

HYDROMETEOR IDENTIFICATION WITH ELLIPTICAL POLARIZATION RADAR: APPLICATIONS FOR GLACIOGENIC CLOUD SEEDING

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Abstract. Polarization capabilities of 8.66-mm-wavelength radar and corresponding hydrometeor depolarization calculations now provide the basis for identifying individual types of ice and liquid hydrometeors within clouds and precipitation. Model results and radar measurements with correlated snow crystal samples from winter stratiform and convective orographic clouds are examined. The results are interpreted to illustrate how this method for estimating hydrometeor types can be applied to monitor cloud evolution and evaluate the potentials and effects of glaciogenic cloud seeding. This information is derived from hydrometeor depolarizations (and related cloud reflectivities) associated with cloud phase transition, snow crystal growth habit, graupel development, snow crystal aggregation, the presence and nature of the melting level, and the distinction of rain from drizzle below the melting level. The radar's polarization capability offers the opportunity to monitor the development of hydrometeors in the volume of cloud affected by seeding or by natural processes. Spatial gradients in hydrometeor types, rates at which a volume is transformed, and the form of precipitation can also be estimated. These features are all important for verifying the changes in hydrometeors introduced by seeding and interpreting the rates and mechanisms by which the changes occur.

1. INTRODUCTION

Measurement capabilities and applied theory are now allowing hydrometeor types to be identified with 8.66-mm-wavelength elliptical polarization radar (Kropfli et al., 1995). Snow crystals of the various growth habits depolarize backscattered millimeter-wavelength radiation according to their aspect ratio, orientation, and bulk density, and the polarization state of incident waves (Matrosov, 1991a,b). The effect of the hydrometeors on the incident radiation can be measured as an elliptical depolarization ratio (EDR). The variation of EDR with radar elevation angle or specific polarization state translates to an estimate of the hydrometeor type. Polarization signatures of drizzle, thick plates, dendrites, aggregates of dendrites, needles, and graupel have been measured during recent field programs with the NOAA Environmental Technology Laboratory (ETL) K_a-band (8.66 mm) polarization-agile Doppler radar. These signatures have shown very good quantitative agreement with theoretical models and corresponding snow crystal samples (Matrosov et al., 1996; Reinking et al., 1993, 1995a,b, 1996). Observations of snow crystals at the surface and aloft for *in situ* verification were obtained with a National Center for Atmospheric

Research (NCAR) microphysics van and a Weather Modification, Inc., Cheyenne II aircraft. In this paper, the previously reported measurements and new depolarization data are interpreted in the context of cloud modification. The results illustrate how ice and liquid hydrometeor identification using elliptical polarization radar can be applied in real-time monitoring or *post facto* analysis to evaluate the potential for and effects of precipitation enhancement by cloud seeding.

2. CAPABILITIES IN MEASUREMENT AND THEORY

The K_a-band radar will detect clouds with reflectivities, Z_e, down to about -30 dBZ at 10-km range or, equivalently, 35-μm droplets in concentrations as small as several hundred per liter. The Z_e is the direct result of size and concentration. EDR varies according to shape, density, and orientation but is only indirectly linked to particle size. Depolarization is calculated as the *ratio* of the power (or Z_e) returned in the cross channel to that returned in the main channel. The K_a-band radar's polarization can be cycled through a continuum of elliptical states by rotating a phase-retarding plate

(PRP) at fixed elevation and azimuth or fixed at a specific state for range-height-indicator (RHI) scans at fixed azimuth. Two PRPs are now in use: one induces a phase shift very near 79.5° ; the other, 95° . The K_a -band radar data for this study were obtained during the 1995 Arizona Program during which the 95° PRP was used (Bruitjes et al., 1994; Reinking, 1995), and the 1993 and 1994 Winter Icing and Storms Projects (WISP) during which the 79.5° PRP was used (Rasmussen et al., 1992). A PRP with a phase shift nearer 90° (e.g., 95°) increases the dynamic range of possible depolarization ratios. However, clouds with small but measurable reflectivity will sometimes produce no depolarization signature due to weak power return in the cross channel. Greater detuning from 90° (e.g., 79.5°) provides greater power return in the cross channel to observe depolarization in low-reflectivity clouds (Matrosov, 1991b); however, the expense is a smaller dynamic range.

Horizontal linear (LDR) and circular (CDR) depolarization ratios define the theoretical limits of EDR. Matrosov (1991a) modeled the radar elevation angle (β) dependency of LDR and CDR at K_a -band for ice crystals of various habits. Refined calculations for a 79.5° PRP were developed by Matrosov et al. (1996) and compared with those for a 95° PRP by Matrosov et al. (1995). The PRP resolves the transmitted beam into two components, retards the phase of one relative to the other, and then reunites the two components (Shurcliff, 1962). A 90° -phase-shift PRP produces circular polarization from incident horizontal polarization when its "slow axis" is rotated 45° . In practice, when the phase shift differs from 90° slightly, all the polarizations produced by such a PRP will be elliptical (except for 0° and 90° rotation). The largest ellipticity coefficient (or the most circular polarization state possible) is then produced at 45° rotation of the PRP "slow axis" relative to the incident electric field. For the respective PRPs in use, we define the corresponding elliptical depolarization ratio as $EDR_{45-79.5}$ or EDR_{45-95} , for which ellipticities of the transmitted signal are $\epsilon_{79.5} = 0.832$ and $\epsilon_{95} = 0.916$. These most circular states are used for RHI scans. Depolarization, as we define it, increases as the absolute value of EDR decreases toward zero. For spherical hydrometeors, depolarization is independent of β , such that $EDR_{45-79.5} \approx -15$ dB and $EDR_{45-95} \approx -21$ dB; near-spherical drizzle provides experimental reference calibrations. The "cross-talk" limit of the radar's antenna is about ± 35 dB; the effect of this is accounted for in the models of Matrosov et al. (1996).

Ensembles of remote sensors will always provide the most complete depiction of cloud state. For example, the microwave radiometer, particularly one that is steerable (Hogg et al., 1983), will provide an excellent, continuous estimate of path-integrated available liquid water (LW); when combined with the K_a -band radar, the ensemble provides more information for monitoring the cloud evolution and seeding potential than either instrument used separately. The radiometer was used in WISP and the 1995 Arizona Program. Also in Arizona, K_a -band radar measurements were often coordinated with measurements from a second radar, the NOAA/ETL circular-polarization X-band (3.2 cm) Doppler radar. Chaff was released in clouds by the Cheyenne II aircraft and tracked with the X-band radar using the Tracking Air with Circular-Polarization Radar (TRACIR) technique (Martner and Kropfli, 1989; Martner et al., 1992). Such chaff tracking measures the transport and dispersion (as that of co-released seeding material) *within* clouds with reflectivities as large as ~ 30 or 35 dBZ (e.g., Reinking and Martner, 1996). The addition of temperature soundings, measured continuously with a radio acoustic sounding system (RASS) or intermittently with a radiosonde, will significantly aid the translation of the radar depolarization measurements, given the temperature dependence of specific ice crystal habits.

3. APPLICATIONS FOR GLACIOGENIC CLOUD SEEDING

Several microphysical features of clouds affect the potential and realized results of precipitation enhancement. Those features that might be observed with elliptical polarization radar may be broadly categorized as cloud phase transition, snow crystal growth habit, graupel, snow crystal aggregation, and the melting level. Included are the volume of cloud affected, corresponding spatial gradients among hydrometeor types and mixes thereof, rates at which a volume is transformed, and the form of precipitation. These features are all important to verify through observations and models: the kinds of hydrometeors nature is producing, the changes in hydrometeors introduced by seeding, and rates and mechanisms by which the changes occur through the affected cloud volume. The following experimental and theoretical results and designs illustrate how these factors may be monitored and measured. Gradients in hydrometeor development occurred primarily in the vertical in the observed winter storms. Therefore, where theory has been compared to measurements, we have used radar data from fixed altitudes.

3.1 Cloud Phase Transition

A capability for observing phase transitions as they progress is of obvious significance to glaciogenic cloud seeding. Methods for monitoring phase changes in stratiform and cumuliform clouds are discussed here.

3.1.1 Stratiform cloud

An observed transition from droplet-dominated fog to fog with measurable crystals (Fig. 1 to Fig. 2) illustrates the capability for observing phase transitions in stratiform clouds. Figure 1 shows the reflectivity and depolarization from a fog that produced occasional light drizzle during a WISP project (7 February 1994). The fog was about 0.7 km deep with reflectivities between ~ -12 and -25 dBZ,

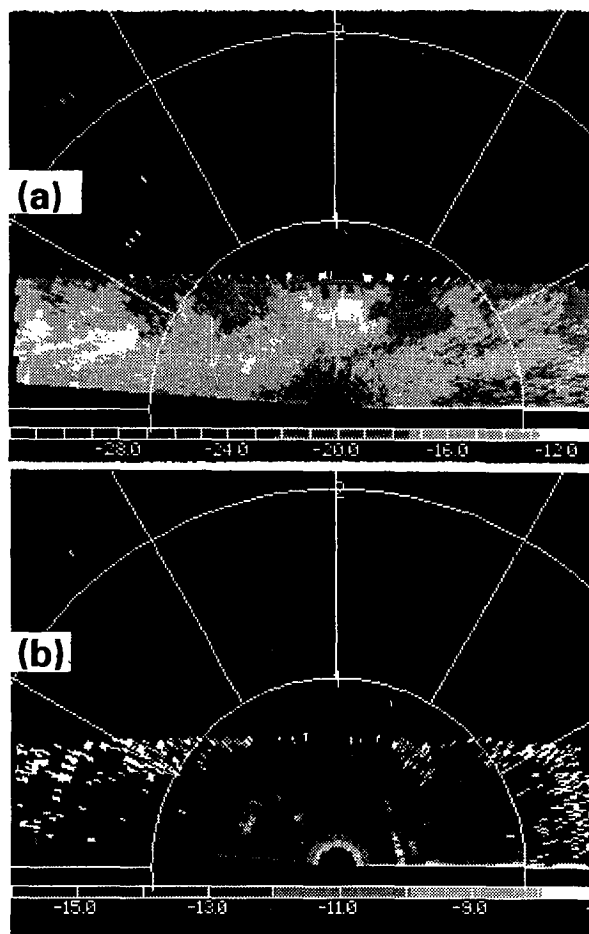


Figure 1. RHI scan through fog and drizzle: (a) reflectivity, Z_e (dBZ); (b) depolarization ratio, $EDR_{45-79.5}$ (dB). $EDR_{45-79.5}$ is invariant with radar elevation angle, indicating spherical hydrometeors; 1317 MST 7 February 1994, 1-km range rings (WISP-94).

in which the temperature increased from about -16°C near the ground to -9°C near cloud top. After about 1300 UTC, depolarizations of -14 to -15 dB were measured with the 79.5° PRP at all radar elevation angles, indicating a dominance of spherical hydrometeors, i.e., large cloud droplets or drizzle at the time of measurement. In this low-reflectivity cloud (Fig. 1), a slight reduction in depolarization with range (~ 1 dB/km) is detectable. ETL's microwave radiometer measured 0.07–0.13 mm of liquid over the period reported here, so the cloud was not glaciated. Some drizzle occurred intermittently over the ensuing half hour at the radar site.

The fog persisted, and eventually depolarizations definitive of planar hydrometeors, with EDR strongly dependent on β (~ -15 dB at $\beta = 90^\circ$ and ~ -7 dB at $\beta = 15^\circ$), developed with sparse dendrite crystals reaching the surface in concentrations that were visually estimated to be maximally only 1 m^{-3} (Fig. 2). The reflectivity within the fog became

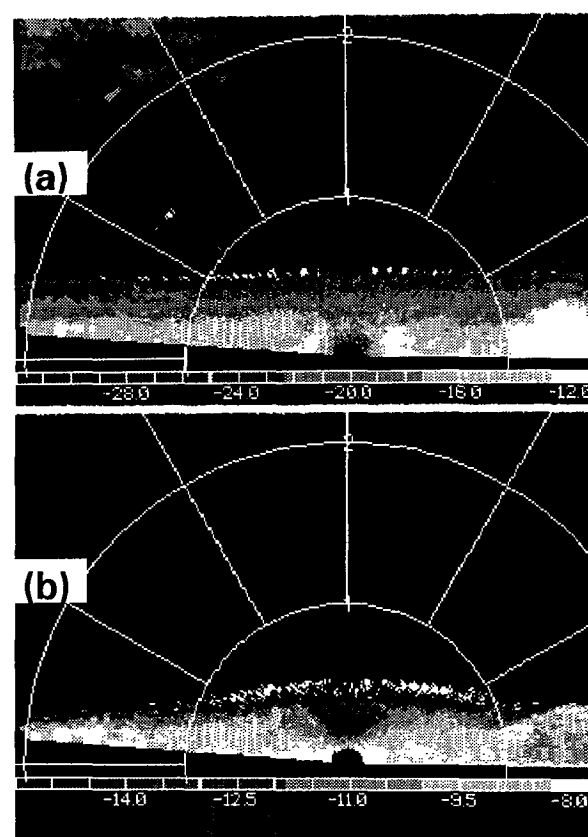


Figure 2. RHI of (a) Z_e (dBZ) and (b) $EDR_{45-79.5}$ (dB) at 1701 MST in the fog of Fig. 1. Here $EDR_{45-79.5}$ varies systematically with radar elevation angle, β ; values of about -15 dB at $\beta = 90^\circ$ and -8 dB at $\beta = 15^\circ$ indicate planar crystals.

horizontally stratified at this time (~1700 UTC) and ranged from about -30 dBZ at cloud top to -12 dBZ near the surface. The reflectivity profile, which is sensitive to hydrometeor size, followed the growth of the crystals during settling. EDR, however, becomes measurable when the cross-channel receiver power exceeds the sensitivity of the radar system. In this case, EDR became measurable at the altitude of about 300 m AGL, where the reflectivity abruptly increased from about -22 dBZ to about -15 dBZ. The sensitivity of the radar extends to reflectivities well below -15 dBZ at close range, so 300 m AGL is the indicated snow crystal generation level.

In summation, crystals were detected *and identified* in clouds with very low reflectivity in concentrations at least as low as the order of 1 m^{-3} . The radar is expected to be capable of detecting depolarization of crystals in much lower concentrations. A cloud with an ice particle concentration of 1 m^{-3} and measurable LW is usually considered to be seedable. Many mountain experiments demonstrate that long-duration fluxes of very minor LW, such as noted for this cloud, could provide important precipitation if seeded (Reynolds, 1988; Reinking and Meitín, 1989).

3.1.2 Cumuliform cloud

The RHI scan, which utilizes the elevation dependency of EDR (e.g., Figs. 1 and 2), is the most direct means for monitoring phase transitions and identifying hydrometeors when observing a reasonably uniform cloud cover. Isolated cumuli do not present this uniformity and spatial continuity, so the depolarization is more readily measured by rotating the PRP at a fixed elevation angle pointed at some part of a convective cloud. An advantage of this method, with the configuration of the NOAA/ETL radar, is that it produces measurements of LDR and the limiting EDR at the same time.

Winter orographic cumuli developing over the ridge of Arizona's Black Hills were observed on 27 February 1995 using this method. The photograph in Fig. 3a shows the growth stages of the cumuli, which were producing "hard" liquid turrets and then glaciating. Rotations of the 95° PRP provided the EDR data in Figs. 3b,c. A quarter turn of the PRP is represented by one cycle in the data; the extreme positive and negative values represent LDR and EDR_{45-95} , respectively, which are 45° apart in the PRP rotation. The beam was fixed at $\beta = 10.6^\circ$ for the first rotations to penetrate a hard cumulus turret

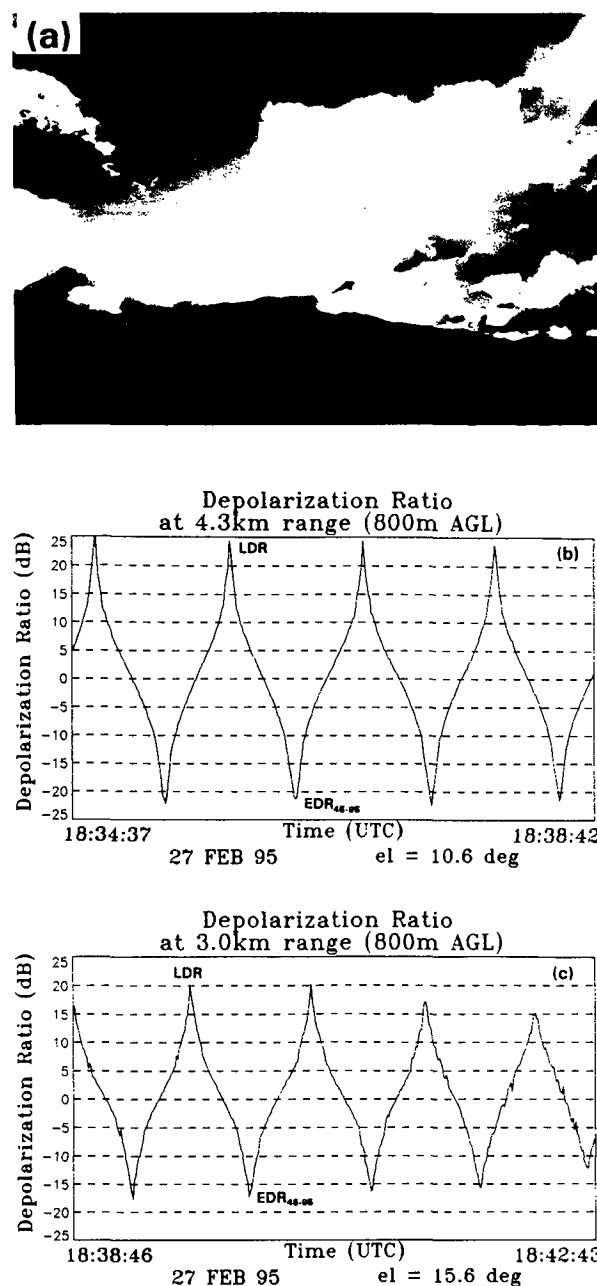


Figure 3. (a) Photograph of orographic cumulus, 1850 UTC 27 February 1995, Arizona. (b) EDR from PRP rotations at $\beta = 10.6^\circ$ and 800 m AGL in the liquid part of a cumulus; the limiting values are LDR and EDR_{45-95} . (c) Same as (b), but at $\beta = 15.6^\circ$ and 800 m AGL in the downwind, glaciating part of the cumulus (the Arizona Program-1995).

above the mountain ridge (Fig. 3b). Consistently, $EDR_{45-95} \approx -21$ dB, the theoretical value for spheres, which shows that the cloud was composed of droplets. These droplets may have been slightly deformed and canted. The +25 dB values of LDR suggest this. LDR is more sensitive than EDR to drop distortions, whereas it would reach about +35 dB with true spheres. (LDR, however, has disadvantages for ice hydrometeor observations in that it is relatively more confused by the randomness of canting angles of crystals; the use of LDR to identify either liquid or ice hydrometeors requires more investigation). The PRP spins in Fig. 3c, for the same altitude, immediately followed those in Fig. 3b, but were set at $\beta = 15.6^\circ$ to penetrate the cloud downwind where visual inspection indicated that it was beginning to glaciate. Here the dynamic range between EDR_{45-95} and LDR is steadily decreasing, such that the depolarization is increasing with time from that of spherical droplets to that of nonspherical ice particles. EDR_{45-95} approaches -12 dB, which approximates Matrosov et al.'s (1995) modeled value for short columns (axis ratio $a/b = 0.7$) at the noted β .

This case illustrates how phase transitions in cumuli, such as those induced by seeding, can be observed with polarization radar by using the rotating PRP at fixed elevations and fixed azimuth, rather than RHI scans, to follow the evolution of the spatially confined target.

3.1.3 Tracking a seeded volume

If tracking of chaff (and seeding aerosol) is coordinated with the kinds of monitoring noted above, the chaff volume tagged and identified in X-band circular depolarization might be targeted with the K_a -band radar to monitor phase change. Thus, a particular volume or parcel of cloud might be followed through its kinematic and microphysical history. The chaff is cut to half the X-band wavelength (i.e., to 1.5 cm). One design for an appropriate experiment is as follows. This design is most applicable to orographic/layer clouds and has three components: (1) mark an aircraft-seeded line or circle in clouds with dashes of chaff; (2) track the chaff in X-band to follow the seeded path and monitor reflectivity trends in the gaps between the chaff marks but through the seeded path; and (3) use K_a -band polarization RHIs through the seeded path, but between the chaff marks, to monitor the production of ice in an otherwise predominantly liquid cloud. This experiment was tried in Arizona; it worked in that a seeded circle could be tracked by following two

dashes of chaff and the reflectivity was enhanced by ice from seeding in the gaps between the chaff dashes. Ice was found in the K_a -band depolarizations, but the target was at a range too distant (with a signal too weak) to separate it from other ice in the vicinity of the seeded circle. This first try at a difficult but very informative experiment should be repeated. The experiment is detailed by Buintjes et al. (1996).

3.2 Snow Crystal Growth Habit

The elementary dependence of snow crystal habit on nucleation and growth temperature is fundamental information for determining cloud seeding potential, cloud water budgets, and precipitation efficiency prior to or resulting from seeding. The habit identification translates semiquantitatively to cloud temperature, cloud saturation, ice mass growth rates, and the related spectrum of microphysical information that ranges from particle fallspeed to probability of aggregation. Once pristine crystal habits are nucleated, the first—and sometimes only—stage of ice particle-size spectrum evolution occurs as these crystals form by vapor depositional growth. A cloud's steady production of crystals of a certain habit, as monitored in EDR and coupled with intensifying reflectivity, implies a rate of glaciation. The depolarization and reflectivity, linked to microwave radiometer measurements of the cloud liquid quantity and trend, can be used to interpret the microphysical cause of the buildup or depletion of seedable water, particularly when all of the measurements are integrated with the physics of a cloud numerical model.

Measured depolarizations, matched to theory and field samples of corresponding snow crystals, offer a means to identify various snow crystal types, as shown by Matrosov et al. (1996) and Reinking et al. (1993, 1995a,b, 1996). The clearest separation of crystal types is between planar and columnar growth habits. This is illustrated experimentally by depolarization measurements from the WISP case (25 February 1994) presented in Fig. 4 and by Reinking et al. (1996), in which columnar crystals were generated in an upper cloud layer (depolarization curve, $EDR_{45-79.5}$ versus β , for 1.3 km AGL) and dendrites were generated in an underlying layer (depolarization curve labeled 0.5 km AGL). The third curve, also from the lower cloud layer but for an intermediate altitude (0.9 km AGL), shows the transition zone where the settling columns mixed with the nucleating and growing dendrites, which lower in the cloud dominated the depolarization signature.

Figure 4 shows that the dynamic range of the depolarization curve between 0° and 90° radar elevation is much greater for the planar than for the columnar types; this facilitates the differentiation. The depolarizations by both columnar and planar crystals show an elevation angle dependence, but columnar crystals depolarize less than the planar types at low elevation angles and more at the high elevation angles, as illustrated in Fig. 4.

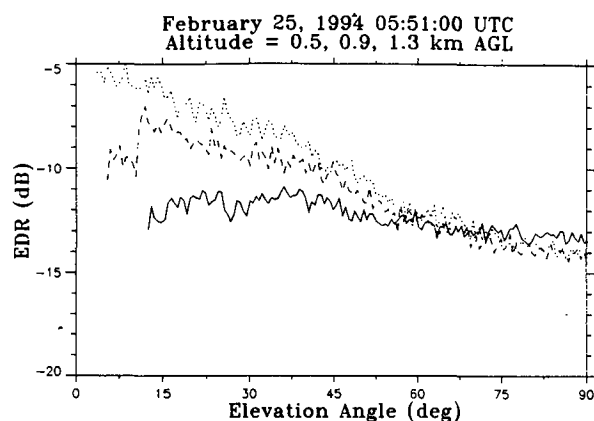


Figure 4. $EDR_{45-79.5}$ vs. β for three altitudes (kilometers AGL) in clouds (0551 UTC 25 February 1994, WISP-94). The 1.3-km AGL curve (solid) indicates a dominance of columnar ice crystals; the 0.5-km curve (dotted) indicates planar crystals. The 0.9-km curve (dashed) indicates a mix of the two types, which was verified with samples at the radar site. Irregular signals at low elevation angles are due to weak clouds.

Even some of the comparatively similar individual planar crystal habits (e.g., plates versus dendrites) or columnar crystal habits (e.g., needles versus long columns) can be distinguished in theory because of their density differences. The effects of density were modeled by Matrosov et al. (1996). In practice, the K_a -band polarization radar can measure some of these distinguishing differences. Reinking et al. (1996) presented some theoretical model scenarios that illustrate some readily distinguishable and some not-so-distinguishable crystal types, according as aspect ratio and bulk density combine to depolarize the incident radiation.

3.3 Graupel

Graupel is of particular importance to cloud seeding, not only because it can efficiently collect and remove a lot of water and thus may be a desirable seeding product, but also because it can induce

ice crystal multiplication via the Hallett-Mossop mechanism when falling through a population of droplets larger than about $25 \mu\text{m}$ (Hallett and Mossop, 1974). Substantial LW is required to facilitate riming and droplet growth to those sizes, so multiplication will likely occur where cloud LW is large. Multiplication occurring naturally will likely reduce the seeding potential below that of ice-free clouds but could enhance precipitation if induced by seeding in ice-free clouds or those in which LW is replenished faster than consumed by existing ice.

Reinking et al. (1995a, 1996) provided some experimental evidence from WISP that graupel of randomly irregular shape will produce some depolarization, such that EDR is independent of elevation angle but offset a few decibels toward zero from spheres, thus distinguishing it from drizzle. The cover of this *Journal of Weather Modification* issue illustrates the EDR from an RHI scan through a convective shower of graupel falling into a layer cloud of thick plates; the plate and graupel signatures, and the signatures where the two kinds of hydrometeors were mixed, were all separately distinguishable. These results were supported by snow crystal samples taken at the ground (Reinking et al., 1995b, 1996).

3.4 Snow Crystal Aggregation

The graupel case just noted has further importance. The snow crystal samples demonstrated that the graupel particles of 1.0–1.5 mm diameter and $\sim 1.5 \text{ m s}^{-1}$ fallspeed were effective scavengers of the 200–300 μm thick plates of $\sim 0.3 \text{ m s}^{-1}$ fallspeed (Reinking et al., 1995b, 1996). This is a mode of aggregation, although not the classic type that occurs among like-kind particles. The graupel of comparatively high fall velocity served to more rapidly and perhaps more efficiently precipitate the water substance of the plates. The process in effect was a seeder-feeder mechanism, which is sometimes usefully stimulated by cloud seeding.

When fallspeeds (v_f) of ice crystals growing initially by deposition become sufficiently different to induce collisions, a second stage of size spectral evolution can commence, which is dominated by aggregation (Mitchell, 1988). The case with graupel ($v_f \approx 1.25 - 1.50 \text{ m s}^{-1}$) and thick plates ($v_f \approx 0.2 - 0.3 \text{ m s}^{-1}$) is an extreme example. More commonly, aggregation will occur among crystals of like kind, and with smaller fallspeed differences. The efficiency of like-kind aggregation depends on habit and

therefore temperature. If, for example, dendrites are assigned a weighted aggregation probability of 1.0, then the corresponding probabilities for needle, sector, hollow column, and hexagonal plate aggregation are, respectively, 0.6, 0.4, 0.1–0.25, and 0.1 (Mitchell, 1988). Thus, by identifying the basic snow crystal habit in the depolarization signature, one can determine the relative probability that aggregation will occur.

Once unaggregated crystals are identified, the evolution of the depolarizations can track the space-time progression of aggregation within the cloud volume, as demonstrated by Reinking et al. (1993) and Matrosov et al. (1996). The distinct elevation-angle dependency of EDR for dendrites, for example, is progressively degraded such that the curve is flattened with increasing aggregation (Fig. 5). In this WISP Instrument Test (WISPIT) case of 11 March 1993, the aggregation proceeded uniformly through the entire 2.8-km depth of the cloud, except for the top 10% where the crystals were being generated. Other cases show a more gradual gradient with aggregation increasing downward.

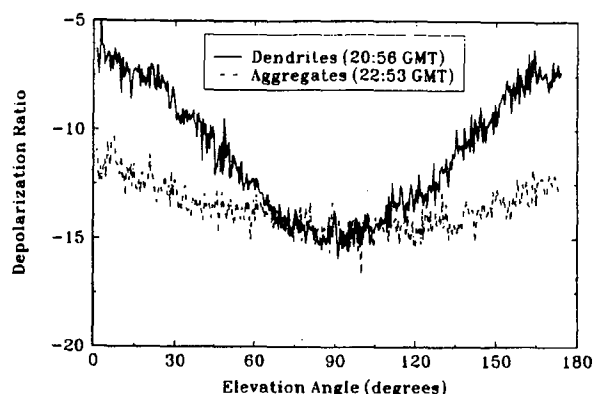


Figure 5. $EDR_{45-79.5}$ vs. radar elevation angle for single dendrites (top curve) and aggregates of dendrites (bottom curve), respectively, 2056 and 2253 UTC 11 March 1993 (WISPIT).

Aggregation enhances crystal fallspeed. Single dendrites, for example, can reach fallspeeds of only about 0.2 m s^{-1} , whereas aggregates of dendrites fall about 1 m s^{-1} (Pruppacher and Klett, 1978). Therefore, in moderate-to-strong horizontal advection, aggregation is a factor in targeting snowfall induced by seeding, and it will decrease the likelihood of sublimation losses in storm outflows. Consequently, one might prefer to seed from aircraft at specific temperatures to produce primarily dendrites at -15°C or needles at -5°C , which will aggregate with greater statistical probability than plates or solid columns that

nucleate between -8° and -12°C or -18° and -20°C . There are tradeoffs; for example, a plate will fall faster than a dendrite with the same growth time and may therefore rim and reach graupel fallspeeds faster. Numerical models can be used to resolve the optimal approach.

The key is that snow crystal growth by deposition, riming, and aggregation interact. Particularly in a population of small, pristine crystals, aggregation will substantially increase particle fallspeed and will, in turn, enhance riming efficiency to produce snowfall rates considerably greater than those by diffusion and riming alone, according to Mitchell (1990). The reasons for this are that (1) aggregation does but riming does not increase collector particle cross section substantially, and (2) the increase in fallspeed is commonly greater when aggregation occurs. The snowfall rate increase due to the addition of aggregation was approximately 60% in Mitchell's case study. He concluded that aggregation is crucial in determining cloud water removal and snowfall rates. It follows that seeding that leads to aggregation can sometimes increase snowfall, so a capability for observing this process is important.

3.5 Melting Level

The altitude of the melting layer (or bright band) is a basic radar observation that is observed with greatest clarity in depolarization. For winter scenarios, a simple depolarization measurement of this altitude defines whether a storm will precipitate rain or snow on existing snowpack. If the measurement shows rain on snow, a flood potential is indicated. This measurement should improve the real-time application and accuracy of seeding suspension criteria. This observation also provides direct verification for cloud models, which must get the types of hydrometeors and atmospheric temperature structure right to simulate the melting layer. For example, bulk water models that produce particles with excessively high terminal velocities will not properly produce a melting layer. Obviously, it is as important to correctly model the hydrometeors and temperature as it is to correctly represent the melting layer, so observations of the latter provide an accuracy check on all three.

The melting layer commonly can be observed by radar in the difference between snow and rain fall velocity or reflectivity; however, a melting layer will normally be evident in a depolarization signature even when it is not evident in reflectivity (Fig. 6 and

Dennis and Hitschfield, 1990). With the 37.5-m range resolution of the K_a -band radar, not only the altitude but also the thickness and fine structure of the melting layer can be monitored.

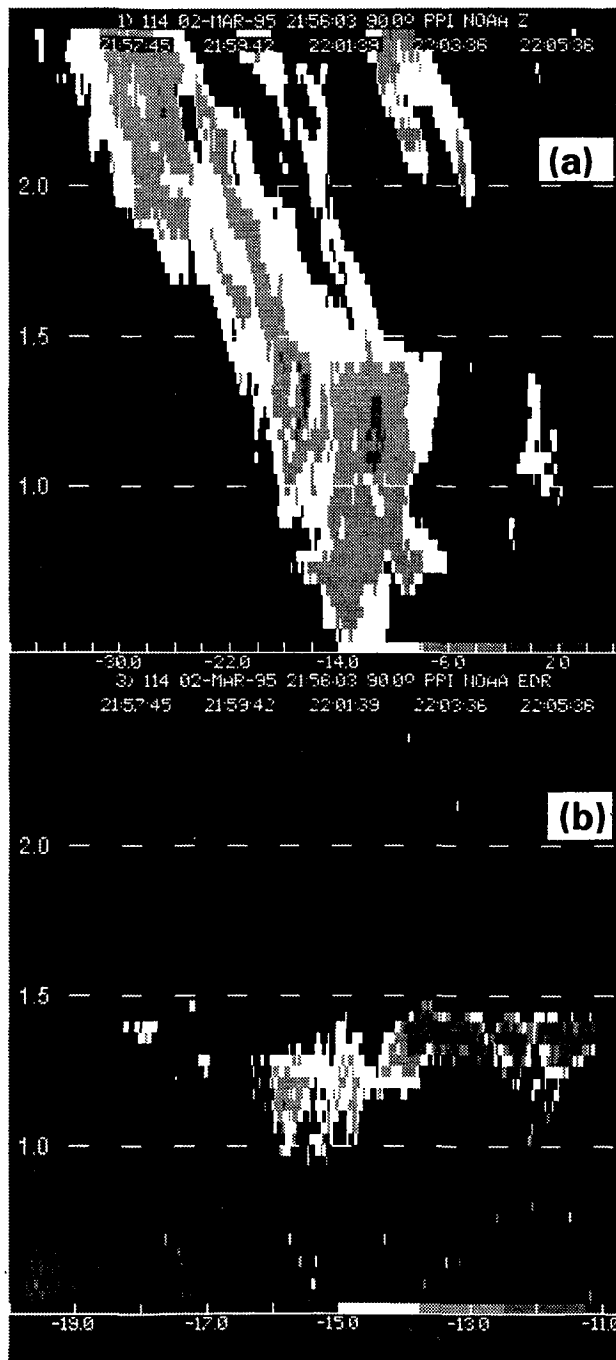


Figure 6. Time-altitude record of (a) Z_e (dBZ) and (b) EDR_{45-95} (dB) for a melting layer associated with drizzle and light rain. Vertically pointing radar, 2158–2207 UTC 2 March 1995 (the Arizona Program); time increases to the right.

Figure 6 shows Arizona 1995 data from light stratocumulus snow showers that produced drizzle and rain on 2 March 1995. The lightest drizzle (from a -16 dBZ cloud) was confirmed by an EDR_{45-95} value of ~ -20 dB. This drizzle fell from a melting layer 0.1 km thick; EDR_{45-95} was approximately -10 to -12 dB at the core of this layer and approximately -15 dB within its upper and lower boundaries. An adjacent, heavier shower from -6 dBZ mixed-phase clouds produced light rain with $EDR_{45-95} \approx -17.5$ dB (also in Fig. 6). Here the melting layer was about 0.5 km thick, and the depolarization within it ($EDR_{45-95} \approx -13$ dB at the core) was about 2 – 3 dB greater than that for the light rain below it, which at -17 dB showed slight depolarization. This case suggests that for given thermodynamics, (1) the melting layer thickness can be related to the rain rate beneath it (this is an extension of the dependence of the thickness on hydrometeor type and fallspeed; see Dennis and Hitschfield, 1990, p. 103), and (2) the depolarization within the melting layer will decrease as rain rate increases (possibly because more melting is required and liquid-surfaced ice particles persist through a deeper layer).

Massive (1 – 3 cm) aggregates, sampled on high ground, were monitored with the radar as they fell from 20 -dBZ cloud through a melting layer and produced a soaking rain on 28 February 1994, during WISP-94 (Fig. 7). The aggregates produced a bright band 0.3 – 0.4 km deep. Through a vertical distance of 150 – 200 m, $EDR_{45-79.5}$ increased from ~ -14 dB in the falling aggregates to -9 dB in the upper boundary and -4 dB at the core of the melting layer; through another 150 – 200 m, $EDR_{45-79.5}$ decreased to -9 dB in the lower boundary and -12 to -13 dB in the resultant rain.

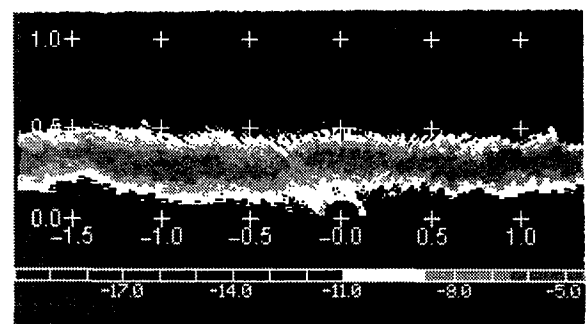


Figure 7. RHI of $EDR_{45-79.5}$ for a melting layer associated with 1 – 3 cm aggregates (2023 UTC 28 February 1994, WISP) and heavy rain; 0.5 -km grid.

The depolarization values in Figs. 6 and 7 cannot be compared directly because different PRPs were used, but this does not affect similarities. The gradients from the melting layer core to its boundaries were substantial (2–5 dB) in these cases. Notably, the bright band was shallower for the massive aggregate/heavy rain case than for the light rain shower, although both were several times deeper than that for the drizzle. The two cases occurred in different storms and locales with differing atmospheric thermodynamics. In each case, the rain depolarized the transmitted radiation 2–3 dB more than the nondepolarizing, quasi-spherical drizzle. This differentiation in itself is a qualitative measure of precipitation intensity as well as relative hydrometeor size.

4. AN ILLUSTRATIVE CASE OF TRANSITIONS WITHIN A CLOUD VOLUME

This case was not selected for the purity of its polarization signatures. Rather, it was chosen to illustrate how many of the concepts discussed so far in this paper may be integrated and applied to decipher the physics of a cloud and its seeding potential. An ensemble of the K_a -band radar, scanning microwave radiometer, and rawinsondes were used to obtain the measurements. The transitions within the cloud volume are described and their significance is interpreted.

The reflectivity and depolarization of a precipitating orographic gravity wave cloud that occurred in Arizona on 4 March 1995 are shown in Fig. 8. The cloud reflectivity (Fig. 8a) was ~6–10 dBZ through a 2-km depth of the wave upwind but only through a 0.5-km depth downwind, except beyond 10-km range where precipitation and a melting layer were observed. Local soundings indicated that the temperatures were of the order of -3° to -4°C in the core of the wave cloud at its trough and as cold as -7° to -9°C in the core at the wave crest.

Spatial gradients in depolarization (EDR_{45-95}) along the wind within the cloud are easily noted in Fig. 8b (the ridge of Arizona's Black Hills and the upwind direction from the radar are to the left and the Mogollon Rim is downwind to the right). Scanning microwave radiometer RHIs during this wave event showed that the path-integrated LW content decreased to a minimal value (≤ 0.1 mm) in the downslope flow from the ridge toward the valley remote-sensing site (antenna elevation angles of 15° to about 60° in Fig. 9), but increased to ~0.5 mm with subsequent lift

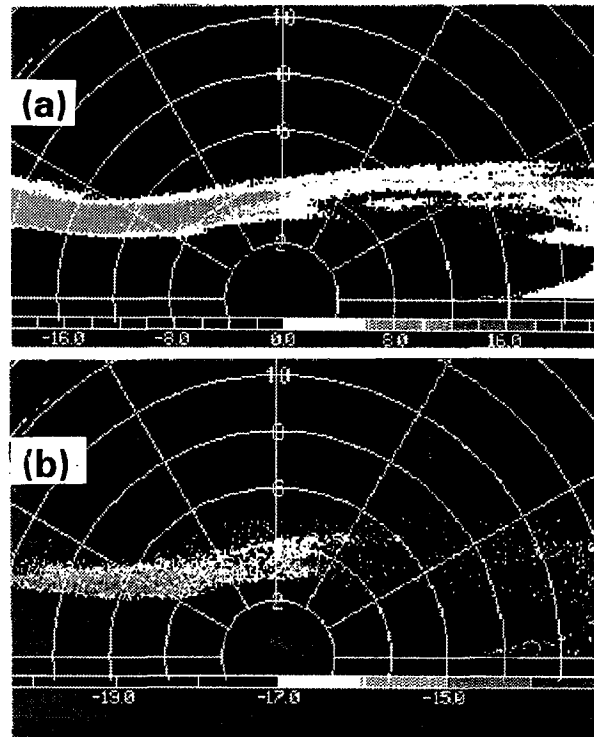


Figure 8. Spatial variation of (a) Z_e (dBZ) and (b) EDR_{45-95} (dB) within a gravity wave cloud; wind blew left to right; azimuth 250° is to the right; 2-km range rings, 0430 UTC 4 March 1995 (the Arizona Program).

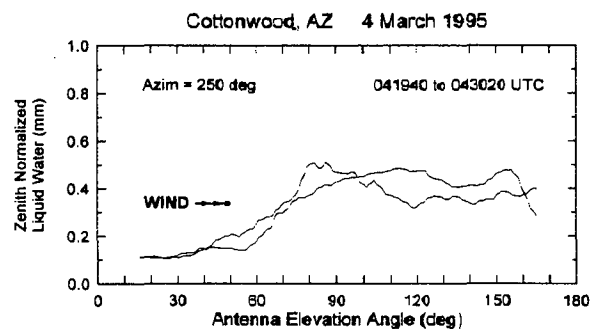


Figure 9. Scanning microwave radiometer measurements of path-integrated cloud liquid water (millimeters) normalized to zenith and corresponding to the wave cloud depicted in Fig. 8. The 250° azimuth (left in figure) pointed into the wind.

in the wave's updraft directly upwind and over the site (angles 60° – 90° and greater, Fig. 9). A half millimeter of liquid is very substantial for winter orographic clouds. However, this was a low value for this wave case, which at other times persistently produced 1–2 mm. In related studies, the vertically integrated LW has been observed to reach values of the order of 1 mm, but the probability of it exceeding 0.2 mm was found to be only 10% in the continental interior orographic storms in Utah and about 20% in the marine-influenced storms over the Sierra Nevada in California (Snider et al., 1986; Reynolds, 1988; Reinking and Meitin, 1989). Notably, the mountain slope component of LW, but no wave component, was accounted for in those studies.

Upwind of zenith at low elevation angles, the hydrometeors significantly depolarized the radar signal. Here, EDR_{45-95} was approximately -15 dB to -17 dB near $\beta = 15^{\circ}$ and -14.5 to -15.5 dB near $\beta = 25^{\circ}$. Thus, the depolarization increased slightly with increasing elevation angle. Contrary to this trend, above 25° , EDR_{45-95} steadily became more negative (depolarization decreased) toward and through zenith to values averaging approximately -17 dB, and subsequently to $\sim$ -18 to -20 dB between $\beta \approx 45^{\circ}$ and 30° in the downwind part of the wave. At the observed downwind extreme, depolarization slightly increased again (the average was $EDR_{45-95} \approx$ -18 dB, with -16 dB minima).

The following interpretation of the spatial variations in depolarization seems most likely. Ice crystals of regular growth habits, which could substantially depolarize the radar's signal, had formed in flow over the upwind ridge and were descending in the leeside flow. The depolarization ratios and their trend to less negative values at increasing elevation angles (between 15° and 25°) indicate that these crystals were columnar, according to the theoretical model of Matrosov et al. (1995). The cloud temperatures support this deduction. The crystals probably were sublimating and therefore deforming in the downslope flow (where the measured LW was within the noise level); thus, the greater depolarizations expected from columnar crystals that advected to higher elevation angles (as measured with the 95° PRP) did not materialize.

The Arizona seeding hypothesis is designed to seed the wave from the foehn gap. The depolarization measurements reveal that the gap depleted the LW but did not open in this case; rather, the wave trough provided a conduit for the crystals

from upwind to feed into the high LW of the updraft. This influenced but did not maximize the natural precipitation efficiency; the LW that formed in the updraft persisted downwind. (The gap did open in other cases to cut off the crystal input through this conduit).

The crystals, modified by evaporation in the subsidence, subsequently rimed with LW formed in the updraft. They continued to become more spherical, and emerged from the wave as irregular, graupel-like hydrometeors corresponding to the measured $EDR_{45-95} \approx$ -18 to -20 dB (recall that for spheres, $EDR_{45-95} \approx$ -21 dB). The graupel were smaller and/or less numerous than the evaporating crystals in the influx; this accounts for the diminished reflectivity in part of the cloud downwind beyond zenith. During further transit, the irregular ice or new ice nucleated in the relatively cold wave crest used an additional 5 min to grow in the enhanced LW. Some new crystals of regular growth habit and shape formed to sparsely enhance depolarization to as much as -15 dB at the downwind reaches of the radar scan (range 12.5 km). This growth led to precipitation. However, the downwind extension of LW in quantities of 0.3–0.4 mm (through 165° elevation, Fig. 9) indicates that cloud ice was not consuming all the water available for precipitation. Thus, a positive seeding potential was indicated.

The height of the melting layer, noted by its depolarization, was about 1.5 km AGL (relative to the Verde Valley radar site). The elevation difference between the radar site and the downwind Mogollon Rim is about 1.1 km; therefore, with the melting layer information alone, it is confirmed that liquid precipitation was falling on an existing snowpack. The depolarization measurements indicate a modest melting layer $\sim$ 300 m deep that would be associated with more than drizzle; also, depolarizations below the melting layer confirm the precipitation as drops of rain ($\sim$ -17 to -18 dB), as distinguished from drizzle ($\sim$ -21 dB). The reflectivity of $\sim$ 8 dBZ exceeds that of many winter ice clouds but is a small value for rain. Together the radar data indicate a light rain rate. At the same time, a rate of 0.2 mm h^{-1} was measured at a surface site within the precipitation zone indicated by the radar in Fig. 8. The rain eventually became persistent; it intensified significantly on the slopes of the rim between 0100 UTC and 1500 UTC on 6 March. One focused seeding experiment was conducted at 0600 UTC with a 5-min release of 10-AgI flares to test methods, dispersion, and ice generation. The dispersion was tracked with chaff

and occurred through a restricted volume just 10 km wide. The various data verify with certainty that, as intended, this had no measurable influence on either area-wide or storm total precipitation (Bruitjes et al., 1996). Nevertheless, seeding suspension criteria were continuously evaluated during the heavy rain period. Subsequently in this stage of what had become a prolonged event, flood warnings were issued. At approximately 0730 UTC on 6 March 1995, seeding experiments were appropriately suspended. As the rain on the snowpack continued for several more hours, flash floods swept from the Mogollon Rim into the Verde Valley.

5. CONCLUSIONS

A number of microphysical features of cloud evolution that are fundamental to glaciogenic cloud seeding can be identified and followed through transitions using K_a -band (8.66 mm) elliptical polarization radar. Initial cloud phase change, which is a transition in hydrometeor shape and density, can be detected. The various forms of cloud particles and precipitation are separable according to their characteristic values of the EDR. Drizzle, rain, pristine ice crystals, graupel, and aggregates have been examined. Clouds of pristine ice crystals are readily differentiated from drizzle, rain, graupel, and aggregates. Columnar and planar growth habits of the pristine crystals are readily separable, and EDR signatures provide some capability for identification of the individual columnar and planar habits, thus defining associated in-cloud thermodynamics and crystal mass growth rates according to known microphysics.

Differentiating among drizzle, rain, graupel, and aggregates is more challenging. However, drizzle, which is quasi-spherical, does not depolarize, whereas the other hydrometeors in this group do. The presence of a melting layer, lucidly depicted in depolarization, will readily separate drizzle and rain from graupel and aggregates. Aggregates, no matter how large, usually show at least a slight dependence of depolarization on elevation angle, apparently because a few single crystals are normally present in the mix. Graupel has not shown this dependence, but rather a 1–3 dB offset in depolarization from drizzle.

In all, the prevalence of individual hydrometeor types, spatial and temporal gradients among them, transitions from pristine crystals to aggregates, and influences of graupel within a cloud volume can be monitored. Therefore, continuous depolarization

measurements can provide information on the potential for seeding and rates and mechanisms by which the cloud converts its water to precipitation under seeded versus unseeded conditions.

Observations of the melting layer, lucidly depicted, have significant application in seeding suspension criteria associated with rain on snow. The fine structure of the melting layer, including thickness and the depolarization values within it, can be monitored with clarity to provide some qualitative information on precipitation intensity, which adds to information from reflectivity. Rain falling from the melting layer can be distinguished from drizzle by differences in the magnitude of depolarization.

All of these measured microphysical features have application in cloud model validation, and in turn, cloud models can integrate radar measurements with the appropriate physics to more quantitatively define cloud seeding potential and effects.

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