Comparison of Raindrop Size Distributions Measured by Radar Wind Profiler and by Airplane

R. R. ROGERS

Department of Meteorology, McGill University, Montreal, Quebec, Canada

D. BAUMGARDNER

National Center for Atmospheric Research, Boulder, Colorado

S. A. ETHIER

Department of Meteorology, McGill University, Montreal, Quebec, Canada

D. A. CARTER AND W. L. ECKLUND

Aeronomy Laboratory, National Oceanic and Atmospheric Administration, Boulder, Colorado

(Manuscript received 20 November 1991, in final form 21 March 1992)

ABSTRACT

Wind profilers are radars that operate in the VHF and UHF bands and are designed for detecting the weak echoes reflected by the optically clear atmosphere. An unexpected application of wind profilers has been the revival of an old method of estimating drop size distributions in rain from the Doppler spectrum of the received signal. Originally attempted with radars operating at microwave frequencies, the method showed early promise but was seriously limited in application because of the crucial sensitivity of the estimated drop sizes to the vertical air velocity, a quantity generally unknown and, at that time, unmeasurable. Profilers have solved this problem through their ability to measure, under appropriate conditions, both air motions and drop motions. This paper compares the drop sizes measured by a UHF profiler at two altitudes in a shower with those measured simultaneously by an instrumented airplane. The agreement is satisfactory, lending support to this new application of wind profilers.

1. Introduction

The drop size distribution in rain is fundamental for understanding the formation and development of precipitation, the interactions of rain with atmospheric pollutants, and the effects of rain on radiative transfer, including the closely related subject of the remote sensing of rain. Standard instruments are available and are widely used for measuring raindrop sizes at the ground (Joss and Waldvogel 1967; Sheppard 1990) and aloft by airplane (Knollenberg 1981). A long-standing challenge has been to observe the drop size distribution and its variation in time and altitude remotely with ground-based equipment. Doppler radar was recognized from the start as having the potential of providing this measurement (Barratt and Browne 1953; Probert-Jones 1960; Rogers and Pilié 1962). The spectrum of Doppler-shifted frequencies measured by a radar pointed vertically in rain is determined by the numbers of drops that are moving towards the radar in different intervals of velocity. If velocity can be related to size, then the drop size distribution may be calculated from the Doppler spectrum. The velocity of the drops relative to the radar depends on 1) the fall speed of the drops relative to the air and 2) the vertical air velocity. The fall speed relative to air is an accurately known function of drop size (Gunn and Kinzer 1949; Hosking and Stow 1991), but the vertical air velocity is usually unknown. In the only previous comparison of radar-estimated drop sizes with aircraft measurements, Probert-Jones (1960) reported good agreement in warm-frontal rain, assuming zero vertical air velocity. But the classical problem in radar drop sizing, explained by Rogers (1967), is that the calculated drop size distribution is extremely sensitive to the vertical air velocity, a quantity neither known nor readily measurable by radar or any other means of remote sensing. This problem has now been solved by wind-profiling radars (Röttger and Larsen 1990; Gage 1990) that are able, under certain conditions, to detect reflections from the clear air as well as from drops. Bimodal Doppler spectra in which the clear-air and precipitation modes are clearly distinguishable enable the measurement of drop fall speed relative to the air and hence the drop size. Wakasugi et al. (1986), Gossard (1988), Gossard et al. (1990), and Currier (1990) have reported

Corresponding author address: Dr. R. R. Rogers, Department of Meteorology, McGill University, 805 Sherbrooke St. West, Montreal, Quebec, H3A 2K6 Canada.

on drop size distributions estimated using this technique. The recent Hawaiian Rainband Project (HARP) provided an opportunity to assess the accuracy of the technique by comparison with independent and simultaneous measurements of drop sizes aloft.

2. The HARP wind profiler

As part of the Hawaiian Rainband Project in the summer of 1990, a boundary-layer profiler was operated at a site 20 km southeast of Hilo, Hawaii, for the period from 11 July to 19 August. Designed especially for HARP, the profiler is a Doppler radar with a transmitted frequency of 915 MHz (33-cm wavelength), similar to other small UHF radars built by the National Oceanic and Atmospheric Administration Aeronomy Laboratory for measuring the wind speed and direction in the lower troposphere and described by Ecklund et al. (1988; 1990). It employs a single, flat, steerable microstrip-array antenna and produces a beamwidth of 9°. Ordinarily, a sequence of pointing directions including the vertical and several directions at 15° off the vertical was used to permit wind profiling and frequent observations of rain development directly overhead. At 33-cm wavelength, the profiler has about the same sensitivity to very light rain as to moderately reflective clear air. For example, rain with a reflectivity factor of 5 dBZ produces an echo as strong as clear air with a refractive-index structure parameter C_n^2 of 10^{-13} $m^{-2/3}$. With a resolution of 105 m in range and 30 s in time, the profiler was ordinarily able to measure Doppler spectra from the clear air up to and including the altitude of the trade wind inversion, which ranged from about 2 to 4 km, and from rain and clouds with reflectivity factors as weak as -10 dBZ. The spectral resolution was 0.34 m s^{-1}. This corresponds to a drop size resolution of 0.09 mm for drops with diameters in the interval from 0.1 to 1.2 mm—the interval in which the terminal fall velocity is approximately a linear function of diameter (Rogers and Yau 1989)—and to less resolution at larger sizes. When heavy rain was present, it would dominate the Doppler spectrum, but in light rain, the spectrum was sometimes bimodal with clearly identifiable peaks corresponding to rain and the clear air.

3. Drop sizes measured by profiler and airplane

On 10 August, the National Center for Atmospheric Research Electra research aircraft made two passes over the profiler site, collecting data on temperature, humidity, wind, and cloud and rain characteristics, including the drop size distribution in rain. This was not a carefully planned experiment, but only a brief opportunity made possible by some unused flight time at the end of an experiment focused on the dynamics and microphysics of offshore rainbands. The drops were measured with a PMS (Particle Measuring Systems, Inc.) 2D precipitation probe, an instrument described by Knollenberg (1981). The volume of atmosphere sampled for each measured distribution averaged about 90 L, as determined by the sampling area of 8 cm^2, the accumulation time of 1 s, and the average airplane speed of 110 m s^{-1}. The diameter resolution of the measurements is 0.1 mm and thus nearly the same as that of the profiler. Figure 1 shows the pattern in time and height of the radar reflectivity over the profiler that includes the time of the Electra measurements. Regions denoted by the letters A and B on the figure indicate the time and altitude of the airplane during its passes

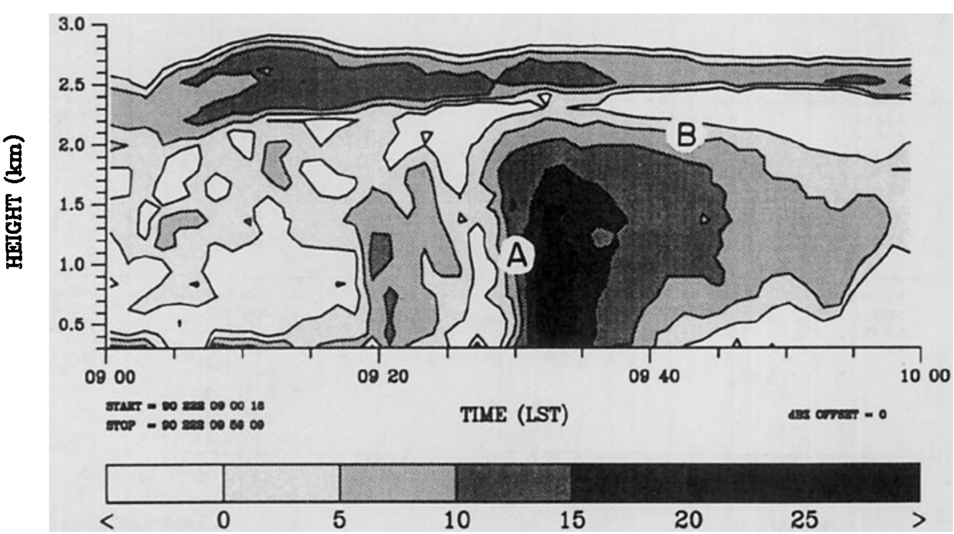

FIG. 1. Time–height pattern of reflectivity measured in the vertically pointing profiler beam. The outer contour corresponds to a reflectivity factor of 0 dBZ; the contour interval is 5 dBZ. A rainshower is evident in the pattern starting at 0927 LST and extending upwards to about 2 km. Regions A and B are locations in time and altitude of the Electra overflights. The layer of high reflectivity at around 2.5 km is the trade wind inversion.

near the profiler. The position of the Electra was determined by an inertial navigation system whose drift was corrected using measurements from the Global Positioning System. Airplane echoes were evident in the profiler data at 0928:24 and 0948:56 LST. The airplane was probably in the main lobe at the first of these times and in a sidelobe at the second time.

The reflectivity pattern of Fig. 1 is fairly typical of much of the HARP data in showery conditions. The rain is approximately defined by the 0-dBZ contour that extends upwards to about 2 km at times between 0927 and 1000 LST. Echoes within this contour arise mainly from scattering by drops. The maximum reflectivity factor in the rain is 25 dBZ at 0932 and just below 1 km. The reflective layer centered at about 2.5 km is the trade-wind inversion, which is evident to the radar because of the strong vertical gradient of refractive index at that level (Rogers et al. 1992).

There were 14 in situ samples of drop size distribution in region A when the airplane was within a 2-km horizontal distance of the profiler and 32 samples in region B. The distributions in each region were remarkably consistent. Examples are shown in Figs. 2a and 2b. The measurements at the higher altitude (region B) indicate exponentially distributed drizzle drops, none exceeding 0.5 mm in diameter. Reflectivity factors computed from the distributions average close to 0 dBZ. At the lower altitude (Fig. 2a), the distributions all show a mode at about 0.5 mm against a background of exponentially distributed smaller drops. Reflectivities are around 20 dBZ.

Figure 3 is an example of a Doppler spectrum measured by the profiler at a time and height near region A and the drop size distribution derived from it. The sign convention is that positive velocities are toward the radar. The distribution $N(D)$ is obtained from the spectrum $s(v)$ by

$$N(D) = \frac{Z}{D^6}\left(\frac{dV_t}{dD}\right)s[V_t(D) - U], \qquad (1)$$

where Z is the reflectivity factor determined from the intensity of the received signal, D is the drop diameter, U is the updraft speed, and V_t is the fall velocity of a drop of diameter D relative to the air. The velocity v of the drop towards the radar is $V_t - U$. The Doppler spectrum $s(v)$ in this expression has the clear-air components removed and is normalized to unit area. In the example shown, the spectral mode arising from the clear air is weak but discernible, indicating an updraft velocity of 1.1 m s^{-1}. Here $N(D)$ is calculated for all spectral components at velocities greater than -0.7 m s^{-1}, the location of the minimum between the clear-air mode and the drop mode.

For the dependence of fall velocity on diameter, the following analytical approximation to the tabulated data of Gunn and Kinzer (1949) is used:

$$V_t(D) = \begin{cases} KD(1 - e^{-kD}), & D \leqslant D_0 \\ A - Be^{-CD}, & D \geqslant D_0, \end{cases} \qquad (2)$$

where $K = 4$ (m s^{-1}) mm^{-1}, $k = 12$ mm^{-1}, $A = 9.65$ m s^{-1}, $B = 10.43$ m s^{-1}, $C = 0.6$ mm^{-1}, and D_0

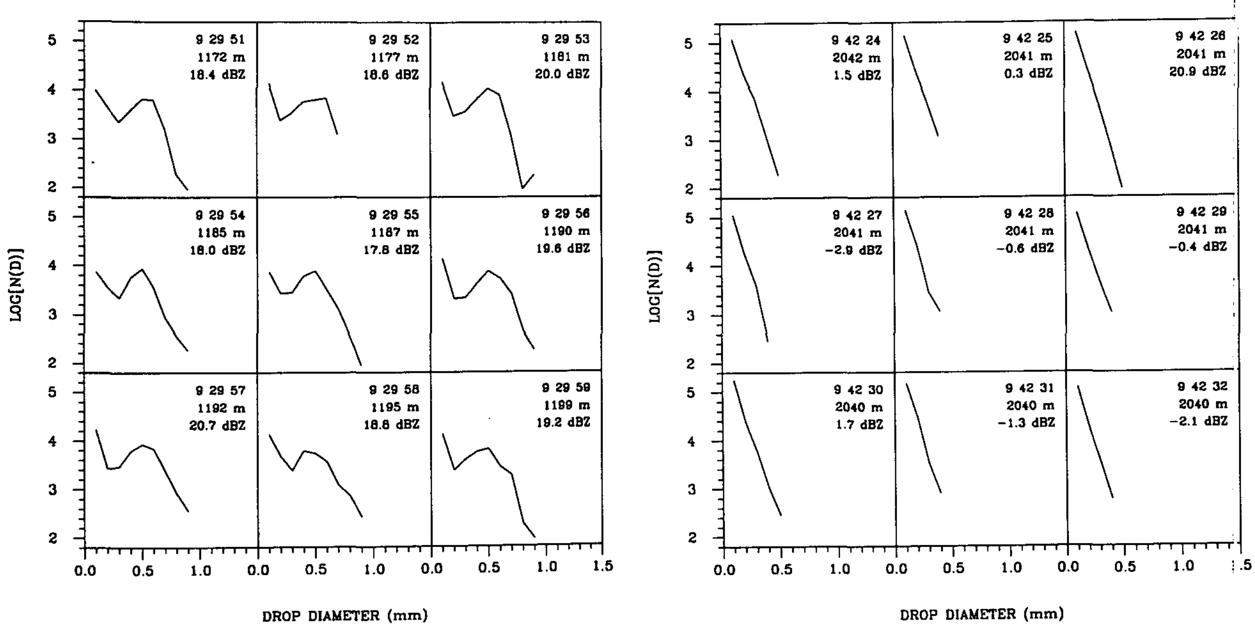

FIG. 2. (a) Examples of consecutive 1-s airplane measurements of drop size distributions in region A. Time, altitude, and reflectivity factor are indicated on each distribution; $N(D)$ has units per millimeter per cubic meter. (b) Same as (a) but for region B. The anomalous reflectivity of 20.9 dBZ at 0942:26 LST is accounted for by a (possibly spurious) count of one drop at 1.6-mm diameter, which is off the scale of this plot.

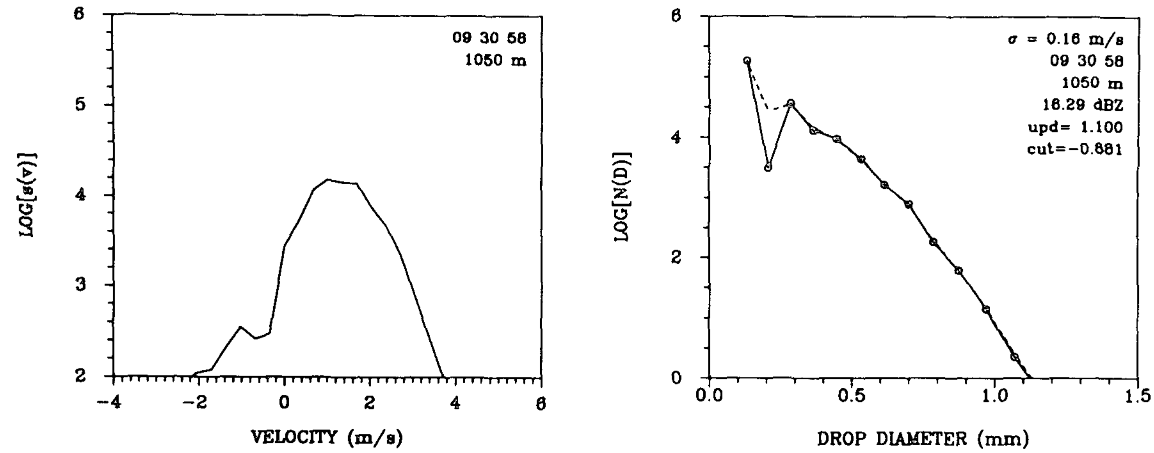

FIG. 3. Doppler spectrum (left) and corresponding drop size distribution (right) measured in region A in the vertical beam of the profiler. Positive velocities in the spectrum represent downward motion. The dashed curve is the distribution obtained from the original spectrum; the solid curve is from the spectrum deconvolved by a Gaussian distribution with a standard deviation of 0.16 m s^{-1}, using the method of Fourier transforms. The reflectivity of the clear air in this example is much weaker than that of the rain, but the clear-air contribution is still recognizable in the spectrum as the low-velocity mode. Indicated on the distribution are the time, altitude, and reflectivity, as well as the updraft velocity, the cutoff velocity separating the air mode from the drop mode, and the standard deviation of the Gaussian function used for deconvolution.

= 0.745 mm. This approximation is continuous and has a continuous first derivative at D_0 = 0.745 mm, which is an important consideration in applying (1). The approximation also approaches the quadratic form of Stokes' Law in the limit for small D. Over the whole range of the Gunn and Kinzer data (i.e., for diameters from 0.1 to 5.8 mm), the maximum percentage deviation of (2) from the data is a 3.7% overestimation at the smallest diameter of 0.1 mm. For the ten data points up to 1 mm, the root-mean-square percentage deviation is 2.4%. Because this is the relevant range for the small drops in these comparisons, the accuracy of the approximation is adequate for our purposes. Incidentally, the root-mean-square deviation of (2) from the Gunn and Kinzer data over the whole range to 5.8 mm is only 1.66%, so the formula is also applicable for drop size distributions broader than those encountered in this experiment.

As a refinement in applying (1), the Doppler spectrum of the drops may be deconvolved by the Doppler spectrum of the air. In principle, this sharpens the drop spectrum by removing the effects of spurious Doppler spreading by turbulence and by the cross-beam wind component [explained by Gossard (1990)], though in practice, deconvolution is a noisy process that should only be applied with caution. The example in Fig. 3 shows the effect on the drop size distribution of deconvolving the Doppler spectrum of the drops by a Gaussian function with a standard deviation of 0.16 m s^{-1}—a function similar to but somewhat narrower than the actual air spectrum. This "mild" deconvolution compensates for some of the unwanted spreading without introducing excessive noise. Our experience, like that of others, indicates that attempting a full deconvolution produces a noisy and unrealistic result. The problem of optimum deconvolution is a subject in its own right and is beyond the scope of this paper.

The application of wind profilers to drop sizing is not limited to measurements in the vertically pointing beam. As long as there is a distinguishable air mode in the Doppler spectrum, the drop size distribution can be estimated from measurements in an off-vertical beam using equations in the form of (1) and (2) but with some of the terms interpreted differently. In particular, in (1), U becomes the radial component of the three-dimensional air velocity, defined as positive away from the radar, and V_t becomes the radial component of the terminal velocity, namely the V_t given by (2) multiplied by $\cos\phi$, where ϕ is the zenith angle.

Figure 4 gives two examples of drop size distributions near region A obtained from profiler measurements in the beam pointed at a 15° zenith angle in the eastward direction. These distributions and the one in Fig. 3 are similar to each other and in fairly close agreement with the airplane measurements. They indicate distributions of small drops extending to 1 mm or slightly larger size and with a mode or a shoulder at about 0.5 mm. In each of these cases, the Doppler spectrum had a weak but recognizable clear-air mode. These and the example of Fig. 3 are the measurements with detectable clear-air spectral modes that were closest in time and altitude to the airplane measurements.

Figure 5 shows drop size distributions measured in the vertical beam of the profiler at and just below the level of the airplane measurements in region B. Because the reflectivity of the clear air was very weak at this higher altitude, the Doppler spectra did not have a discernible clear-air mode. The spectra measured at the same time at the lower altitudes of 1260, 1365, and 1470 m did have identifiable clear-air modes, however,

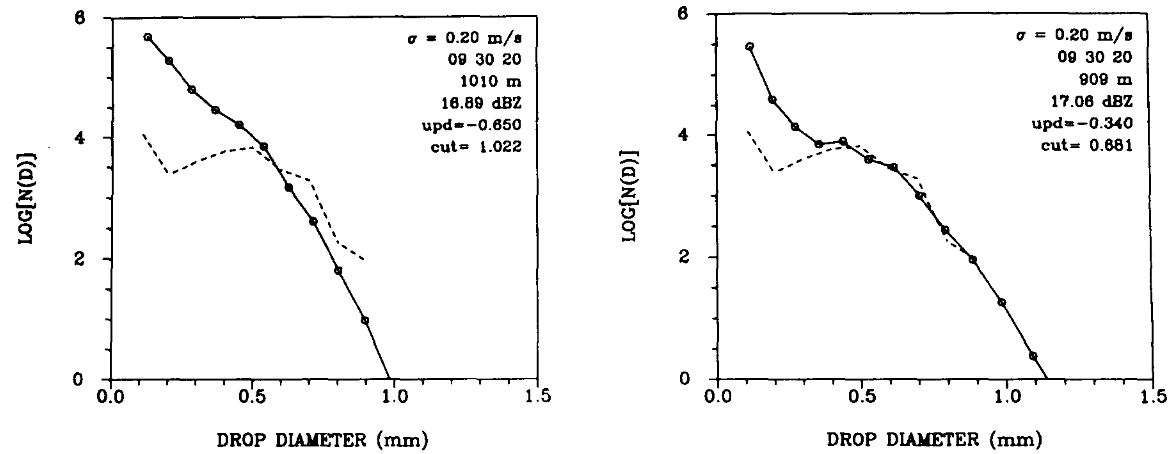

FIG. 4. Two more examples of drop size distributions measured by profiler near region A, in this case, the beam tipped 15° from the zenith in the eastward direction. Both spectra were deconvolved as explained for Fig. 3. Shown dashed for comparison on both plots is the airplane-measured distribution at 0929:59 LST from Fig. 2a. The profiler distributions extend to lower values of $N(D)$ than the airplane measurements because of the much larger sample volume.

and these all showed the vertical air velocity to be negligibly small. These measurements were extrapolated upward, and an updraft of zero was assumed in region B to determine the distributions in Fig. 5.

4. Conclusions

The agreement between airplane and profiler measurements is encouragingly close—in fact, closer than might have been expected from the vastly different resolutions of the two methods. While the volume sampled in an airplane measurement is only 90 L, the volume sampled by the profiler is approximately 2×10^6 m^3 in region A and 8×10^6 m^3 in region B. A profiler sample thus contains on the order of 10^8 times more drops than an airplane sample. This accounts for the ability of the profiler to sense drops in much smaller concentration than the airplane. Though not ideal because the set of airplane measurements in region B happened to be just above the level of detectable clear-air spectral components, this comparison supports the application of wind-profiling radars to the measurement of drop size distributions in rain.

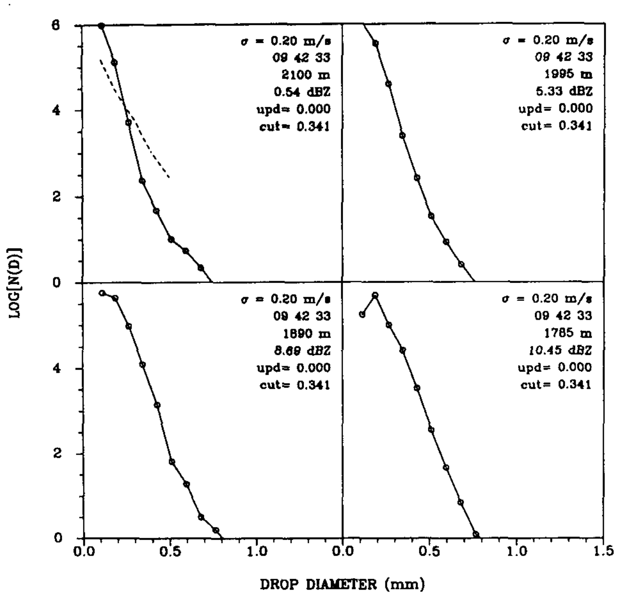

FIG. 5. Drop size distributions measured by profiler at and below region B. Shown dashed for comparison with the uppermost distribution is the airplane measurement at 0942:30 LST from Fig. 2b.

Acknowledgments. The HARP experiment of 1990 was the fourth major collaborative field study of rain on the Island of Hawaii. Pat Squires, then with CSIRO, was a key player in the first of the series: Project Shower of 1953. The fundamental work by Squires and Warner on cloud microphysics has strongly influenced our own work and that of all the other researchers who have attempted to understand the fascinating details of Hawaiian rain. We thank K. S. Gage of NOAA for arranging to build the profiler and take it to Hawaii and E. E. Gossard of the University of Colorado for convincing us of the usefulness of profilers in precipitation studies. Our research was made possible by grants from the National Science Foundation and the Office of Naval Research of the United States and the Atmospheric Environment Service of Canada.

REFERENCES

Barratt, P., and I. C. Browne, 1953: A new method of measuring vertical air currents. *Quart. J. Roy. Meteor. Soc.*, **79,** 550.

Currier, P. E., 1990: Precipitation measurement using a dual frequency doppler system. Ph.D. thesis, University of Colorado, 242 pp.

Ecklund, W. L., D. A. Carter, and B. B. Balsley, 1988: A UHF wind profiler for the boundary layer: Brief description and initial results. *J. Atmos. Oceanic Technol.*, **5,** 432–441.

——, ——, ——, P. E. Currier, J. L. Green, B. L. Weber, and K. S. Gage, 1990: Field tests of a lower troposphere wind profiler. *Radio Sci.,* **25,** 899–906.

Gage, K. S., 1990: Radar observations of the free atmosphere: Structure and dynamics. *Radar in Meteorology,* D. Atlas, Ed., American Meteorological Society, 534–565.

Gossard, E. E., 1988: Measuring drop-size distribution in clouds with a clear-air-sensing Doppler radar. *J. Atmos. Oceanic Technol.,* **5,** 640–649.

——, 1990: Radar research on the atmospheric boundary layer. *Radar in Meteorology,* D. Atlas, Ed., American Meteorological Society, 477–527.

——, R. G. Strauch, and R. R. Rogers, 1990: Evolution of dropsize distribution in liquid precipitation observed by ground-based Doppler radar. *J. Atmos. Oceanic Technol.,* **7,** 815–828.

Gunn, R., and G. D. Kinzer, 1949: The terminal velocity of fall for water drops in stagnant air. *J. Meteor.,* **6,** 243–248.

Hosking, J. G., and C. D. Stow, 1991: Ground-based measurement of raindrop fallspeeds. *J. Atmos. Oceanic Technol.,* **8,** 137–147.

Joss, J., and A. Waldvogel, 1967: Ein Spektrograph für Niederschlagstropfen mit automatisher Auswertung. *Pure Appl. Geophys.,* **68,** 240–246.

Knollenberg, R. G., 1981: Techniques for probing cloud microstructure. *Clouds, Their Formation, Optical Properties and Effects,* P. V. Hobbs and A. Deepak, Eds., Academic Press, 15–91.

Probert-Jones, J. R., 1960: Meteorological use of pulsed Doppler radar. *Nature,* **186,** 271–273.

Rogers, R. R., 1967: Doppler radar investigation of Hawaiian rain. *Tellus,* **19,** 432–455.

——, and R. J. Pilié, 1962: Radar measurements of drop-size distribution. *J. Atmos. Sci.,* **19,** 503–508.

——, and M. K. Yau, 1989: *A Short Course in Cloud Physics.* 3d ed. Pergamon Press, 293 pp.

——, C. A. Knight, J. D. Tuttle, W. L. Ecklund, D. A. Carter, and S. A. Ethier, 1992: Radar reflectivity of the clear air at wavelengths of 5.5 and 33 cm. *Radio Sci.,* **27,** 645–659.

Röttger, J., and M. F. Larsen, 1990: UHF/VHF radar techniques for atmospheric research and wind profiler applications. *Radar in Meteorology,* D. Atlas, Ed., American Meteorological Society, 235–281.

Sheppard, B. E., 1990: Effect of irregularities in the diameter classification of raindrops by the Joss–Waldvogel disdrometer. *J. Atmos. Oceanic Technol.,* **7,** 180–183.

Wakasugi, K., A. Mizutani, M. Matsuo, S. Fukao, and S. Kato, 1986: A direct method for deriving drop-size distribution and vertical air velocities from VHF Doppler radar spectra. *J. Atmos. Oceanic Technol.,* **3,** 623–629.