CHAPTER 12

How Good Are Our Conceptual Models of Orographic Cloud Seeding?

ARTHUR L. RANGNO

Department of Atmospheric Sciences, University of Washington, Seattle, Washington

ABSTRACT

Some of the complexities of clouds and precipitation that have been encountered in field projects are reviewed. These complexities highlight areas of cloud microstructure and precipitation development that need to be better understood before adequate conceptual or numerical models of orographic cloud seeding can be developed. Some concerns about cloud sampling with regard to the evolutionary behavior of supercooled clouds from water to ice are also discussed.

12.1. Introduction

Recent advances in airborne instrumentation and remote-sensing techniques combined with intensive field programs with dense observation networks are providing much new information on clouds and precipitation, particularly in the mountainous regions of the western United States. These observations have revealed great complexity, but with them comes the promise that many of these complexities will be understood. However, as yet our description of the coupled processes governing air motions and cloud microphysics relating to clouds upwind, over, and downwind of mountains is still inadequate in many ways. This represents a fundamental drawback to the formulation of conceptual and numerical models for use in the prediction of the effects of artificially seeding orographic clouds.

In this paper, the term "orographic" will be used to refer to those cloud systems that form solely as the result of air rising over terrain, which are seen as quasi-stationary clouds of variable coverage on satellite imagery. The term "orographically enhanced" will be used to refer to long-lived cloud systems associated with fronts and troughs that are trackable on satellite imagery prior to impinging on mountain barriers. The discussion will be largely confined to observations made in wintertime clouds in the mountainous regions of the western United States and will attempt to present a holistic view of the seeding potential of these clouds in mountainous regions.

12.2. Brief summary of recent findings

12.2.1. Ice particle concentrations in clouds

Of paramount concern with regard to mounting a cloud seeding effort is knowing what concentrations of ice particles are going to develop naturally in clouds. A logical early expectation in this regard was that ice particle concentrations in clouds, at least in the Rockies, might only be a function of the concentrations of ice nuclei (Grant, 1968) and, therefore, of cloud-top temperature.

However, much of the field work in the West has revealed unexpectedly high ice-particle concentrations in clouds across a wide range of cloud-top temperatures, locations, and types of clouds (Koenig, 1968; Hobbs, 1969; Auer et al., 1969; Cooper and Saunders, 1980; Heymsfield, 1977; Grant et al., 1982). These reports showed that the occurrence of ice at unexpectedly high cloud-top temperatures (some $\geqslant -10°C$) and in unexpectedly high concentrations (10s to 100s L^{-1} at $> -20°C$) was not restricted to warm-based summertime cumulus clouds (Coons and Gunn, 1951; Braham, 1964), but that it could also occur in wintertime clouds with relatively cold bases.

Several factors have been suggested that can at least partially explain these surpluses of ice particles above those expected based on ice nuclei concentrations (Fletcher, 1962). The two most important of these include the production of "secondary" ice particles, for example, through the breakup of fragile crystals (Hobbs and Farber, 1972; Vardiman, 1978) or through a riming and splintering process (Mossop and Hallett, 1974; Hallett and Mossop, 1974). How these various factors might affect the distribution of ice in orographic clouds will be mentioned in section 12.3.

On the other hand, some studies have appeared to support a temperature-dependent activation of ice nuclei, although not necessarily from cloud top. Cooper and Vali (1981), for example, suggested that a temperature-dependent activation of ice nuclei was evident in their data obtained in two lenticularlike clouds over Elk Mountain, Wyoming, and in the rear portion of a large-scale stratiform cloud system (orographically enhanced) over southwestern Colorado. The temperature effect they reported, however, depended on the accuracy of their estimated origins of the crystals, which were not necessarily from cloud top. In addition, they found that concentrations of ice particles were about an order of magnitude higher in the southwestern Colorado cloud than that at Elk Mountain, despite similar temperatures.

In a further summary of ice particle concentrations in cap clouds in Wyoming, Vali et al. (1982) found that a temperature-dependent effect (again, not necessarily associated with cloud top) was indicated in ice particle concentrations in the upwind-to-crest portion of these clouds. Still higher concentrations were found downwind of the crest, although it was not mentioned whether these higher ice particle concentrations also supported a temperature-dependent effect. Thus, while a temperature-dependent effect was reported in these studies, it is not clear that ice particle concentrations were well predicted by cloud-top temperature.

Mossop (1985) summarized the results of ice particle concentrations in clouds reported by a number of workers. He reported that, with few exceptions, ice particle concentrations in stratiform clouds were dependent on cloud-top temperature when those clouds did not meet the criteria for the riming and splintering production of secondary ice particles. In cumuliform clouds, high ice particle concentrations relative to expected ice nuclei concentrations also appeared to be restricted, with one exception, to those cases where the criteria for riming and splintering were met.

However, there have also been reports of ice particle concentrations in clouds that are *lower* by orders of magnitude than expected ice nucleus concentrations. For example, Stewart (1967) encountered ice-free altocumulus clouds at temperatures below $-25°C$. Heymsfield (1977) stated that in the stratiform clouds that he sampled, ice particle concentrations were, without exception, less than expected ice nuclei concentrations at temperatures below $-20°C$. Measurable liquid water was encountered at temperatures as low as $-35°C$. Grant et al. (1982) reported ice particle concentrations of only about $2\ L^{-1}$ in a liquid-topped, stratiform cloud in the Rockies at $-31.8°C$. Hobbs and Rangno (1985) encountered altocumulus with tops at $-26°C$ with only a $1\ L^{-1}$ ice particle concentration. In addition, Sassen (1984) reported liquid water at $-36°C$ over Utah.

In these cases, the presence of liquid water at these temperatures in clouds with characteristically slight updrafts is surprising since Fletcher's (1962) summary of ice nucleus measurements indicates that tens to thousands per liter of ice particles should be present in these clouds, and, if ice nuclei indeed act as sublimation nuclei, they should have formed long before the water phase occurred and therefore been present in more than enough numbers to have prevented the development of liquid water.

Finally, we note that there is no significant correlation between ice particle concentrations and cloud-top temperature in the large dataset of Hobbs and Rangno (1985). About 25% of their observations of unexpectedly high ice-particle concentrations were in clouds that did not appear to meet the criteria for ice multiplication specified by Hallet and Mossop (1974). A somewhat larger percentage appears not to meet the revised riming–splintering criteria (Mossop, 1978) that specifies the presence of droplets $\leqslant 13\ \mu m$ diameter in concentrations of $\geqslant 100\ cm^{-3}$. Hobbs and Rangno (1985) found a moderate to strong correlation between a measure of the broadness of the cloud-top droplet spectrum and maximum ice particle concentrations over a range of cloud-top temperatures from $-6°$ to about $-26°C$ in both stratiform and cumuliform clouds. This finding may shed light on why cloud-top temperature alone cannot, in general, predict ice particle concentrations, since they may first be dependent on the broadening of the droplet spectrum.

In summary, ice particle concentration surpluses relative to expected ice nucleus concentrations can be found on the warm ($\geqslant -15°C$) cloud-top side of Fletcher's summary curve while large deficits in ice particle concentrations, often combined with the presence of liquid water, can be found on the cold ($\leqslant -25°C$) cloud-top side. These observations suggest a tendency for a leveling of ice particle concentrations in clouds over a wide range of cloud-top temperatures, a phenomenon that was first noted by Weickmann (1957).

12.2.2. Ice nuclei

Some suggested sources for ice nuclei include meteorites, soils, bacteria and spores, and emissions from industrial plants, automobiles, and volcanoes. Kumai (1951, 1961) found that most of the ice particles he examined contained clay particles at their centers. In an example of the potential importance of soil as ice nuclei, Marwitz et al. (1976) reported that the highest natural ice nuclei concentration they measured over southwestern Colorado was on a windy, dusty day associated with a cold front. This maximum ice nuclei count was even greater than the maxima they had measured on all other days above AgI ground seeding sources!

If ice nuclei are important in producing ice particles found in clouds overhead, then numerical models should predict the concentrations of ice nuclei in any given situation. For example, it is conceivable from the Marwitz et al. report that seeding may be unnecessary when it is windy upwind of the region to be seeded but necessary when it is calm or after a regionwide heavy rain or snowfall may have subdued surface sources of ice nuclei. To date, however, and in spite of the potential importance of ice nuclei, ice nuclei concentrations on a day-to-day basis are seemingly unpredictable and their role in ice particle formation is as yet unclear.

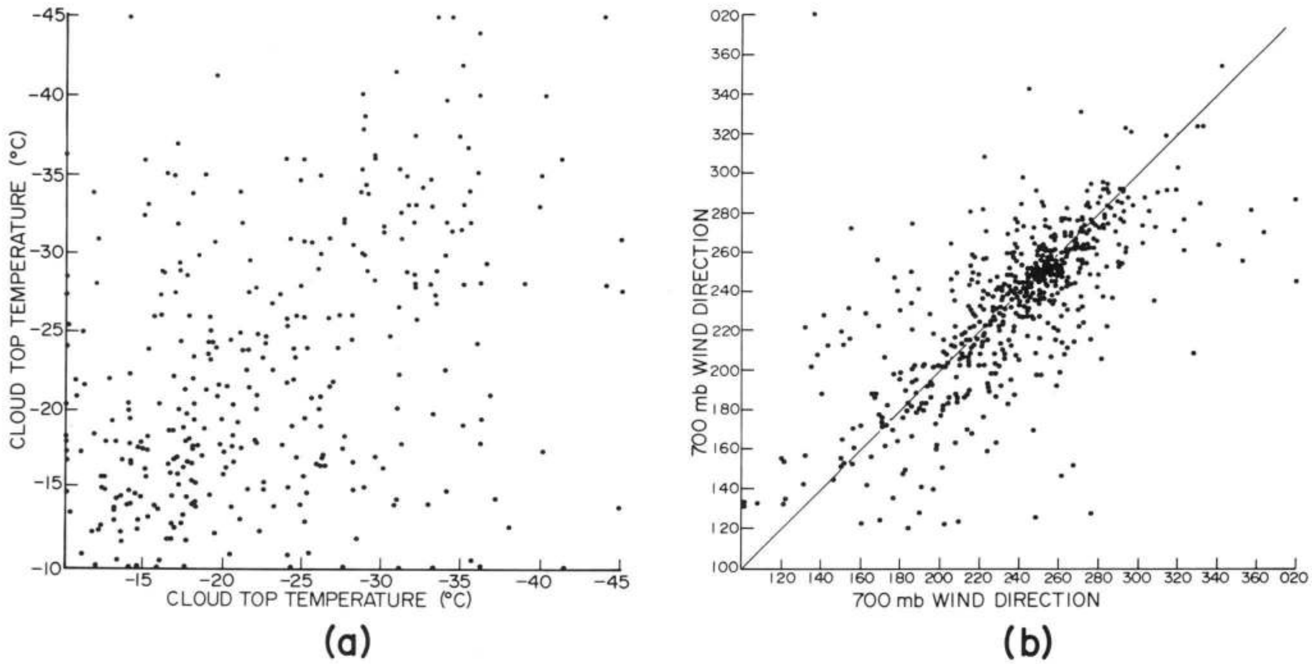

FIG. 12.1. Scatter diagrams of consecutive measurements at Durango, CO, of (a) rawinsonde cloud-top temperatures and (b) 700-mb wind direction for soundings ≤4 h apart. (After Elliott et al., 1976.)

12.2.3. Variability of clouds and winds during storms

While cloud-top temperature as a predictor of ice particle concentrations in clouds is questionable, it is nevertheless useful to examine its variability as a measure of how rapidly project operations may have to be adjusted in order to cope with changing seeding potential. Figure 12.1a is a scatter diagram of cloud-top temperatures (after Elliott et al., 1976) over the Rockies derived from pairs of rawinsondes[1] launched not more than four hours apart during the Colorado River Basin Pilot Project (CRBPP). Elliott et al. used the point where the dewpoint passes ice saturation to deduce cloud tops. The data shown are limited to the period after 1 February 1973, when shielded relative humidity elements apparently came into use (Hill, 1980a). The cloud top of the first sounding is plotted along the ordinate and the cloud top of the second along the abscissa. As can be seen, the serial correlation is poor, illustrating the large, short-term variability in clouds and the forecast and treatment challenge it presents to cloud seeding experiments.

Figure 12.1b is a similar plot of 700-mb wind directions. While the serial correlation is better than in the case of cloud-top temperatures, in about one-third (36%) of the soundings, the 700-mb wind did not remain within 40° of the azimuth value measured ≤4 h earlier.

The purpose of this discussion is to point out that quasi-steady-state situations in mountainous regions (which are often the foundation of conceptual or numerical models) appear to be rare. The cloud variability encountered poses a severe challenge in the forecasting of seeding opportunities and another challenge in the treatment strategy should seeding opportunities be short-lived.

This variability also suggests that experiment strategy may have to be changed in future experiments to better retain homogeneity between control and seeded samples. For example, in an experiment reported by Hobbs et al. (1981) it was shown that long-lived, dry-ice-produced precipitation plumes in stratocumulus clouds, trackable from the cloud to the ground by a vertically pointed 8.6 mm radar that served as the target, can produce a precipitation signal at the ground at or very near forecasted times. Such plume intercepts could be compared directly with the natural precipitation, if any, that immediately preceded and followed the seeding plume intercept while cloud conditions were relatively uniform.

12.2.4. Temporal variations in precipitation

Snowfall intensity in the Rockies varies over periods of minutes (Medenwaldt and Rangno, 1973; Vardiman and Hartzell, 1976). Figure 12.2 shows a plot of visual observations of daytime snowfall intensity over a five-month period (made by the author in Durango, Colorado). It reveals a distribution of intensity changes with time analogous to the relationship between the concentration of precipitating particles and their size reported by Marshall and Palmer (1948). This finding was unexpected be-

[1]Rawinsonde-derived cloud-top temperatures can be questioned in some situations, and more comparisons are needed with satellite data and vertically pointed cloud-sensing radars.

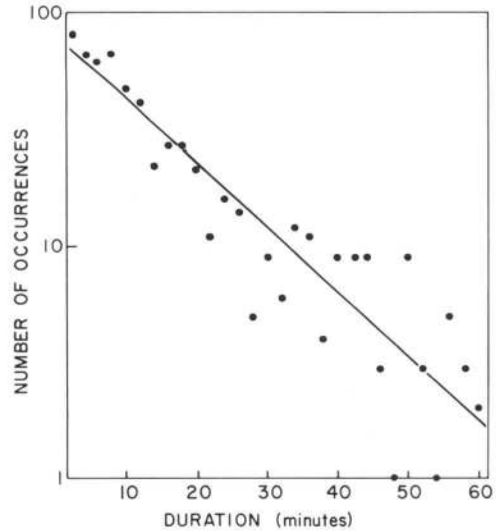

FIG. 12.2. Duration of any initial, visually determined intensity of snowfall using five categories of intensity: flurries (no accumulation), very light snow, light snow, moderate snow, and heavy snow.

cause it was believed that snowfall in the Rockies was relatively consistent from hour to hour (Grant et al., 1969).

The rapid changes in snowfall intensity indicate microscale structural differences in clouds and precipitation that are not yet included in our models of wintertime clouds in mountains. What effect will seeding have on these structures? What do they say about how ice is formed and organized in the clouds overhead? Perhaps these microscale structures are unimportant in the overall assessment of seeding potential, but this cannot be assumed a priori.

12.2.5. Fine-scale gradients in orographic precipitation

It has long been known that extremely fine-scale precipitation gradients exist even over modest terrain in stratiform and convective precipitation situations (e.g., Godske et al., 1957; Huff et al., 1975). Figure 12.3, for example, shows the distribution of precipitation for three winters in the San Juan Mountains of southwestern Colorado when the 700-mb wind is southwesterly through

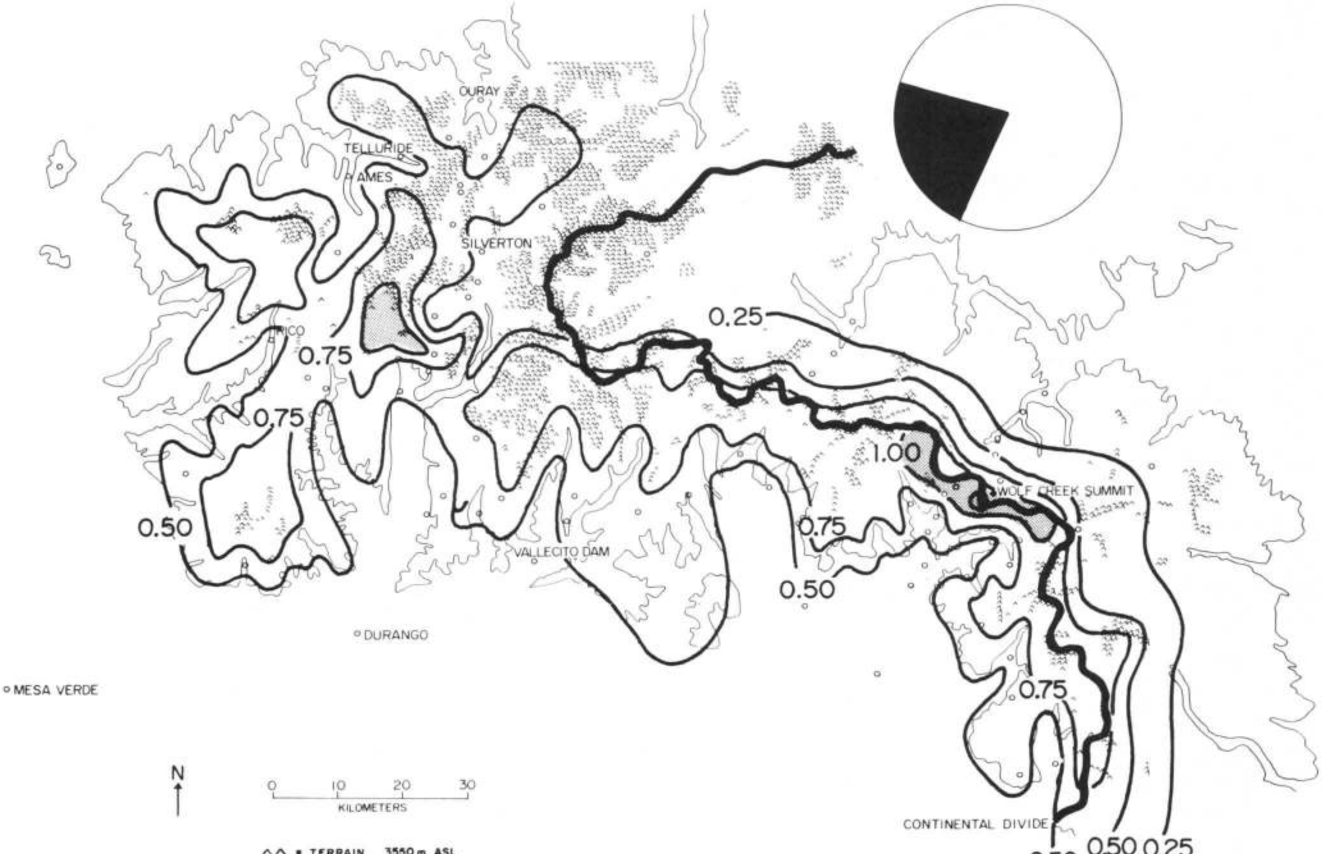

FIG. 12.3. Distribution of precipitation in the San Juan Mountains of southwestern Colorado with the 700-mb wind direction between 206° and 285° (darkened portion of azimuth circle upper right). Isohyets (solid bold lines) are in the percentages of the precipitation observed at Wolf Creek Pass summit. Shading indicates ≥100% of Wolf Creek Pass summit precipitation. Solid light line indicates terrain ≥2670 m (9000 ft) MSL and peaks (⌃) terrain ≥3560 m (12 000 ft) MSL. Gage locations are indicated by small circles.

westerly (Rangno, 1979). An extremely large (>100%) increase in precipitation occurs from the base of Wolf Creek Pass to the summit, a distance of only about 10 km. An examination of the hourly precipitation records of the CRBPP by the author shows that about 70% of this greater total at the summit as compared with the base of Wolf Creek Pass is due to greater intensity (more or heavier ice particles or both) and about 30% is due to greater duration (precipitating cloud was not over the lowest slopes, or the precipitation was not initiated far enough upwind of the lower slopes to fall out on them).

These large gradients in precipitation over extremely small distances, both parallel to and across the barrier, both upwind and downwind of the crest, cannot be produced by changes in cloud-top height resulting from the passage of the cloud system over the barrier. This is because any changes in cloud-top microstructure immediately upwind and over the barrier could not affect snowfall at the ground during the limited time the cloud traverses the barrier. The atmosphere can produce these small-scale gradients in precipitation only through rapid changes in the mass of precipitating particles close to ground level. Bergeron (1950) described a similar situation as a "seeder–feeder" process where ice crystals (or raindrops) fell into and accreted the droplets of a lower-level cloud, and he observed that such a process was capable of producing such fine-scale precipitation gradients in terrain of even modest topography.

What would be the effect of seeding on these fine-scale precipitation patterns? Trajectory changes are likely due to seeding (e.g., Hobbs et al., 1973), and the question of the redistribution of precipitation without an increase is raised. Would some of the heavier precipitation falling onto sharply rising windward slopes be shifted to the next downwind canyon and eventually to the lee of the crest, where some of the precipitation could be lost due to evaporation? This is an old question, but nevertheless one that still needs a firm answer.

12.2.6. Targeting clouds with modification potential

Conceptual or numerical models of clouds for artificial modification purposes should also incorporate the capability to predict if, when, where, and in what concentrations the seeding agent will arrive in targeted clouds. Of the questions that beset cloud seeding, the targeting problem may prove to be the most intractable. Airborne studies of surface-released seeding agents in and near mountains has revealed that they often do not reach the clouds due to intervening stable layers of air (Rhea et al., 1969; Hobbs et al., 1975; Marwitz et al., 1976). Severe targeting problems also develop when there is a large directional wind shear with increasing height (e.g., Elliott et al., 1978), a situation that usually occurs in storms. Or, even when there is strong vertical dispersion, the seeding agent may nevertheless arrive in the clouds at locations inappropriate to produce an effect in the target (Hobbs et al., 1975). Thus, the opportunities for using ground generators to seed wintertime clouds has been found to be limited in many instances. In addition, no studies have been conducted concerning vertical dispersion of ground-released seeding agents at night over snowy surfaces, a situation that often occurs during cloud seeding operations in mountainous regions.

Even when there is good vertical dispersion in the presence of suitable clouds, however, the nature of turbulence is to produce puffs of higher concentrations interspersed with regions of lower concentrations of the source plume while meandering (e.g., Hobbs et al., 1975). For an update on the complexities of dispersion, the reader is referred to Weill (1985), Wyngaard (1985), and Briggs (1985).

Is it safe to assume, when conducting a cloud seeding experiment, that the actual, chaotic plume will produce the same effects in the clouds as those calculated using a time- or space-averaged plume? Or will some portions of some clouds be continuously underseeded while other portions are overseeded? Should precipitation modification projects or experiments be carried out without firm answers to these questions?

12.3. Conceptual models of clouds over mountains

While uncertainties still exist concerning the structure and evolution of orographic and orographically enhanced clouds, it is useful to summarize some common aspects found in field work. Composites derived from recent observations of orographic and orographically enhanced cloud systems are presented in Figs. 12.4 and 12.5, respectively. These composites are for two generic cloud regimes. The transition between these two regimes is often marked by frontal zones (e.g., Hobbs, 1975). It should be kept in mind that conditions (cloud tops, cloud coverage, stability, and winds) within these regimes change continuously. Observations used to derive these models originate with vertically pointed cloud-sensing radars (e.g., Plank et al., 1955), aircraft observations (e.g., Hobbs and Atkinson, 1976), and satellite and surface observations. Further refinement of these two basic types of meteorological situations affecting mountainous regions has been presented by Heggli et al. (1983) and Heggli and Reynolds (1985).

The class of clouds most pertinent to this discussion is orographic (Fig. 12.4). Cloud coverage over a barrier may range from solid overcast to scattered (less than one-half sky coverage) near the end of a storm episode. Liquid water maxima may vary from 0.1–0.2 m^{-3} in thin, stable stratiform clouds (Ac and Sc len) over the high elevations of the Rockies (e.g., Cooper and Vali, 1981) to $\geqslant$1.5 g m^{-3} in postfrontal cumulus and small cumulonimbus near the West Coast (Heggli et al., 1983; Hobbs and Rangno,

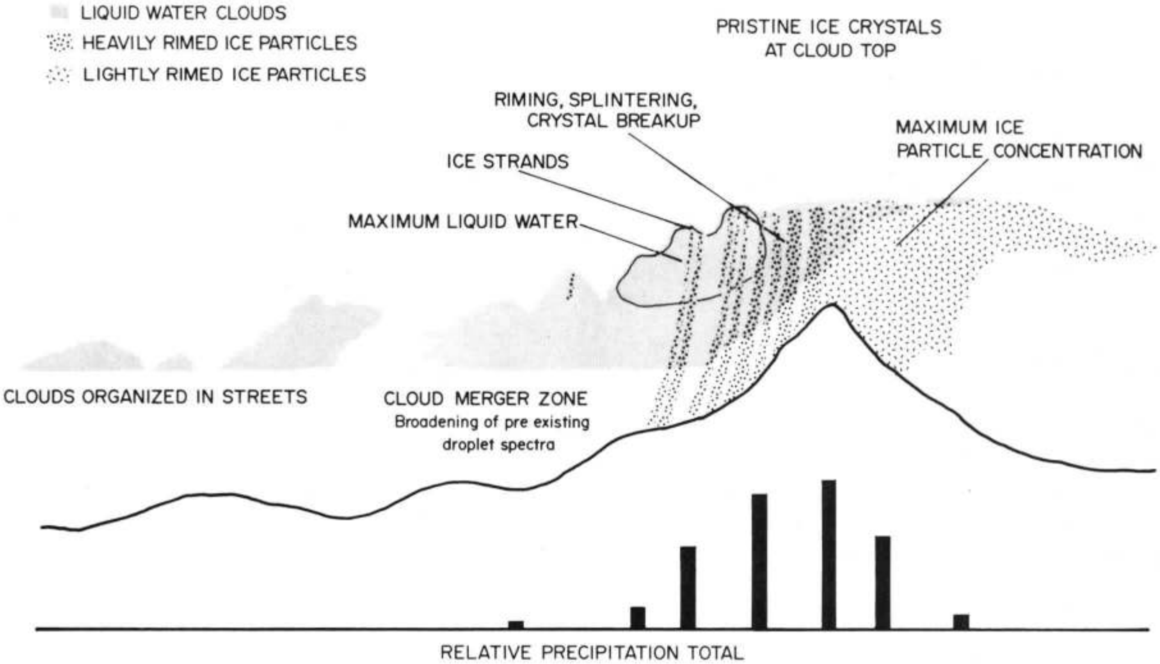

FIG. 12.4. A conceptual model of an orographic cloud system composed of stratocumulus and cumulus.

1985). Due to low areawide cloud coverage, strong diurnal effects are superimposed on orographic clouds that can convert nighttime laminar cloud forms into convective forms during the daytime. Cloud bases rise during the day and descend at night, while cloud tops are likely to do the opposite. While such changes are likely to affect cloud microstructure, no studies of the diurnal effects that are superimposed on orographic clouds have been conducted.

In convective situations, such as that depicted in Fig. 12.4, clouds and showers may align themselves in cloud

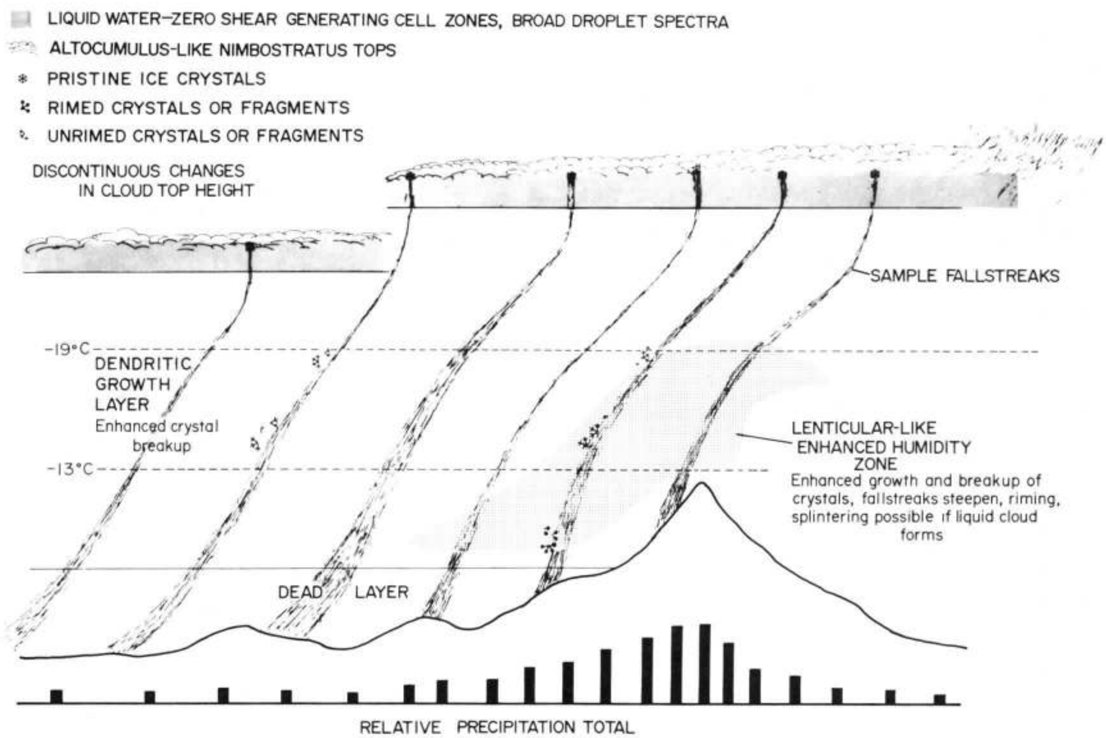

FIG. 12.5. Schematic of an orographically enhanced, precipitating cloud system composed of nimbostratus.

streets emanating from favored spawning points well upwind of the barrier (Heggli et al., 1983). Ice particle concentrations are often highly variable in time and space (e.g., Rauber et al., 1984). Usually, heavily rimed ice particles are found at the ground at upslope and crest locations (e.g., Hobbs, 1975; Vardiman and Hartzell, 1976) with a sharply diminished degree of riming in the lee.

In the stable stratiform case (lenticularlike clouds), ice particle concentrations develop near the upwind edge of the liquid cloud and within a few minutes reach a maximum that remains rather steady (Cooper and Vali, 1981) up to the crest. Since the supply of moisture is continually changing, the upwind edges of these clouds can recede downwind and expand upwind in a matter of minutes. At the same time, layers can be added or subtracted to the bases and tops of these clouds within minutes.

The onset of ice particles in clouds near the West Coast is at cloud-top temperatures between $-4°$ and $-10°C$ (e.g., Hobbs, 1974; Hobbs and Atkinson, 1976; Heggli et al., 1983), but they do not appear until temperatures of about $-10°$ to $-15°C$ are reached in the Rockies (e.g., Auer et al., 1969; Vali et al., 1982). However, data from the CRBPP suggest that only a small fraction (less than 10%) of wintertime clouds in the Rockies have tops that are less than $-15°C$ (Elliott et al., 1976). Hence, most substantial orographic clouds that develop there usually contain some ice particles.

A characteristic of supercooled orographic clouds is their tendency to become increasingly infected with ice particles as they transit a barrier. This facet of orographic clouds has no doubt been visually observed by many, but only on a few occasions has it been mentioned (Hobbs, 1975; Rangno et al., 1977; DeMott and Grant, 1984).

Eight factors, some of which may act in concert, may aid the increase in ice particle concentrations over and downwind of a mountain crest in orographic clouds:

- the ice-forming process is time dependent (e.g., Koenig, 1963) and ice particles would be expected to continue to increase as the orographic cloud progressed across a barrier;
- enhancement of contact nucleation activity (Young, 1974) in the lee side evaporating portions of liquid cloud;
- evaporative cooling, potentially up to several degrees (e.g., Cerni and Cooper, 1980), could lead to an enhancement of temperature-dependent ice-forming mechanisms;
- the lifting and resultant cooling of cloud top as it traverses the barrier also leads to increased activation of temperature-dependent ice nuclei (e.g., Fletcher, 1962);
- the riming and splintering production of small secondary ice particles (Hallett and Mossop, 1974) on the windward slopes that carry to the lee side;
- the mechanical breakup of fragile crystals (e.g., Hobbs and Farber, 1972) whose small fragments could carry to the lee side;
- the blow-off of ice particles from the mountain surface into the downwind cloud (Rogers and Vali, 1981);
- enhanced development of ice in strong uplift zones preceding the crest due to increased values of ice supersaturation (DeMott et al., 1982).

This near-crest and post-crest ice-enhancing region needs to be examined more closely in future research so that it can be incorporated into models of orographic clouds and precipitation and so that what effects, if any, it has on downwind barriers can be discovered.

There is an unambiguous situation, however, where liquid water escapes a mountain barrier and a potential to increase precipitation is present as envisioned by Ludlam (1955). This occurs when there is a flow of nonprecipitating supercooled cloud that satisfies the following criteria: appreciable cloud depth, for the purpose of appreciable crystal growth, say 1 km; cloud base not too far (<1 km) above mountaintop; of such upwind extent that seeding could cause an appreciable crystal fallout on the barrier; and a cloud-top temperature $\leqslant -5°C$.

To date, however, documentation of this situation is either absent or it has only been inferred to exist (preferentially with 500-mb temperatures $\geqslant -20°C$) from apparent duration-of-snowfall effects in the evaluation of cloud seeding experiments (Chappell et al., 1971; Mielke et al., 1981). Documentation of this situation would establish a baseline of seeding potential.

Figure 12.5 shows a schematic of orographically enhanced clouds. In these systems, maximum ice particle concentrations are generally a few tens per liter across a wide range of cloud-top temperatures (Hobbs and Atkinson, 1976; Cooper and Saunders, 1980; Hobbs and Rangno, 1985). Liquid water content is modest (<0.5 g m^{-3}) or absent. Often, and perhaps characteristically, the liquid water present is confined to a thin layer at cloud top (e.g., Cunningham, 1957; Walsh, 1980) where small, pristine crystals are encountered (Hobbs and Rangno, 1985). The highest concentrations of ice particles, characteristically, occur at elevations below cloud top and can be due to several factors, including the breakup of fragile crystals, riming-splintering processes, or the direct formation of new crystals.

Lapse rates are often stable in the lowest layers, particularly in the earliest stages of these storm episodes (e.g., Marwitz, 1980), and a "dead" layer a kilometer or more deep in which the flow is light and variable or diverted around the barrier can be present.

Due to the low amounts of supercooled water and to plentiful ice particle concentrations, particularly in and near major bands (e.g., Heymsfield, 1977; Heggli et al., 1983), it appears unlikely in this phase of a storm that precipitation can be increased through seeding, although a possible exception to this conclusion will be considered in the following section.

12.4. Some further comments relating to sampling and seeding

Exciting new measurements in mountainous regions such as in the Sierra Cooperative Pilot Project and in the Colorado Orographic Seeding Experiment are becoming available. These new observations, which include those from scanning radiometers, polarizing lidars, and vertically pointed cloud-sensing radars, will provide much larger and more complete studies of clouds than have been available in the past. If put into climatological perspective, and not just restricted to one location upwind of a mountain, these measuring systems can provide vital information from which decisions can be made on how to conduct, when to conduct, or whether to conduct, cloud seeding operations.

However, the chance for an erroneous assessment of cloud seeding potential still exists, in spite of these new measurements, and it may even be quite high. Some possible scenarios in which erroneous interpretations could be made are outlined below. Wintertime cumuli are included in these discussions because they are being considered as targets for cloud seeding in the Sierra Nevada (Heggli et al., 1983).

Figure 12.6 shows a wintertime cumulonimbus calvus near St. George, Utah. A new cumuli turret, reaching its maximum height at point A, does not outwardly display signs of ice particles and may contain ≥ 1.5 g m^{-3} of liquid water. At point B, a maturing turret that has ceased rising appears to consist only of ice particles. Let us suppose that an aircraft flew only in the small, newly formed cumuli that had not yet produced ice and only in the newest, virtually ice-free portions of the larger cumuli. A researcher examining this data would conclude that there were few ice particles in these supercooled clouds and that there was considerable potential for seeding on this day.

Conversely, the data from another aircraft flown on the same day, but which flew through only the glaciated portions of these cumuli, would lead another researcher to conclude that such clouds were composed solely of ice particles and had no seeding potential.

A third aircraft that flew through both the ice-free and the glaciated portions of these clouds would have collected data that would cause the researcher to choose between two conclusions: a seeding potential was present in some of the clouds, since some regions with large amounts of supercooled water were encountered that had few ice particles, or, there was no seeding potential because it was assumed that all of the liquid water encountered was subsequently consumed by the ice particles that developed later and comprised the glaciated cloud. Clearly, a seeding experiment based on the perception of an unambiguous and appreciable seeding potential on this day would lead to disappointing results. While ice particles could certainly be initiated by seeding earlier in the cloud's lifetime, the ability to produce a strong precipitation signal above the background of natural precipitation at the ground from these clouds is doubtful. A seeding potential would have

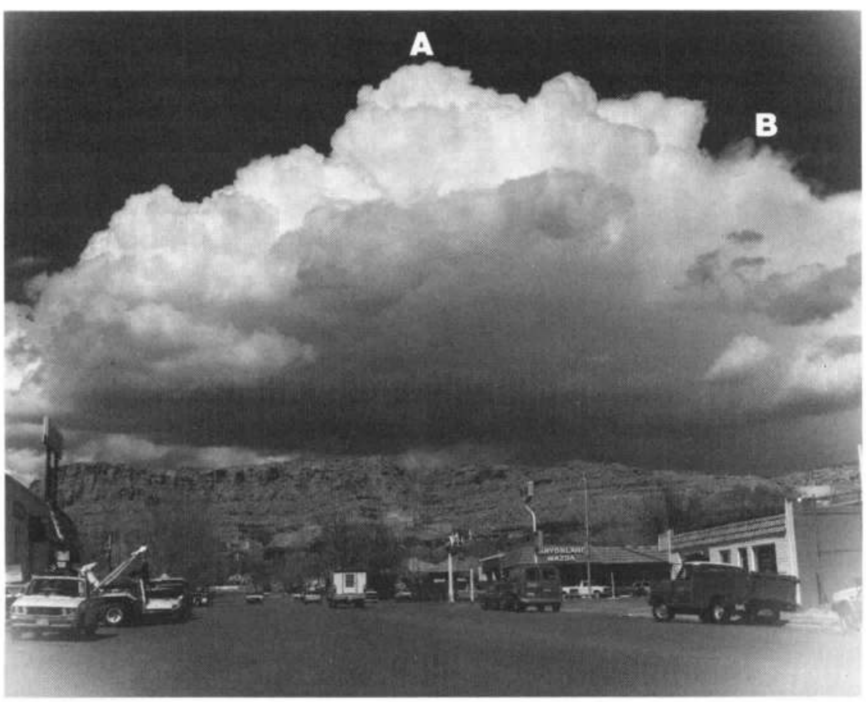

FIG. 12.6. Wintertime cumulonimbus calvus near St. George, UT. Point A is a cloud in the early portion of development. Point B is a glaciated portion of cloud.

been unambiguously detected only if it had been shown that ice particle concentrations remained virtually nonexistent in these clouds and if the clouds lasted long enough or were wide enough for appreciable-size precipitation particles to develop and fall within them. The problem of cloud lifetime in precipitation development has been recently addressed by Schemenauer and Isaac (1984) and Cooper and Lawson (1984).

A situation similar to that described above for cumuli exists for orographic clouds. Figure 12.7 shows a postfrontal, orographic cloud with modest embedded convection in westerly flow over the La Plata Mountains of southwestern Colorado. The features of interest are the liquid-appearing cloud at point A on the upwind slopes of the La Plata Mountains and the glaciated cloud at the downwind end of the cloud at point B. This behavior mimics that of cumulus life cycles described above, except that the water-to-ice evolution is spread over a greater horizontal extent and usually in thinner clouds.

The risk of an erroneous assessment of cloud seeding potential in orographic situations lies in making incomplete measurements or in making interpretations that do not take into account the evolutionary behavior of cloud microstructure as it traverses a barrier. For example, if orographic cloud observations are confined only to the region of the windward slopes, liquid water and variable, but often low, concentrations of rimed ice particles would be observed. Without additional measurements farther downwind it might be concluded that there exists a potential to increase overall precipitation on the barrier when, in fact, the possibility of redistributing the precipitation without an overall increase exists. Of course, in some cases this may be a desirable goal (e.g., Hobbs et al., 1973).

A seeding scenario within synoptic-scale cloud systems has been suggested by Hobbs and Matejka (1980) based on the banded structure of these precipitation areas. In some cases, the regions between bands incorporate supercooled liquid water (Matejka et al., 1978; Rauber et al., 1984). This supercooled water may appear to present an unambiguous seeding opportunity. However, the presence of differential wind speeds and directions between layers where supercooled water might have been detected and, say, ice-spawning, upper-level generating cells, raises the possibility that supercooled water intercepted at one point could be overtaken and depleted by the fallstreaks from a new crop of generating cells. Also, low-level convergence in and near rainbands can advect supercooled water into the high ice particle rainband regions. Or it might be that these supercooled clouds would eventually transform into high ice-producing clouds that feed the rainband.

These are two possibilities that could make the opportunity to increase precipitation in the "between-band" situation less than it might otherwise be. These regions between rainbands should be tracked for several hours to determine whether the regions of supercooled water always remain liquid and do not contribute significantly to pre-

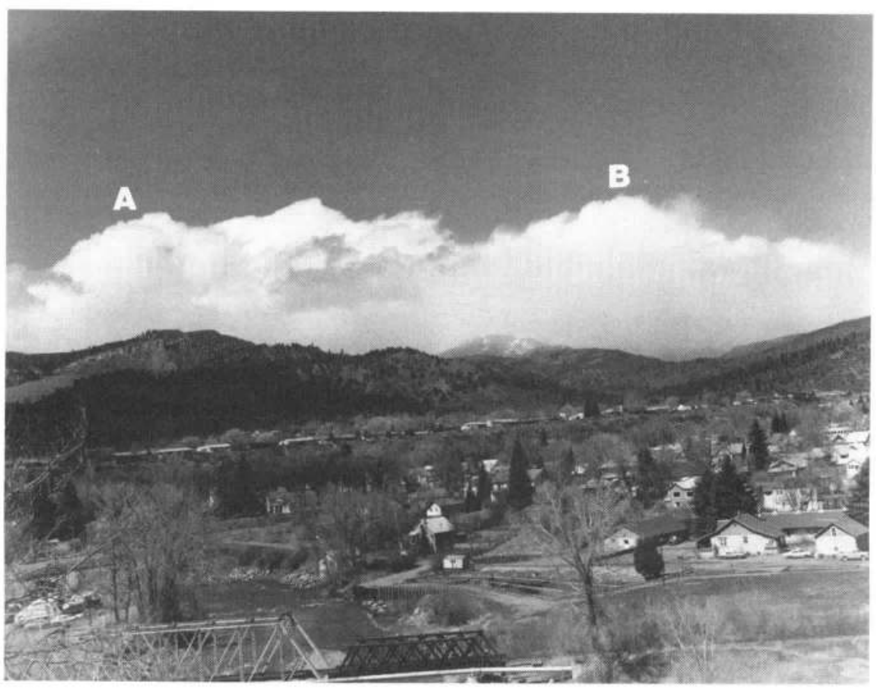

FIG. 12.7. Orographic clouds with modest convection topping the La Plata Mountains of southwestern Colorado. Point A shows a cloud top ascending the barrier. Point B is the glaciated "exhaust" portion of the cloud.

cipitation at the ground, say, through riming or through a transformation into an ice-producing cloud. Similar comments can be made about the occasionally high values of liquid water that have been encountered in frontal zones (e.g., Hobbs and Persson, 1982).

Also, for the reasons discussed above, aircraft icing reports alone cannot be used to infer seeding opportunities as has been suggested (Hill, 1980b).

Contributing to the problems of sampling and interpretation discussed before is the problem of what days to sample on: Which days would be the most climatologically representative? This is a particularly vexing problem since large field programs are generally planned far in advance and are conducted over miniscule time periods relative to climatological scales. No one, for example, would want to assess the precipitation modification potential in California or Colorado based on the frequency of clouds and storms that occur during an El Niño winter or to deduce the structure of all cumuli based on only those days when the cloud bases were $<0°C$. Yet rarely is the question of representativeness addressed in field reports—a point made by Johnson (1982). It is recommended that statements relevant to climatological perspective be included in field studies.

12.5. Conclusions

While good progress has been made during the past few years in understanding the nature of the precipitation process in mountainous regions, several outstanding issues remain that hinder the development of more reliable conceptual and numerical models of orographic clouds and precipitation. Those discussed in this paper include the following:

- There is not yet a widely agreed upon parameter that will predict the maximum ice particle concentrations that occur in most stratiform or convective clouds. For example, some studies have indicated that ice particle concentrations in building cumuli and in the upslope portions of lenticularlike, orographic cap clouds are temperature-dependent. In a more recent study, the broadness of the cloud-top droplet spectrum was found to predict ice particle concentrations with no temperature dependence indicated within the cloud-top temperature range of about $-6°$ to $-26°C$.
- An apparent ice-enhancing process near and downwind of mountain crests exists. Its cause should be elucidated. This will help in answering the question of whether increasing ice particle concentrations over the windward slopes of a mountain range should have a negative or positive effect in the immediate lee.
- Knowledge of the history, evolution, and role of supercooled liquid water regions within storms and over mountain barriers remains scant and needs more attention.
- There is a need for the documentation of the occurrence of substantial, nonprecipitating, supercooled clouds over mountain barriers.
- The microphysical processes that produce microscale temporal and spatial distributions of precipitation need attention.
- Diurnal effects, for example, the rise of cloud bases during the daytime and lowering at night, can produce systematic effects on the microstructure of orographic clouds. The extent of these effects is unknown.
- Considerable uncertainty exists on how to seed clouds. The presence of suitable clouds is useless if ground-released nucleants are inhibited by stable layers, are mistargeted, or arrive in the clouds in inappropriate concentrations, or when a seeding aircraft can provide only a thin strip of treatment in a sea of cloud.
- Statements on the representativeness of field data based on climatological studies should accompany short-term field measurements in order to put them into perspective.
- As a note of practicality, it should be demonstrated prior to field programs that conditions amenable to seeding can be forecast far enough in advance to permit treatment.

Acknowledgments. The author wishes to thank Professor Peter V. Hobbs for his advice, without which this paper would have been impossible. In addition, this paper benefitted greatly from the many comments and suggestions of Professor Roscoe R. Braham, Jr., as well as those of the anonymous reviewers.

REFERENCES

Auer, A. H., D. L. Veal and J. D. Marwitz, 1969: Observations of ice crystals and ice nuclei observations in stable cap clouds. *J. Atmos. Sci.*, **26**, 1342–1343.

Bergeron, T., 1950: Über der mechanismus der ausgiebigen Niederschlage. *Ber. Dtsch. Wetterdienstes*, **12**, 225–232.

Braham, R. R., Jr., 1964: What is the role of ice in summer rain showers? *J. Atmos. Sci.*, **21**, 640–645.

Briggs, G. A., 1985: Analytical parameterizations of diffusion: The convective boundary layer. *J. Climate Appl. Meteor.*, **24**, 1167–1186.

Cerni, T. A., and W. A. Cooper, 1980: Ice crystal concentrations in isolated cumulus clouds of Montana. *Preprints Eighth Int. Conf. on Cloud Physics*, Clermont-Ferrand, WMO, 195–198.

Chappell, C. F., L. O. Grant and P. W. Mielke, Jr., 1971: Cloud seeding effects on precipitation intensity and duration of wintertime orographic clouds. *J. Appl. Meteor.*, **10**, 1006–1010.

Coons, R. D., and R. Gunn, 1951: Relation of artificial cloud modification to the production of precipitation. *Compendium of Meteorology*, Amer. Meteor. Soc., 235–241.

Cooper, W. A., and C. P. R. Saunders, 1980: Winter storms over the San Juan Mountains. Part II: Microphysical processes. *J. Appl. Meteor.*, **19**, 927–941.

——, and G. Vali, 1981: The origin of ice in mountain cap clouds. *J. Atmos. Sci.*, **38**, 1244–1259.

——, and R. P. Lawson, 1984: Physical interpretation of results from the HIPLEX-1 experiment. *J. Climate Appl. Meteor.*, **23**, 523–540.

Cunningham, R. M., 1957: A discussion of generating cell observations with respect to the existence of freezing or sublimation nuclei. *Artificial Stimulation of Rain,* H. Weickmann, Ed., Pergamon, 267–270.

DeMott, P. J., and L. O. Grant, 1984: Development of ice crystal concentrations in stably stratified orographic cloud systems. *Preprints Ninth Int. Conf. on Cloud Physics,* Tallin, USSR, Acad. Sci. USSR, 191–194.

——, W. G. Finnegan and L. O. Grant, 1982: A study of ice crystal formation in a stably stratified orographic cloud. *Preprints Conf. on Cloud Physics,* Chicago, Amer. Meteor. Soc., 488–490.

Elliott, R. D., R. W. Shaffer, A. Court and J. F. Hannaford, 1976: *Colorado River Basin Pilot Project Comprehensive Evaluation Report.* Final Report to the Bureau of Reclamation, Aerometric Research, Inc., Goleta, CA, 641 pp.*

——, ——, —— and ——, 1978: Randomized cloud seeding in the San Juan Mountains, Colorado. *J. Climate Appl. Meteor.,* **17,** 1298–1318.

Fletcher, N. H., 1962: *The Physics of Rainclouds.* Cambridge University Press, 240–241.

Godske, C. L., T. Bergeron, J. Bjerknes and R. C. Bundgaard, 1957: *Dynamic Meteorology and Weather Forecasting.* Amer. Meteor. Soc. and the Carnegie Institution, 608–610.

Grant, L. O., 1968: The role of ice nuclei in the formation of precipitation. *Proc. First Int. Conf. on Cloud Physics,* Toronto, Amer. Meteor. Soc., 305–310.

——, C. F. Chappell, L. F. Crow, P. W. Mielke, Jr., J. L. Rasmussen, W. E. Shobe, H. Stockwell and R. A. Wykstra, 1969: An operational adaptation program of weather modification for the Colorado River basin. Interim Report to the Bureau of Reclamation, Colorado State University, 69 pp.*

——, P. J. DeMott and R. M. Rauber, 1982: An inventory of ice crystal concentrations in a series of stable orographic storms. *Preprints Conf. on Cloud Physics,* Chicago, Amer. Meteor. Soc., 584–587.

Hallett, J., and S. C. Mossop, 1974: Production of secondary particles during the riming process. *Nature,* **249,** 26–28.

Heggli, M. F., and D. W. Reynolds, 1985: Radiometric observations of supercooled liquid water within a split front over the Sierra Nevada. *J. Climate Appl. Meteor.,* **24,** 1258–1261.

——, L. Vardiman, R. Stewart and A. Huggins, 1983: Supercooled liquid water and ice crystal concentrations within Sierra Nevada winter storms. *J. Climate Appl. Meteor.,* **22,** 1875–1886.

Heymsfield, A. J., 1977: Precipitation development in stratiform ice clouds: A microphysical and dynamical study. *J. Atmos. Sci.,* **34,** 367–381.

Hill, G. E., 1980a: Reexamination of cloud-top temperatures used as criteria of cloud seeding effects in experiments on winter orographic clouds. *J. Climate Appl. Meteor.,* **19,** 1167–1175.

——, 1980b: Seeding-opportunity recognition in winter orographic clouds. *J. Climate Appl. Meteor.,* **22,** 1371–1381.

Hobbs, P. V., 1969: Ice multiplication in clouds. *J. Atmos. Sci.,* **26,** 315–318.

——, 1974: High concentrations of ice particles in a layer cloud. *Nature,* **251,** p. 694.

——, 1975: The nature of winter clouds and precipitation in the Cascade Mountains and their modification by artificial seeding. Part I. Natural conditions. *J. Appl. Meteor.,* **14,** 783–804.

——, and R. Farber, 1972: Fragmentation of ice particles in clouds. *J. Rech. Atmos.,* **6,** 245–258.

——, and D. G. Atkinson, 1976: The concentrations of ice particles in orographic clouds and cyclonic storms over the Cascade Mountains. *J. Atmos. Sci.,* **33,** 1362–1374.

——, and T. J. Matejka, 1980: Precipitation efficiencies and the potential for artifically modifying extratropical cyclones. *Preprints Third WMO Conf. on Weather Modification,* Clermont-Ferrand, WMO, 9–15.

——, and P. Ola G. Persson, 1982: The mesoscale and microscale structure and organization of clouds and precipitation in midlatitude cyclones. V. The substructure of narrow cold frontal rainbands. *J. Atmos. Sci.,* **39,** 280–295.

——, and A. L. Rangno, 1985: Ice particle concentrations in clouds. *J. Atmos. Sci.,* **42,** 2523–2549.

——, R. C. Easter and A. B. Fraser, 1973: A theoretical study of the flow of air and fallout of solid precipitation over mountainous terrain. Part II. Microphysics. *J. Atmos. Sci.,* **30,** 813–823.

——, L. F. Radke, J. R. Fleming and D. G. Atkinson, 1975: Airborne ice nucleus and cloud microstructure measurements in natural and artifically seeded situations over the San Juan Mountains of Colorado. Final Report to the Bureau of Reclamation, University of Washington, 89 pp.**

——, J. H. Lyons, J. D. Locatelli, K. R. Biswas, L. F. Radke, R. W. Weiss, Sr. and A. L. Rangno, 1981: Radar detection of cloud-seeding effects. *Science,* **213,** 1250–1252.

Huff, F. A., S. A. Changnon, Jr. and D. M. A. Jones, 1975: Precipitation increases in the low hills of southern Illinois. Part I. Climatic and network studies. *Mon. Wea. Rev.,* **103,** 823–829.

Johnson, D. B., 1982: Geographical variations in cloud-base temperature. *Preprints Conf. on Cloud Physics,* Chicago, Amer. Meteor. Soc., 187–189.

Koenig, L. R., 1963: The glaciating behavior of small cumulonimbus clouds. *J. Atmos. Sci.,* **20,** 29–47.

——, 1968: Some observations suggesting ice multiplication in the atmosphere. *J. Atmos. Sci.,* **25,** 460–463.

Kumai, M., 1951: Electron-microscope study of snow-crystal nuclei. *J. Meteor.,* **8,** 151–156.

——, 1961: Snow crystals and the identification of the nuclei in the northern United States of America. *J. Meteor.,* **18,** 139–150.

Ludlam, F. H., 1955: Artificial snowfall from mountain clouds. *Tellus,* **7,** 277–290.

Marshall, J. S., and W. McK. Palmer, 1948: The distribution of raindrops with size. *J. Meteor.,* **5,** 165–166.

Marwitz, J. D., 1980: Winter storms over the San Juan Mountains. Part I. Dynamical processes. *J. Appl. Meteor.,* **19,** 913–926.

——, W. A. Cooper and C. P. R. Saunders, 1976: *Structure and Seedability of San Juan Storms.* Final Report to the Bureau of Reclamation, University of Wyoming, 324 pp.*

Matejka, T. M., R. A. Houze, Jr. and P. V. Hobbs, 1978: Microphysical and dynamical structure of mesoscale cloud features in extratropical cyclones. *Preprints Conf. on Cloud Physics and Atmospheric Electricity,* Issaquah, Amer. Meteor. Soc., 292–299.

Medenwaldt, R. A., and A. L. Rangno, 1973: *Colorado River Basin Pilot Project Comprehensive Atmospheric Data Report, 1972–1973 Season.* Report to the Bureau of Reclamation, E. G. & G., Inc., Durango, CO, 376 pp.*

Mielke, P. W., Jr., G. W. Brier, L. O. Grant, G. J. Mulvey and P. N. Rosenweig, 1981: A statistical reanalysis of the replicated Climax I and II wintertime orographic cloud seeding experiments. *J. Appl. Meteor.,* **20,** 643–659.

Mossop, S. C., 1978: The influence of drop size distribution on the production of secondary ice particles during graupel growth. *Quart. J. Roy. Meteor. Soc.,* **104,** 323–330.

——, 1985: The origin and concentration of ice crystals in clouds. *Bull. Amer. Meteor. Soc.,* **66,** 264–273.

——, and J. Hallett, 1974: Ice crystal concentration in cumulus clouds: Influence of the drop spectrum. *Science,* **186,** 632–634.

* Available from the Bureau of Reclamation, Denver, CO 80225.

** Available from the Cloud and Aerosol Research Group, University of Washington, Seattle, WA 98195.

Plank, V. G., D. Atlas and W. H. Paulsen, 1955: The nature and detectability of clouds and precipitation by 1.25 cm radar. *J. Meteor.*, **12,** 358–378.

Rangno, A. L., 1979: A reanalysis of the Wolf Creek Pass cloud seeding experiment. *J. Appl. Meteor.*, **18,** 579–604.

——, P. V. Hobbs and L. F. Radke, 1977: Tracer and diffusion and cloud microphysical studies in the American River basin. Final Report to the Bureau of Reclamation, University of Washington, 64 pp.**

Rauber, R. M., L. O. Grant, D. Feng and J. B. Snyder, 1984: The spatial and temporal distribution of supercooled water during wintertime storms over the northern Colorado Rockies. Atmos. Sci. Paper No. 382, Colorado State University, 83 pp.

Rhea, J. O., L. G. Davis and P. T. Willis, 1969: *The Park Range Project.* Final Report to the Bureau of Reclamation, E. G. & G., Inc., Steamboat Springs, 288 pp.*

Rogers, D. C., and G. Vali, 1981: Observational studies of ice crystal production by mountain surfaces inside winter orographic clouds. *Preprints Eighth Conf. on Planned and Inadvertent Weather Modification,* Reno, Amer. Meteor. Soc., 54–55.

Sassen, K., 1984: Deep orographic cloud structure and composition derived from comprehensive remote sensing measurements. *J. Climate Appl. Meteor.*, **23,** 568–583.

Schemenauer, R. S., and G. A. Isaac, 1984: The importance of cloud top lifetime in the description of natural cloud characteristics. *J. Climate Appl. Meteor.*, **23,** 267–279.

Stewart, J. B., 1967: A preliminary study of the occurrence of ice crystals in layer clouds. *Meteor. Mag.*, **96,** 23–27.

Vali, G., D. C. Rogers and T. L. Deshler, 1982: Ice crystal and ice nucleus measurements in cap clouds. *Preprints Conf. on Cloud Physics,* Chicago, Amer. Meteor. Soc., 333–334.

Vardiman, L., 1978: The generation of secondary ice particles in clouds by crystal–crystal collision. *J. Atmos. Sci.*, **35,** 2168–2180.

——, and C. L. Hartzell, 1976: *Investigation of Precipitating Ice Crystals from Natural and Seeded Winter Orographic Clouds.* Final Report to the Bureau of Reclamation, Western Scientific Services, Inc., Fort Collins, 129 pp.*

Walsh, P. A., 1980: Cloud droplet distributions in wintertime Rocky Mountain clouds. *Preprints Eighth Int. Conf. on Cloud Physics,* Clermont-Ferrand, WMO, 183–185.

Weickmann, H., 1957: The snow crystal as aerological sonde. *Artificial Stimulation of Rain,* H. Weickmann, Ed., Pergamon, 315–326.

Weill, J. C., 1985: Updating applied diffusion models. *J. Climate Appl. Meteor.*, **24,** 1111–1130.

Wyngaard, J. C., 1985: Structure of the planetary boundary layer and implications for its modeling. *J. Climate Appl. Meteor.*, **24,** 1131–1142.

Young, K. C., 1974: The role of contact nucleation in ice phase initiation in clouds. *J. Atmos. Sci.*, **31,** 768–776.