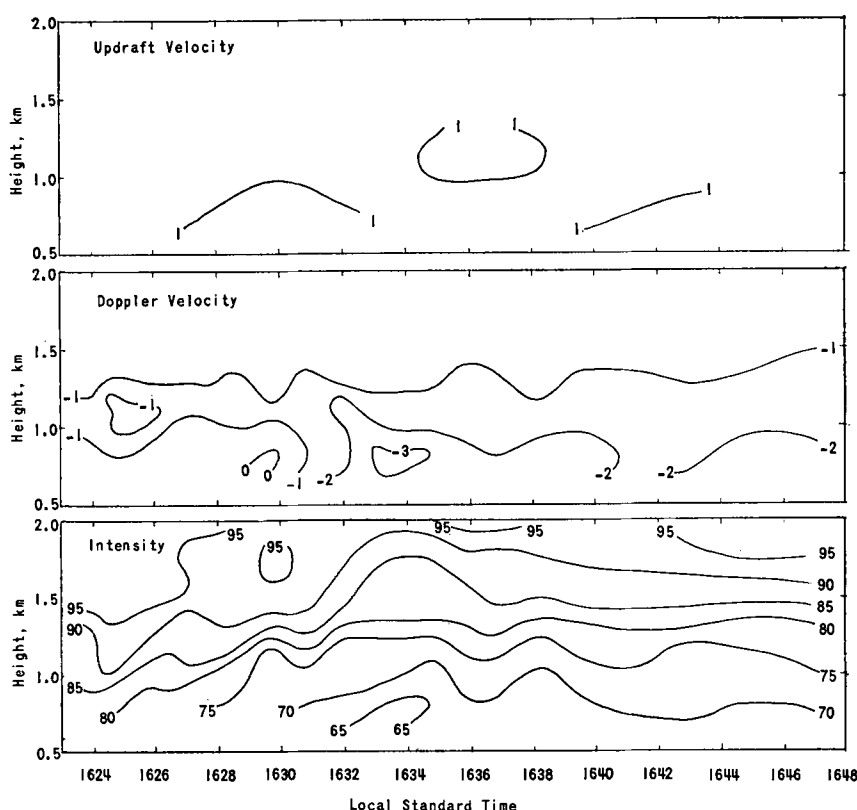


Fig. 6. Height-time pattern, second series August 5. Intensities in - dbm, normalized to 1 km. Velocities in m/sec, positive upward.

is indicative of the stratiform character of the precipitation. Maximum downward Doppler velocities of 3 m/sec occurred at the same place. From 1640 onward, the intensity pattern is extremely uniform, showing a steady fall off of reflectivity with altitude of about 20 db/km. Updrafts were light, if detectable at all.

The horizontally-stratified pattern, with intensity and downward Doppler velocity increasing progressively with downward distance from echo top, is consistent with a continuous precipitation mechanism in which small raindrops near the echo top slowly descend through cloud and smaller raindrops, growing by accretion and coalescence. Such a process results in an increase in average drop size with distance fallen, with corresponding increases in signal intensity and Doppler velocity. In a steady state condition, a vertical gradient of reflectivity is established by this process. Relations between the precipitation growth mechanism

and the steady-state reflectivity profile have been considered by Wexler & Austin (1954) for the case of snow and by Rigby *et al.*, (1954), Hardy (1963), and others for rain. In no case have theoretical or experimentally observed gradients been as large as the 20 db/km value noted in Hawaiian stratiform rain. The strong reflectivity gradient thus might be a unique characteristic of Hawaiian-type rain; it is discussed further in a subsequent section.

(3) *August 12.* Approximately an inch of rain fell on August 12. Half of this occurred between 0700 and 0800 and most of the remainder fell after 1830; in the period between there was continuous light orographic rain at the radar site. In Fig. 7 is presented an *h-t* pattern beginning at 1829, when the light rain had just changed over to moderate intensity. Being similar to the latter case discussed from August 5, this pattern is fairly typical of the many situations in which a system of showers, in being

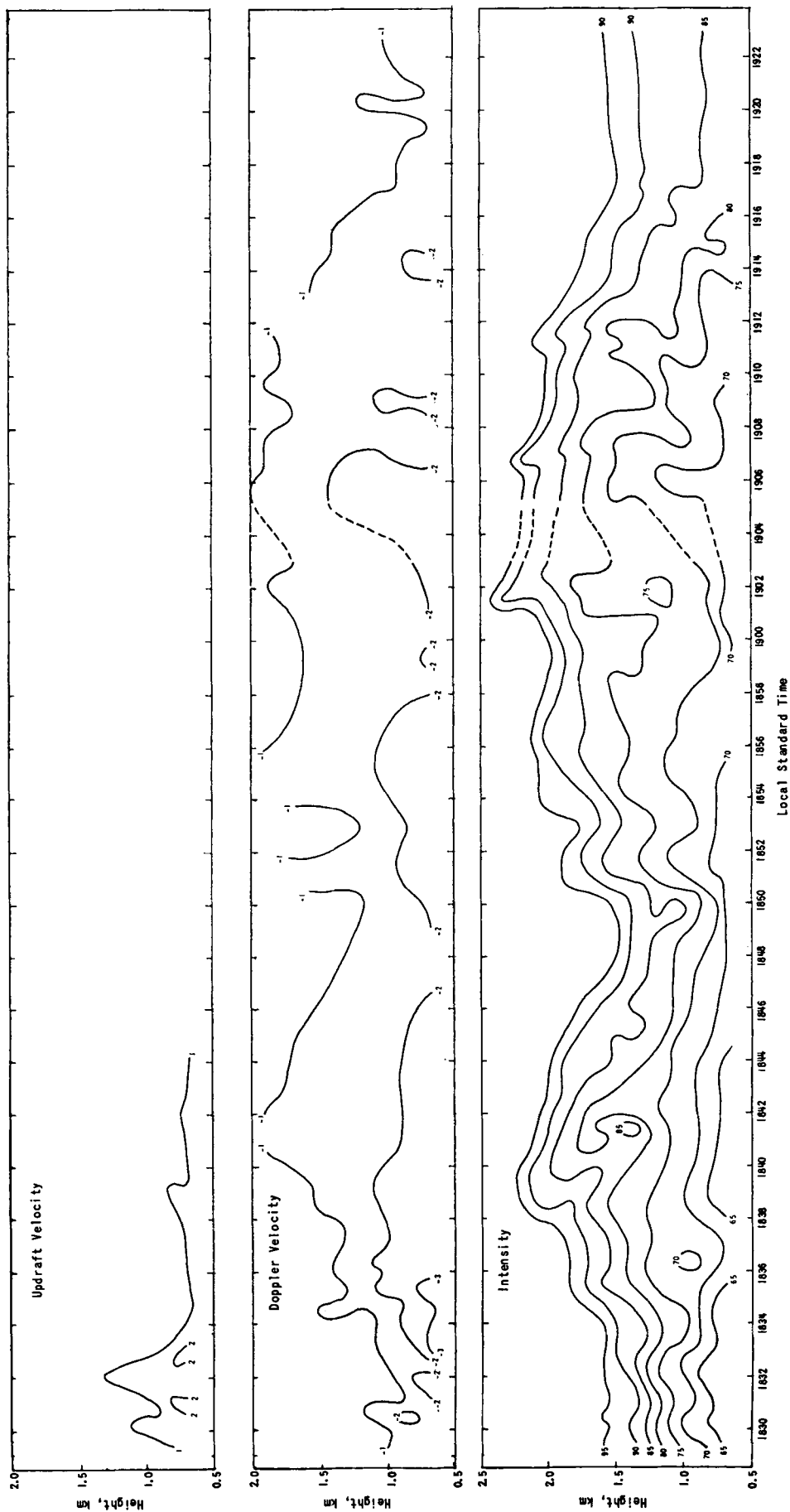


Fig. 7. Height-time pattern, August 12. Intensities in -dbm, normalized to 1 km. Velocities in m/sec, positive upward.

squeezed when moving upslope, displayed a stratiform character by the time it reached the nine-mile site. Rainfall was heaviest at the ground during the first 20 min of this pattern and had a value of about 0.4 in/hr. Total accumulation over the approximate hour of record was just under 0.2 in. Within the area of general rain there were embedded two regions of greater echo height, appearing in Fig. 7 as broad waves in the intensity pattern, which are thought to be indicative of the convective origin of the precipitation. The only updrafts of significance, however, occurred at low levels in the leading portion of the system (at approximately 1831). Again, this example suggests that a generating process might be associated with the system's upslope movement.

(4) *August 18.* This was the first stormy day after nearly a week of clear weather. There were showers at Hilo in the morning and a strong and cool east wind blew at the nine-mile site most of the afternoon. It was at about 1730 LST before light rain began at the site, by which time a low-level downslope wind had developed. Initial Doppler measurements indicated that the direction of motion was from the northeast, which was somewhat contrary to visual observations of clouds overhead; but it was not uncommon in late afternoon for low level clouds overhead and west of the site to be moving eastward with the katabatic wind, while clouds and echoes to the east of the site still moved with the northeast trades.

At 1800 there was observed a line of intense echoes oriented approximately NW-SE and extending in radar azimuth from north to east. The horizontal Doppler velocities in this line were everywhere toward the radar and exhibited strong time and space variability. As the line approached, there was time available to take detailed sets of horizontal measurements. With the antenna fixed at a selected azimuth, range profiles were taken at a series of elevation angles, usually starting at 0° and moving up in steps of about 5° . A set of these range profiles at a given azimuth lies in a vertical measurement plane through the echo. Data in the plane can be analyzed in the same manner that $h-t$ patterns are constructed, except in this case the coordinates of the pattern are height versus range. The time required for one such set of range profiles was about 5 min. Although these $h-r$ patterns contain whatever time variability

that occurs over the measurement period, they can be interpreted primarily as spatial patterns because the time required for measurements is short.

In Fig. 8*a* are given $h-r$ patterns of intensity and Doppler velocity, measured at 1810 looking due northeast. Four range profiles, at elevation angles of 0° , 5.6° , 11.2° , and 16.8° , were used for constructing these patterns. The path of the range gate for each profile in range-height coordinates is indicated on the patterns; it was along these paths that data points were obtained. The patterns indicate an apparently simple structure: echo top was at about 2 km (above the radar plane) and the horizontal extent was from about 5.5 to 8 km. The intensity maxima (corresponding to $Z \approx 10^{4.5}$ mm⁶/m³) and the velocity maxima of about 4 m/sec were located in the echo center. Part of the echo tilt was caused by the shower's drift toward the radar during the time required for measurement. It should be noted that Doppler velocities at non-zero elevation angles contain a contribution from raindrop fall velocities. No correction for fall velocity was applied to the $h-r$ patterns, so the velocities at elevations above horizontal are somewhat higher (toward the radar) than the actual horizontal velocity of motion.

At 1815 another set of range profiles was taken at azimuth 50° . The resulting pattern is shown in Fig. 8*b*. In addition to the elevation angles used for the pattern of Fig. 8*a*, a profile was also taken at 22.4° in this set. As before, range gate paths are indicated on the pattern. The intensity pattern again displays a fairly well organized structure. Velocities were more irregular, however, and some arbitrariness was used in drawing the velocity contours. (The forward tilt is not present in this pattern because the profiles were taken in reverse sequence from the previous case, i.e. from top to bottom.)

Patterns from azimuths of 56° , 34° , and 73° are given in Figs. 8*c*, *d*, and *e*. The sequence corresponds to the order in which the measurements were taken. The echo character is seen to vary with azimuth, but there are still some features common to all $h-r$ patterns. Maximum intensities for all directions were in the order of -50 to -60 dbm, corresponding to Z values of $10^{4.5}$ to $10^{5.5}$. Horizontal echo extent was always, 2–4 km, although the vertical extents were higher at those azimuths closest to due northeast.

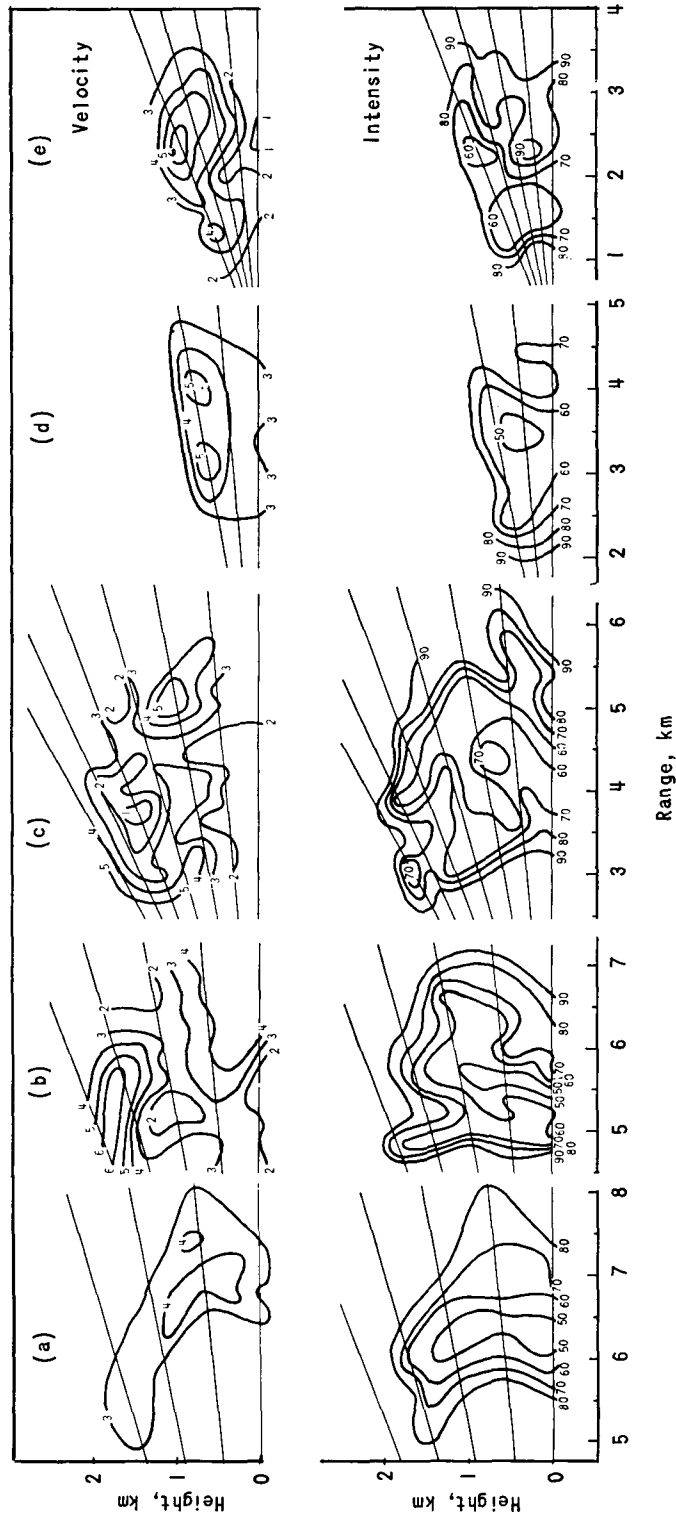


Fig. 8. Height-range patterns at various azimuths, August 18. Intensities in -dbm, normalized to 1 km. Velocities in m/sec, positive approaching.

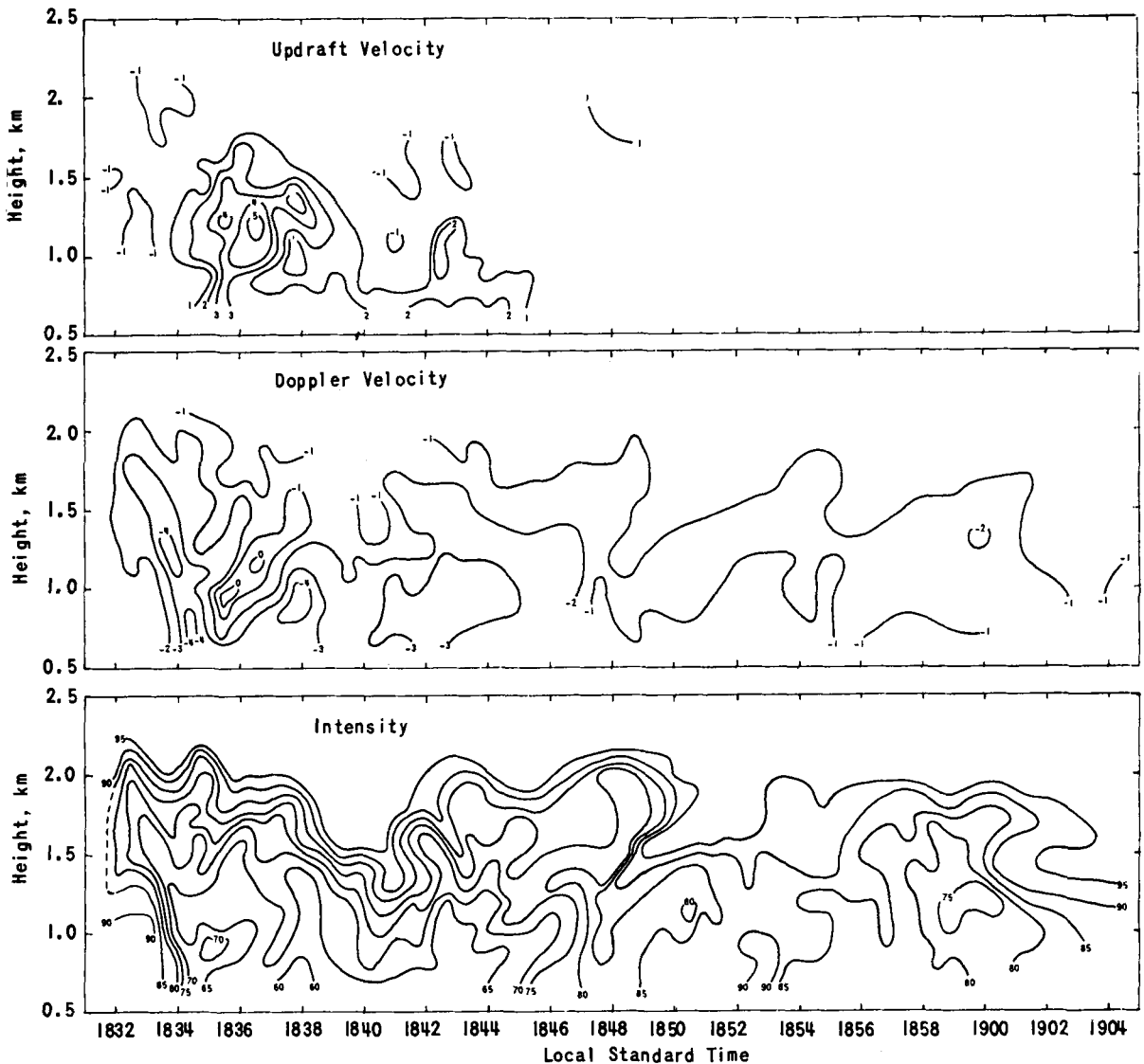


Fig. 9. Height-time pattern, August 18. Intensities in -dbm, normalized to 1 km. Velocities in m/sec, positive upward.

Vertical height profiles were begun at 1832 immediately after completion of the range profiles of Fig. 8. The resulting h - t pattern is given in Fig. 9. It is surprising, since only a single convective line was observed to be approaching the radar site a few minutes prior to the vertical profiles, that three convection cells are apparent in the h - t pattern. Whether the explanation of this rests in time variations of the echo or in previously unobserved spatial variations, the

appearance of the pattern suggests that the three cells, having quite similar shapes, differ primarily in stage of development.¹ Rain started 2 min after the first vertical profile and was heaviest at 1840, which was immediately after

¹ The PPI data obtained by the M-33 radar show that the line was beginning to dissipate as it moved over the Doppler radar site, which is consistent with the changes observed in the Doppler data.

the apparent passage of the first cell. It continued to rain during the entire measurement sequence but total accumulation was only 0.03 in. In addition to being associated with the heaviest rain, the first cell contained the strongest signals. Most noteworthy is the pattern of updraft estimates, with the first cell containing values of 4 m/sec, the second cell only 2 m/sec, and the last cell with undetectable updrafts. This pattern is a good illustration of the connection between the age of a shower and its degree of convective activity. It shows again that an echo's h - t pattern might be entirely stratiform in character although the shower had a convective origin.

(b) Drop-size distributions

The methods of estimating updrafts and drop-size distributions from Doppler radar data are closely associated with each other; neither one can be determined independently of the other. The updraft estimates given above were determined by making assumptions about the drop-size distribution. Likewise, to obtain drop-size distribution requires assumptions about the wind structure.

While it was possible to obtain satisfactory estimates of updraft velocity from measurements of only the signal intensity and the Doppler velocity, the complete Doppler spectrum is needed for obtaining drop-size distributions. Given a spectrum obtained from a vertically-pointing radar, it is necessary to establish a one-to-one correspondence between velocities in the spectrum and raindrop sizes. In the applications (cf. Caton, 1966), this is usually accomplished by simply assuming that there is no vertical air motion in the region sampled and that the raindrops fall with their still-air terminal velocities. Then the drop-size distribution is obtained from the spectrum of terminal velocities by assuming that the raindrops satisfy the Rayleigh scattering criterion. More generally, if the updraft velocity is known and is uniform throughout the sampled volume, the measured spectrum may first be displaced along its velocity scale by the amount of the updraft to yield the spectrum of terminal fall velocities. Then drop sizes may be inferred using the Rayleigh assumption, as before.

For the general case, the relationship between drop-size distribution and the Doppler spectrum, assuming Rayleigh scattering, may be shown to be

$$N(D) = \frac{Z S(V+w) dV}{I D^6 dD},$$

where $N(D)dD$ is the number of drops, per unit volume of air, with diameters in the interval between D and $D+dD$. As before $S(u)$ denotes the Doppler spectrum, where in this case $u = V + w$ with V the terminal fall velocity of a drop with diameter D and w the updraft velocity. Z is the reflectivity factor and I is the signal intensity.

For the Hawaii data, it was found that the zero updraft assumption was not always valid, even in the nonconvective appearing echoes. Consequently, it was often necessary to specify a nonzero value of w in order to determine $N(D)$ from a measured spectrum. This really amounts to guessing a value of w , since w and $N(D)$ cannot be determined independently. Several considerations were taken into account in establishing a suitable value of w for drop-size determinations. First of all, only the spectra obtained from stratiform echoes were analyzed, to insure that updrafts would be small. A given spectrum was then examined to determine whether the zero updraft assumption was reasonable. In some instances, positive (upward) velocities were evident, clearly indicating the presence of an updraft in the beam. In these cases, an updraft velocity was assumed to be present, at least sufficient in magnitude to account for all the positive velocity components in the spectrum. Then a second value for updraft velocity was calculated on the basis of the mean Doppler velocity and signal intensity, using the procedure for estimating updrafts discussed above. This updraft estimate is usually correct to within about 1 m/sec. Finally, the initial guess for w , for the purpose of computing $N(D)$, was obtained as a compromise between this calculated value and the value needed to account for all upward velocities in the spectrum. Utilizing the initial guess for w gave an initial form for $N(D)$, from which were computed by integration both the total drop number density N and the liquid water content M . These quantities turn out to be extremely sensitive to the value used for w . For a given spectrum, a twofold change in liquid water content might be associated with a change in w of only one quarter of a meter per second. This strong sensitivity may be explained as follows: Since the value used for updraft determines the

correspondence between a given velocity component in the spectrum and a dropsize, changing the value of w will cause $N(D)$ to be translated one way or the other along the diameter axis. The *shape* of $N(D)$, without regard to the absolute scale of N , is relatively insensitive to these changes; in the case of the small drops of which Hawaiian rain usually consists, an increase in w of 1 m/sec causes a displacement of $N(D)$ toward larger drop sizes of only about 0.25 mm on the diameter axis. Since w can usually be estimated to within an accuracy of 1 m/sec, it follows that the relative form of $N(D)$ can be determined rather adequately. The problem arises with the D^{-6} factor in the relation between $N(D)$ and the spectrum. As a consequence of this strong weighting factor, the absolute scale of N is highly sensitive to the value chosen for w . For an increase of 1 m/sec in w , the computed form of $N(D)$ might be lowered on the ordinate scale by as much as two orders of magnitude. It is for this reason that the derived values of number density and liquid water content are so sensitive to the choice of updraft velocity.

In principle, this strong sensitivity provides the key for determining updraft velocity; for even an approximate independent measurement of liquid water content or number density (or, for that matter, rainfall rate) in the radar-sampled volume would enable the fairly accurate determination of w , via the Doppler spectrum. Such independent measurements unfortunately are difficult to come by. About the best that one can do, in the face of this problem, is to adjust the initial guessed value of w until the derived number density and liquid water content, for a given spectrum, become reasonable. By "reasonable" here, one could require that liquid water content, for example, be within approximately a factor two or three of 1 gm/m³, to be in accord with the aircraft measurements obtained in penetrations of clouds similar to those observed by radar. (Cf. the paper by Jiusto in this series.) In addition, one could select w values which would insure that the liquid water content not fluctuate too severely in time or space for a given sequence of Doppler spectrum measurements. These were the rather arbitrary criteria which were used in determining the final estimates of updraft velocities, and hence drop-size distributions. Values of $N(D)$ thus determined ob-

viously may not be employed to examine the structure of cloud liquid water content or number density, since these quantities are, in effect, assumed to begin with in order to obtain a suitable determination of $N(D)$. At best, the $N(D)$ curves can be taken as descriptive of the general shapes of actual drop-size distributions; it is not feasible to use them for quantitative assessment of cloud physical processes. For this reason, none of the drop-size distributions that were computed will be included here.¹ Instead, only the general features of the results will be discussed.

Of most interest were the drop-size distributions measured at a sequence of altitudes in an echo over a short period of time. In the stratiform echoes, a fairly regular progression in $N(D)$ was observed when descending from echo top. At highest levels, the population of drops evidently consisted of large numbers of droplets (number densities in the order of 10⁶ m⁻³), all smaller than about 0.4 mm diameter. At lower altitudes the distribution would broaden somewhat and shift to larger drop-sizes. At the lowest altitudes (600 or 700 m) the distribution might have a maximum at a diameter of 0.5 mm, with a small number of drops as large as 1 mm in evidence. Number densities might be some hundred times less than at the echo top, as best as could be determined. These low-level radar observations were not inconsistent with droplet samples collected at the surface on gelatincoated slides.

In the more irregular, less stable-appearing echoes, a systematic change of $N(D)$ with altitude was not generally observed. The variability in these cases might be real, in consequence of vertical mixing, or might only be apparent, as a result of a nonuniform updraft in the radar beam.

In short, for echoes that appeared to be associated with fairly steady, continuous rain, the drop-size measurements were consistent with a regime of steady-state growth by gravitational accretion and coalescence. Owing, however, to limitations inherent in the accuracy of the measurement technique, the observations are not suitable for quantitative droplet growth study.

¹ The interested reader is referred to the original project report (Rogers & Jiusto, 1966) for example distribution curves.

Summary and discussion

Radar observations from thirteen days were the basis of this study of Hawaiian rain. All rains investigated fell from clouds that were entirely warmer than freezing. Rainfall rates at the ground varied from unmeasurably small values up to about 40 mm/hr. Associated with the rain was a variable amount of convective activity, with updrafts ranging up to 7 m/sec. Echo tops were usually at heights between 2 and 3 km above the radar site (at an elevation of 750 m). Signal intensities were often rather high, in extreme cases corresponding to Z values of 10^5 mm⁶/m³.

In the nonconvective cases signal intensity decreased in a regular manner with increasing altitude, a behavior consistent with the growth of raindrops by accretion and coalescence in a gravitational field. Radar reflectivity would characteristically diminish from about $10^{2.5}$ mm⁶/m³ at 600 m to less than 10^0 mm⁶/m³ at echo top (usually between 2 and 2.5 km). The vertical gradient of reflectivity consequently was often quite strong, values of 20 db/km not being uncommon. As well as could be determined, raindrop size distributions in such cases were also consistent with the coalescence process, exhibiting higher concentrations of small drops and lower concentrations of large drops with increasing altitude. Drop number density could evidently be as high as 10^6 m⁻³, often with virtually no drops larger than 1 mm.

Drop-size data in the more convective cases showed evidence of mixing between different altitudes, somewhat obscuring the apparent gravitational growth mechanism. Signal intensities were usually higher and drop sizes larger in these cases, probably indicating that droplet growth is enhanced by convection. Within a given convective shower, regions of high reflectivity were often observed to correspond to regions of updraft maxima.

As storms passed overhead, it was not uncommon to observe a transition within a single large echo system, with initially convective rain changing to stratiform rain.¹ Such observations suggest that the systems generate and propagate by convective means. Depending upon where

and at what stage of development a system is observed, therefore, it might appear to be convective, or stable, or anywhere in between. It is likely that some of the stable-appearing orographic rain might have been convective in origin, but in its dissipating stage when observed. The often observed tendency for echoes to become more stable in character while moving inland is in accord with the observations made by Squires & Warner (1957) in their flights through Hawaiian orographic cloud. They noted the tendency for the cloud to become more cumuliform and turbulent toward the coast.

The example h - t profiles have demonstrated the varieties of echo type associated with Hawaiian rain. To some extent the variability among the data might be explained by differences in the mesoscale wind and stability structure. Even with the variability, certain characteristics distinguish all these echoes from those typical of the Continental United States. The most apparent difference is in echo height. While continental showers are rarely less than 3 km in extent, none of the Hawaii echoes were observed to extend higher than approximately 3.5 km above the ground. This difference between marine showers and mid-latitude continental showers has long been known. It was somewhat surprising, however, to observe that the radar reflectivity of Hawaiian showers is occasionally just as high as that of large continental thunderstorms. Generally speaking, the highest reflectivities were observed in echoes with the greatest vertical development. These were also the echoes which exhibited the strongest updrafts. This association between reflectivity, echo thickness, and updraft structure is in accord with growth by coalescence.

Another distinguishing characteristic of the echoes is the strong vertical gradient of reflectivity usually observed in stratiform rain cases. The existence of such gradients is a consequence of precipitation growth by accretion of cloud droplets and coalescence between raindrops. Wexler & Austin (1954) made use of measured reflectivity profiles in snow to analyze the accretional growth of snow flakes. On the basis of a steady state model of the continuous rain process, Rigby *et al.*, (1954) computed the changes in drop-size distribution with distance fallen as a result of accretion and coalescence, finding that the distributions most rapidly modified are those consisting of large numbers of

¹ Aoyagi *et al.* (1966) have recently pointed out the striking similarity between this convective structure and the structure of snow showers on the coast of the Sea of Japan.

small drops. They showed that the accretion process dominates in these cases, and for one example found that a reflectivity gradient of 6 db/km was expected from accretion alone.

More recently, Hardy (1963) has carried out a series of more exact calculations than those of Rigby *et al.* His results show again that steep initial distributions, i.e., those consisting of numerous small drops, "which are common to the rain formed through orographic lifting in Hawaii", are rapidly modified with distance fallen. Hardy adds that such distributions constitute an efficient configuration for depleting cloud liquid water content. Evidently associated with this efficiency is the strong reflectivity gradient.

It would be interesting to know whether the magnitudes of the observed reflectivity gradients are consistent with simple accretional growth. Unfortunately, Hardy does not present any quantitative estimates of the gradients expected for the special case of Hawaiian rain. Values he presents in discussion of the reflectivity structure below the melting layer are nowhere near the 20 db/km values observed in Hawaiian rain. The 6 db/km value of Rigby *et al.*, is also for a drop-size distribution more appropriate for continental than tropical marine rain. From the generalizations presented in each of these reports, it is reasonable to expect stronger gradients for Hawaiian rain, where drop sizes are smaller than in the examples considered. Moreover, it seems to be possible, by means of a decidedly approximate calculation, to show that the extreme observed gradients are not significantly in disagreement with the simple accretional growth process.

From the continuous-growth equation, assuming zero updraft, the increase in diameter of a raindrop with distance fallen is found to be

$$\frac{dD}{dh} = \frac{EM_c}{2\rho}, \quad (1)$$

where E is collection efficiency, M is cloud liquid water content, and ρ is the density of water. Now, for the purposes of this calculation, let us assume that all the raindrops at a given altitude are of the same size, and that their number density is N . Then the reflectivity factor is given by

$$Z = ND^6, \quad (2)$$

where the contribution of cloud droplets to reflectivity is neglected. The vertical gradient of reflectivity is obtained by assuming a steady state process in which drop size and number may vary with height, but not with time at a given height. Then

$$\frac{1}{Z} \frac{\partial Z}{\partial h} = \frac{1}{Z} \frac{\partial}{\partial h} (ND^6) = \frac{1}{N} \frac{\partial N}{\partial h} + \frac{3EM_c}{D\rho}. \quad (3)$$

Not much can be inferred about the first term on the right hand side, except that it will be negative as a consequence of the increase in fall velocity of the growing drops. Clearly an upper estimate of the value of $1/Z \partial Z / \partial h$ can be obtained by setting this first term equal to zero and solving (3) for selected values of E , M_c , and D .

A lower estimate of the reflectivity gradient may be obtained by introducing the liquid water content M , in the form of raindrops, in place of number density. Since

$$M = \pi \rho N D^3 / 6$$

it follows that

$$Z = \frac{1}{N} \left(\frac{6M}{\pi \rho} \right)^2. \quad (4)$$

Differentiating (4) logarithmically with respect to distance fallen and substituting for $1/N \partial N / \partial h$ in (3), one obtains

$$\frac{1}{Z} \frac{\partial Z}{\partial h} = \frac{1}{M} \frac{\partial M}{\partial h} + \frac{3EM_c}{2D\rho}. \quad (5)$$

In this equation the first term on the right hand side will be positive because the effect upon M of the decrease in N with distance fallen is more than offset by the increase in diameter. Thus a lower estimate for the value of reflectivity gradient can be determined by setting the liquid water content term in (5) equal to zero. On the basis of these considerations, it follows from (3) and (5) that

$$\frac{3EM_c}{2D\rho} \leq \frac{1}{Z} \frac{\partial Z}{\partial h} \leq 3 \frac{EM_c}{D\rho}. \quad (6)$$

Typical values of the parameters in this inequality are $M_c \approx 1$ gm/m³, $E \approx 0.8$, $D \approx 0.5$ mm. These values incorporated in (6) lead to the result that the upper and lower bounds of vertical reflectivity gradient are 21 db/km and

10.5 db/km. Coalescence between raindrops, although not as important as accretion, could give an additional contribution to the gradient. Attenuation of the beam could also account for some of the apparently strong gradients, but judging from horizontal measurements in similar echoes, attenuation effects over the short ranges of these experiments are negligible. The large observed values of reflectivity gradient therefore do not appear to be in serious disagreement with a simple gravitational accretion growth process.

Just as stratiform Hawaiian rain satisfies Hardy's criterion for efficiency in removing cloud liquid water content, so also do many of the convective rain observations fulfill Wexler's (1960) criterion of precipitation efficiency. The latter criterion expresses the ratio of the amount of precipitation that falls from a cloud to the amount of water made available to the cloud

by condensation in adiabatic ascent. Wexler explains that the most efficient clouds, in this sense, are cumulus embedded in widespread stratus. It will be recalled that some of the h - t patterns indicate exactly this kind of cloud structure. Hawaiian rain, therefore, in many instances, is efficient in both the large scale sense of Wexler and in the microphysical sense of Hardy.

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ИССЛЕДОВАНИЕ ГАВАЙСКИХ ДОЖДЕЙ ДОППЛЕРОВСКИМ РАДАРом

Измерения с помощью доплеровского радара указывают на присутствие в ряде дождей на о. Гавайи конвекции различной интенсивности, причем вертикальные скорости воздуха менялись от неподдающихся измерениям малых величин в некоторых случаях до величин в 7 м/сек в случаях более развитой конвекции. Эхо, соответствующие слабым вертикальным движениям, обычно имеют слоистую структуру, будучи отложены в координатах высота-время: контуры интенсивности сигнала и доплеровской скорости стремятся быть приблизительно горизонтальными. В этих случаях интенсивность уменьшалась с высотой и не были необычными такие сильные вертикальные градиенты, как 20 дб/км.

В случае эхо, соответствующих сильным вертикальным движениям, интенсивности сигналов, глубина эхо и величина осадков на поверхности стремятся быть больше. Наблю-

дались такие большие коэффициенты радиолокационной отражаемости как 10^5 мм⁶/м³. Хотя эхо, отложенные в координатах высота-время, образовывали менее правильную картину, чем эхо, соответствующие слабым вертикальным скоростям, имелась тенденция для ведущей (вверх по склону) части конвективного эхо быть более активной, чем задняя часть. По мере продвижения эхо далее вверх по склону, они стремятся стать более слоистыми.

Тот факт, что умеренный дождь мог иногда идти из тонких слоистых облаков, взятый в сопоставлении с сильными вертикальными градиентами коэффициента радиолокационного отражения, наблюдаемыми в этих случаях, указывает на высокую эффективность процессов образования осадков в орографических облаках на о. Гавайи. Наблюдения эхо указывают, что конвективные движения усиливают процессы образования осадков.