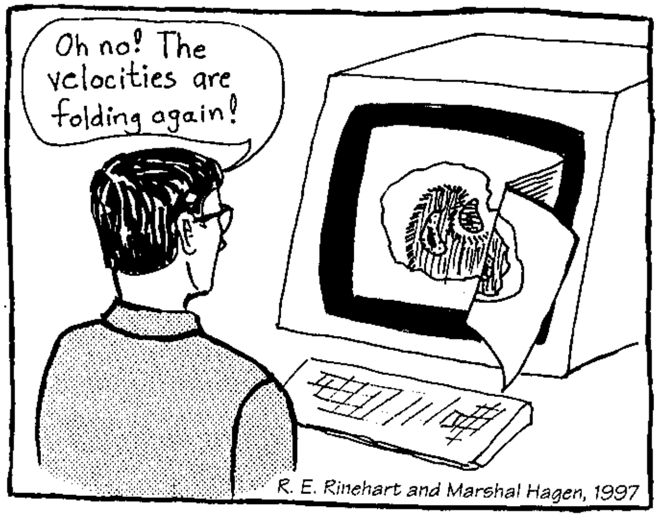
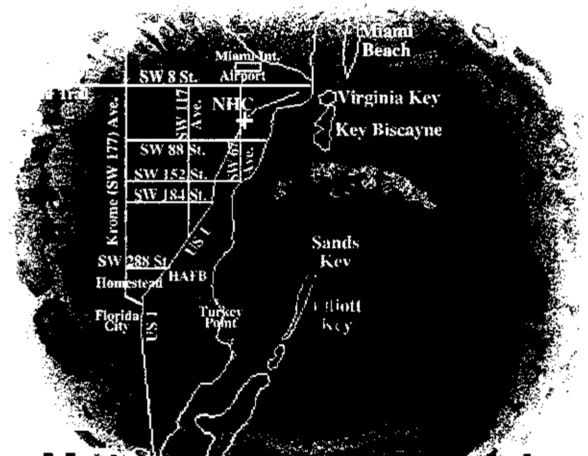
directions. In reality, though, few people probably use spectrum width information because reflectivity, velocity and polarization products are more important for meteorologists to keep track of and use.





Meteorological Targets

The ability to detect storms and other weather phenomena is perhaps one of the most valuable uses of radar. Severe storm and tornado warnings, hurricane observations, flood warnings and wind shear warnings are very frequently based on radar observations and likely result in the saving of lives and property every year. In this chapter we will explore how radar detects various weather events and some of the constraints and limitations in this activity. Since most of the meteorologically detectable echoes come from hydrometeors, we will begin with those and then consider other sources of echo that provide meteorological information.

Clouds

Clouds are usually not detectable by most radars but can, under some circumstances and with some radars, provide detectable echoes. For this discussion, let us distinguish between precipitating and non-precipitating clouds. If a cloud is not precipitating, either the particles within the cloud are too small to fall downward, the cloud is a new cloud that has not yet had sufficient time to produce precipitation-sized particles, or there is a constant, steady upward air motion that keeps the cloud particles suspended at approximately a constant altitude.

Clouds are composed of very small water droplets, ice crystals or both, depending upon the temperature and other factors. Many clouds that start out as liquid hydrometeors eventually change into either all ice clouds or a combination of both ice and supercooled liquid water droplets (i.e., liquid droplets whose temperatures are colder than 0°C; supercooled droplets can exist at temperatures between 0°C and -40°C; liquid water does not exist below -40°C).

The sizes and concentrations of cloud droplets have been studied for many years. The size distribution within the cloud depends upon the cloud type, age and height within a cloud and the geographic location. In general, the farther from cloud base, the larger the droplets are. As a cloud gets older, droplets usually get larger. There is also a distinct difference between clouds that form over or near oceans and those that form well inland. Maritime clouds tend to have fewer droplets per unit volume than do continental clouds. Droplet sizes for both types range from perhaps 5 μm to 100 μm or more. Figure 8.1 (Fletcher, 1966) shows the mean droplet size distributions for various cloud types. Drop size distributions from other places and cloud types would differ from these in detail but would probably have similar general characteristics.

The radar reflectivity factor for clouds is generally quite weak. Since $z = \Sigma N_i D_i^6$, we can calculate reflectivities that might result from some clouds. Table 8.1 gives the reflectivity that would come from a continental cloud.

The overall reflectivity of -17 dBz is quite weak and would not be detected by most weather radars beyond a few kilometers; some radars could not detect echo this weak at any range. Weather radars on board most aircraft are not sensitive enough to detect clouds.

There is one interesting point from the above table which should be mentioned. This is the fact that the contribution to reflectivity from the small droplets, even though they outnumber the larger drops by one or more orders of

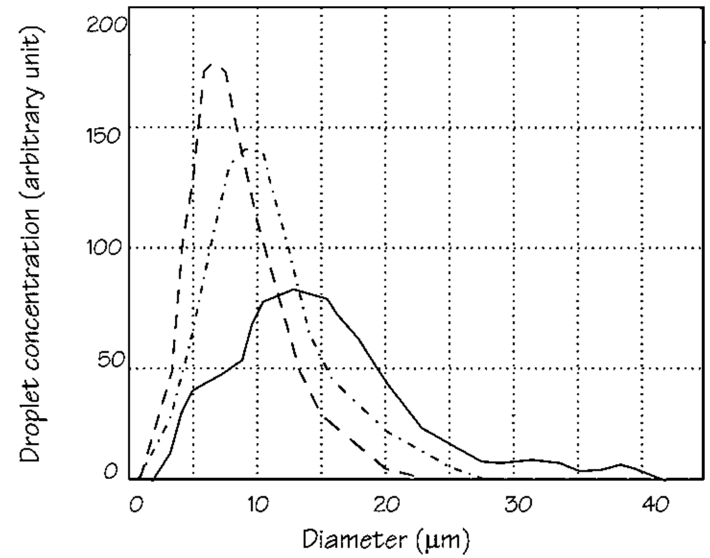


Figure 8.1 The mean droplet size distributions for various cloud types. Cumulus congestus (solid); altostratus (dash, dot, dash); stratus (dash). Based on Fletcher, 1966.

Table 8.1 Contribution to radar reflectivity factor by each individual cloud droplet size. Size spectra is based on the continental type cloud of Fig. 8.1.

Dia (μm)	No./cm^3	$N D^6$ (mm^6/m^3)
5	100	$1.56 \cdot 10^{-6}$
10	100	$1.00 \cdot 10^{-4}$
15	50	$5.69 \cdot 10^{-4}$
20	25	$1.60 \cdot 10^{-3}$
25	10	$2.44 \cdot 10^{-3}$
30	5	$9.19 \cdot 10^{-3}$
35	1	$4.01 \cdot 10^{-3}$
	Total =	$1.80 \cdot 10^{-2}$ mm^6/m^3
		= -17.4 dBz

magnitude, is generally negligible. Most of the reflectivity comes from the largest droplets. This is a consequence of the diameter-to-the-*sixth*-power term in the equation for reflectivity. The same feature is true of raindrop size distributions.

Rain

Rain is very easily detected by most radars. Rain can come in a wide variety of intensities, from light drizzle (the "Oregon mist" of the west coast) to the near-blinding downpours in severe thunderstorms. The measurement of rain by radar is one of the more important quantitative uses of radar. In the following sections we examine some of the properties of rain and its detectability by radar.

Raindrop size distributions

Raindrop size distributions have also been studied extensively for more than 50 years. There have been a number of techniques developed to sample these distributions. One of the most-used techniques was a raindrop camera which photographed a volume of rain with enough resolution that individual raindrops could be measured. From these or other size distributions, rainrate (e.g., mm/h), liquid water content (e.g., g/m^3) and radar reflectivity (mm^6/m^3) are easily calculated.

Figure 8.2 shows three raindrop size distributions collected at Ottawa and used by J. S. Marshall and W. McK. Palmer (1948); these are possibly the most widely known raindrop size distributions in all of meteorology, certainly in radar meteorology. They were the basis for the relationship known as the Marshall-Palmer distribution. This is a convenient relationship which gives an approximate size distribution for raindrops as a function of rainrate. As such, it is useful for various analytical exercises and for deriving other relationships.

The Marshall-Palmer relationship is given by the following:

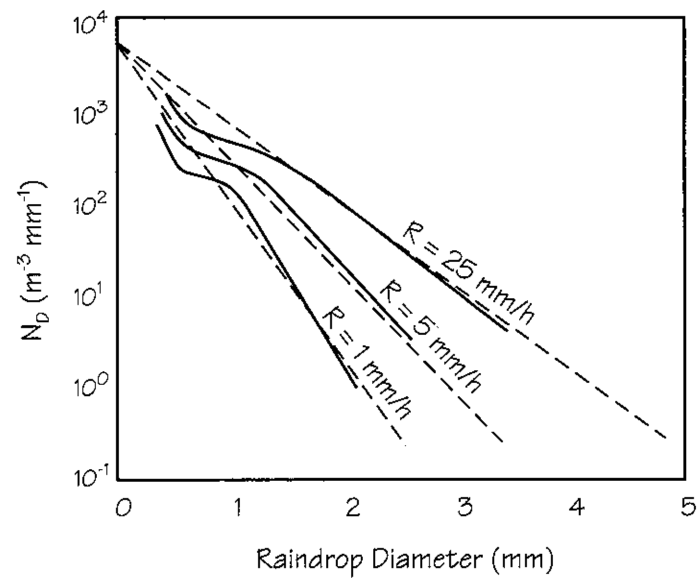


Figure 8.2 Marshall and Palmer drop-size distribution functions (dashed lines) compared with the results of Laws and Parsons (solid lines). Based on Marshall and Palmer, 1948.

$$N_d = N_0 e^{-\lambda D} \tag{8.1}$$

where N_0 = 8000/(m^3 mm) is the point where all of the dashed lines converge on Fig. 8.2, D is drop diameter (mm), and λ is given by

$$\lambda = 4.1 R^{-0.21} \tag{8.2}$$

where R is rainrate in mm/h.

Using this relationship and a specific rainrate, we can calculate the number of drops per unit volume and per unit drop size interval for any particular raindrop size. The size distribution can then be used to calculate the radar reflectivity or liquid water content of the rain.

Z-R relationships

From the above discussion it should be obvious that there is a relationship between rainrate and radar reflectivity. Experimentally measured drop-size distributions have been extensively used to calculate both radar reflectivity and rainrate. By plotting rainrate against reflectivity or by correlating these statistically, we can determine the relationship between these two parameters. The most commonly used mathematical relationship between reflectivity and rainrate is the empirical power-law relationship

$$z = AR^b \tag{8.3}$$

where R is the rainfall rate (mm/h), z is the radar reflectivity factor (mm^6/m^3), and A and b are empirical constants. In reality, however, we measure reflectivity and use it to calculate the rainrate, so the equation could perhaps be more appropriately written as $R = \alpha z^{\beta}$, where α and β are again emperical constants. Battan (1973) lists more than 60 experimentally determined *Z-R* relationships; many more have been determined since then. While each applies to a specific time and place and rainstorm, there is often little difference from one to another. Unless a specific *Z-R* relationship exists for a given situation, there are only three or four that really need to be used. And even these are not dramatically different.

The most commonly used *Z-R* relationship is also due to Marshall and Palmer. It is $z = 200\ R^{1.6}$. This has formed the basis for much research and been widely used to calculate rainfall amounts from radar data. Indeed, as will be discussed in Chapter 10, radar is a very useful way to measure rainfall over large areas; *Z-R* relationships are the backbone of this activity.

Radars can provide quantitative information on rainfall with excellent resolution. The radar reflectivity factor

from rain varies from perhaps 20 dBz (100 mm^6/m^3) to more than 50 dBz (100000 mm^6/m^3). Reflectivities as high as 75 dBz have been measured in storms, but reflectivities higher than about 55 dBz are frequently associated with hail; the higher the reflectivity, the more likely it is that hail is present and the larger the hail is likely to be.

Radar signal processors can resolve moderately small differences in reflectivity. Many radars have a dynamic range (the difference between the strongest and weakest powers that can be detected, usually expressed in decibels) on the order of 80 to 90 dB; they frequently divide this range into 256 parts, giving a resolution on the order of 1/3 dB per measurement interval.

DVIP levels

This much resolution is not always available, however. When digital signal processing was first applied to radar, it was not possible to produce nearly as many levels of intensity as we now take for granted. Some of the early signal processor had only a handful of intensity levels.

The National Weather Service was able to divide the possible range of storm reflectivities into a relatively small number of intervals and get some very useful results. The processor used to do this was named the digital video integrator processor or DVIP. They divided storm intensities into six intervals but made the division on the basis of rainrates rather than radar reflectivity factor. Further, the divisions were made before the general change from English to metric units; rainrates in in/h were used for these divisions. Table 8.2 shows these intensity levels and the corresponding rainrates and reflectivities.

The same intensity levels used with NWS radars are used in a slightly modified form in aircraft radars. Radars on board aircraft use only four levels of intensity. These are levels 1, 2, 3 and 5.

When examining a radar display using DVIP levels,

Table 8.2 DVIP intensity levels based on rainrates and the corresponding radar reflectivity factors. Reflectivities are based on $z = 200\ R^{1.6}$. Note that R is mm/h in the Z-R relationship while it is in in/h below.

DVIP level	R (in/h)	Z (dBz)
1	0.1	29.5
2	0.25	35.9
3	0.5	40.7
4	1.25	47.0
5	2.5	51.9
6	4.0	55.1

be sure to recognize that if an echo shows at a given level, it means that the intensity shown is at least that level, probably stronger. For example, if an echo shows with a DVIP level of 3 on an aircraft radar, it means that the storm has a rainrate of at least 0.5 in/h ($Z \geq 40.7$ dBz) but less than 2.5 in/h ($Z \leq 51.9$ dBz). As a matter of safety, it would be prudent to assume that the detected reflectivity or rainrate is closer to the upper limit (i.e., the next DVIP level) rather than near the lower limit.

Snow

On American television weathercasts, it is quite common in the wintertime to hear forecasters say that radar doesn't detect snow very well. In some ways they are right, I suppose, but in another way they may be laying the blame on the radar instead of where it belongs – on the storms. Snow storms try to hide from radar as much as possible, it seems. Let's see why this is the case.

In reality, snow is easily detectable by radar. But there are some important differences between snow and rain, however. One of the major differences is that the precipitation rate for snow is usually much less than it is for rain.

This comparison is based on the "water equivalent" precipitation rate. That is, the rate of snow is sometimes converted to liquid and then measured similarly to rainfall in mm/h of melted water. The maximum (saturation) amount of moisture in the atmosphere is a very strong function of temperature; at warm temperatures in the atmosphere there can be much more water vapor than at cold temperatures. One consequence of this is that the heaviest snowfalls (and rainfalls) occur at the warmest temperatures. The heaviest snows often fall when the temperature at the surface is just above the melting temperature of ice, i.e., 33° to 36°F. Of course, the temperature above the surface is colder than this; otherwise the precipitation would fall as rain instead of snow.

The second major difference between snow and rain is that the dielectric constant of ice is less than the dielectric constant of water. The $|K|^2$ term in the radar equation for beam-filling meteorological targets has a value of 0.93 for water but only 0.197 for ice. [Note that both of these numbers depend slightly upon radar frequency and/or temperature; for most meteorological radars and temperatures, we can ignore these variations.] Because of this difference, the power received back from snow and ice is about 7 dB less than it would be if a radar were looking at liquid precipitation. While ice crystals are usually larger than cloud droplets, ice crystals appear to a radar as though they are solid ice spheres of the same mass (Battan, 1973). Consequently, their larger size is not as great an advantage as it would seem because their density is usually much lower than that of pure water or solid ice.

Let's consider a numerical example. Color Figures 13 and 14 show two snow situations near Denver, Colorado. Radar reflectivities for these storms were on the order of 20 to 25 dBz. For the WSR-88D radar, the radar constant from Appendix C is 64.9 dB; the minimum detectable signal power is -113 dBm. Equation 5.19 can be used to calculate the maximum range at which an echo of 20 dBz can be detected.

Equation 5.19 is

$$Z = C_3 + P_r + 20\log(r)$$

Substituting in values and solving for range r gives $r_{max} = 2540$ km. That is much farther than the tallest thunderstorm has ever been detected! So, low reflectivity is *not* a reason why snow cannot be detected! But echo height is.

The primary reason snow is not always detectable by radar is the shallow height of typical snow storms. Snow storms are usually much lower than most rain storms, especially the very tall thunderstorms that produce rain and hail. Snow storms are often very widespread in area but may only extend a few thousand meters above the surface. A storm that is 1500 m high would be below the radar beam at distances beyond about 120 km (assuming that standard refraction applies and that the radar used a minimum elevation angle of 0.5°).

All of these effects tend to make snow less easily detectable than rain. Whereas rain can usually be detected to very long ranges with a radar, snow is usually not seen out to the maximum range of a radar.

When the NEXRAD network was established in the United States, thunderstorms were the main concern with less thought given for the detection of snow storms. The distance between radars in the existing network works fine for thunderstorm detection, but they are too far apart to detect all of the snow that falls across the country. On snowy days, it is quite easy to see gaps in the radar coverage when it is known that the snow is falling between radar sites.

Bright band

It is a fact of meteorology that much precipitation forms through an ice or "cold" process rather than as an all-water or "warm" process. Much of the rain that falls to the ground begins as ice or snow. In the transition from snow

aloft to rain at the surface, some changes take place that have interesting consequences on what a radar sees.

As mentioned before, the reflectivity from ice is less than that from water for particles of the same diameter (or approximately the same mass). There is one other difference between snow and rain that must be mentioned, and this is terminal velocity. The terminal velocity of a freely falling object is the constant velocity that occurs when there is a balance between the force of gravity pulling it downward and the aerodynamic drag acting to slow it down. The terminal velocity of a particle depends upon its density and shape as well as the density and viscosity of the atmosphere. Spheres and other smooth objects fall faster than rough objects (of equal mass). Dense objects fall faster than light objects (of equal size). Objects fall faster high in the atmosphere where density is less than near the Earth's surface where atmospheric density is higher.

So, with that as background, what happens when snow falls and melts, becoming rain? Above the melting level in the atmosphere (i.e., above the 0°C isotherm), snow will fall at a relatively slow terminal velocity. As soon as it reaches the melting level it will begin melting. Since the snow is falling through temperatures slightly above freezing temperatures, it will start to melt, but from the outside toward the inside. This means that the extremities of the snow will melt first. When enough melting has taken place, the snowflake will have started to develop a water coating while still remaining moderately large and irregularly shaped. Thus, to the radar a melting snowflake will start to look like a large, slowly falling water droplet. Since the power received by the radar is proportional to $|K|^2$, the change from ice to water initially increases the reflectivity by as much as 7 dB.

As the water-coated snow continues to fall and melt, its size decreases and its terminal velocity increases. A consequence of the first effect is that the reflectivity decreases somewhat, depending upon the change in effective diameter between the snow and the water drop. The results of the second effect is that the drops leaving below the melting level move faster than those coming into it. This decreases the number density or concentration of snowflakes (number per cubic meter) somewhat, and this further decreases the reflectivity in this region.

If we combine all of these effects, the reflectivity has the following characteristics. (See Figs. 8.3 and 8.4.) If we start with a given reflectivity in the snow above the melting level, there is on the order of 5 to 15 dB increase in reflectivity from the snow to the maximum signal received. Below this maximum the reflectivity will decrease 5 to 10 dB.

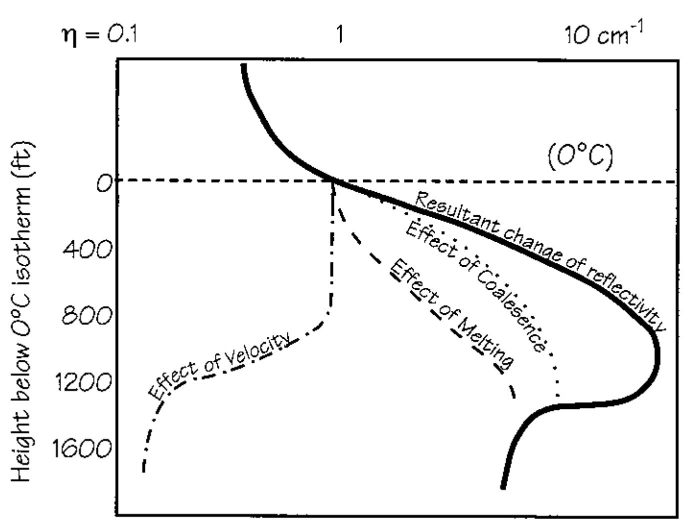


Figure 8.3 Schematic drawing showing the effects of particle coalescence, melting, and changes in the terminal velocity on radar reflectivity through the bright band. Based on Austin and Bemis, 1950. Zero height is the melting level. Radar reflectivity η is given along the top of the figure.

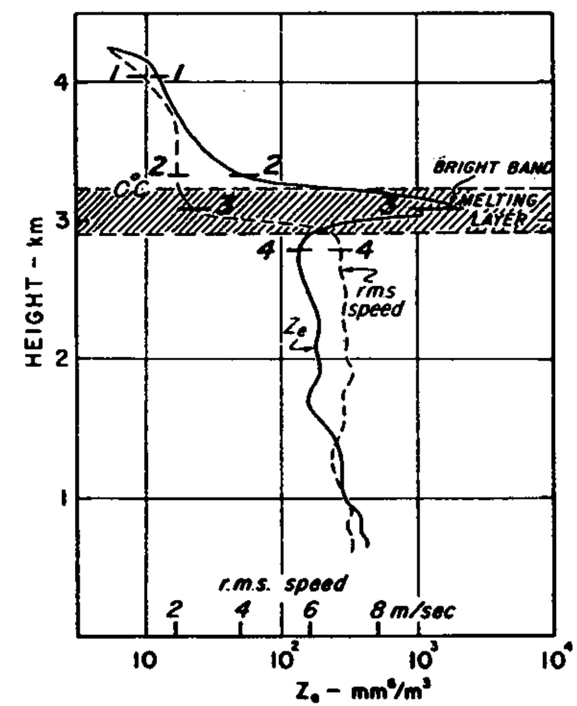


Figure 8.4 Simultaneous profiles of reflectivity factor z and root-mean-square particle fall speeds in light (1 mm/h), steady precipitation with a bright band. Based on Lhermitte and Atlas, 1963.

In any case, the reflectivity below the maximum is usually higher than it was in the snow above.

The appearance of this phenomenon on a radar display depends upon the kind of display being used. On an RHI, it would be a horizontal layer of enhanced reflectivity just below the melting level. On older analog (i.e., monochrome) display tubes, this level was usually brighter than other regions; hence the name "bright" band. On modern displays that show reflectivities using color, the bright band would be a colored band of higher reflectivity.

On a PPI display, especially for a radar located at ground level and a bright band located somewhere above the surface, it is necessary to tilt the antenna up somewhat so the beam will intersect the bright band at a shallow angle. Then, as the antenna scans around in azimuth, the radar will display a ring-like region of enhanced reflectivity. The melting level will be at the height corresponding to the most distant part of the bright-band region.

Bright bands occur primarily during stratiform or stable situations. When strong convection is present, the same physics apply, but the transition between snow and rain is often so chaotic as to be undetectable most of the time. When echoes are widespread and acting more sedately, it is much easier to detect bright bands. During the decaying portions of thunderstorms, however, bright bands will often be detected; their presence is usually an indication that the storm (or at least that portion of the storm containing the bright band) is dying out.

I was surprised to observe bright bands in decaying thunderstorms in Kenya while working on a hail-suppression cloud-seeding project. The radar was located so close to the equator that almost half of its coverage area was north and the other half south of the Equator. When a bright band became evident in a storm, it usually meant that the storm was no longer a hail threat to the region; operations for the day usually ended about the time the bright band appeared.

Hail

Hail is defined as precipitation in the form of ice that has a diameter of at least 5 mm. It almost always occurs in thunderstorms but can fall from rainstorms that do not produce lightning and thunder; this is moderately rare, however. On the other hand, many thunderstorms produce lightning and thunder but no hail. Some people estimate that 85% of all thunderstorms contain hail at least during part of their lives.

Hail ranges from 5 mm to about 10 cm in diameter. On the other hand, the former United States' record hailstone that fell near Coffeeville, Kansas, on 14 September 1972, was 14 cm across its longest dimension (see Fig. 8.5). An even larger one fell on 22 June 2003 near Aurora, Nebraska. It had a diameter of 7 in (18 cm) and a circumference of 18.75 in (47.6 cm). There was a 1-kg hailstone reported in Bangladesh in 1986, and that's even heavier than either the Kansas or the Nebraska stones. Really large hailstones are rather rare, however, so they don't usually occur in large numbers.

Figure 8.5 Photograph of a plaster-cast model of the hailstone that fell near Coffeeville, Kansas, 14 September 1972. Until recently, this was the largest hailstone ever collected in the United States. The scale at the top is in centimeters. Nancy Knight, NCAR, was kind enough to give me this replica years ago.

Just as cloud droplets and water droplets have different sized particles present at one time, hail also falls with a size distribution that depends upon the storm that produced it. Because hail can vary so much from the smallest to the largest stones and stones fall at velocities that depend upon their sizes, it is not unusual for the largest stones to fall out first, followed by smaller and smaller stones. The gravitational sorting that produces this, along with the fact that most hailstorms are moving along at moderate speeds, can combine to make one point on the ground have large hail while nearby locations will have much smaller hail.

The terminal velocity of hail, as mentioned, depends upon the hailstone diameter. It also depends upon the shape of the hail (i.e., its "drag coefficient") and on the density of the air. Measurements and/or calculations of hail terminal velocity have usually found that hailstone terminal velocity can be expressed in a power relationship of the form $V_t = A\,D^{0.5}$, where D is hailstone diameter (usually in cm) and V_t is in m/s; A is an empirical constant. One set of measurements found a value of 11.45 for A (Matson and Huggins, 1979). This applies at ground level; higher in the atmosphere, where air density is lower, the terminal velocity would be proportionally higher.

The reflectivity from hail depends upon whether the outside surface is wet or dry or if there is any water enclosed in the hail (i.e., spongy hail). Dry hail has a lower reflectivity than wet hail of the same size. This is again a consequence of the different dielectric constants of ice and water. The reflectivity from a single hailstone (or collection of stones falling through similar conditions at the same location) can change as it falls from above the melting level to below the melting level.

Another complication for hail is that it is often large enough that Rayleigh scattering conditions do not apply. That is, the hailstones are in the Mie region. For 3-cm and 5-cm wavelength radars, almost all hail is in the Mie region;

small hail detected by 10-cm wavelength radars would still be in the Rayleigh region, but large hail would be in the Mie region. Chapter 10 discusses the use of dual-wavelength radar (i.e., a set of two collocated radars with different wavelengths) specifically for the detection of hail.

One consequence of having hail in the Mie region is that the backscattering cross-sectional area of a hailstone can actually increase as the hailstone melts and gets smaller. This effect might not be detectable with a radar, however, because the radar pulse volume is usually so large that hundreds if not thousands of individual hailstones will be contributing to the received power at the same time; the effects of an individual stone would be less important because of the large number of stones present.

Finally, one characteristic of hail when it is falling is that it often tumbles. Raindrops, because they are liquid, have a shape that is determined by aerodynamic and gravitational forces (plus surface tension). On average, they have a shape that depends mostly on diameter. Hailstones, on the other hand, are mostly solid. They can and sometimes do tumble. Others, however, do not tumble. Figure 8.5 is a photograph of a plastercast copy of the hailstone that fell at Coffeeville, Kansas. It appeared to have had a preferred orientation. The bottom of this stone was moderately smooth while the top had fingers which were likely formed when water flowed from the bottom to the top and refroze.

Besides falling with a preferred orientation or tumbling as they fall, hailstones can also follow at least one other trajectory. During a hailstorm that occurred on Fathers' Day, 1976, in my house in Boulder, Colorado, I went outside to try to catch some of the largest hailstones I had ever seen falling. [I was unsuccessful in this attempt, but, hey, it was worth trying.] The largest stone was 32 mm across. They were visible falling from moderately high distances, so I had a few seconds to get under each one. I saw two that followed a helical path. When I picked these two up, they were shaped somewhat like coins. They were round horizontally but moderately flat. The paths they fell with were akin to some coins dropped into water where they, too, spiral in their downward trajectory. I have never seen this again, nor have I ever heard of anyone else who has witnessed this kind of trajectory. Smyth *et al.* (1999) reported some polarization radar data which suggested that hail was falling with its elongated axis apparently horizontally oriented. So, keep your eyes open, especially if you live where hail is common.

Attenuation

Electromagnetic radiation passing through any medium is reduced in power by an amount that depends upon the kind of material present and its density. Some materials reduce or attenuate the radiation more than others. In free space where there is no material (as in the nearly empty space between the Earth and the moon, for example) there is no attenuation; anywhere within the atmosphere there has to be at least a little attenuation. Because attenuation can have such important consequences on the use of radar, let's examine it and its causes in more detail.

Atmospheric Attenuation

The cloud- and precipitation-free atmosphere still contains nitrogen, oxygen, water vapor, and other gases in lower amounts. Nitrogen and many other gases cause no significant attenuation at radar wavelengths. Oxygen and water vapor do cause attenuation. Figure 8.6 (Bean and Dutton, 1968) show the attenuation from both oxygen and water vapor as a function of frequency. From this figure it is obvious that attenuation is not much of a problem at frequencies below about 10 GHz. However, when the water vapor is higher than the amount used for the figure, the attenuation will be higher.

One interesting point to make about the water vapor attenuation near 20 GHz is that this is about in the middle

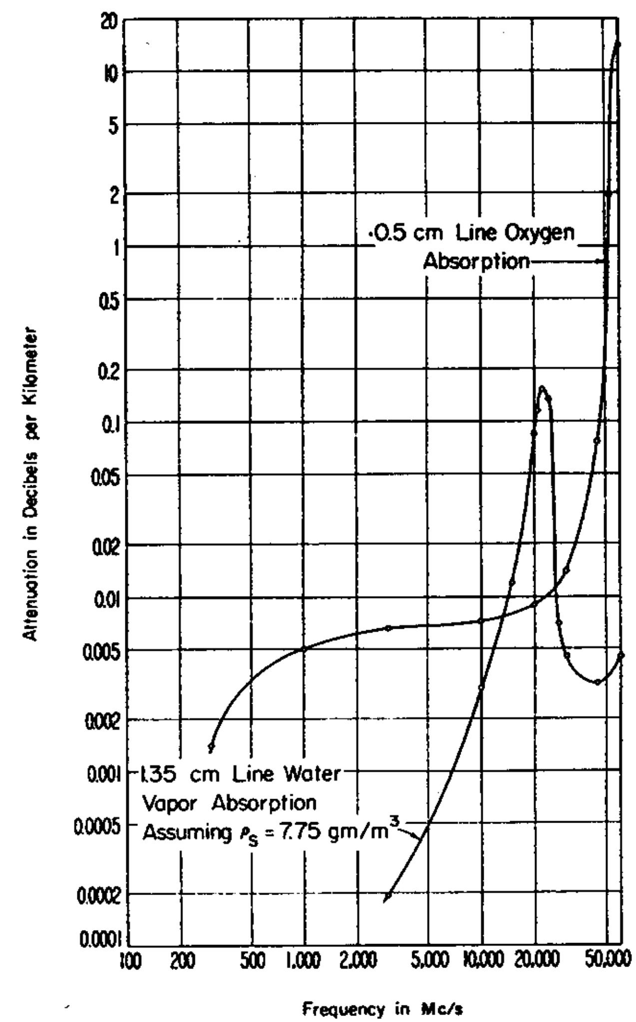


Figure 8.6 Atmospheric attenuation from water vapor and oxygen at standard pressure (1013.25 mb) as a function of frequency. The water vapor curve assumes an absolute humidity of 7.75 g/m^3. From Bean and Dutton (1968)

of the K-band of radar frequencies. However, to avoid excessive attenuation, most meteorological radars operate under or over this frequency. This gives rise to the K_u and K_a bands, respectively.

Note that the attenuation given along the ordinate is in dB/km. An attenuation of 0.01 dB/km over a 100 km path will produce a total of 1 dB of attenuation. Also, since radar works by transmitting a signal and receiving an echo back, the path traveled by the radar waves will be twice this distance, producing for our example a total of 2 dB of attenuation.

Because of Earth's curvature, the height of a radar beam will usually get higher as the distance from the radar increases. Most of the attenuation suffered by radar waves will thus be close to the radar. Figure 8.7 shows the attenuation for two-way radar propagation as a function of radar frequency for elevation angles of 0° and 5°.

In conclusion, the attenuation caused by the atmosphere is usually quite small and is often neglected. If extremely accurate measurements are needed, atmospheric attenuation can be corrected for by increasing reflectivities as a function of range and elevation angle.

Cloud Attenuation

Attenuation by clouds is considerably more variable than that from the atmosphere because clouds themselves are more variable, ranging from nonexistent to very thick clouds. Further, it depends upon whether the clouds are composed of water droplets or ice particles. Except for very long paths through ice clouds, the attenuation through ice clouds is probably negligible.

Table 8.2 gives the attenuation rates for clouds as a function of radar wavelength, cloud temperature, and whether it is water or ice. For ice situations, attenuation rates range from about 0.0006 to 0.009 (dB/km)/(g/m^3). Again, we can see that there is more attenuation at higher frequencies

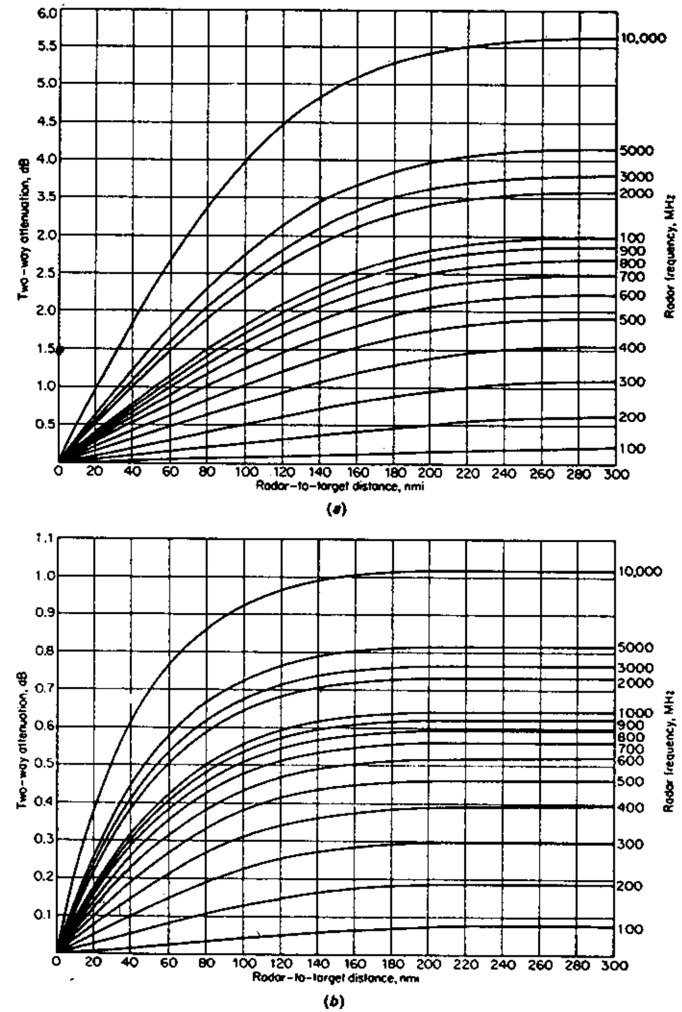


Figure 8.7 Attenuation for two-way radar propagation as a function of range and frequency for elevation angles of 0° (a) and 5° (b). From Skolnik, 1980, Introduction to Radar Systems, with permission of McGraw-Hill, Inc.

*Table 8.2. One-way attenuation coefficient K_1 in clouds in (dB/km)/(g/m³). * indicates "extrapolated". From Gunn and East, 1954.*

	Temp. (°C)	0.9	1.24	1.8	3.2
		Wavelength (cm)			
Water	20	0.647	0.311	0.128	0.0483
	10	0.681	0.406	0.179	0.0630
	0	0.99	0.532	0.267	0.0858
	-8	1.25	0.684	0.34*	0.122*
Ice	0	0.00874	0.00635	0.00436	0.00246
	-10	0.00291	0.00211	0.00146	0.00081
	-20	0.00200	0.00145	0.00100	0.00056

(shorter wavelengths) than at lower frequencies. Also, the low attenuation rates for ice clouds are clearly evident.

For water clouds, the amount of attenuation cannot be ignored for most radar wavelengths if the clouds are at all dense and/or extensive. For example, a cloud with a liquid water content of 4 g/m^3, a temperature of 20°C and a one-way path length of 25 km using a 3.2-cm wavelength radar would have a total attenuation of 10 dB.

Rain Attenuation

As should be expected, attenuation by rain is even stronger than that from clouds. Tables 8.3 and 8.4 give the attenuation rates for rain as a function of rainrate and radar reflectivity, respectively. For X-band radars, the attenuation rates are high enough that severe attenuation can occur in many rain situations, especially thunderstorms. For example, using the Mueller-Jones relationship at 3.21-cm wavelength, a 10-km one-way path through a thunderstorm having a 100-mm/h precipitation rate, the attenuation would be 11.6 dB. Longer paths and/or heavier rainfalls would produce even more severe attenuation.

As another example, if an X-band radar detects a

Table 8.3 One-way rain attenuation K' in (dB/km)/(mm/h). From Wexler and Atlas, 1963.

λ (cm)	M-P (at 0°C)	Modified M-P (0°C)	Mueller-Jones (0°C)	Gunn & East (18°C)
0.62	0.50-0.37	0.52	0.66	
0.86	0.27	0.31	0.39	
1.24	$0.117R^{0.07}$	$0.31R^{0.07}$	0.18	$0.12R^{0.06}$
1.8				$0.045R^{0.11}$
1.87	$0.0045R^{0.10}$	$0.050R^{0.10}$	0.065	
3.21	$0.011R^{0.15}$	$0.013R^{0.15}$	0.018	
4.67	0.005-0.007*	0.0053	0.0058	$0.0074R^{0.31}$
5.5	0.003-0.004*	0.0031	0.0033	
5.7				$0.0022R^{0.17}$
10	0.0009-0.0007*	0.00082	0.00092	0.0003

**First value applies at 2 mm/h, second at 50 mm/h, and there is a "smooth transition" between them.*

Table 8.4 Attenuation by rain expressed in terms of z (mm⁶/m³). Except for those from McCormick, the values are based on the modified MP data in Table 8.3 and a Z-R relationship of $z = 300\,R^{1.5}$.

Frequency (GHz)	λ (cm)	k_p *(dB/km)*
15.0	2.0	$7.15\ 10^{-4} z^{0.725}$*
9.3	3.21	$2.9\ 10^{-4} z^{0.72}$
8.0	3.75	$1.16\ 10^{-4} z^{0.806}$*
5.5	5.5	$1.12\ 10^{-4} z^{0.62}$
3.0	10.0	$3.0\ 10^{-5} z^{0.62}$

*From McCormick, 1970

storm having a DVIP level of 6 (using Z = 55.1 dBz) of 10-km extent, it would also produce an attenuation of 11.6 dB. This is probably not enough attenuation that a weak or moderate storm beyond the first storm would go undetected, but it is enough that the distant storm's intensity would be significantly underestimated.

Snow Attenuation

While snow causes more attenuation than clouds, the total amount of attenuation caused by snow is usually negligible. The low attenuation in snow is a result of the same factors discussed for general snow detection, namely, the dielectric constant effect, the lower melted-precipitation rates in snow as compared to rain, and the generally low clouds that produce snow.

Table 8.5 gives the attenuation rates for four different wavelengths and three different precipitation rates. A 50-km two-way path through snow at 10 mm/h precipitation rate would produce only 2 dB of attenuation. Except for all but the most critical measurements, snow attenuation could probably be ignored.

Table 8.5 One-way attenuation coefficients (dB/km) by low-density snow at 0°C calculated from: $k_s = 3.5{\cdot}10^{-2}\, R^2/\lambda^4 + 2.2{\cdot}10^{-3}\, R/\lambda$ (Battan, 1973).

λ	**Precipitation rate** R **(mm/h)**		
(cm)	1	10	**100**
1.8	0.0046	0.344	33.5
3.2	0.0010	0.040	3.41
5.4	0.00045	0.0082	0.45
10.0	0.00022	0.0026	0.057

Hail Attenuation

It is much more difficult to quantitatively estimate the attenuation from hail. Hail is quite variable in duration, extent and intensity. In reality, hail is such a rare atmospheric phenomena that it is usually not present at all. When hail is

present in a storm, it usually accompanies rain, and often very heavy rain. The net effect of both rain and hail is to produce even more attenuation than without the hail.

Correcting for attenuation

In the preceding sections we found that it is possible to estimate the amount of attenuation that might be taking place if we know the conditions (i.e., rainrate or reflectivity) correctly. Armed with that knowledge, can we then correct for the lost signal? The answer is an unequivocal "maybe" or "sometimes"!

With the exception of atmospheric attenuation by gases, we usually don't really know the total extent of the area that is causing attenuation. Gas attenuation is quite straightforward. Given the path of the radar signal, we should be easily able to calculate and correct for gas attenuation. Even water vapor, which is quite variable in time and space, can often be determined sufficiently accurately that we can correct for losses through it. And further, the losses from these are typically fairly small and are often ignored.

Correcting for attenuation in cloud and precipitation is not as easy. Clouds are typically not detectable on radar, yet the attenuation in them may exceed that from gases in the atmosphere. Precipitation, on the other hand, is usually clearly displayed on a radar, at least to the far edge of the storm *as seen by the radar!* And therein lies the problem. Once the attenuation exceeds a certain amount such that no signal is coming back through a storm, more attenuation will only reduce the maximum range of echo detection in that direction. So, what the radar sees on the far side of the storm may have already been affected by attenuation. As long as there is some echo detected, it may be possible to estimate how much stronger the echo would have been if there had been no attenuation between the radar and that point. Attempting to correct for attenuation can give very incorrect estimates of reflectivity or rain rate. Be careful if you try this that your final estimate doesn't start to approach <u>infinity</u>!

For certain purposes, however, it might still be important to try to correct for attenuation, provided enough is known about the situation to do so safely. Rainfall estimates, for example, could be improved if the attenuation were accurately known and accounted for. There are techniques to estimate the echo lost from attenuation. But once the echo is completely gone, it is impossible to recover anything beyond that point unless some other assumptions or information is available to fill in the missing information. For example, having another radar (or two or three) looking at a storm from another direction can sometimes provide sufficient information to correct for attenuation losses. This is essentially what is done when data from a network of radars is combined by using the strongest return at any point from any of the radars in the network.

Recognizing the presence of attenuation

Can you recognize when attenuation is a problem with a radar? Sometimes. It depends upon the kind, intensity and extent of the precipitation being detected. During thunderstorms or very strong echoes, it is almost certain that some attenuation will take place.

Sometimes it is obvious that attenuation is a problem. When very strong attenuation takes place there will often be a radial region on the far side of an echo that contains very weak echo or none at all. This attenuation "shadow," as it has been called, is evidence that the storm producing the attenuation is very intense. It may not be clear, however, that the region on the far side of the storm is attenuated. Some innocuous storms have shapes that might naturally produce the same kind of pattern. How can you tell the difference? From a ground-based radar it may be difficult if there are no storms on the far side of the nearest storm. If more distant storms are present, they may give a clue about what the nearer storm is really doing. Then again, maybe they won't.

Color Figure 11 shows a region of attenuation on the far side of the strong, tornado-producing thunderstorm 20 to 40 km to the southwest of the UND radar. The reduced signal is evident in both the reflectivity and the velocity portions of this figure. It is difficult to estimate the exact magnitude of the signal loss, but it is not difficult to see that some of the echo is missing. By comparing the reflectivity of the clear-air echo in the general area with that being detected beyond the storm, it would appear that the tornadic storm produced as much as 10 to 15 dB of attenuation, and possibly more.

In an airplane it is possible to operate a radar to improve your chances of recognizing attenuating situations. To do this, you can tilt your antenna down so that the ground is being detected at a range near the maximum range of the radar. Then, when a storm exists between the aircraft and the normal maximum range of the radar, the absence of any ground clutter behind the nearby storm would be clear evidence that attenuation is present. This technique has been suggested by Archie Trammell (1989) in his training courses on the use of airborne radar. His suggestion of flying with the antenna pointed just low enough to give ground clutter at a range near the maximum range of the radar is an excellent way to insure that your radar is working properly as well as giving warning of strongly-attenuating situations. Adjusting the elevation angle of the antenna in this manner is sometimes called "tilt management."

Other Meteorological Targets

Besides those already discussed, there are certainly other targets of meteorological interest. Among those not yet discussed are tornadoes, hurricanes, mesoscale convective complexes, and various wind phenomena. Wind phenomena will be discussed in the next chapter on clear-air targets.

Ever since Don Staggs of the Illinois State Water Survey detected the first tornado with a radar in 1953, tornadoes have been detected and extensively studied by radar. While tornadoes are quite rare at any given point on Earth, in the central United States it is not at all uncommon to have tornadoes *within radar range* of a single radar almost every year. Tornadoes are so prevalent within the United States, in fact, that the WSR-88D (NEXRAD) radars were given the capability to measure Doppler velocities primarily so they would be better able to detect tornadoes. NEXRAD algorithms are being used specifically for the detection of mesocyclones and the tornadoes themselves.

Color Figure 11 shows an example of a tornadic vortex signature (TVS) detected by the UND C-band radar in the Kansas City, Missouri, area. The TVS is the small region located at approximately 35 km range and 215° azimuth. At that point there is a very small region of quite variable velocities associated with a tornado. On the reflectivity image there is an appendage of echo off on the southwest side of the main storm which, while not being the perfect example of the classic "hook echo," does have hook-echo characteristics. As mentioned earlier, however, this is also on the edge of a region where attenuation is taking place, so that may contribute to the apparent shape of this echo. The ability to zoom in on small details of storms using newer radars makes it significantly easier to detect some of these small-scale features.

Hurricanes are also amenable to study by radar – provided they are close enough to the radar to be detectable. Hurricanes generally form hundreds or thousands of kilometers away from land. Ground-based radars have detectable ranges on the order of 400 km or so. Consequently, hurricanes are not detected until they have moved close enough to land to already be a problem. On the other hand, airborne Doppler weather radars have been used very effectively in studying hurricanes (a case of Muhammad going to the mountain instead of waiting for the mountain to come to him!). And the TRMM satellite can also detect hurricanes while they are far from shore. Once a hurricane gets close enough to

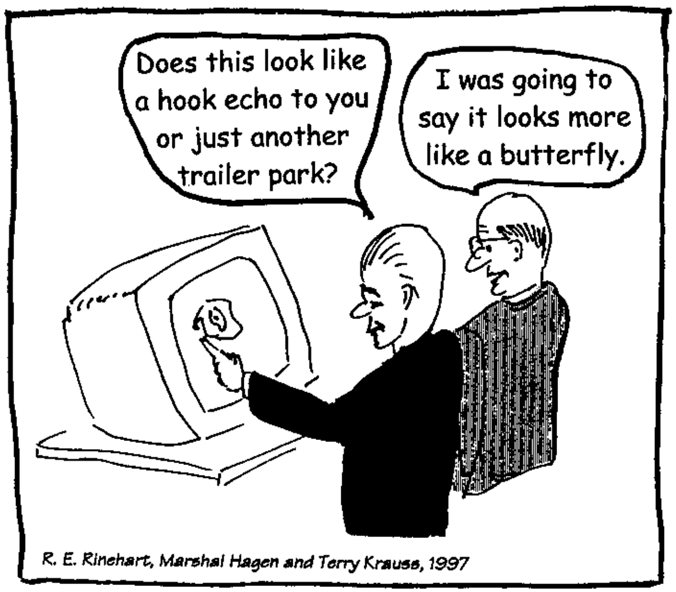
shore for a ground-based radar to detect the, radar can be very useful in measuring the copious amounts of rain that often fall during hurricanes (if it survives the storm!). Also, many hurricanes spawn one or more tornadoes shortly after they make landfall; these can also be detected and warnings issued using the aforementioned detection algorithms.

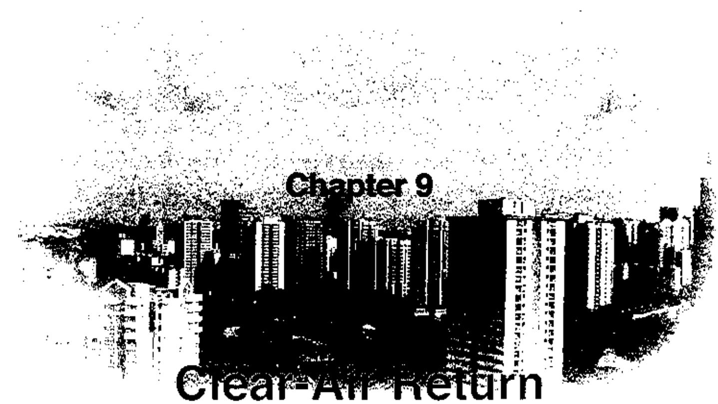
Figure 8.8 is one of my all-time favorite hurricane pictures from a radar. In fact, it is the very last full-circle image collected by the Miami WSR-57 radar before the radar was destroyed by Hurricane Andrew in 1992! The picture clearly shows the eye of Andrew as it moved onshore near

Figure 8.8 PPI from the Miami, Florida, National Weather Service WSR-57 radar just before Hurricane Andrew destroyed the radar (0835 UTC, 0435 EDT, 24 August 1992).

Homestead Air Force Base in Florida. This radar has since been replaced by a WSR-88D radar.

Mesoscale convective complexes are, as the last word in the name implies, complexes of numerous storms acting as a giant unit. As such, the various components are just as detectable as they would be if they occurred individually. MCC's are sometimes so large, however, that a single radar does not have the ability to cover the entire event; they are just too extensive. Data from more than one radar are routinely combined, fortunately, so that entire events can be monitored by ground-based radar networks. It is now quite easy to see these on displays that mosaic data from many radars onto a single image.





Chapter 9
Clear-Air Return

Meteorological information can come from nonmeteorological as well as meteorological targets. When we think of weather radar, we usually think in terms of the radar detecting echoes from weather. However, since radars can receive detectable power from insects and other targets, it is often possible to learn about the weather from these nonmeteorological targets. In fact, as will be seen, some important wind phenomena are detectable largely because of clear-air echo.

As a historical note, the detection of echoes in the optically clear air began fairly early in the use of weather radar. Since there was nothing visible to the human eye or even through binoculars and telescopes, these unidentified echoes were given the names of "angels" or "ghosts." Angels were discrete, point targets (most likely birds, in many cases) while those that were more nebulous and diffuse and seemed to cover an area were called ghosts. There were many papers in the early history of radar meteorology related to angel and ghost echoes. There were a number of experiments in the early 1960's in which individual birds and insects were tracked by radars, demonstrating that these were likely the source of most of the unknown echoes. The last paper in the radar conference preprint volumes using "angel" in the title was published in 1970. There was even one paper published on an unknown echo detected over wa-