# CLOUD PROPERTIES MEASURED FROM AIRCRAFT – AN ASSESSMENT

Gabor Vali

Department of Atmospheric Science, University of Wyoming, Laramie, WY, USA

### Introduction

Cloud and precipitation studies continue to have a very strong empirical component. In situ observations from aircraft<sup>1</sup> are indispensable for documenting the composition of clouds and thereby providing diagnoses of the processes within them. The importance of such observations is well demonstrated by the progress they helped to achieve over the last 40–50 years. Advances were quite spectacular over this period both in aircraft capabilities and in the instrumentation carried by them. Over the past decade or so, and at an increasing pace, aircraft are also equipped with remote sensing instruments for observations of cloud–scale characteristics. These observations may stand on their own, but are most useful when combined with the information derived from the direct sampling probes. Furthermore, in most studies the aircraft observations are combined with data from ground–based and space–borne remote sensors (radars, lidars, radiometers, etc.); the latter provide the larger–scale context into which the aircraft measurements are embedded.

For many of the measurements made by in situ probes on aircraft there is, simply, no substitute. While a number of remotely sensed, and hence integrative, measurements can be used to interpret cloud composition in terms of the phase, size and shape of the hydrometeors, these interpretations are neither firm nor detailed enough for process studies. Hence, the evolution of a cloud, or the formation of precipitation within it, can only be diagnosed but not understood on the basis of remotely sensed data. Conversely, the interpretation of in situ data in terms of radiative properties and other impacts has proven to be severely limited by two factors: insufficient sampling rates of the in situ probes, and the complexities of three–dimensional cloud structure. It is the result of this difference and complementarity that makes advances in cloud studies dependent on both approaches to observations.

The scope of the measurements now required for descriptions of clouds – whether from process, impact or climatology perspectives – has increased enormously over the past decades. This demand is driven by increased understanding, and by the capacity of models both to utilize detailed inputs and to predict detailed outcomes. Also, it is increasingly clear that clouds cannot be studied in isolation, but must be treated in coupling with their environment. The question often debated in the '70s, whether dynamics or microphysics is most critical for an understanding of cloud and precipitation evolution, has been answered, for the majority of cases, by a 'both'. Measurements of parameters relevant to both types of processes are thus considered indispensable for further progress.

When considering aircraft instruments for aerosol and cloud observations, it is evident that these systems have also benefitted greatly from the miniaturization of electronic, and to a lesser degree, optical and me-

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<sup>1.</sup> Although not exclusively so, aircraft are certainly the principal means to obtaining in situ cloud observations. Blimps, kites, balloons, rockets, dropsondes are some of the other means; the instrumentation requirements for use on these vehicles are often different from those for aircraft. These approaches are note discussed in this paper.

chanical components. Technological advances opened many new possibilities, and promise more. So, both the demands and the possibilities grow – at uneven pace, and subject to limitations in ideas and in resources. Looking at the progress made in different directions, the most evident advances were in the capabilities of data systems. Regarding instrumentation, it appears that the most significant advances over the past decade or so have been in remote sensing, spurred by the space program and by military technology. Aircraft measurements also benefitted from advances in navigation (the GPS, and the laser–ring INS). In situ probes for hydrometeors have gotten a major jump in development in the 1970's with the Particle Measuring Systems (PMS) devices. For over twenty years after that, much work has been dedicated to the improvement, calibration and evaluation of the various PMS probes, but very few new devices were introduced. Only in the '90s is there a renewed push to develop a new generation of hydrometeor probes. Another recent factor in instrumentation is the more open exchange between scientists from the two blocks of the cold war.

In the following sections of this paper brief summaries are presented of what current measurement capabilities exist, principally for cloud composition and structure. Cloud composition is considered from the physical and not from the chemical point of view, i.e. number, size, mass, shape, etc. of the hydrometeors. Linked to composition are the state parameters, and also condensation and ice nuclei. Cloud structure here means the spatial distribution of its content and the air motions that shape that distribution. Other characteristics of clouds, such as optical, electrical and chemical properties, as well as trace gas, turbulent flux and other measurements, are important elements of the overall effort to understand cloud processes and cloud impacts, but are not discussed in this paper.

## 1. Cloud composition – condensed phase.

#### Water and ice mass measurements.

The mass concentration of condensed water (liquid water content, LWC; ice water content, IWC; and total water content, TWC) is still a difficult parameter to observe with accuracies better than several tens of percent. Major difficulties arise from the need that the instruments respond to a large range of particle sizes, that they have large sampling rates, that they measure mass concentration over a range of more than two orders of magnitude (from near 0.02 g m<sup>-3</sup> to above 4 g m<sup>-3</sup>), that they be independent of phase of the hydrometeors or distinguish between ice and water, and that they have rapid response times. Particle counting and sizing instruments, to be discussed in following section, can also be used to obtain integral mass concentrations. This is, however, an inherently imprecise approach, because of the third power of size which enters the calculation, and because, for ice particles, there is the additional variable of density. Also, the sampling rates of the spectrometers are at best marginal for sampling the larger particles which contribute much of the total water content.

### Cloud droplets.

While, strictly speaking, LWC should include the mass of all liquid hydrometeors, in practice, LWC is often used to refer to the mass concentration of cloud droplets (< 50 µm diameter). This usage is justified by the fact that cloud droplets usually contribute much more to the total mass than drizzle drops or raindrops. Nonetheless, it is important to state clearly what is meant in cases where there is the possibility of ambiguity.

With the two most widely used instruments (Johnson–Williams, JW<sup>2</sup>, and CSIRO–King<sup>3</sup> probes), LWC is derived from the cooling effect exerted by cloud droplets impinging on the heated sensor element that is exposed to the airflow outside the aircraft. A third version is described by Nevzorov (1980) and Korolev et al. (1996). The Johnson–Williams probe references the sensor element to another element that is cooled by the

<sup>2.</sup> The Johnson–Williams probes are no longer made.

<sup>3.</sup> There are many versions of this probe in use. Some were manufactured 'in-house' by the users and there are at least two commercial firms marketing such probes.

air but is not exposed to cloud droplets. The main features of the CSIRO–King probe are: the temperature of the sensor is kept constant (at about 160°C), buffer coils protect the sensor from heat loss to the supporting structure, and LWC is derived from the power needed to maintain the sensor temperature constant. Cooling due to the air alone is predicted from heat–transfer equations and is set using measurements taken in clear air (free of cloud). The Nevzorov probe uses a reference element that is not exposed to cloud and the active element is compensated (using alternating current) for the power needed to keep the elements at constant temperature; the additional power needed to react to cooling by cloud droplets is measured by a separate direct current signal. All these instruments have detection limits of about 0.02 g m<sup>-3</sup>, and have decreasing detection efficiencies for larger droplets (>30  $\mu$ m diameter for the JW, and >60  $\mu$ m for the CSIRO–King and Nevzorov probes). This limitation is due to the increasing probability that larger droplets disintegrate on impact, and that some of their mass gets blown off the sensor prior to full evaporation. The CSIRO and Nevzorov probes can provide measurements up to about 10 Hz frequency. The accuracy of these devices appears to be about 10% under good conditions but larger errors are not uncommon and not well understood.

A device originally designed to serve as a detector of aircraft icing (Rosemount icing rate detector), has been added to several cloud physics aircraft, and its quantitative use explored in the laboratory by Baumgardner and Rodi (1989). Field results are also quite encouraging (Heymsfield and Miloshevich, 1989), especially for the detection of very low LWC values. This device can operate only at temperatures below 0°C, since it measures the mass of the accumulated ice on the probe tip. The instrument becomes inoperable at high LWC (0.5...1 g m<sup>-3</sup>, depending on temperature and airspeed), because of incomplete freezing of the impacted water mass and because of the large fraction of time spent de–icing the sensor rod.

Optical response, rather than direct collection of cloud droplets, is the basis of two LWC measuring instruments. They operate on similar principles: scattered light from a volume containing a large number of droplets is analyzed with a special optical filter to yield various moments of the droplet size spectrum, among them the third moment, i.e. mass. The Gerber–PVM (particle volume monitor) instrument (Gerber et al., 1994) has been extensively tested both in wind–tunnels and in flight and has shown good stability and accuracy. It is limited to measuring droplets of <  $50 \,\mu$ m diameter by the design of the optical filter. The instrument of Lawson and Cormack (1995), the 'SPEC–CDS cloud droplet spectrometer', is designed to measure to larger droplet sizes; so far it has been used little on aircraft, but has seen extensive ground testing (on Mt. Washington).

Another approach to in situ mass concentration measurements is that of Ruskin (1967), and of Nicholls et al. (1990). In these instruments air is passed through an evaporator, so that all hydrometeors (liquid and solid) are vaporized, and the total water vapor content of the air is determined using a Lyman– $\alpha$  absorption hygrometer. The devices have a reported sensitivity of 0.005 g kg<sup>-1</sup> and an accuracy of 0.15 g kg<sup>-1</sup>. The mass concentration of the condensed phases is obtained by subtracting the saturated vapor content from the measured total. Air is taken into the instrument through a tube of several centimeter in diameter. The instrument is located either inside the fuselage or in some external housing. In one particularly productive version of this principle (Ström and Heintzenberg, 1994), the sample is taken with a counterflow virtual impactor (CVI) which separates droplets or crystals above a threshold size (say 2  $\mu$ m) from the airstream and thus concentrates them; the result is a higher water to air mass ratio and greater sensitivity. With the addition of a particle counter, the residues of the evaporated droplets or crystals can also be counted and thus their number concentration in the cloud determined (assuming that each droplet or crystal leaves a single non–volatile residue). In general, these evaporator measurements rely on a simple and direct principle and that makes the data reliable and unambiguous. The CVI application is specially promising.

#### Drizzle, rain, ice and total mass.

Adequate sampling of larger drops<sup>4</sup> and crystals requires large sampling rates, because of the usual strong decreases in number concentrations with increasing sizes.

<sup>4.</sup> The definition adopted throughout this paper is: 'droplets' are  $< 50 \,\mu$ m in diameter, 'drops' are  $> 50 \,\mu$ m in diameter. Drops may further be classified as 'drizzle drops' and 'raindrops', with 500  $\mu$ m as the approximate demarcation point.

Because larger drops and crystals have a tendency to break into fragments upon impact at aircraft speeds, a simple scaling up of the cloud droplet instruments is not practical. Yet, the principle of mass measurement via the power needed to evaporate the hydrometeors has been applied in at least two heated–element collector devices for larger drops and crystals. Principally for the measurement of IWC, an instrument based on the principle of the CSIRO probe was built by King and Turvey (1986). The heated element was made much larger (about 1 cm width) and was given a convex cylindrical shape to allow particles to remain in contact with the element long enough to be evaporated and not be blown off the probe. Some field tests showed this approach to be promising, but it has not been given sufficient attention yet. Nevzorov (1980) introduced a similar probe, using a convex depression in the end of a cylinder to hold the impinging drops and crystals. Korolev et al. (1996) present promising results obtained with this probe, though the size–dependence of the collection efficiency is not yet published. In general, due to their larger physical size and possibly low collection efficiencies for small objects, these collector devices may in fact not be sensitive to cloud droplets, so that to obtain the TWC the data from them may have to be combined with that from a probe for cloud droplets. Uncertainties in the overlap in size sensitivity between the two types of probes requires attention.

The CVI measurement mentioned earlier is also applicable to larger hydrometeors; in fact it yields TWC regardless of the size (above the threshold) and phase of the hydrometeors. CVI measurements of IWC in cirrus were reported by Ström and Heintzenberg (1994). A recent comparison (Gerber et al., 1997) showed that both the PVM and the CVI performed well in measuring IWC in cold wave clouds. Ice particles were nearly spherical and <50  $\mu$ m in size, thus within the range of the PVM. With maxima near IWC=0.01 g m<sup>-3</sup>, the two instruments agreed within 5%.

#### Remote sensing.

Aircraft–carried microwave radiometers are very attractive additions to in situ measurements. A feasibility study and some initial testing of a 'tomography' method to derive the three–dimensional distribution of LWC was reported by Warner et al. (1985) and Warner and Drake (1988). Principally for the validation of microwave measurements from satellites (SSM/I, TRMM), similar aircraft–mounted instruments have also come into use. For example, an airborne microwave imaging radiometer operating at 37 and 90 GHz is intended for the measurement of liquid water path. This measurement is limited to where the emissivity of the underlying surface is known, which in practice means over the ocean. Heymsfield et al. (1996b) and Olson et al. (1996) explore the use of a downward looking 10.7 GHz microwave radiometer for determining rain amount, and higher frequencies for the detection of ice regions. These instruments were mounted on the NASA ER–2 aircraft, along with a radar (to be discussed later). The radiometric data yield vertical precipitation profiles over the ocean; over land, radar data also need to be incorporated in the retrieval. The focus of these studies so far has been on intense rain events from tropical storms. In general, airborne microwave radiance measurements appear to offer significant new information, especially if combined with direct sampling of the clouds, so that the radiative transfer calculations can be constrained by specific data rather than the more general assumptions usually employed.

Based on analyses of measured droplet spectra from non-precipitating cumuli in Florida, Paluch et al. (1996) concluded that the correlation between LWC and radar reflectivity is strong enough, with altitude-dependent constants, that radar backscatter measurements yield quantitative LWC fields. For the narrow range of cloud conditions where this method can be expected to work, radar depiction of the spatial distribution of LWC is bound to be a useful addition to other observations.

While not strictly a LWC measurement, this is a reasonable point to mention instruments that measure other integrative cloud properties. One example of this group of instruments is the ATTEX extinction probe. It provides a relatively long path (fuselage to tail) measurement of the optical extinction in clouds which would otherwise be a quantity derived from the droplet spectra. Other short–path instruments have also been built for this purpose.

#### Calibration.

One of the main reasons for uncertainties with LWC measurements is, of course, the lack of a 'standard'. Theoretical prediction of performance is a crucial initial step. Laboratory calibrations are the next step for any device, and those should include wind tunnel tests. These are important, but still can never quite reproduce the range of conditions (airspeed, size distributions, water content, interferences like ice, rain, etc.) actually expected in clouds. Intercomparisons of instruments is the next level of evaluation. Finally, the reasonableness of the results in actual cloud observations is used to judge performance. The most acute test is the comparison of the measured values with that expected for adiabatic lift. Selection of cloud volumes where the adiabatic assumption is likely to hold is the key here. Examples of such comparisons have shown in recent years that not far cloud base cumuli may have adiabatic cores; the larger the cloud, the further from the base may adiabacity hold and the higher the LWC will be. Even so, such comparisons have their limitations too: opportunities for them will not occur at very cold temperatures, for example. Final assessment of a probe's performance will thus remain based on a combination of all the means of calibration, and require continuous vigilance.

#### Cloud droplet spectra.

### <u>FSSP</u>

The most widely used tool for droplet size measurements is the light–scattering device developed by Particle Measuring Systems, Inc. (PMS), called the Forward Scattering Spectrometer Probe, FSSP–100 (Knollenberg, 1976, 1981). FSSP's have been in use now for over 20 years so that the possibilities and limitations of the device are quite well known. There are several FSSP models in use, with differences in the elctro–optical definition of the sample volume and in data processing. When examining data from these probes, these possible differences must be born in mind. Calibrations and corrections applicable to one probe may not apply to another.

The FSSP is mounted external to the aircraft so that cloud droplets can freely stream through its sampling aperture. The instrument counts and sizes individual droplets by the light they scatter as they traverse a laser beam. Optical and electronic means are used to define the sample volume. The output of the instrument normally consists of droplet counts in fifteen size channels. The upper limit of the measurement range is adjustable from 7.5 to 15, 30 or 45  $\mu$ m droplet diameter. The sampling volume is large enough to provide adequate statistical accuracy for up to 10 Hz sampling rate in most clouds. Clouds with low droplet concentrations (<50 cm<sup>-3</sup>) and/or broad size distributions may not be adequately sampled.

The performance and the accuracy of the FSSP probe did received a great deal of scrutiny. Main factors influencing probe performance are: droplet scattering and optics geometry, beam size and homogeneity, depth–of–field definition, electronics response time and velocity rejection. The latter is the means of electronic discrimination of droplets passing only the edge of the laser beam. Main items requiring calibration are: sizing and sample volume, both as functions of airspeed and of each other, and coincidence corrections.

Theoretical analyses of the probe were made by Hovenac and Lock (1993) and by Jaenicke and Hanusch (1993), focussing attention on the sensitivity of probe performance to various parameters. Experimental tests were reported by Baumgardner et al. (1985), Cooper (1988), (Brenguier and Amodei (1989), Baumgardner and Spowart (1990), Kim and Boatman (1990), Kim et al. (1990), Korolev et al. (1991), Brenguier et al. (1993) and many others. Unfortunately, experimental difficulties, variations in conditions and differences among the probes often led to disparate results. Even so, these studies led to better ways of processing and analyzing the data, to improvements in the hardware configuration, and to an appreciation of the degree of confidence, and also caution, the data deserve. One important practical result is the method developed by Brenguier et al. (1993) for monitoring instrument performance by examination of the internal consistency and stability of the data. The application of this method requires that the probes be equipped with output lines for "activity", "total strobe" and "total reset". Injection of glass beads of known sizes into the sample volume remains the relatively reliable and simple method of 'first order' size calibration. Even simpler to use is the rotating pinhole device described by Hovenac and Hirleman (1991); this calibrator can be readily attached to the probe and give repeatable checks of size calibration. A commercial device operating on the same principle is also available. For the evaluation of coincidence corrections, Brenguier (1989) devised a random pulse generator.

The bead tests and the rotating disk calibrations are very useful for assuring repeatability of a probe's performance. They do not give assurance about absolute errors or about compatibility of data from different probes. Those are the issues still not fully resolved. A test and calibration of an FSSP with monodisperse water droplets of known sizes was done recently by Wendisch et al. (1996). The tests were made at the fall velocities of the droplets for the upper range settings of the probe with droplets ranging from 20–45  $\mu$ m diameter. With droplets passing the center of the beam, the tests revealed better than 15% agreement with the factory specifications, and also with the calculated response of the instrument. The same result held anywhere along the length of the beam. However, when letting droplets pass off–center, the detected size fell off toward the edges of the beam, by about 20% at the limit normally defined by the velocity rejection circuitry of the probe. Both this effect and problems arising from the electronic response of the probe lead to undersizing of the droplets; the error increases with velocity. They suggest errors in LWC of up to 41% for 100 m s<sup>-1</sup> air velocity. Based on these results Wendisch et al. derived distortion matrices for sampling at different aircraft velocities.

Overall, it is clear that the calibration of the FSSP is not a trivial task and there is more to be learned from future tests. Procedures are well established at least for maintaining the instrument in good working condition; absolute calibrations are still a problem. It is important that presentation of results of FSSP measurements be always accompanied by some description of the calibration of the device.

The latest hardware improvement of the FSSP–100 probes consists of the addition of a slit on the detector surface. This step improves the definition of the depth–of–field, at the cost of reducing the sampling rate.

An improved method of evaluating droplet concentrations from FSSP data is given by Pawlowska et al. (1997). Their estimation takes advantage of the known properties of Poisson processes to remove the noisiness that results from random sampling, without degrading the detection of sharp gradients.

The capabilities of the FSSP have been significantly extended in versions built by Brenguier (1993) and Baumgardner et al. (1993). The main difference between Brenguier's Fast–FSSP and conventional FSSP probes is that individual events are recorded rather than counting them in size bins. This permits the spatial distribution of droplets to be studied, specifically to obtain concentration data with centimeter resolution. Baumgardner's method achieves a similar purpose by recording the time intervals between droplet samples. High– resolution data from these probes has stimulated a number of investigations of cloud entrainment.

There are conflicting reports on the performance of the FSSP in the presence of ice particles. Since the calibration of the probe assumes spherical scatterers, substantial deviations can be expected for crystalline shapes. Irregular behavior of the FSSP in the presence of larger crystals was widely reported (e.g. Gardiner and Hallett, 1985). Yet, Gayet et al. (1996) found that for small near–spherical ice particles in aircraft contrails the size distributions from the FSSP formed reasonable extensions of measurements for larger sizes by a 2D probe.

#### Other probes

As already mentioned in connection with LWC measurements, the two probes (Gerber's PVM and the SPEC–CDS) in which scattered light from a large volume containing many drops is analyzed yield information not only on the total volume of the scatterers but also on other characteristics of the droplet distribution. In the PVM the second moment of the size–distribution, i.e. total surface area, is also measured using a separate detector and a different mask. The volume and surface data yield values of the effective radius in a more direct fashion than when derived from the size distribution. The effective radius obtained from the PVM can also provide a check on, or used as a correction of the size spectra obtained from the FSSP. A systematic difference,

believed to be due to the velocity-dependence of the FSSP response, was noted by Gerber et al. (1994) and Gerber (1996).

### Drop and crystal spectra.

For hydrometeors of drizzle and precipitation sizes (  $>50 \mu$ m ) the optical array probes of Particle Measuring Systems Inc. (PMS) continue to be the most widely used. Both in the one–dimensional (1D) and in imaging (2D) configuration, these probes utilize laser illumination and linear arrays of photodetectors.

The 1D probe's output is similar to that of the FSSP: numbers of particles in 15 size channels. With a resolution (channel width) of 12.5 or 25  $\mu$ m, the probe provides a bridge between the FSSP and the 2D probes. Particles that shadow the edges of the detector array, and which are therefore incompletely registered, are rejected. The influence of these rejections on the actual sample volume can only be estimated in a statistical sense based on the observed size spectra. Perhaps the most severe limitation of the 1D probes is that little can be done to reliably eliminate artifacts (particle breakup, shedding from the probe tips, etc.) from the data. Other error sources are the variation of depth of field with particle size, and the ambiguity caused by out–of–focus particles. For these reasons, the use of data from the 1D probes is somewhat risky and requires careful judgement. On the other hand, their use is favored by the instrument's relative simplicity and moderate data output rate.

The 2D imaging probes are perhaps the most powerful tools now in use for cloud particle studies; the availability of these probes had a rather significant impact on cloud physics. Various forms of the probes exist. The two most widely used versions have array–element sizes of 25 and 200  $\mu$ m, respectively, and contain 32–element detectors. By rapid sampling of the array, the shadow image of the particle passing through the laser beam is recorded. The sample volume of the 25  $\mu$ m resolution probe is ~50 mm<sup>2</sup> times the air velocity. The so–called 'gray–scale' model of the probe records the shadow intensity at two threshold levels. The basic design of these instruments didn't change much over the years. Most development work addressed calibration, shape recognition and artifact rejection methods in the processing software.

The first task that is addressed in the image analysis routines is to eliminate 'artifact' records. Contamination of the data can result from drops or crystals that hit the probe tips and break apart, from water streaming off the probe apertures, from water or ice on the mirrors, from electronic errors, etc. Each of these artifact image types need to be recognized through special algorithms. While there is no fully satisfactory way to accomplish the artifact rejection, procedures that have been designed do take care of the large majority of the problems.

Determination of the phase of the sampled particles (water vs. ice) was found to be best addressed through analyses of the 2D image shapes (quasi-round and smooth vs. irregular or faceted). This discrimination is useful for images that are at least about 6 times larger than the individual array elements and becomes more reliable the larger the drops are. Attempts to distinguish ice and water particles by the amount of depolarization they produce as they pass through the laser light failed to produce convincing results.

Extension of the water/ice discrimination problem is the recognition of ice crystal shapes (crystal habits). This is essential for the estimation of crystal surface area and mass, for prediction of absorption and scattering, for the estimation of the degree of riming on the crystals and for deducing the crystal's growth history. Accordingly, this question received considerable attention. Common to all image analysis attempts is the need to deal with partial images (particles at the edge of the sample volume, or larger than the sample volume). For water drops, this was addressed by Cooper (1980), and similar algorithms relying on crystal symmetry are incorporated in some of the other classification procedures. Such procedures, for classifying images, have been devised by Duroure (1982), Holroyd, (1987) and others. Heymsfield and Baumgardner (1985) report the results of intercomparisons among the various schemes. The success of these classification methods has not been critically tested due to the lack of independent determinations of crystal habits (other than visual inspection of the images) or of total crystal mass. In practice, subjective judgements still prevail. Recently, Korolev devel-

oped a statistical approach in which population characteristics rather than details about individual images are used to classify the data.

Beyond hydrometeor phase and shape, total drop or ice crystal concentration and size distributions are the main data to be derived from the optical array probes. The sample volumes of the probes are defined, to a first approximation, by the aperture width and by the total array width - both physical dimensions, subject to minimal uncertainties. However, for small particles the sample volume is dependent on both the size of the particle (because of depth-of-field limitations) and on velocity (because of electronics response), The combined effects of these factors is reported by Baumgardner and Korolev (1996). The size range affected is <150 (120)  $\mu$ m for the 50% shadow threshold probes and <240 (180)  $\mu$ m for 67% shadow threshold; the first figure is for 160 m s<sup>-1</sup> air velocity, the figure in parentheses is for 100 m s<sup>-1</sup>. Below these sizes the depth-of-field reduces by a factor 10 at roughly half the sizes given above. Another important source of error in the image data arises from the fact that a diffraction pattern rather than a geometric shadow is produced. For spherical objects, Korolev et al. (1991) and Korolev et al. (1996) derived a distortion matrix which can be used to correct observed size distributions. The effect of this correction is to increase the concentration of drops >100 µm and to decrease the concentration of smaller drops. For example, images nominally assigned to 500 µm drops arise with 10% probability from 600 µm drops, with 42% probability from 575 µm drops, with 28% probability from 550 µm drops, and the remainder from 500 and 525 µm drops. How similar effects alter the data recorded for ice crystals is not yet known. A further uncertainty, already mentioned, arises from approximations and assumptions involved with any treatment of partial images. Regardless of whether partial images are accepted or rejected, the sample volume has to be adjusted and a different value used for every size. Yet another type of problem was identified by Rauber and Heggli (1988): high concentrations of large cloud droplets, or drizzle drops may saturate the probe electronics with zero area images and thus lead to an under-recording of the larger crystals.

In-flight evaluations of the performance of the 1D and 2D probes are made on the basis of compatibility of the data with other probes (e.g. FSSP) over limited overlap ranges. There are no reference standards. But, the relative simplicity of the probes (at least for larger particles), and the fact that image quality can be quite well judged by inspection, provide fair confidence in the data. This is evident in the widespread use of these imaging probes.

Another approach to recording particle images is laser holography. Holography offers the major advantage of a large and undisturbed sample volume in which several particles might be present at the same time. Because the sample volume is far removed from mechanical parts, disturbance on the airflow is minimal. The method has the inherent disadvantage of intermittent sampling. Successful field uses of a holographic system ware reported by Brown (1989) and Lawson et al. (1996). The resolution achieved appears to be comparable or better than the imaging probes (25  $\mu$ m), and even droplets down to 12  $\mu$ m diameter have been successfully detected and sized.

Detection of ice particles in clouds was shown (Jones et al., 1989) to be possible with a rather simple, light–weight device which records the electric charge generated when ice particles collide with an ice–coated sensor wire. The large sample volume provides for high spatial resolution in the data, but no sizing information is obtained. The detection efficiency of the probe may vary with cloud conditions, influencing the reliability of the measured concentrations. Simplicity is an attractive feature of this device. With more experience, the probe might become a useful adjunct to other measurements.

Direct collections of ice particles by impaction was widely used before the availability of the imaging probes. Microscopic examination of the collections provided details on crystal shape and degree of riming that is not obtainable from the imaging probes. The decelerator collections of 'oil-hexane' slides on the Wyoming KingAir, and the Formvar replicator of Hallett still have their uses. In the former, crystals are impacted onto an oil-covered slide at 1/10th the aircraft speed, and the slides are saved for examination on the ground. In the replicator, crystals impact through a small aperture on a moving ribbon coated with Formvar; the Formvar preserves a 'cast' of the crystals (or drops). The latest instrument in which impacted crystals are examined under

high magnification is the DRI 'cloudscope': a forward–facing glass disk is viewed from the rear by a video microscope, crystals are recorded as they impact and then evaporate due to dynamic heating of the air. The instrument is housed in a PMS cylinder and has been used on the NASA DC–8 during the SUCCESS project.

Advancing from shadow imaging, holography and collection methods, direct optical imaging of crystals in the free air represents the most desirable method as far as detail and resolution are concerned. An instrument that accomplishes this without an excessive sacrifice in sample volume has been built and flight tested by Lawson (SPEC Cloud Particle Imager). Pulsed lasers illuminate crystals as they enter the sample volume of the instrument and its image is recorded by a CCD camera of 10<sup>6</sup> pixels. Data rate is kept manageable by real–time selection of image pixels for recording. Crystals or drops are sampled from a 5.3 mm<sup>2</sup> area across the airflow. Since the instrument is mounted external to the aircraft, and crystals flow through an open tube, disturbance is minimal so that the recorded crystal shapes are truly faithful. Growth patterns and riming are clearly discernible.

## 2. Cloud composition – state parameters

### Static pressure

While aircraft manufacturers provide carefully designed static ports, points where the aircraft's motion creates no pressure perturbations, any modification of the aircraft profile (by instruments mounted on the fuse-lage) can render the ports inaccurate. Tower fly–by comparisons are one way of checking the validity of the static pressure measurement. A more complete check can be obtained by the 'trailing cone' technique: a long tube is reeled out during flight and stabilized by a cone which is carefully constructed not to create pressure perturbations (Ikhtiari and Marth, 1964). Pressure measurements with this tube during various aircraft maneuvers allow the static port measurement to be checked for all likely flight configurations.

#### Air temperature.

The measurement of temperature from aircraft is complicated by aerodynamic heating, the need to protect the sensing element from physical damage and, in clouds, the possibility of wetting.

Based on wind-tunnel experiments and on flight tests, an acceptable solution to the first problem is to apply a fraction, r, of the theoretical heating effect which is calculated for an adiabatic compression of air at the velocity of the aircraft. The numerical value of r, called the recovery factor, is dependent on the housing design. With the instruments most frequently used, the Rosemount and reverse–flow probes, measurements with 0.5°C absolute accuracy, 0.005°C relative precision and about 1...10 m spatial resolution (depending on the housing design and airspeed) are possible.

Major consideration for many applications is the time response of the temperature probe. Best response can be achieved with small completely exposed sensors. This is not an option for in-cloud measurements, both because mechanical protection is needed and because of sensor wetting. Solutions to these problems invariably reduce the time response. The time and frequency response of one of the most commonly used instruments (Rosemount 102 probe) has been shown by Payne et al. (1994) to have a fast and a slow components. The former reflects the response of the sensing wire, while the latter reflects the heat transfer into the support structure. However, this analysis considered only the sensing element, not the housing, so it predicts better response than is indicated by in-flight data. Friehe and Khelif (1993) replaced the normal platinum wire sensing element in a Rosemount 102 probe with a small thermistor bead. Theoretical analysis of this system was presented by Fuehrer et al. (1994), showing the dependence of response on bead size and on the number and size of lead wires (which increase heat transfer). Response starts to degrade significantly for frequencies >1 Hz.

Sensor wetting can be a serious problem, since the cooling effect of evaporation (in subsaturated air) can amount to several °C. The reverse flow housing (Rodi and Spyers–Duran, 1972) directs airflow to the tem-

perature sensor in such a way that cloud elements are separated from it. The Rosemount 102 thermometer also has some, but much less complete separation of cloud elements. Even so, investigations revealed that these thermometers can become wet in clouds with high liquid water contents (Lawson and Cooper, 1990). There appears to be no problem in supercooled clouds, since water then freezes onto the probe housing rather than flowing on its surface and getting blown onto the sensor element. No full solution to wetting, for high liquid water contents, has been found so far; recent attempts (Lawson and Rodi, 1992; Haman, 1992) show only partial success. It is therefore, important to be aware of the problem and recognize its symptoms when analyzing temperature data from cloud penetrations.

The fast–response temperature sensor constructed by Haman, appear to resist wetting on relatively slow aircraft (<40 m s<sup>-1</sup>) (Haman et al., 1997). The long, open sensing wire is mounted on a vane and is protected by a rod upstream from it. Response time is near  $10^{-4}$  s, so that centimeter–scale structures can be observed.

A non-contact approach to temperature measurements in clouds was suggested by Nelson (1982) and tested by Lawson and Cooper (1990). This device is a short-path radiometric detector of emissions from  $CO_2$  at 4.255  $\mu$ m wavelength. Since this instrument has no parts directly exposed to the airstream, the only possible wetting problem is that of the windows and this can be prevented. The radiometer is also free of the uncertainties associated with determinations of the recovery coefficients of immersion sensors, and it has a response time which is only limited by electronics, rather than by airflow considerations. These advantages are significant, but so far only the prototype instrument was built and tested.

### Humidity.

The simple assumption that within clouds in the presence of liquid droplets (crystals) the humidity is 100% with respect to water (ice) is only a first approximation. Examinations of fine–scale structure at cloud boundaries, explorations of the region between ice and water saturation, plus the need for accurate measurements in the surroundings of clouds, including the inflow, necessitate humidity measurements of good accuracy and rapid response. The currently available instruments do not fully satisfy these needs.

The chilled–mirror type dew–point instrument, most widely used on research aircraft today, provides reasonably good data in clear air, but it has slow response (order of seconds), and is only accurate to perhaps  $\pm 1^{\circ}$ C. It is not reliable within clouds, so the solution usually adopted is to set RH=100% for periods when the aircraft is in cloud, as identified by droplet probes or other means. There is also uncertainty associated with dew point temperatures <0°C since there is no reliable way to predict whether the condensate on the mirror is liquid water or ice and so the measurement actually represents either the dew point or the frost point. Much below 0°C, let's say by about  $-10^{\circ}$ C, it is fairly certain that the coating is ice, but even this may be wrong. The potential error from this uncertainty increases with decreasing dew point temperature, but is fortunately only 1°C for  $-10^{\circ}$ C dew point. The simplest assumption, that all values below 0°C represent the frost point, is thus an acceptable one for most needs, and in fact does not introduce larger errors than the instrumental inaccuracies. The commercial versions of this device are limited in the lowest temperatures it can reach, therefore a special cryogenic unit was built at NCAR for use in the upper troposphere. Heymsfield and Miloshevich (1993) found the response time of this device to be about 2 s and its accuracy to be  $\pm 5\%$  in relative humidity.

One alternative to the chilled–mirror instruments is the Lyman–alpha hygrometer which measures absorption at the Lyman–alpha UV wavelength through a known path–length of the order of centimeter. Several of these instruments have been installed on aircraft, and provided fast–response data on humidity fluctuations. Problems with the stability and calibration of the device (Eloranta et al., 1989; Lind and Shaw, 1991) have not been fully resolved.

Infrared absorption offers another alternative for hygrometry (e.g. Cerni, 1994). IR at 2.6 μm wavelength is employed in the commercially produced Licor hygrometer which has been adapted to aircraft use. With carefully configured (high flow rate, turbulent) intake and tubing leading to the instrument it provides data with up to 10 Hz response. A second channel just on the shoulder of the main absorption line is used to correct for signal variations of extraneous origins. A instrument based on the same principle but with an open sampling path is described by Auble and Meyers (1992). The sample path is folded by a mirror between the source and the detectors and is longer than that of the Licor device. The source is chopped with a rotating disk that has filters for both the center and shoulder wavelengths.

## 3. Cloud composition – nuclei

### Cloud condensation nuclei, CCN.

The two most frequently used instruments for CCN measurements are the 'traditional' static diffusion chambers and the more complex continuous–flow chambers.

The static diffusion chamber is a batch–processing instrument and measurements are taken at one supersaturation at a time. These limitations result in a sampling frequency of about 30 s per datum and about 3 min for obtaining a complete spectrum. This type of instrument has a lower operating limit of about 0.2% supersaturation. For general airmass characterization these sampling rates are acceptable. Much of the literature on CCN originated from measurements with static diffusion chambers (cf. Twomey, 1977; Götz et al., 1991), and reliability, relatively small size and moderate cost lead to widespread use of these devices. For aircraft use the additional requirement of adequate pressure isolation from the interior of the aircraft cabin has to be addressed. There are approximately 2 to 5 such instruments in current use; the latest version installed on the Wyoming King Air (Wechsler, private communication) uses solid state laser illumination and a photodetector. Progress has also been made in the development of calibration methods for these instruments (Gras, 1995; Parente de Oliviera and Vali, 1995; Delene et al., 1997) so that measured concentrations are probably accurate to within 10 or 20%.

Instruments of greater capabilities (CCN spectrometers) have been developed by Fukuta and Saxena (1979) by Radke et al. (1981) and by Hudson (1989). The common advantage of these continuous–flow devices is that a full CCN spectrum can be obtained in about 30 s. In the first two of these devices air is drawn through one or more chambers in a steady flow and is subjected to a range of supersaturations; the number of droplets that form is then counted by optical single–particle counters (OPC's). In the instrument of Hudson, the sizes of the emerging droplets are used to infer the supersaturation at which they became activated. This design leads to simpler construction, yields essentially instantaneous spectra and extends the measurements to 0.01% supersaturation. Results from this instrument have been reported by Hudson (1993), Hudson and Li (1995) and Hudson and Svensson (1995).

A common concern for all types of CCN measurements from aircraft is that the air intake and ducts do not significantly bias the aerosol content of the sample air. Isokinetic intakes are needed, and loss in the tubing has to be minimized. Evaluation of how well these conditions are satisfied is an outstanding problem.

#### Ice nuclei, IN.

Measurements of IN have been attempted in the past with a large variety of instruments (e.g. Vali, 1985) but none have been proven to be fully acceptable. In general, one would like to measure the concentrations of ice nuclei of different modes of activity, and over a range of conditions for each mode. Clearly, that is a daunting task.

Some use is still being made of filter sampling. This method produces somewhat ambiguous results, but is relatively simple and therefore represents a compromise between making no measurements at all and attempting to use some more elaborate and still not fully satisfactory technique. Filter samples are taken during flights, over periods of ten minutes or longer. Subsequent exposure of the filters (in a laboratory chamber) to known supercooling and supersaturation yields a count of IN. The use of several filters in parallel can provide a spectrum.

Another approach to obtaining IN measurements is to take samples of air in the cloud inflow, or other region of interest, in some large container (metallized mylar bag, for example) and deliver the sample within as short a time as possible to a laboratory cloud chamber (e.g. Rogers and DeMott, 1995; DeMott et al., 1995). This approach can only yield a few data points per flight, requires a large effort, and has to be accompanied by careful studies of potential losses of IN in the container. The method inherits all the advantages and limitations of cloud chamber measurements of IN.

A major effort to develop an airborne IN counter is the Continuous Flow Diffusion chamber (CFD) of Rogers (1988, 1993, 1994). Two chilled cylinders with vertical axes form an annulus in which temperature and supersaturation can be independently controlled. Sample air enters at the top and is sheathed in clean air. Ice crystals initiated in the chamber are detected and sized at the outlet by optical scattering. This instrument has been operated in two flight programs so far, and is reported to have functioned well even at very cold ambient temperatures. It is to be hoped that the promising initial results will be confirmed by further use of the instrument.

### Tracers.

While analyses of cloud processes is most meaningful in Lagrangian reference frames, aircraft can provide only intermittent snapshot measurements within a parcel. Hence, the use of tracers is a natural complement to airborne observations.

Virtual tracers may be constructed by computing the positions of air parcels under the assumption of uniform and stationary winds. Such 'pointers' have proven to be quite useful in identifying how to intercept repeatedly a given air parcel.

Several direct tracer methods have been developed in various combinations of release from the ground and detection by an airborne instrument and vice versa, or with both release and detection done by the aircraft. The two most productive methods are the release of radar–reflective chaff from the aircraft and tracking of the chaff with a ground–based radar, and the use of sulfur hexaflouride (SF<sub>6</sub>). Chaff release requires a dispensing apparatus and a chute to the outside of the aircraft. Chaff has been successfully tracked for periods exceeding 30 min, documenting the transport and spreading of the chaff through the cloud volume (Martner et al., 1992). Spatial and temporal resolution depend on the radar characteristics and scanning sequence used. The SF<sub>6</sub> gas is released form pressurized tanks; detection is with a fast–response electron–capture device. The analyzer has a response time of near 1 s with a lag–time of a few seconds, so that spatial resolution of few hundreds of meters can be achieved. Stith and Benner (1987) demonstrated the potential of this technique and Stith et al. (1990) show interesting results on the transport of seeding material in cumuli.

### 4. Cloud structure – air motions

#### Navigation and winds.

This topic is mentioned here only for the sake of perspective without the detail it deserves. The fundamentals of the topic are well presented by Lenschow (1986).

Major developments in the measurements of position, velocity and attitude with respect to both air-relative and ground-relative coordinate systems, came over recent years from the implementation of the Global Positioning System and from the availability of laser-ring intertial navigation systems. Advances have also been important in the sensing of airflow with respect the aircraft.

The GPS made inexpensive and much improved position–keeping a reality. This alone is a major advance in the location of aircraft observations with respect to satellite images, ground–based radars or other observing systems. Accuracies of the order of 100 m are routinely possible wherever the coverage of GPS

signals is good. Further improvements in accuracy are expected from differential–GPS methods (DGPS), but this is much more severely limited in coverage than the basic GPS reception. The DGPS approach uses a stationary GPS receiver near the location of the desired measurement to correct for propagation errors, computer inaccuracies, and intentional degradation of GPS signal by the military for security reasons. The reference signals are superimposed on broadcasts from some FM stations; this makes DGPS available inexpensively, but in limited areas. Highly accurate, portable DGPS systems are available at greater expense.

The accuracy of the DGPS approach is so high that it also makes possible determination of an aircraft's pitch and roll angles. By mounting GPS receivers on the wingtips and on the nose (for example), the attitude of the aircraft can be derived, making the use of INS unnecessary.

The laser–ring INS technology also represents an advance toward affordability. The initial investment, and, even more significantly, the maintenance costs are less than for mechanical gyroscope systems. From the user's point of view the laser INS provide the same information (accelerations and their integrals) as the mechanical units. The fundamental limitations (oscillations, drifts) are also nearly the same.

The GPS and INS systems are complementary in a number of ways, so that combining data from the two systems offers many advantages. For positioning purposes, the INS is independent of external signals and so more reliable, but it is subject to oscillations and cumulative errors with flight duration. An approach for optimal use of the combined data set is given by Matejka and Lewis (1997). For velocity determinations and flux measurements, the GPS provides better long–wavelength data while the INS is best for high–frequency data. A method for taking advantage of this complementarity by use of a Kalman filter is described by Leach and MacPherson (1991). If no GPS is available for correcting the INS data, the method of Rodi et al. (1991) in which corrections are based on the DME navigation signals, may be helpful.

Measurements of the airflow with respect to the aircraft are obtained from vane systems or from differential pressure probes. For the latter, either a Rosemount 858 probe, or an array of five pressure holes in the radome on the nose of the aircraft are used (Brown et al., 1983). The principle is the same in both cases – sensing the pressure differentials that develop due to asymmetric flow across the surface of the probe; these pressures change as a function of angle between the aircraft axis and the air flow. A thorough examination of this angle–dependence has led Rodi (1997) to revise the relevant constant previously used from 2.2 to 1.7; this correction means that previous calculations of the attack angle and the side slip were 25% too low and that vertical velocity estimates were in error by about 1 m s<sup>-1</sup>. The correction most seriously affects the high–frequency components of air motion.

### 5. Cloud structure – airborne radars

Meteorological research radars have been carried on aircraft for many years. Most are operating at X or C–band frequencies and were essentially intended to provide the same type of information as ground–based weather radars, but with the mobility of the airborne platforms. Best known examples of such radars are the NOAA WP–3 hurricane tracking radars and the NCAR Eldora radar (Jorgensen et al., 1994; Hildebrand et al., 1994). These radars are usually operated while flying alongside a storm system with the observation range extending to 100 km or more. Another installation, the EDOP radar on the NASA ER–2 aircraft (Heymsfield et al., 1996a) is downward looking and observes storms from an altitude near 20 km; it obtains data from only one plane (curtain) but it is well suited to the measurement of vertical velocities. Both Eldora and EDOP have dual beams at 30–40° angles to each other so that two components of the wind are measured (assuming stationarity), and to permit the application of some scheme for attenuation correction.

The aforementioned radar systems have their main applications in the study of larger clouds (Cb, MCS, etc.) and may be referred to as storm radars. In contrast, the W–band radar on the Wyoming KingAir may be viewed as a cloud radar, since it is most useful for observations at short ranges (to 5 km) and it is usually operated while the aircraft is penetrating the clouds and is collecting in situ measurements. In that way, the radar data are directly referenced to the in situ observations. The Wyoming cloud radar (Pazmany et al., 1994) operates at 3 mm wavelength; that wavelength is more seriously attenuated by vapor and by liquid water than the longer wavelengths used in most other radars, but it permits a 0.7° beamwidth to be achieved with an antenna that could be readily accommodated on the twin–engine aircraft. The antenna points horizontally but the beam can be deflected vertically upward by means of a reflector plate housed externally. Spatial resolution is defined by the 30–m pulse length. Reflectivity, velocity and polarization data are recorded. Results from this radar are given by Vali et al. (1995), Galloway et al. (1997) and Vali et al. (1997).

The airborne radars also provide measurements of air velocity in clouds and precipitation. Scanning radars such as Eldora derive the wind field from staggered measurements at two different view angles, similarly to dual–Doppler methods. The fixed–beam radars yield only velocities along the beam direction, however, by varying the view direction through aircraft maneuvers and by assuming horizontal homogeneity in the wind field, the VAD method can be applied to yield estimates of the all three ground relative components of the wind. The Airborne VAD (AVAD) method, principally applicable to stratiform clouds, is described by Leon and Vali (1997). The main advantage of the AVAD method is that is can describe the wind field from measurements taken from a relatively small volume (few km) in just a few minutes, so that the homogeneity assumption is not very restrictive. Also, the regions of observation can be selected with all the flexibility of the airborne platform.

A potentially important adjunct to the wind measurements inside clouds by radars is the Doppler lidar system (Schwiesow et al., 1989) that was flown on the NCAR King Air. The lidar data could be very useful for the examination of air motions at cloud boundaries.

Common to Doppler radars and lidars are the limitations that arise from inaccurate removal of the component of aircraft motion in the beam direction, and from errors in the 3–D positioning of the observed volume. These errors are directly related to the positioning of the aircraft with respect to the ground, and to determinations of it's attitude. Computational inaccuracies also become a factor if the aircraft motion is not rectilinear and the sampled volume has to be re–gridded in some fashion.

# 6. Some conclusions (opinions)

- Greater detail and refinement of cloud observations are indispensable for the understanding of cloud processes. Such understanding is, in turn, essential to the ability of forecast and climate models to represent the impacts of clouds. Detail can be superfluous — the n-th decimal place syndrome. The criteria to avoid irrelevance are acuity and accuracy. Acuity is guided by intuition, honed by model calculations, and aided by serendipity. Accuracy is gained by patient work, and by invention.
- 2. The development of instruments for in situ measurements of hydrometeors has been very slow in the last decade or so. Capabilities in that area have fallen behind remotely sensed observations.
- 3. There is a specially pressing need for improvements in the measurement of hydrometeors in the size range between 30 and 300 μm, the region that brackets the gap between the useful ranges of FSSP and 2D–C probes. Size measurement is the primary concern, but phase and crystal shape information are also needed. It is important to establish whether the significantly steeper slope observed in many size distributions over this size range is an artifact, or is real. This problem strongly manifests itself in studies of stratus and cirrus.
- 4. Promising results for LWC measurements (PVM, CDS, Nevzorov–probe, CVI + hygrometer) should be followed by further evaluations, until a *de facto* standard, or standards, for the next several years emerge. Reliable LWC measurements have been long awaited, and that goal appears to be now within reach. Establishing confidence in LWC data will be a significant step.
- Positioning of observing aircraft in 3–D to roughly 10 m accuracy is becoming feasible. Horizontal and vertical winds are reliable to the order of [cm s<sup>-1</sup>] for wavelengths of tens of meters, and to order of [m s<sup>-1</sup>] for kilometer scales.

- 6. Temperature measurements in clouds with high LWC (>2 g m<sup>-3</sup>) and at the boundaries of such clouds can not be made reliably with current sensors so such data need careful use.
- 7. Humidity measurements fall short of resolving the variations in and around clouds that are one of the key factors in understanding cloud initiation (e.g. cirrus) and entrainment (stratus, cumulus).
- 8. Ice nucleus measurements could significantly aid in the understanding of cloud evolution, first of all in cirrus, wave clouds and nimbostratus, but also in other cloud types; however, much more effort will be required to develop credible instrumentation. Cloud condensation nucleus counters are providing the necessary data, and are to be trusted; the small number and cost of available units is a hindrance.
- 9. Great opportunities exist for advances in observations of cloud composition and structure through the combined use of in situ measurements and of passive and active radiation measurements at optical, infrared and microwave wavelengths. Instrumentation for many of these measurements of the nearby cloud environment can be carried on aircraft making the in situ measurements. Since many of these instruments have been developed to simulate and validate satellite observations, linking observations obtained with them to in situ observations is a step toward linking satellite and in situ observations, and toward the development of improved parameterizations of cloud processes and of structures in large–scale models.

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