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OCCURRENCE AND DETECTION OF SUPERCOOLED WATER IN THE ATMOSPHERE

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A brief introduction to supercooling is followed by a general discussion of clouds and fog. The next consideration is the freezing of supercooled drops on objects, and the primary concern here is aircraft icing. Climatological information on icing conditions is included for the Federal Republic of Germany. Next the relative merits of different wavelengths for remote sensing of clouds are examined. The final section is a summary and conclusions.

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1. INTRODUCTION

Water drops with temperatures above 0°C are in the liquid state, and they often remain liquid when the temperature decreases below the melting point of ice. Liquid water with a temperature below 0°C is said to be supercooled. The probability of water remaining supercooled decreases as temperature decreases. Supercooled clouds in the atmosphere are normal rather than exceptional at atmospheric temperatures between 0°C and -10°C.

The freezing temperature of water drops depends upon several factors. At low temperatures, smaller and purer drops are more likely to remain supercooled than larger drops which contain more impurities. In laboratory studies, pure water droplets only a few micrometers in diameter freeze spontaneously by homogeneous nucleation at about -40°C (Hobbs, 1974) or -41°C (Anderson et al., 1980). Homogeneous nucleation of larger pure water drops takes place at higher temperatures. When freezing nuclei are present, water freezes at higher temperatures. The laboratory study by Anderson et al. (1980) showed a memory effect in re-evaporation nuclei formed from the residue of frozen and later evaporated drops. Re-evaporation nuclei raised the threshold of ice nucleiation by approximately 2°C. The atmosphere typically contains particles which act as freezing nuclei between -15°C and -20°C. In one study (Jursa, 1985), 90 percent of clouds with supercooled water were warmer than -20°C. A few studies indicate that water in the atmosphere may occasionally contain impurities which have the opposite effect and permit it to supercool well below -40°C. Seagraves (1981) reported supercooling to -46°C in the atmosphere, and Curran and Wu (1982) found -47°C. Laboratory experiments by Hoffer (1961) demonstrated that droplets saturated with soluble salts commonly found in the atmosphere froze at much lower temperatures than the freezing point of pure water drops.

Supercooled droplets can remain liquid indefinitely if the temperature does not become too low and if the droplets are not mechanically disturbed. Supercooled droplets freeze rapidly if they are disturbed by collision with an object.

This report is divided into five parts. A general discussion of clouds and fog follows this introduction. Section 3 deals with the freezing of supercooled drops on objects. The main concern in section 3 is aircraft icing. Section 3 includes some climatological information. Section 4 deals with remote sensing of potential icing condtions. The final section contains a summary and conclusions.

2. CLOUDS AND FOG

2.1 Types of Clouds. The ten genera of clouds are usually divided into four groups: clouds with vertical development, low clouds, middle clouds, and high clouds. Information on cloud classification can be found in the Handbook of Meteorology (Berry et al., 1945) and in standard texts such as Haynes (1947) or Petterssen (1958). The International Cloud Atlas of the World Meteorological Organization (1956) contains many pictures and some discussion about types of clouds.

2.1.1 Clouds with Vertical Development. Cumulus and cumulonimbus are clouds with vertical development. These are dense clouds associated with strong vertical motion produced either by thermal buoyancy or by hydrodynamic instability. There are two ways that sufficiently intense thermal instability for development of convection currents in the atmospheric boundary layer can be produced. A steep lapse rate of temperature can be established when a cold air mass moves over a relatively warmer surface which continuously supplies heat energy at the base of the atmosphere. Strong solar heating at the surface also produces steep lapse rates in the boundary layer, and the convection resulting from this energy input is strongest in the afternoon. Hydrodynamic pressure forces often generate vertical accelerations comparable in strength to ordinary thermal buoyancy forces. A good example is the surface convergence and upward motion often produced up to 300 miles ahead of a strong and fast-moving cold front. A line of clouds can exist ahead of a cold front when the boundary layer is moist and convergence is strong. Further discussion of hydrodynamic instability may be found in Newton and Newton (1959) and Carbone (1982). There are also complex terrain-induced dynamical interactions with large scale weather patterns in some regions. They produce a nocturnal maximum of thunderstorms from central Texas to western Minnesota (Carbone et al., 1990).

Cumulonimbus generally evolves from well-developed large cumulus by a process of continuous transformation. Cumulonimbus clouds typically have a greater vertical extent and may have tops as high as high clouds. Cumulus and cumulonimbus may have bases as low as 500 m, but the presence of clouds at higher levels is not an impediment to development. Cumulus has the ability to penetrate other cloud layers. The cumulus retains its identity as long as it remains primarily vertically developed, is physically separate from the other cloud, and has a tower or dome-shaped summit. A discussion of the relation of convective and stratiform clouds associated with squall lines has been given by Lafore and Moncrieff (1989) who also included numerous references to earlier work.

Cumulonimbus clouds are commonly associated with thunderstorms, squall lines, or hurricanes and are quite frequently 10-15 km deep (Byers, 1965; Park et al., 1974; and Simpson and Dennis, 1974). However, the cumulonimbus clouds studied by Shishkin (1965) typically had a vertical thickness of only 4 km.

The thickness of cumulus clouds covers a considerable range. Plank (1969) measured thicknesses of Florida cumulus in the range 0.08-6.44 km. In other locations, Imianitov (1965), Betts (1973), Hill (1974) and Kapoor et al. (1976) found smaller ranges which were all within the larger range discovered by Plank. Perkey (1974) examined one cumulus at several times and measured thicknesses from 1.50 km to 5.25 km. Simpson and Dennis (1974) concluded that a medium tropical cumulus was 5.5 km deep, and the depth of a small trade cumulus was 2.5 km.

In the tropics most cloud top temperatures are above 0°C. Cumulus clouds are capped by the trade wind inversion. In one study from March through July on the windward coast of Hawaii, cloud top heights were generally less than 3.0 km, and bases were near 0.6 km (Takahashi, 1981). A typical temperature at 3 km is 10°C. Warm clouds in this study and in a later study in Hawaii (Takahashi et al., 1989) produced rain, and some were a part of long-lasting trade-wind rainbands.

Many cumulus in middle latitudes also contain little or no supercooled water. For example, in North Dakota on 10 June 1987, Stith and Politovich (1989) studied a cumulus cloud with a base near 2 km and a depth of less than 1 km. Temperature at the base of this cloud was 10.6°C, and temperatures near the top were 6-8°C.

Outside the tropics, cumulonimbus clouds usually reach such great heights that the temperature at the top is well below 0°C. The upper portions of cumulonimbus clouds in middle and high latitudes will normally consist of ice crystals or a mixture of ice crystals and water drops (Petterssen, 1958).

The range of values of liquid water content in cumulus and cumulonimbus clouds is quite large. The median of 342 observations from Lewis (1951) was 0.34 g/m^3, and the maximum was 1.71 g/m^3. Falcone et al. (1979) considered typical cumulus models with liquid water contents of 0.57 g/m^3 and 1.00 g/m^3. Values in precipitating cumulonimbus can be much higher. According to one study of Oklahoma thunderstorms (Jursa, 1985), liquid water contents greater than 9.4 g/m^3 occur in 25 percent of these storms. There is no conclusive evidence concerning the upper limit which can exist in severe storms, but it is probably 30 g/m^3 or higher. The very highest liquid water contents will usually be at temperatures above 0°C.

2.1.2 High Clouds. High clouds include the three genera of cirriform clouds: cirrus, cirrocumulus, and cirrostratus. Cirriform clouds form in air which is more stable than air which produces cumulus clouds. High clouds are most commonly found in the warm air above a wedge of cold air associated with a frontal system. Bases are above 6 km in the tropics, 5 km in middle latitudes, and 3 km in polar regions. Tops are seldom higher than 1.2 km to 1.5 km below the tropopause.

Cirriform clouds consist mostly of ice crystals which are fairly widely dispersed so that the clouds are optically thin. Some cirrus clouds do not produce a visual sensation in a human. These subvisual cirrus can be detected by lidar backscattering. Sassen, Griffin, and Dodd (1989) concluded that subvisual cirrus are fairly common.

A small amount of water is occasionally present in cirriform clouds. For example, Sassen, Starr, and Uttal (1989) detected supercooled droplets ≥5 μm in diameter near the base of cirrus clouds at temperatures from −21°C to −36°C.

Depths of high clouds vary from a fraction of a kilometer to several kilometers. Baranov (1965) analyzed 1370 measurements of high clouds from European USSR and 459 observations from Leningrad. Throughout the year at both locations, approximately half of the high cloud thicknesses were within the range 1.1–3.0 km. Thirteen percent were less than 0.6 km thick, and two percent were thicker than 5.0 km. The maximum thickness in Baranov's investigation was approximately 7.0 km. Studies by Guzzi et al. (1974) and Fritz and Rao (1967) were consistent with the findings of Baronov. More recently in the United States, Sassen et al. (1989) studied one widespread and well-developed overcast cirrostratus layer with a thickness of 6 km. This is exceptional. Pandey et al. (1983) recommended a typical thickness of 2 km for cirrostratus at all latitudes. A standard reference (Jursa, 1985) has stated that the average vertical extent of high clouds is 2.1 to 2.2 km and does not vary significantly with season.

Ice water contents of high clouds are typically in the range 0.03–0.30 g/m^3 (Jursa, 1985).

2.1.3 Middle Clouds. The category of middle clouds includes altocumulus and altostratus. In middle latitudes, these clouds most often form as a result of the slow ascent of extensive layers of air in association with the motion of frontal systems. Layers of altostratus are ordinarily quite widespread and extend over distances of hundreds of kilometers. Frequently, altocumulus clouds are found around the edges of an extensive layer of altostratus.

The lower boundaries of altocumulus and altostratus are often indistinct. Virga (wisps or streaks of water or ice particles falling from a cloud but evaporating before reaching the surface of the earth) are frequently seen with both types of clouds. Altostratus clouds are more likely to produce precipitation which actually reaches the ground.

The minimum height of the base of middle clouds is approximately 2 km at all latitudes, but the maximum height of the tops varies with latitude. Tops reach maximum heights of 4 km in polar regions, 7 km in temperate regions, and 8 km in the tropics.

Considerable variation exists in the thicknesses of middle clouds. In a rather extensive study of middle level clouds, Platt and Bartusek (1974) measured vertical cloud thicknesses of 0.2 km to 1.8 km with a mean of approximately 1 km. Pandey et al.'s (1983) recommended model thickness of 0.5 km for both altocumulus and altostratus falls near the lower end of this range. Falcone et al. (1979) suggested a typical thickness of 0.5 km for altostratus. Cagle and Halpine's (1970) range of ordinarily expected thicknesses of altostratus is 0.3 km to 2.1 km. Huschke (1959) stated that the vertical thickness of altostratus can vary from several hundred feet to thousands of feet. According to a summary table in the Handbook of Aviation Meteorology (Meteorological Office of Great Britain, 1971), altocumulus clouds are usually thin, but altostratus may develop to thicknesses of about 15000 feet (4.57 km).

Middle level clouds may consist of water, ice or a mixture of water and ice. A major portion of altocumulus usually consists of small liquid water drops. Altostratus clouds invariably contain at least some ice crystals, and they may consist mostly of ice crystals. Measurements listed by Lewis (1951) show median and maximum liquid water contents of 0.10 g/m^3 and 0.41 g/m^3 for cloud layers consisting of altocumulus or of both altocumulus and altostratus. Falcone et al. (1979) recommended 0.41 g/m^3 as a typical liquid water content for altostratus.

2.1.4 Low Clouds. Low clouds include stratus, stratocumulus, and nimbostratus. All three types are frequently associated with frontal systems in middle latitudes. They may occur both ahead of and behind the intersection of the front with the surface of the earth. Nimbostratus clouds are most often formed as a result of the slow ascent of extensive layers of air to sufficiently high levels for condensation to occur. Air in which stratus clouds form is more stable than air within other clouds at low levels. Stratus clouds form from the cooling of a layer of air. An example occurs when a warm and humid air mass flows over colder land. Slow lifting of a fog layer is a common form of stratus formation. The normal location of low clouds is somewhere between the surface and 2 km.

Nimbostratus has the largest vertical extent of the three types of low clouds, and it is the most likely to produce precipitation. Nimbostratus clouds are often so thick that they extend above 2 km, and they frequently merge with a layer of altostratus. Robinson (1959) measured thicknesses of nimbostratus from 1.8 km to 3.0 km. Mason (1971) discussed warm tropical nimbostratus clouds with thicknesses of 1.4 km and 2.1 km. According to the Glossary of Meteorology (Huschke, 1959), nimbostratus clouds are usually many thousands of feet thick.

Most of the literature on thicknesses of stratiform clouds is for stratus or stratocumulus or does not break the information down according to type. A study of 10/10 stratiform clouds in Germany was done by Essenwanger and Haggard (1960, 1962) who found a median vertical thickness of 1.1 km. Approximately two percent in this study were less than 0.15 km thick, and nearly all were less than 3.6 km thick. Robinson's (1959) numerous observations of stratocumulus thicknesses gave a range of 0.2 km to 1.5 km. Salomonson and Marlatt (1968) made a limited number of measurements of stratus in four locations and found a range of thickness of 0.15 km to 0.85 km. Based on several hundred soundings in the Arctic, the average depth of stratus was between 0.2 km and 0.3 km, and the average depth of stratocumulus was between 0.3 km and 0.4 km according to Voskresenskii and Dergach (1965). Kaveney et al. (1977) studied large numbers of stratus and stratocumulus in different air masses when no middle or high clouds were present. They measured thicknesses from 0.1 km to 2.6 km in polar continental air masses and from 0.2 km to 1.7 km in maritime tropical air masses. Pandey et al. (1983) suggested that a good model thickness for low stratus (below 1 km) is 0.5 km. They recommended a model thickness of 0.33 km for a very low stratocumulus and 0.66 km for a layer centered near 1 km. In summary, the evidence is that a typical vertical extent of low clouds in middle latitudes is approximately 1 km. Stratus and stratocumulus are very often thinner, and nimbostratus clouds are usually thicker.

Low level clouds most frequently consist entirely of liquid water. The probabilities that drops are supercooled or that some ice crystals are present depend upon season and geographical location. According to Lewis (1951), the median of 327 observations of liquid water content in stratus and stratocumulus was 0.22 g/m^3, and the maximum was 0.80 g/m^3. According to Jursa (1985), a typical peak value of liquid water content in a stratocumulus cloud is 0.8 g/m^3, but values as high as 1.3 g/m^3 have been observed in the densest part of some individual stratocumulus. Falcone et al. (1979) discussed some different cloud models of stratus and stratocumulus with mean liquid water contents within the range 0.15-0.55 g/m^3. The liquid water contents of two models of nimbostratus from Falcone et al. were 0.61 g/m^3 and 0.65 g/m^3.

2.2 Fog. Fog exists when the atmosphere contains a suspended aggregate of very small water drops (or ice crystals) which reduce horizontal visibility to 1 km or less near the surface. Fog is really just a cloud at the surface. It has long been known that fog is often similar to stratus. More recently, Welch and Wielicki (1986) have shown that many fogs have considerable structure and are much more like stratocumulus clouds. Fog may contain a small fraction of a gram to 1 g/m^3 of liquid water.

Fog is formed by a variety of meteorological processes, but there are two main types. When the ground loses heat at night by radiational cooling, radiation fog forms if the adjacent air cools below the dew point. Radiational cooling is at least a minor factor in the formation of most types of fog. Advection fog is fog which forms when warm, moist air moves over water or land with a lower temperature. When the ground warms in the morning, fog at the surface may lift and leave a stratus layer aloft. The reverse process often occurs at night when a stratus layer builds down to the surface. More detailed discussions of fog formation may be found in Berry et al. (1945), Byers (1965) and Willett and Sanders (1959).

The vertical extent of fog can be a few meters or more than 1 km. Vertical depths up to 100 m are quite common throughout the world. Sometimes coastal fogs and valley fogs are several hundred meters thick. Coastal fogs more typically have depths near 200 m. Many of the deeper and more persistent coastal fogs are really combined fog-stratus layers. Numerous references and other information on fogs have been summarized in a survey article by Stewart and Essenwanger (1982). Newer studies by Fitzjarrald and Lala (1989) and Findlater et al. (1989) are consistent with the earlier conclusions.

3. FREEZING ON OBJECTS

3.1 Types of Icing. More than one process can cause icing on an object. Icing can be caused by the impingement and subsequent freezing of supercooled drops on a surface. A layer of ice can form by direct deposition of water vapor when the temperature is below 0°C.

3.1.1 Clear Ice. A relatively transparent layer of ice formed by the freezing of large supercooled drops upon contact with a surface is called clear ice or sometimes glaze ice. Latent heat is released as a drop freezes, and therefore an entire large drop cannot freeze instantaneously. Large drops coalesce before freezing, and air pockets are small and not numerous. The ice is particularly smooth and clear if the drops are not highly supercooled. Density of such ice may be as high as 0.8 g/cm^3 or 0.9 g/cm^3 (Huschke, 1959).

3.1.2 Rime Ice. Rapid freezing of very small and highly supercooled drops upon contact with a surface causes air to be trapped between them. The resulting opaque granular mass is called rime ice. Density of rime ice is less than the density of clear ice. Rime ice may have densities as low as 0.2-0.3 g/cm^3 (Huschke, 1959).

3.1.3 Hoarfrost. Ice deposited directly from the vapor phase onto the surface of an object is known as hoarfrost. The process of formation of hoarfrost is similiar to the process by which dew is formed except that the temperature of the object is below 0°C. Hoarfrost is more fluffy and feathery than rime.

3.2 Aircraft Icing. The following references summarize much information about aircraft icing: Rodert (1951); Brun (1957); Cagle and Halpine (1970); Meteorological Office of Great Britain (1971); and Australian Bureau of Meteorology (1977). A more recent update on the topic is given by Hansman (1989).

Icing of an aircraft engine can occur in the air intake even when the ambient air temperature is above 0°C. Adiabatic expansion in the air intake causes a lowering of temperature so that condensation and ice formation can occur under appropriate circumstances. In piston engines with carburetors, additional cooling results from evaporation of gasoline. If there is no carburetor heater, icing can occur in clear air at temperatures above 25°C if the relative humidity is high.

It must be reiterated that a low temperature does not guarantee the absence of supercooled drops. According to Sassen, Starr, and Uttal (1989), airframe icing at temperatures from -42°C to -51°C has been reported. Impurities in drops can permit excessive supercooling. It is also possible for icing to occur in updrafts of air with droplets which have not yet reached a temperature in equilibrium with the air temperature at the new level.

3.2.1 General Discussion. According to one set of records (Cagle and Halpine, 1970), the relative frequencies of types of airframe icing are as follows: rime, 72%; clear-rime mixture, 17%; clear, 10%; and frost in flight, 1%. The infrequent formation of hoarfrost during flight can occur in a situation where an aircraft flies rapidly from a very cold region into a warm and moist layer. Hoarfrost forms when an aircraft is parked out-of-doors on a clear night if the temperature falls below the dew point and the air temperature is below 0°C.

The hazard of icing to aviation has decreased considerably since World War II. This is partly due to improved anti-icing devices. Higher altitudes and greater speeds of modern jet aircraft are also factors. Higher aircraft are usually above the worst icing conditions. Faster airplanes are subject to greater kinetic heating (George, 1960; Australian Bureau of Meteorology, 1977; and Lewis, 1951). Kinetic heating is frictional heating over most of an airframe but is the result of compressional heating at the leading edges. Kinetic heating varies as the square of the airspeed. The temperature increment from kinetic heating of a wing may be as much as 25°C for a speed of 500 knots (257 m/sec). The kinetic temperature rise in clouds is less than this value for clear air, but it could still be as much as 20°C. The increment of kinetic temperature should not exceed 1°C at a speed of 100 knots. Because of the decreased concern about icing for much of the aviation community, newer studies on the topic are limited. Many of the best studies are in the older literature. Much of the newer work is being done by investigators whose motivation is to learn about cloud physics. Many of those workers who are presently concerned primarily with aircraft are evaluating icing effects on helicopters.

Flying high where some ice crystals are present is often an advantage even if the altitude is not above all supercooled drops. There is evidence that the presence of a few large ice particles among supercooled drops can decrease the icing rate. In a study in Spain, Bain and Gayet (1982) measured icing rate as a function of liquid water content (LWC) in cumuliform clouds with temperatures from -8°C to -21°C. When large (diameter > 200 μm) ice particles were present in concentrations greater than 5 per liter, the icing rate was less than half as large as the rate without the ice particles. This was a larger reduction than was found earlier in the laboratory.

Sand et al. (1984) did an extensive study of airframe icing encountered by a research aircraft during 1091 flight hours through different kinds of clouds in different locations and seasons. Sand et al. did not find the significant effect of a few ice crystals on accumulation as a function of LWC that Bain and Gayet (1982) found. Most icing encounters led to a reduction in the rate of climb that increased linearly with the path integral of the supercooled LWC. Two cases of clear icing departed significantly from this linear relationship. This clear icing produced disproportionately large reductions in the rate of climb compared to reductions caused by similar total accretions of rime ice. These two unusual cases were apparently associated with the presence of small numbers of very large drops. Politovich (1989) showed that the presence of a few very large drops with diameters 30–400 μm is not sufficiently unusual to be considered rare. Icing conditions are considered moderate for supercooled LWC between 0.5 g/m^3 and 1 0 g/m^3. Sand et al. found that icing conditions were highly variable. Although average LWC > 0.5 g/m^3 could exist over paths as long as 200 km, <u>continuous</u> LWC > 0.5 g/m^3 was never encountered over a distance longer than approximately 13 km.

Avoidance of icing conditions requires knowledge of both vertical and horizontal variations in order to decide upon the best evasive action. In middle latitudes under cold winter conditions, a temperature inversion often exists above clouds with low ceilings (Guttman and Jeck, 1987) even though temperatures decrease with height below and in cloud layers. Low-level clouds capped by temperature inversions often build downward to the surface and form one thick layer of fog-stratus. An aircraft in this situation could run into serious icing conditions farther along the flight path if an attempt were made to fly under the cloud layer. A temperature profile over a considerable depth would be needed to know of this situation and to determine that the best evasive action would be to fly above the cloud. In other situations, ascending to higher altitudes within a cloud may be desirable if temperatures decrease with altitude fast enough that most or all water particles at the higher levels are expected to be frozen.

3.2.2 Effects of Icing. Effects on aerodynamics are crucial. Ice formation on an airframe changes its contours so that the pattern of the air current is altered. Formation of ice on the leading edge of an airfoil does not interfere with flight as much as formation of ice farther back. Clear ice from large droplets has more ice farther back than rime ice from small droplets. Because of this serious aerodynamic consideration, clear ice is usually considered more hazardous than rime ice even though rime ice is associated with greater small-scaled roughness. Accretion of any type of ice on wings is undesirable because it produces increased drag and decreased lift. Another aerodynamic problem is the change of shape of a propeller when ice forms on the blades. The load of ice on a propeller can be so unbalanced that stress and vibration cause serious trouble.

Under appropriate circumstances, extreme aerodynamic instability can be created by modification to the shape of the wing of an aircraft. Telford (1988) has documented a case where an experienced pilot and co-pilot crashed in a research aircraft. Everyone on the airplane died, but investigators were able to recover much of the data. According to the original plan, if icing seemed to be approaching a dangerous level, the aircraft would descend to a level where the ice would melt. This plan was followed, and the worst icing seemed to be over when unusual lift behavior intensified after the aircraft began descent. The worst problems occurred in clear air below the cloud base when accumulated ice started to melt. Apparently ice on the leading edge was melting and refreezing farther back on the wing where the temperature was lower. Telford's conclusion was that the resulting instability caused the crash.

There are other effects of icing. Instruments can malfunction. Control surfaces can be jammed. Icing on windshields and canopies can interfere with a pilot's vision. Finally, any kind of ice adds weight to the aircraft, but other trouble usually develops before the excess weight becomes critical.

3.2.3 Climatological Information. All available estimates of climatological frequencies of icing are based on assumptions and are subject to limitations. The various estimates give widely divergent frequencies of aircraft icing. The two most often quoted references are Ingram and Gullion (1964) and Heath and Cantrell (1972). Figure 1 was obtained by averaging data from these two reports. Both contain hemispheric maps for the Northern Hemisphere. Information for a given location is hard to determine because of the scale of the maps. Furthermore, local effects at some sites may cause additional deviations from the large-scale features of the maps. To obtain information for altitude levels, one must interpolate between pressure levels in the report by Heath and Cantrell (1972). Figure 1 is an estimate of mean frequency of icing over the Federal Republic of Germany (West Germany). Encounters with icing conditions at 15000 ft are less frequent than icing encounters at 5000 ft in winter because cloud particles are much more likely to be frozen at 15000 ft in winter. In summer, clouds are not normally supercooled at 5000 ft. Aircraft icing is infrequent at 15000 ft and non-existent at 5000 ft in summer in West Germany.

Figure 1 should be considered to give conservative estimates of icing probability. The maximum of 15 percent at 5000 ft in January is lower than some other estimates. Furthermore, probabilities may be higher at levels below 5000 ft. Rapp (1979) states that those aircraft flying low over rugged terrain in the southern two-thirds of Germany frequently encounter icing conditions from October through April. Rapp's estimate is that at 3000 ft in southern Germany in January, icing conditions should be expected about 50 percent of the time.

Information on supercooled fog in Germany shows that aircraft icing conditions sometimes occur at the surface. Essenwanger (1977) examined data for 0600 GMT at ten stations in Germany during a ten-year period. The average probability of fog with temperatures ≤32°F at 0600 GMT during December through February was 6.2 percent.

There was considerable variation from station to station, and the highest frequency was the 13.6 percent at Fuerstenfeldbruck. Although overall occurrence of fog was larger during fall, the mean frequency of fogs with temperatures ≤32°F at 0600 GMT was only 2.4 percent for September through November. The corresponding frequency for March through May was 1.2 percent.

The probablility of cool fog in Germany is much less if all hours are combined. Weinstein (1975) examined a few stations in Germany. During the peak month the average frequency of occurrence of supercooled fog was less than 2.0 percent at all stations. Weinstein's study showed that there is considerable variation from year to year at some stations. At Hahn, the average frequency of occurrence of supercooled fog during the peak month was 1.7 percent, but it was 10.8 percent during this month in the worst year based on 14 years of data. At Bitburg, the average during the peak month was 0.9 percent, but is was 14.9 percent during this month in the worst year based on 15 years of data.

Much low-level icing is associated with frontal passage in winter. Ryerson (1988) found that approximately half of all severe icing conditions on two New England mountaintops occurred during and immediately after a frontal passage.

Jackson (1980) published a very thorough icing climatology for northern Europe. The climatology consists of statistics for liquid water content (LWC) greater than of equal to 0.1 g m^{-3}, 0.3 g m^{-3}, 0.5 g m^{-3}, and 1.0 g m^{-3} at different temperatures ≤32°F. These statistics are based on the Smith-Feddes model of liquid water content. This model uses temperature, cloud type, relative height above the cloud base, cloud amount in the layer, and whether precipitation is or is not occurring at the surface below the point. A recent discussion of the accuracy of the Smith-Feddes model has been given in Haines et al. (1989). A computer program generated a zig-zag path through the area. The algorithm was adjusted to use the 20 statute mile encounter as the basic unit of horizontal measure to conform with the standard of the National Advisory Committee for Aeronautics (NACA).

Jackson's (1980) tables for Germany give probabilities of encountering 20, 40, 60,...,320 statute miles of the given conditions. Tables 1-4 are extracted from Jackson's tables. The probabilities in Tables 1-4 are from the columns for standard encounters with a horizontal extent of at least 20 statute miles for the years 1973-1976.

Table 1 contains probabilities of an encounter with liquid water content greater than or equal to 0.1 g m^{-3} at temperatures less than or equal to 32°F (0°C) in central Germany during November through February. The heights of the midpoints and boundaries of the layers are in feet above mean sea level. The lowest boundary is 100 ft because most of the terrain is at least this high in central Germany. The designation Zulu time was used by Jackson. This is the same as Greenwich Mean Time (GMT). Central European Time (CET) is one hour later than GMT. The probability of encountering at least 0.1 g m^{-3} at temperatures ≤32°F is greater than 20 percent in the layers centered at 2850 ft, 4350 ft, 5850 ft, and 8350 ft at 0100 CET and 1300 CET. Icing encounters are slightly more probable in the early afternoon than an hour after midnight in these layers and in the layer centered at 12100 ft. At lower levels, icing encounters are more probable in the middle of the night than in the early afternoon.

Table 2 is similar to Table 1 except that it is for the German coastal area. Here probabilities are greater than 20 percent at both times in layers centered at 2750 ft, 4250 ft, 5750 ft, and 8250 ft.

Table 3 contains probabilities of specified liquid water contents at temperatures ≤27°F in central Germany during the months November through March. An aircraft with 5°F of kinetic heating would still encounter icing at these temperatures. Probabilities of LWC ≥ 0.1 g m^{-3} reach 20 percent only in the layer centered at 8350 ft. Probabilities of high LWC ≥ 1.0 g m^{-3} at all levels to 14100 ft are less than 2.0 percent at 0100 CET but reach 5.5 percent at 5850 ft at 1300 CET.

Table 4 contains probabilities of specified liquid water contents at temperatures ≤27°F during November through April in the German coastal area. No probabilities are as high as 20 percent in this table. Probabilities of LWC ≥ 0.1 g m^{-3} are greater than 10.0 percent in layers centered at 4250 ft, 5750 ft, and 8250 ft at both 0100 CET and 1300 CET and at 12000 ft at 1300 CET. Probabilities of large LWC ≥ 1.0 g m^{-3} are actually a little larger in coastal areas of Germany than in central Germany.

Table 5 is also extracted from Jackson (1980). It contains probabilities of encountering specified liquid water contents at temperatures ≤27°F near Leningrad during December. Magnitudes of probabilities are comparable to those in winter in coastal areas of Germany. The most distinctive feature of the December data for Leningrad is the relatively low probability of small LWC at 5950 ft with much higher probabilities above and below. Jackson's study indicates that aircraft icing is significant in Leningrad as well as in Germany.

Even the best climatological information or meteorological forecasts may not be reliable indicators of aircraft icing if excessive air traffic exists in the area. Rangno and Hobbs (1983) presented evidence that the passage of an aircraft through supercooled clouds can produce high concentrations of ice particles. In their test at -8°C, a cylindrical volume along the flight path contained large numbers of ice particles of a uniform size distribution. The passage of an aircraft through a supercooled cloud causes localized changes somewhat similar to deliberate seeding of clouds.

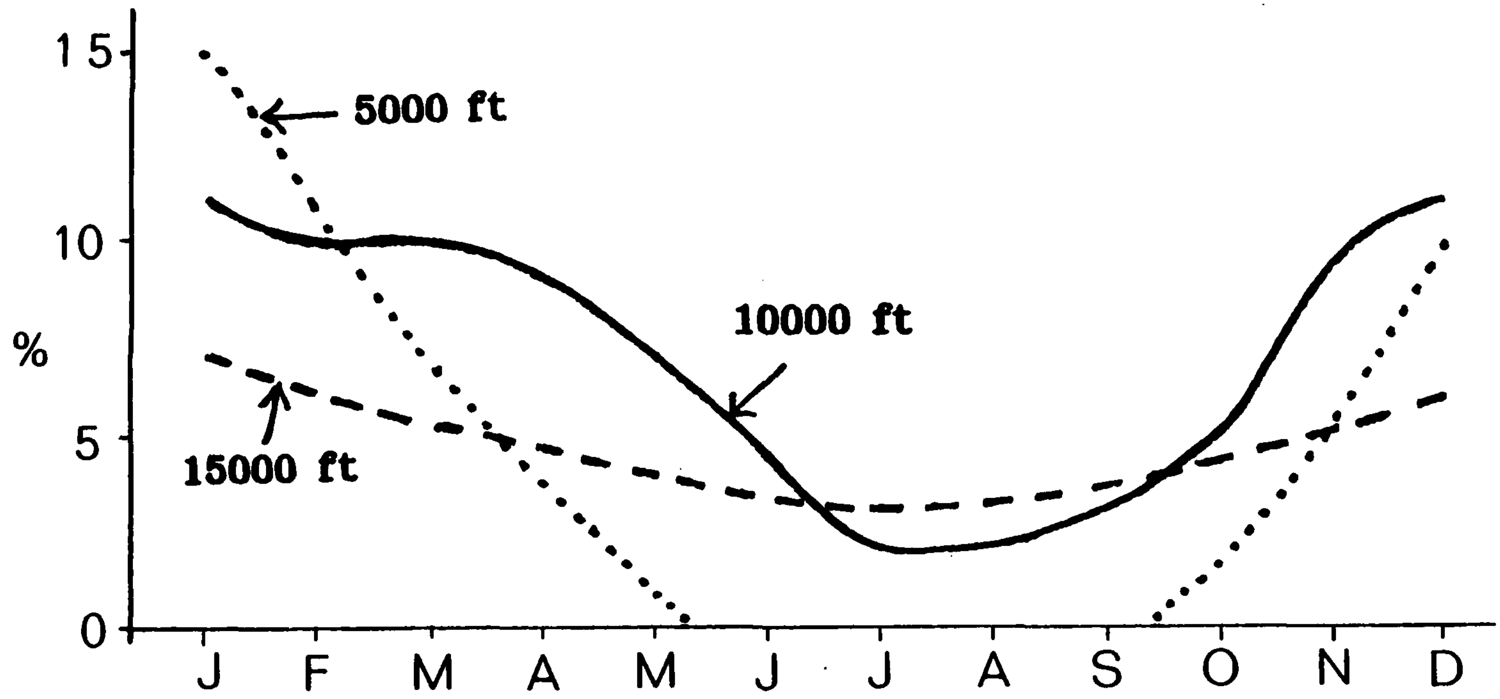

Figure 1. Mean estimate of probability of icing over West Germany from maps by Ingram and Gullion (1964) and Heath and Cantrell (1972).

Table 1. Probability of LWC ≥0.1 Grams per Cubic Meter at Temperatures ≤32°F in Central Germany during November through February

Layer Boundary	Layer Midpoint	00 Z (01 CET)	12 Z (13 CET)
14100			
	12100	9.64	11.69
10100			
	8350	27.96	30.39
6600			
	5850	26.95	29.46
5100			
	4350	27.75	30.52
3600			
	2850	22.63	24.81
2100			
	1600	14.22	12.35
1100			
	900	7.18	5.80
700			
	550	4.68	3.25
400			
	325	3.26	2.03
250			
	175	2.53	1.45
100			

Table 2. Probability of LWC ≥0.1 Grams per Cubic Meter at Temperatures ≤32°F in the German Coastal Area during November through February

Layer Boundary	Layer Midpoint	00 Z (01 CET)	12 Z (13 CET)
14000			
	12000	10.86	12.66
10000			
	8250	23.28	24.19
6500			
	5750	22.91	24.19
5000			
	4250	28.59	30.61
3500			
	2750	25.31	25.96
2000			
	1500	15.53	13.56
1000			
	800	6.74	4.65
600			
	450	3.59	2.17
300			
	225	2.17	1.10
150			
	75	1.90	0.69
0			

Table 3. Probability of Specified Liquid Water Contents at Temperatures ≤27°F in Central Germany during November through March

Layer Boundary	Layer Midpoint	LWC≥0.1 $g\ m^{-3}$		LWC≥0.3 $g\ m^{-3}$		LWC≥0.5 $g\ m^{-3}$		LWC≥1.0 $g\ m^{-3}$	
		00 Z	12 Z	00 Z	12 Z	00 Z	12 Z	00 Z	12 Z
14100									
	12100	8.86	10.68	0.34	0.44	0.28	0.41	0.28	0.36
10100									
	8350	21.64	24.30	2.68	3.96	0.25	0.39	0.25	0.39
6600									
	5850	15.22	15.82	1.89	5.54	1.89	5.54	1.89	5.54
5100									
	4350	13.06	13.92	1.39	4.53	1.35	4.51	1.35	4.51
3600									
	2850	9.24	8.44	1.43	2.65	1.00	2.43	1.00	2.43
2100									
	1600	4.18	2.29	0.72	0.75	0.38	0.55	0.35	0.55
1100									
	900	1.41	0.72	0.16	0.15	0.09	0.08	0.09	0.08
700									
	550	0.75	0.28	0.07	0.04	0.06	0.03	0.06	0.03
400									
	325	0.44	0.15	0.05	0.02	0.05	0.01	0.05	0.01
250									
	175	0.35	0.10	0.04	0.01	0.04	0.00	0.04	0.00
100									

Table 4. Probability of Specified Liquid Water Contents at Temperatures ≤27°F in the German Coastal Area during November through April

Layer Boundary	Layer Midpoint	LWC≥0.1 g m^{-3}		LWC≥0.3 g m^{-3}		LWC≥0.5 g m^{-3}		LWC≥1.0 g m^{-3}	
		00 Z	12 Z	00 Z	12 Z	00 Z	12 Z	00 Z	12 Z
14000									
	12000	9.33	11.69	0.90	2.35	0.85	2.33	0.85	2.33
10000									
	8250	17.32	18.35	3.40	5.23	0.97	2.81	0.97	2.81
6500									
	5750	10.75	14.13	2.99	7.50	2.59	7.50	2.58	7.50
5000									
	4250	12.46	15.53	2.44	6.41	2.42	6.39	2.42	6.39
3500									
	2750	8.07	7.89	1.75	2.99	1.51	2.79	1.51	2.79
2000									
	1500	2.58	1.87	0.44	0.54	0.34	0.36	0.34	0.36
1000									
	800	0.71	0.47	0.07	0.07	0.03	0.02	0.03	0.02
600									
	450	0.36	0.22	0.02	0.02	0.01	0.01	0.01	0.01
300									
	225	0.20	0.10	0.00	0.02	0.00	0.01	0.00	0.01
150									
	75	0.15	0.06	0.00	0.02	0.00	0.01	0.00	0.01
0									

Table 5. Probability of Specified Liquid Water Contents at Temperatures ≤27°F near Leningrad during December

Layer Boundary	Layer Midpoint	LWC≥0.1 g m^{-3} 00 Z	LWC≥0.1 g m^{-3} 12 Z	LWC≥0.3 g m^{-3} 00 Z	LWC≥0.3 g m^{-3} 12 Z	LWC≥0.5 g m^{-3} 00 Z	LWC≥0.5 g m^{-3} 12 Z	LWC≥1.0 g m^{-3} 00 Z	LWC≥1.0 g m^{-3} 12 Z
14200									
	12200	4.10	8.72	2.46	3.36	2.46	3.36	2.46	3.36
10200									
	8450	20.49	25.21	3.28	5.04	2.46	3.36	2.46	3.36
6700									
	5950	4.92	8.40	2.46	5.04	2.46	5.04	2.46	5.04
5200									
	4450	14.75	19.33	2.46	7.56	2.46	7.56	2.46	7.56
3700									
	2950	14.75	26.05	1.64	5.88	1.64	5.04	1.64	5.04
2200									
	1700	18.03	21.85	0.00	6.72	0.00	5.04	0.00	5.04
1200									
	1000	15.57	13.45	0.00	2.52	0.00	1.68	0.00	1.68
800									
	650	9.02	3.36	0.00	0.00	0.00	0.00	0.00	0.00
500									
	425	2.48	0.84	0.00	0.00	0.00	0.00	0.00	0.00
350									
	275	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
200									

4. REMOTE SENSING

The ability to determine cloud characteristics by remote sensing would permit an aircraft to select the least hazardous path. This task is complicated because simple measurements of the magnitude of backscattered energy do not distinguish between water and ice particles and do not determine drop sizes. General information on atmospheric propagation is given below.

Although both scattering and absorption cause attenuation by removing energy from a propagating beam, there are substantial differences between the phenomena. Absorption is explained by quantum theory, and is associated with changes in the internal energy states of molecules. Molecular absorption has a very strong and rapidly varying dependence on wavelength. Scattering is explained in terms of the wave theory of light, and no net change in internal molecular energy is associated with scattering. Dependence of scattering on wavelength is much smoother and slowly varying.

This section contains overviews of atmospheric absorption and scattering at different wavelengths. One subsection is devoted to the effect of scattering on polarization. Discussion of the importance of these factors in cloud probing is also included.

There are many sources of more detailed information than can be given here. McCartney (1976) and Hobbs and Deepak (1981) supply information on clouds and their optical properties in considerable detail. Bissonnette et al. (1988) investigated backscattering in artificial clouds in the laboratory. The main emphasis is on remote sensing of ice clouds in Wu (1987), Platt (1979), Platt and Dilley (1979), and Platt et al. (1987). Propagation over a large number of wavelengths is included in Gomez (1981). References concerning propagation at millimeter and submillimeter wavelenths are Essenwanger and Stewart (1980), Button and Wiltse (1981), Liebe (1983), and Liebe et al. (1989). Absorption of laser wavelengths is discussed by McCoy et al. (1969). Stewart (1987) summarizes the literature about visible wavelengths. Burch (1968) gives considerable detail on atmospheric absorption from 278 μm to 2 cm. Pasqualucci et al. (1983) examine depolarization of longer wavelengths. References on depolarization of shorter wavelengths are Liou and Lahore (1974), Sassen (1974), Derr et al. (1976), and Pal and Carswell (1973, 1977).

4.1 Absorption. Quantum theory postulates that three kinds of transitions can occur in a molecule which absorbs one photon of radiant energy. These are rotational, vibrational, and rearrangement of electronic sturcture of an atom. If the energy of a photon in a beam of radiant energy matches the energy required for a permitted transition, absorption occurs. Any photon can be absorbed if it has sufficient energy to separate an electron from an atom completely.

The energy of a photon is proportional to the frequency, and the constant of proportionality is called Plank's constant. Electronic transitions are associated with the highest frequencies, and rotational transitions with the lowest frequencies. Vibrational transitions are intermediate.

There exist so-called window regions of weak absorption between strong absorption lines. The absorption which occurs in windows is called continuum absorption. One explanation is that window absorption is caused by the sum of the absorption in the far wings of numerous absorption lines throughout the spectrum. Absorption by water vapor dimers and polymers is sometimes suggested as an explanation for excess absorption in many windows.

In the portion of the spectrum between 20 μm and 700 μm, atmospheric absorption is very strong. Peak absorptions at standard sea-level conditions are thousands of decibels per kilometer in this spectral region, and absorptions in windows are in the range 30-1000 dB/km.

Near millimeter wavelengths, water vapor is the chief absorber, and therefore absorption can be much higher than for standard conditions. Absorptions in windows centered near 730 μm and 880 μm are about 20 dB/km and 11 dB/km, respectively, at standard conditions. Atmospheric absorptions in windows at wavelengths beyond 1 mm are lower. Molecular absorption of 10 dB/km in the window centered near 1.3 mm would be associated with a very high absolute humidity. Atmospheric attenuations are lower in windows at longer wavelengths.

Molecular absorption is always small in much of the spectrum below 20 μm. Molecular absorption is usually negligible at visible wavelengths even when heavy pollution exists. Molecular absorption is also very small in several windows in the infrared. Particularly broad windows exist near 2.2 μm, 4 μm, and 10 μm.

4.2 Scattering. When electromagnetic energy is scattered by a particle, magnitude and direction depend upon the ratio of the diameter of the scattering particle to the wavelength of the scattered energy. When the particle diameter is less than about one-tenth of the wavelength, equal amounts of energy are scattered into the forward and back hemispheres. If a particle is larger, the total scattering is greater, and a larger portion of the energy is scattered in the forward direction. When the particle diameter is comparable to or larger than the wavelength, most of the large total scattering is in the forward lobe.

When the signal is scattered at an angle which is 180° to the direction of propagation of a radar or lidar beam, it is the backscattered energy. This is the energy which is measured by most radar and lidar systems. They use a configuration in which the transmitter and receiver are side by side. Such a configuration is called monostatic.

Cloud particles are so close to each other that single scattering does not account for all of the energy which reaches a lidar or radar receiver. The intensity of the returned signal depends upon the geometry of the optical system as well as upon the characteristics of the cloud. Because of the divergence of the transmitted beam, the multiple-scattering component depends upon the distance between the cloud and the transmitter. Sophisticated techniques can be developed to account for these factors.

4.3 Polarization. Scattering can change polarization characteristics of a beam. Depolarization occurs with multiple scattering or when individual scatterers are anisotropic and/or non-spherical.

Radiant energy backscattered from an isotropic sphere retains the polarization of the incident energy. Small water drops are nearly spherical. Multiple scattering from water drops causes some depolarization which increases as the signal penetrates deeper into a cloud. Linear depolarization observed in water clouds is usually in the range 0.02–0.04.

Depolarization by ice crystals occurs in the backscattered energy rather than in the forward scattered energy. Multiple scattering has little effect on depolarization in ice clouds. Increased depolarization is not observed as a beam penetrates farther in an ice cloud. Linear depolarization of a signal backscattered from an ice cloud is typically near 0.3. The usual range is 0.2–0.8. In special situations depolarization may be greater than unity.

4.4 Discussion. Conventional usage dictates that radar (radio detection and ranging) applies to wavelengths near and beyond one millimeter, and lidar (light detection and ranging) refers to visible and infrared wavelengths. A good reference which compares the relative advantages of lidar and radar is Derr (1978).

Particles in non-precipitating clouds are typically much less than one-tenth of the wavelength of a millimeter wave radar. It follows that Rayleigh scattering occurs, and attenuation of the beam by scattering is small. A radar beam can penetrate several kilometers into a cloud which is not developing precipitation. An 8.6 mm (35 GHz) radar may detect a useful signal from a moderate or thick cloud. Thin clouds may escape detection even though they contain enough supercooled water to cause aircraft icing. Longer wavelengths are more suitable for detecting rain. Shorter wavelength radars near 94 GHz (3.2 mm), 140 GHz (2.1 mm), and 230 GHz (1.3 mm) are still somewhat experimental. Furthermore, water vapor attenuation is not always negligible in windows near 2.1 mm and 1.3 mm.

Wavelengths of lidars are nearer the size of non-precipitating cloud particles. Visible lidars operate at wavelengths shorter than most of the droplet radii. Mie scattering theory applies, and scattering is very large. Penetration depths are usually less than 300 m and often less than 100 m. The major portion of the backscattered signal comes from the 10 m nearest the edge of thicker clouds. The path within 40 m of the edge of thinner clouds contributes most of the information in a backscattered signal. These thinner clouds may contain a few large drops, and they are hazardous to aircraft if they are supercooled.

Several laser sources are available for lidars. In the visible there are ruby, He-Ne, and frequency-doubled Nd-YAG. In the infrared, there are Nd-YAG and carbon dioxide. Because lidar wavelengths are much shorter than radar wavelengths, lidar instrumentation can be smaller than radars. Nd-YAG lidar systems have been used for cloud studies such as the ones by Boers et al. (1988) and Spinhirne et al. (1989).

Two factors must be considered as disadvantages of lidars operating at and near visible wavelengths. In the daytime, solar radiation produces considerable background noise between 0.2 μm and 3 μm. Another problem is that the wavelength range 0.4-1.4 μm is the most hazardous to the human eye. Danger exists when one views a direct beam or specularly reflected energy. Maximum distances at which direct viewing is hazardous are in the range 1-10 km for some typical military lasers according to Sliney and Wolbarsht (1980). In combat situations where binoculars, telescopes, and other optical viewing devices are used, a maximum hazardous distance of 10 km could be increased to 80 km.

Both lidar and radar are capable of producing polarized signals. The larger total signal from the lidar can give more reliable polarized components for thin clouds.

5. SUMMARY AND CONCLUSIONS

The first part of this report discusses clouds and fog. A typical vertical thickness of low and middle clouds is about 1 km, and they are seldom thicker than 4 km. The average thickness of high clouds is approximately 2 km, and the maximum is about 7 km. Clouds with vertical development average about 3 km in vertical thickness, but thicknesses can reach 10-15 km in thunderstorms, squall lines, and hurricanes. Vertical depths up to 100 m are quite common in fogs throughout the world. Coastal fogs typically have depths nearer 200 m. Fogs several hundred meters deep can exist. High clouds usually have little or no supercooled water but consist mostly or entirely of ice crystals. The probabilities that other clouds and fogs contain supercooled water depend upon location and season.

Encounters of an aircraft with supercooled water drops in the atmosphere are the main cause of airframe icing. Direct deposition of ice from the vapor phase in flight occurs only about one percent of the time. Completely frozen ice crystals are not hazardous. Small and highly supercooled drops freeze rapidly and cause rime ice. Large and slightly supercooled drops freeze more slowly and produce clear ice which is sometimes called glaze.

Available estimates of climatological frequencies of aircraft icing are subject to limitations, but it appears that the probability of aircraft icing at low levels in Germany is far from negligible. One investigator believes that icing conditions exist half the time in January at 3000 ft in southern Germany. Other evidence indicates that supercooled water is found at least one-fifth of the time in winter from about 2000 ft to 10000 ft in Germany. Heavy icing is probable near 6000 ft at least 5 percent of the time in early afternoon in winter in Germany.

Desirable wavelengths for remote sensing of clouds in the lower troposphere are limited because of molecular absorption. This absorption is very strong in the spectral range 20-700 μm. Molecular absorption is usually negligible in the visible portion of the spectrum even when heavy pollution is present. Molecular absorption is very small in several window regions between strong absorption lines in the infrared. Absorption in windows at wavelengths beyond 1 mm is small enough that remote sensing is feasible. Instruments which operate at wavelengths near and beyond 1 mm are called radars, and those which utilize visible and infrared energy are called lidars.

Lidar has advantages and disadvantages. First, lidar can detect much thinner clouds because backscattering of the shorter wavelengths by cloud droplets is much larger. Second, lidar instrumentation can be made smaller because of the shorter wavelengths. There are problems with lidars in and near the visible. Eye safety is very important from 0.4 μm to 1.4 μm. Furthermore, solar energy in the daytime can be a significant source of noise at wavelengths up to about 3 μm. These are less serious problems with carbon dioxide lasers near 10 μm.

The main advantage of radar is that the longer wavelengths which scatter less can penetrate more deeply into clouds.

Both lidar and radar can produce polarized signals which are useful tools for discriminating between ice and water particles.

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