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Cloud phase discrimination by near-infrared remote sensing

Pilewskie, Peter Andrew, Ph.D.

The University of Arizona, 1989
CLOUD PHASE DISCRIMINATION
BY NEAR-INFRARED REMOTE SENSING

by
Peter Pilewskie

A Dissertation Submitted to the Faculty of the
DEPARTMENT OF ATMOSPHERIC SCIENCES
In Partial Fulfillment of the Requirements
For the Degree of
DOCTOR OF PHILOSOPHY
In the Graduate College
THE UNIVERSITY OF ARIZONA

1989
As members of the Final Examination Committee, we certify that we have read
the dissertation prepared by Peter Pilewskie
entitled Cloud Phase Discrimination by Near-Infrared Remote Sensing

and recommend that it be accepted as fulfilling the dissertation requirement
for the Degree of Doctor of Philosophy.

Sean A. Twomey  
Benjamin M. Herman  
Kenneth C. Young  
Donald R. Huffman  
John O. Kessler

Final approval and acceptance of this dissertation is contingent upon the
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A ground-based near-infrared spectroradiometer was built and used to measure relative spectral reflectance from cumulus congestus and cumulonimbus clouds during the 1985 and 1986 Arizona summer monsoon seasons. Thermodynamic phase was inferred from spectral features in the regions between 1.55-1.75\(\mu m\) and 2.1-2.3\(\mu m\) where there are distinct differences between absorption in liquid water and ice and absorption by water vapor is very weak.

Although liquid water and ice are nearly transparent in the visible, they absorb weakly in the near-infrared and that absorption is amplified by multiple scattering in clouds. Reflectance measurements are simple to make, requiring neither high spectral resolution nor absolute detector response. Three distinct aspects of differences between absorption in liquid water and ice were used to infer phase: a) Ratio of the signal at 1.65 \(\mu m\) to that at 2.2 \(\mu m\); b) Wavelength of peak signal in the 1.65 \(\mu m\) water vapor transmission window; c) Half-bandwidth of the 2.1-2.3\(\mu m\) feature.

Representative spectra are presented and analyzed on the basis of the predicted behavior of liquid water and ice
cloud absorption. The results are consistent with young cumuli rapidly glaciating as they reach cooler levels, well before evidence of anvil formation or fibrous structure, contrary to the notion that phase can be inferred from visible cloud features.
CHAPTER 1
INTRODUCTION

The thermodynamic phase of water in terrestrial clouds plays an important role in many of the physical and dynamical processes of the atmosphere. Thunderstorm evolution, from the rise of individual thermals into organized cells, to a mature cumulonimbus marked by lightning and precipitation, and finally dissipation, is closely coupled with the phase of the particles which make up the cloud. For example, freezing 1 gram of liquid water in a cubic meter of cloud at -10°C would increase the temperature of that volume by about .5°C through the release of latent heat; for a 2 km diameter turret, this represents an increase in ascent rate of a few meters per second (typical thunderstorm ascent rates are 5 ms\(^{-1}\) to 10 ms\(^{-1}\); see Ludlam, 1980). Such growth surges are commonly observed in developing cumuli.

Ice is important in cloud electrification and precipitation processes. Interaction among ice particles is considered the primary mechanism for microscale charge separation in thunderstorms (Illingworth, 1985), a contention strongly supported by observations that lightning is
confined almost completely to clouds which extend to levels where the temperature is well below freezing. For most cold clouds (a "cold cloud" here meaning at least part of it is colder than 0°C), the initial step in the processes leading to precipitation is the growth of ice crystals at the expense of liquid water droplets (Bergeron, 1935); that growth is occasioned by the fact that the saturation vapor pressure over ice is lower than that over water at the same temperature.

Although the significance of ice nucleation in cloud development is understood, the manner in which ice particles are created in clouds is not. Field studies, for example those by Koenig (1963) and Mossop (1968), have indicated that ice particle concentrations often exceed the concentrations of ice nuclei, sometimes by several orders of magnitude. Laboratory investigations (e.g., Hallet and Mossop, 1974) into the fragmentation or ejection of splinters by a freezing droplet as a possible mechanism for nucleating many more droplets, were found to be inconclusive. It has been suggested by Hobbs (1974) and Young (1978), among others, that the methods for measuring ice nuclei concentrations do not always account for all the modes of nucleation acting in a particular cloud. Nevertheless, until more accurate measurement procedures are developed, the discrepancy between ice particle
concentration in clouds and the corresponding number concentration of ice nuclei remains to be reconciled.

Certainly the ability to determine the phase of water in clouds would assist studies in the previously mentioned areas. It would also have a practical application for aviators concerned with the freezing of super-cooled liquid water onto aircraft, known as icing. Results from aircraft observations of the frequency of super-cooled liquid water clouds versus ice clouds, published in Pruppacher and Klett (1978), show that at \(-10^\circ C\) a cloud is as likely to be composed of supercooled water as it is to be glaciated; this is about the temperature at which aircraft are most likely to experience icing. Pilots and navigators would assume lesser risks having knowledge of cloud phase before flight.

Cloud phase information would also benefit the study and application of precipitation enhancement. Methods of inducing precipitation by seeding super-cooled liquid water with ice-nuclei have been practiced for several decades, but, as recent studies suggest (see, for example, Reynolds, 1988), further research is needed to determine if cloud seeding actually increases precipitation. The seeding of clouds which have already glaciated would be ineffective towards producing precipitation-size particles; a-priori knowledge of cloud phase could assist new investigations in
weather modification and operational methods for cloud seeding programs.

In-situ cloud phase measurements are difficult, costly, and sometimes impractical, hence the need for remote measurements (either from the ground or from aircraft and satellites). This dissertation describes a technique for inferring cloud phase from near-infrared reflectance measurements which could be made at the ground. A passive radiometer was built and used to measure the relative spectrum of reflected radiation from air-mass convective clouds near Tucson during the Arizona summer monsoon season. Phase was determined from spectral features in near-infrared water vapor window regions, where distinct differences exist between the absorption spectra for liquid water and ice. Similar principles were employed by Pollack et al. (1978) to infer the composition of clouds on Venus. Also, near-infrared reflectance measurements to infer terrestrial cloud properties other than phase, such as average droplet radius, optical depth, and liquid water path include the work by Twomey and Cocks (1982), Doherty and Houghton (1984), Stephens and Platt (1987), Foot (1988, a and b), and Paltridge (1988).

Transmission measurements would be similarly effective in cloud phase discrimination provided that intensity levels exceed detector sensitivity. In fact, Gates and Shaw (1960)
were among the first to suggest that ice clouds could be distinguished from water clouds based upon measurements of cloud transmission at the earth's surface. They found what appeared to be anomalously high transmission in a narrow spectral interval centered at 1.82 μm from a cloud they assumed had glaciated and conjectured that the effect was due to weaker absorption by ice compared to liquid water.

Blau et al. (1966) recognized that the ability to distinguish cloud phase from near-infrared spectral measurements would be a potentially powerful meteorological tool. They conducted a series of aircraft measurements of cloud reflectance and confirmed their results with measurements from simulated laboratory clouds of known phase. Differences in the location and magnitude of reflectance minima (absorption maxima) between water and ice spectra obtained in the field agreed with the laboratory studies.

A more recent study by Curren and Wu (1982) used three near-infrared channels from a radiometer on board Skylab to infer the phase of an orographic cloud over New Mexico; a thermal infrared channel at 11.4 μm simultaneously provided an estimate of cloud-top temperature. These authors interpreted their results as indicating super-cooled water at temperatures as low as -47°C, rather unlikely since cloud physicists consider that homogeneous freezing (ie., not
needing a nucleating particle) occurs at about -40°C (Pruppacher and Klett, 1978). Disagreement between measured optical properties of clouds and theory has been common. For example, disparities between droplet size inferred from optical measurements and conventional in-situ measurements have been reported by Rozenberg et al. (1974), Twomey and Cocks (1982), and Foot (1988a), among others. Further examination of these types of problems is necessary before cloud parameters can be faithfully extracted from remote measurements.

Energy considerations do not limit near-infrared cloud reflectance measurements; between 1.5 μm and 1.6 μm, for example, the solar energy flux at the top of the atmosphere is 22.65 Wm⁻², roughly 12% of that between .5 μm and .6 μm (Nikol'skii, 1973), and the albedo of a moderately thick cloud (1 km) in the same spectral region is more than .5. Previous aircraft and satellite studies, however, have often indicated disparity between cloud reflectance measurements and theory. Both broad- and narrow-band measurements of cloud absorption have consistently exceeded theoretical estimates (reviewed by Stephens and Tsay, 1988). This discrepancy has important climatological consequences in the global energy budget. Imbalances between short-wave heating and long-wave cooling in clouds represents an important energy source that drives the atmospheric circulation.
Clouds are also considered to be a strong feedback mechanism in the greenhouse warming of the atmosphere (see Somerville and Iacobelli, 1987). It is clear that more experimental investigations are needed to provide further insight into the interaction of clouds with radiation.

Other methods to distinguish cloud phase have been devised and employed but they usually incorporated active sensors, limiting their potential for use on a global scale (such as on satellites). Harris (1971) suggested the use of polarized laser beam scattering (in the thermal infrared) and showed that polarization ratio of perpendicular to parallel scattered components as a function of scattering angle were strongly phase dependent. Spinhirne et al. (1983) used the depolarization ratio of a returned lidar signal, which is greater in scattering from ice clouds than water clouds, to infer phase. As for near-infrared measurements, the author is unaware of any ground-based cloud reflectance measurements with the specific purpose of determining phase, other than a Soviet patent (Twomey, 1985) describing a potential method.

The present study pertains to the discrimination of ice from liquid water in clouds by ground-based optical remote sensing. In chapter 2, some approximate theoretical methods for computing near-infrared absorption in individual particles as well as thick clouds are presented, along with
a more detailed method for computing cloud reflectance. Water vapor absorption in the near-infrared, and its effect on cloud reflectance spectra, is presented in chapter 3. The instrument, a spectroradiometer, which was built and used to measure cloud reflectance, is described in chapter 4. We will focus attention on representative cloud spectra measured during the summers of 1985 and 1986. These results are presented in chapter 5, along with the procedures used to acquire a collection of spectra from a single cloud. Chapter 6 discusses the limitations of near-infrared reflectance measurements, especially with regard to mixed phase clouds and varying optical depth. Further applications, such as measurements from satellites and ground-based transmission measurements, are suggested.
CHAPTER 2
LIQUID WATER AND ICE ABSORPTION

Most clouds in the lower and middle troposphere are optically thick to the extent that their pattern of scattered radiation differs substantially from the scattering pattern from a single water droplet or ice crystal. Most of the photons emerging from a cloud have been scattered several times, having no memory of the angular dependence or the particle shape dependence of scattering from a single particle. Indeed, single-scattering optical phenomena, such as haloes and coronae, are not visible from thick clouds. Angular and wavelength-dependent features of single scattering are smeared out, and the retrieval of cloud phase from reflectance measurements would be impossible if not for the phase dependence of absorption.

The goal of this chapter is to determine the influence of particle phase and size on cloud reflectance spectra. Approximate methods for calculating single particle absorption and absorption from multiple scattering layers are presented, followed by a more detailed procedure for computing the reflectance spectra from clouds. This treatment considers scattering from plane parallel layers.
only; not only does this simplify the computational problem, but the experimental geometry was such that a plane-parallel assumption is sufficient. The final portion of this chapter discusses the validity of this assumption.

2.1 Single Scattering: Geometric Optics

Liquid water and ice are nearly transparent in the visible but their bulk absorptions increase by about six orders of magnitude in the near-infrared as shown in figure 2.1, where volume absorption coefficient $k_a$ is plotted against wavelength. The data for the curves were derived from studies by Hale and Querry (1973), Irvine and Pollack (1968), and Warren (1984). Typical cloud droplet radii ($r$) are appreciably greater than near-infrared wavelengths and cloud droplets are weak absorbers - in the sense that $k_ar << 1$ - so that the following geometric optics/weak absorption approximation (Bohren and Barkstrom, 1974, and Bohren and Huffman, 1983) can be used to compute absorption by a single droplet:

$$1 - \bar{\omega}_o = .84k_ar$$  \hspace{1cm} (2.1)

where $\bar{\omega}_o$ is the single scattering albedo or the ratio of scattering to total extinction (scattering + absorption), and $k_a$, the volume absorption coefficient, is equal to $4\pi n''/\lambda$, $n''$ being the imaginary part of the refractive index.
Figure 2.1. Bulk absorption coefficient for liquid water (solid curve) and ice (dashed curve).
The constant of proportionality, .84, is actually a function of the real part of the refractive index (taken to be 1.33 here) and particle shape (spherical in this case).

Ice particles in clouds may not be spherical but the proportionality $1-\bar{\omega}_o \propto k_a L$, where $L$ is now some characteristic dimension of the particle, still holds (Twomey, 1983). The reason is that in the geometric optics limit ($r>>\lambda$) the probability distribution of photon path lengths within a particle scale with the dimensions of the particle. That is, if $p(x)$ is the probability of a photon path-length (before eventual emergence from a particle) lying between $x$ and $x + dx$ for a particle of linear dimension $L$, then $p$ is also the probability of photon path-lengths between $2x$ and $2(x + dx)$ for a particle of linear dimension $2L$. Therefore, $p(x/L)$ is invariant with size; when absorption is included, the probability of emergence (or single scattering albedo) is given by

$$\bar{\omega}_o = \int_0^\infty e^{-k_a x} p(x/L) d(x/L) = \int_0^\infty e^{-k_a L u} p(u) du; \quad (2.2)$$

again, $k_a$ is the volume absorption coefficient. In the region of interest, $k_a L \ll 1$ (weak absorption) so that

$$\bar{\omega}_o \approx 1 - k_a L \int_0^\infty u p(u) du, \quad (2.3)$$

and again we have $1-\bar{\omega}_o \propto k_a L$, regardless of particle shape. The above expression, though quite simple, is crucial in
retrieving phase information. Uncertainties in particle shape would certainly reduce the accuracy to which particle size could be inferred since absolute absorption would be needed, but would not change the relative absorption spectrum which is determined by \( k_a \), a function of wavelength and phase.

2.2 Multiple Scattering

Most clouds are optically thick in the near-infrared so that photons, on average, undergo multiple interactions before eventually emerging. For this reason, the probability of their absorption is amplified. For an infinitely deep layer of isotropic scatterers,

\[
a \propto (1-\omega_0)^{1/2},
\]

where \( a \) is fractional absorption (Twomey and Bohren, 1980). The proportionality factor is the \( H \) function (defined in Chandrasekhar, 1950, p. 77 and 97) which is a function of the incident illumination zenith angle and single scattering albedo, but the latter dependence is quite weak when the quantity \( (1-\omega_0) \) is small. Indeed, that is the case for average size cloud droplets in the near-infrared, but the scattering diagram or phase function is far from isotropic. Typically, the average cosine of the phase function (often referred to as the asymmetry parameter, \( g \)) is about .86 for
cloud droplets in the near-infrared, indicating a strong peak in the forward direction.

The anisotropy of scattering by cloud droplets need not limit application of the proportionality (2.4) because the important parameters in multiple scattering, $\tau$, $\bar{\omega}_0$, and $g$, are, to some extent, redundant. It has been shown by van de Hulst and Grossman (1968) that the scattering properties of a cloud with optical depth $\tau$, single scattering albedo $\bar{\omega}_0$, and asymmetry $g$ are similar to those of a cloud with asymmetry $g'$ if the single scattering albedo and optical depth are concurrently scaled to $\bar{\omega}_0'$ and $\tau'$ respectively. The similarity transformation is achieved by weighing the scattering part of the optical depth, $\tau_s = \bar{\omega}_0 \tau$, by the average deviation from forward scattering, $(1-g)$, while leaving the absorbing part of the optical depth, $\tau_a = (1-\bar{\omega}_o) \tau$, unchanged. Therefore, the scattering by two different clouds is similar if

$$\bar{\omega}_0' \tau'(1-g') = \bar{\omega}_0 \tau (1-g) \quad \text{and}$$

$$\tau' = (1-\bar{\omega}_0).$$

If one wishes to scale from a non-isotropic problem to an isotropic problem, then $g' = 0$ and so

$$\bar{\omega}_0' = \bar{\omega}_0 (1-g)/(1-g \bar{\omega}_0) \quad \text{and}$$

$$\tau' = (1-g \bar{\omega}_0).$$

The validity of this similarity transformation is outlined in van de Hulst (1980).
Fractional absorption, $a$, or the reduction (from a conservatively scattering medium) in reflectance is then proportional to $((1-\omega_o)/(1-g\omega_o))^{1/2}$; using equation (2.1), letting $g=.86$, we get

$$a \propto \left[ \frac{-\mu_k \tau}{1-.86(1-.84k_r \tau)} \right]^{1/2}$$

(2.5)

The quantity $(1-\omega_o)^{1/2}$ is plotted versus wavelength in figure 2.2 for infinitely thick clouds of 5 $\mu$m water (solid curve) and ice (dashed curve) spheres. It is apparent that not only does absorption become appreciable but that there are several regions where absorption by ice and liquid water are substantially different, particularly the spectral regions around 1.6 $\mu$m and 2.2 $\mu$m. This can also be seen in the bulk spectral absorption shown in figure 2.1.

2.3 Reflection from Deep Layers: Diffusion Approximation

A "reflection function" relates emerging reflected radiation to the incident illumination. A function $S(\tau;\mu,\phi;\mu_o,\phi_o)$ was defined by Chandrasekhar (1950, p.20) as follows:

$$I(0,\mu,\phi) = (F/4\mu)S(\tau;\mu,\phi;\mu_o,\phi_o),$$

where $I$ is the reflected intensity at $\tau=0$ in the direction $\mu$ (cosine of the zenith angle) and $\phi$ (azimuth angle), and $\pi F$ is the flux in the direction $\mu_o, \phi_o$, incident upon a layer
Figure 2.2. $(1-\omega_0)/\omega_0$ versus wavelength for a deep cloud layer composed of 5 $\mu$m liquid water (solid curve) or ice (dashed curve) spheres. This ordinate is approximately proportional to fractional absorption in the layer.
of optical thickness $\tau$. Similarly, the transmission function is defined by:

$$I(\tau, \mu, \phi) = (F/4\mu)T(\tau; \mu, \phi; \mu_0, \phi_0).$$

The factor $1/\mu$ in both expressions insures symmetry in $S$ and $T$, required by the Helmholtz principle of reciprocity:

$$S(\tau; \mu, \phi; \mu_0, \phi_0) = S(\tau; \mu_0, \phi_0; \mu, \phi)$$
$$T(\tau; \mu, \phi; \mu_0, \phi_0) = T(\tau; \mu_0, \phi_0; \mu, \phi).$$

That is, $S$ and $T$ remain the same if the directions of incidence and emergence are interchanged. Reciprocity and the principle of invariance, which asserts that radiation emergent from a semi-infinite layer is invariant upon the addition or subtraction of arbitrarily thick layers, provide the fundamental basis from which solutions to multiple scattering radiative transfer solutions are derived. Thus, along with energy conservation, they also provide the most powerful computational tests of accuracy.

There are a number of methods for computing reflectance from multiply scattering media, for example, the Monte Carlo method, the doubling method, the method of successive orders, etc., and each has merit depending upon the selected problem. Van de Hulst (1980, vol. I, ch. 4) has reviewed several of the most commonly used techniques and the criteria for which they are best suited; King and Harshvardhan (1984) have tested several approximate methods for varying optical depths and single scattering.
albedos in water clouds. The method preferred in this study, the diffusion approximation, has been reviewed by van de Hulst (1980); it was chosen because if its suitability for thick layers and the relative ease and speed with which spectrally varying quantities can be computed.

The underlying premise is that deep within a layer, far from boundaries or sources, radiation is transported by simple diffusion. Expressions for the radiation emerging from a cloud are derived from properties in the diffusion domain, where intensity is defined by

$$I(\tau,\mu) = s_1P(\mu)e^{\kappa\tau} + s_2P(-\mu)e^{\kappa\tau}.$$  

Here P is the diffusion pattern, s₁ and s₂ define the strengths of the upward and downward diffusion streams, and \(\kappa\) is the diffusion exponent determining the attenuation of the diffusion stream. A diffusion domain exists at levels further than about \(\tau = 2/(1-g)\) from cloud boundaries; for \(g=0.86\) and \(Q_o=0.99\), a diffusion domain exists at levels deeper than about \(\tau=7\) which is a geometric depth of a few hundred meters for a cloud with liquid water content of 0.2g m\(^{-3}\) and average drop radius 5 \(\mu\)m.

In the following notation, all capitalized quantities are matrices, underscored quantities are vectors, and all others are constants. Asterisks indicate transposition.

For a finite layer (in which exists a diffusion domain) of optical depth \(\tau\), reflectance is defined by:
The following definitions apply:

a) If \( \mathbf{u} \) is a downward directed intensity vector whose \( i \)th component, \( u_i \), is the intensity in the direction \( \mu_i, \phi_i \) (\( \mu \) is the cosine of the zenith angle, \( \phi \) the azimuth angle), then \( S\mathbf{u} \) is the intensity vector scattered back in upward directions. To allow for finer subdivision in angle we consider only azimuthally-averaged quantities so that \( S \) is only a function of \( \mu_0 \) (incident) and \( \mu \) (emergent). Solutions for the azimuthally-dependent intensity terms in the diffusion domain exist and are of the form

\[
I(\tau, \mu, \phi) = e^{\kappa \tau} F^0(\mu) \cos(m\phi).
\]

However, by definition, the diffusion approximation is only valid for relatively thick layers which also implies strong damping of the azimuth dependent solutions (van de Hulst, vol II, ch.15). For such layers, reflected intensity varies very little with azimuth, provided that the incident illumination is not close to glancing. \( S \) in this notation is not identical to Chandrasekhar's reflection function and it does not obey reciprocity. However, \( (MS) = (MS)^* \) if \( M \) is a diagonal matrix of elements \( \mu_{ii} \); \( MS \) then, is the azimuthally averaged discrete equivalent of Chandrasekhar's reflection

\[
S\mathbf{u} = S_0\mathbf{u} - \frac{\beta e^{-\kappa z^2}}{1 - \beta^2 e^{-\kappa z^2}} k_0 k_i^2 \mathbf{u}. \tag{2.6}
\]
function. For the computations in this work, 10 incident and 10 emergent directions were used.

b) $S_{e}$ is the scattering matrix for a semi-infinite layer. It is solved iteratively from the following set of equations:

$$(1 + \mu_i / \mu_j) S_{ij} = \omega_0 h_i h_j$$

and

$$h_i = 1 + \Sigma_j S_{ji}.$$  

$\omega_0$ is the scaled single scattering albedo, as outlined previously; indeed, this is just the discrete form of Chandrasekhar's solution for isotropic scattering by use of the $H$ functions. Therefore, $h$ is the discrete form of the $H$ function. $f$ is a normalization factor to insure energy conservation.

c) $\kappa$, as mentioned previously, determines the attenuation of the diffusion stream. It can be solved iteratively from a characteristic equation in $\omega_0$ and $\mu$ when single scattering is isotropic (or scaled to isotropic).

d) $k_i$ is called the injection vector and it defines the strength of the diffusion stream in the diffusion domain, for unit incident illumination in the direction $\mu_i$.

e) $k_e$ is called the escape vector; it defines the angular pattern of the radiation emerging from the medium, by
definition outside the diffusion domain, when a unit diffusion stream is flowing inside, towards the boundary.

f) \(-\beta\) \((\beta > 0)\) is a negative internal reflection coefficient. At the upper boundary of a deep layer an upward propagating diffusion stream is internally reflected. Near the boundary, there is a reduced downward-propagating intensity field and this is equivalent to a downward diffusion stream with strength \(-\beta\).

Figure 2.3 graphically depicts the situation. When \(u\) is incident on a semi-infinite layer, \(S_u\) is reflected. Now consider a boundary at a depth of \(\tau\) such that between \(\tau\) and the upper surface there exists a diffusion domain. The presence of the lower boundary now allows photons, which previously would have been redirected in the back direction, to escape through the bottom. Therein lies the significance of the negative internal reflection. It serves to reduce the amount of radiation escaping through the upper boundary when there is a loss of intensity by transmission through the lower boundary. \(k_i u\) is "injected" downward; it is attenuated to \(e^{-kr}k_i u\); an amount \(e^{-kr}k_{e}k_i u\) escapes through the bottom, but \(-\beta e^{-kr}k_i u\) is reflected (negative); on reaching the upper surface (with strength \(-\beta e^{-kr}k_i u\)) a field \(-\beta e^{-kr}k_i u\) escapes out the top at \(\tau=0\). But part of the upward propagating stream is again reflected, so the process continues ad
Figure 2.3. Reflection from an infinitely thick layer (top) and a layer of thickness $\tau$ (bottom).
infinitum. The sum of "negative" emerging intensities is then:

\[-\beta e^{-\kappa_2 \tau} (1 + \beta^2 e^{-\kappa_2 \tau} + \beta^4 e^{-\kappa_2 \tau} + \beta^6 e^{-\kappa_2 \tau} + \ldots) k_k k_u^*,\]

which sums to

\[\frac{-\beta e^{-\kappa_2 \tau}}{1-\beta^2 e^{-\kappa_2 \tau}} k_k k_u^*.\]

This "negative" intensity field determines the degree to which the reflectance of a finite layer of optical depth \(\tau\) is reduced below that of an infinite layer, and this is expressed in equation 2.6. To remain consistent, \(\tau\) needs to be scaled also because \(\kappa\) and all resulting quantities resulted from a scaled single scattering albedo.

Since single scattering albedo and asymmetry parameter are needed for the reflectance calculations, a Mie scattering routine was used to compute the spectral dependence of those quantities for several different radii. A Gaussian distribution in cloud particle size was assumed, with a dispersion of .2, and quantities were integrated over 2.5 standard deviations on both sides of the mean radius. Twomey, et al. (1967), however, have shown that cloud reflectance is relatively insensitive to size distribution, instead depending mainly on the average radius.

Figures 2.4 to 2.6 show spectral reflectance \(S\) for both liquid water (solid curves) and ice (dashed curves) clouds for different average particle radius, \(\bar{r}\). Only the (10,10)
Figure 2.4. Reflectance, $S$, for water (solid curves) and ice clouds (dashed curves). $\mu_0 = \mu = .95$. $\tau = 5 \mu m$. 
Figure 2.5. Reflectance, $S$, for water (solid curves) and ice clouds (dashed curves). $\mu_o = \mu = .95$. $\bar{T} = 25 \mu m$. 
Figure 2.6. Reflectance, $S$, for water (solid curves) and ice clouds (dashed curves). $\mu_o = \mu = .95$. $\bar{x} = 50 \, \mu m$. 
element, corresponding to $\mu_0=.95$ and $\mu=.95$ (i.e., incident and viewing angles close to normal), is plotted. Reexamination of figure 2.2 reveals that the absorption features from the simple approximation agree well with the reflectance spectra shown here. Again, the features near 1.6 $\mu$m and 2.2 $\mu$m stand out. At 1.65 $\mu$m for example, the reflectance from a water cloud is approximately twice that from an ice cloud for each of the three sizes shown. Nevertheless, one can see also differences which are due to size only; remember that the ratio of reflectance from a cloud of radius $r_1$ to that from a cloud of radius $r_2$ is approximated by $(1-c\sqrt{k_a r_1})/(1-c\sqrt{k_a r_2})$. Therefore, a single narrow-band reflectance value alone would be insufficient to distinguish between a liquid water and ice cloud. Measurement of the continuous cloud spectrum would be needed to see the various shifts of peaks and troughs in going from liquid water to ice. The significance of the spectral features near 1.6 $\mu$m and 2.2 $\mu$m will become more apparent when water vapor absorption is considered in the next chapter.

The dependence of reflectance on optical depth reveals that beyond about 1.3 $\mu$m clouds of optical depth 64 or greater are virtually indistinguishable from a semi-infinite cloud. This allows one to define what is meant by a thick cloud. Van de Hulst (1964) specifies an optical depth of
\[ 2(1-\bar{\omega})^{-1/2} \] as effectively infinite. For both water and ice, smallest values of \((1-\bar{\omega})\) beyond 1.3 \(\mu\)m are around .001, implying that a cloud would reach its asymptotic limit for reflectance at an optical depth around 60, as shown in figures 2.4-2.6. This implies that clouds thicker than about 1 kilometer are essentially infinite in the near-infrared.

2.4 Plane-Parallel Assumption

The results from the preceding sections apply to plane-parallel, horizontally infinite clouds, whereas the reflectance spectra measured in this study were from clouds with finite lateral boundaries. Nevertheless, the situation differs minimally from horizontal plane-parallel layers illuminated from above if one considers the experimental geometry and the dimensions of the clouds observed.

Figure 2.7 illustrates typical shape and size of a cumulonimbus cloud. During the Arizona monsoon, cloud bases are generally near 3 km above sea level, and the most massive cells occasionally reach the tropopause at about 16 km. Observational studies of cumulus cloud growth (see, for example, Saunders, 1960, and Glass and Carlson, 1963) have indicated that the width of individual turrets tend to be about one-half to three-quarters of the turret height. The figure shows several turrets of variable size that join to form a larger cell of considerable horizontal extent, a
Figure 2.7. Shape and dimensions of typical cumulonimbus cloud, showing reference normal for incident and viewing directions.
common occurrence in real cumulus clouds. Also shown is the
direction of solar illumination, \( \mu_o \), and the direction of
measured emergent radiation, \( \mu \), both referenced to the
vertical plane normal. Reflectance measurements were usually
made near normal incidence (5°-15° with respect to the
vertical sides of clouds), and with low solar elevation
angles (15°-30°) so that the geometry was similar to more
commonly made nadir measurements above a horizontal cloud
layer.

Scattering from finite clouds differs from horizontally
infinite layers because photons can escape from the sides of
the former. This could be interpreted as absorption by
someone measuring absolute cloud reflectance, unable to
account for the escaping photons. Since we measured relative
cloud reflectance, there was no attempt to extract absolute
absorption, and so, to an extent, this problem was avoided.
Furthermore, uncertainties arising from the unknown
scattering properties of finite clouds are minimized when
considering the number of photons which actually escape from
the sides of clouds the size we have studied.

Twomey (1985) has computed the photon optical path-
length distribution via a non-linear inverse method for a
horizontally infinite cloud with a vertical optical depth of
128 (\( \approx 2 \) km). The average optical path travelled by reflected
photons was 260, the median was 20, and 90% had travelled
less than 370. The distribution of geometric displacement from point of origin to point of emergence can be found using random-walk statistics; we will consider those photons travelling optical paths less than 370 to determine a limit on the geometric displacement suffered by a majority of photons.

To simulate a simple one-dimensional random walk with equal probability of displacement in the forward or backward directions, the optical path-length is scaled by \((1-g)\), giving the total number of steps with average step length in unscaled optical units of \((3^{1/2}(1-g))^{-1}\). (The factor \(3^{1/2}\) arises because we are looking at the projection of three-dimensional random walk on a one-dimensional axis.) For \(g=.85\), the number of steps \(N\) in our limiting case is \(370 \times .15 \approx 50\) and the average step length is approximately 4 (optical units), or 60 m.

A Monte Carlo calculation (2000 photons) gave the distribution of maximum displacement (not the final displacement since photons exiting through a cloud boundary have no way of getting back in) from the origin after 50 steps. The cumulative probability, \(P\), shown in figure 2.8, indicates the fraction of reflected photons as a function of optical and geometric displacement from point of origin. The distribution is over both forward and backward directions. Only 1% of the photons reach a distance of 2 km (1% also
Figure 2.8. Cumulative probability of maximum displacement (before emergence) of photons reflected from a cloud of optical depth 128. Distance scale is shown in scaled optical units and geometric units.
reach ~2 km); keep in mind that this is a limiting case and that most of the photons emerge after travelling much shorter distances. Since the dimensions of the types of clouds we measured were on the order of 10km x 10km, and since we have just shown that 90% of the photons reflected from a cloud of optical thickness 128 travel less than 2 km, the 10km x 10km clouds are essentially infinite in optical extent and may be considered plane-parallel.
CHAPTER 3

WATER VAPOR ABSORPTION

The previous chapter considered absorption by liquid water and ice clouds only. However, as solar near-infrared radiation travels from the top of the atmosphere to the cloud and then (after several scatterings) on to the observer at the ground, it is attenuated to some degree. Molecular absorption is the primary component of near-infrared atmospheric extinction, and of several absorbing gases, water vapor is by far the most influential, the absorption bands of O₂ and CO₂ being relatively weak in the near-infrared.

The emphasis of this chapter is on absorption by water vapor, specifically, the extent of its influence on near-infrared cloud reflectance spectra measured at the ground. We will show that between the bands of water vapor absorption the atmosphere is effectively transparent. On the other hand, liquid water and ice absorption are still appreciable, yet differ substantially from one another, allowing the measurement of cloud reflectance to distinguish between phases. (The chapter will conclude with a brief discussion of the other components of extinction.)
3.1 Influence of Water Vapor Absorption on Cloud Reflectance Spectra

There are six near-infrared water vapor absorption bands, commonly designated the ρ, σ, τ, Φ, ψ, and Ω bands. Each is actually a group of several overlapping vibrational-rotational bands, listed in table 3.1 (from Goody, 1964). The wavelengths of liquid water and ice absorption maxima are also shown in table 3.1 for comparison. Note that while liquid water absorption maxima are near vapor absorption bands, those of ice tend to lie in the intervening regions. In the region of spectral sensitivity of our instrument (chapter 4) there are also two groups of relatively weaker visible bands centered at .72 μm and .82 μm respectively.

Figure 3.1 shows water vapor transmittance (from McClatchey, et al., 1971) through a vertical path in an atmosphere with 1 g cm⁻² precipitable water. (Precipitable water is defined as vapor density integrated over the depth of the atmosphere; during the Arizona monsoon season it ranges from about 2 to 3 g cm⁻¹.) The three widest "window" regions, where absorption is slight, are shaded in figure 3.1. Once again referring to figures 2.1 and 2.2, bulk absorption and fractional absorption by deep clouds, respectively, note that the windows centered at 1.65 μm and 2.2 μm correspond to regions where liquid water and ice
Table 3.1. Near infrared water vapor absorption bands. Wavelengths (μm) of absorption maxima in liquid water and ice are listed for comparison.

<table>
<thead>
<tr>
<th>Name</th>
<th>No. of Bands</th>
<th>Band Center (μm)</th>
<th>Liquid</th>
<th>Ice</th>
</tr>
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<tr>
<td>ρ</td>
<td>2</td>
<td>.906</td>
<td></td>
<td>.920</td>
</tr>
<tr>
<td>σ</td>
<td>2</td>
<td>.942</td>
<td></td>
<td>.943</td>
</tr>
<tr>
<td>τ</td>
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<td>.97</td>
<td>.972</td>
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<td>1.03</td>
</tr>
<tr>
<td></td>
<td></td>
<td>1.114</td>
<td></td>
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<td></td>
<td></td>
<td>1.141</td>
<td></td>
<td></td>
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<tr>
<td></td>
<td></td>
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<td></td>
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</tr>
<tr>
<td></td>
<td></td>
<td>1.209</td>
<td></td>
<td></td>
</tr>
<tr>
<td>ψ</td>
<td>5</td>
<td>1.343</td>
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<td></td>
<td></td>
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<td>1.389</td>
<td></td>
<td></td>
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<td></td>
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<td>1.455</td>
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</tr>
<tr>
<td></td>
<td></td>
<td>1.476</td>
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<tr>
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</tr>
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<td></td>
<td></td>
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<tr>
<td></td>
<td></td>
<td></td>
<td>1.94</td>
<td>2.00</td>
</tr>
</tbody>
</table>
Figure 3.1. Water vapor (1 g cm\(^{-2}\)) transmittance in the near-infrared. Widow regions are shaded.
absorption are considerable and furthermore, differ substantially.

Cloud reflectance measurements were usually made at viewing angles near 15° above the horizon with low solar elevation angles. Thus, the actual path traversed normally represented a larger water vapor amount than that used for figure 3.1, due to the usual distance to the clouds. To model the problem more realistically, the LOWTRAN 6 computer code (Kneizys, et al., 1983) was used to compute transmittance over paths from sun to cloud, $T_{s-c}$, and cloud to ground, $T_{c-g}$, for a cloud 20 km from an observer. Viewing angle and solar elevation angle were both 18°, so the observer would be looking at a portion of cloud 8 km above the surface; the geometry is shown in figure 3.2. Height dependent vapor density was specified by choosing August 8th (5 pm.), 1986, Tucson radiosonde data since cloud reflectances gathered that day will be presented later. Precipitable water was 3.7 g cm$^{-2}$. Vapor absorbs along the circuitous path inside the cloud as well but we ignore it for the following reason: it was shown in last chapter's treatment of photon path distribution that nearly 90% of the reflected photons from a cloud of optical depth 128 travel distances less than 2 km. This path is negligible compared to the one from cloud to observer, especially since nearly 70% of the water vapor is below typical cloud base levels.
Figure 3.2. Geometry of LOWTRAN calculations.
We assume that the scattering from the cloud is given by $S_a$ since it was shown that beyond 1.3 $\mu$m a cloud with optical depth 64 or greater is virtually indistinguishable from an infinitely deep layer. The product $S_aT_{a-c}T_{c-g}$ was computed and normalized to unity to produce normalized cloud reflectance spectra, plotted in figures 3.3-3.5. Note, in figure 3.3, the size dependence of liquid water absorption in the window regions centered at 1.65 $\mu$m and 2.2 $\mu$m. The ratio of normalized reflectance from, say, the cloud comprised of 5 $\mu$m droplets to the one comprised of 25 $\mu$m droplets is 1.62 for the wavelength 1.65 $\mu$m and 2.62 at 2.20 $\mu$m. This is to be expected, because if we crudely approximate reflectance to be $1-c/r^2$ (taking ratios eliminates the normalizing factor), where $r$ is droplet radius and $c$ is a constant, then, as $r$ increases, reflectance decreases most rapidly at the wavelength of greatest absorption. At 1.65 $\mu$m $k_a$ is 5.1 cm$^{-1}$ and increases to 16.5 cm$^{-1}$ at 2.2 $\mu$m. For ice clouds, in figure 3.4, we see less size dependence in the ratio of reflectance at 1.65 $\mu$m to that at 2.2 $\mu$m (near unity for all sizes) because absorptions in the windows are nearly equal; for ice, $k_a$ is 18.3 cm$^{-1}$ and 16.5 cm$^{-1}$ (almost identical to liquid water absorption at 2.2 $\mu$m) at 1.65 $\mu$m and 2.2 $\mu$m respectively.

More pronounced differences between phase are illustrated in figure 3.5. Most notable is the shift of
Figure 3.3. Normalized cloud reflectance including absorption by water vapor between sun and cloud and between cloud and ground. Clouds are composed of liquid water droplets with average radius: A) 5 \( \mu m \); B) 25 \( \mu m \); C) 50 \( \mu m \).
Figure 3.4. Normalized cloud reflectance including absorption by water vapor between sun and cloud and between cloud and ground. Clouds are composed of ice spheres with average radius: A) 5 μm; B) 25 μm; C) 50 μm.
Figure 3.5. Normalized cloud reflectance including absorption by water vapor between sun and cloud and between cloud and ground. Clouds are composed of: A) 5 μm liquid water droplets; B) 25 μm liquid water droplets; C) 25 μm ice spheres.
about .05 μm in peak signal near 1.65 μm in going from a water to ice cloud. This shift can only be reconciled on the basis of a phase change since a liquid water absorption minimum occurs near the center of the 1.65 μm window. Increasing the average size of liquid water droplets would decrease the signal but would produce no shift. A similar shift is seen in the bulk absorption (figure 2.1); however, for the spectra shown here, water vapor absorption confines the signal maxima to window regions. For liquid water droplets and ice particles of equal size, the magnitude of the peak at 2.25 μm for the ice cloud slightly exceeds that for the water cloud and the crossover point (where liquid absorption becomes greater) is again predicted on the basis of bulk absorption.

The motivation behind normalizing these spectra is two-fold: measured cloud spectra were normalized in a similar fashion so that analysis was restricted to the qualitative features mentioned above; secondly, normalization allows us to eliminate the particle shape dependence of absorption since it enters only as a scaler factor which is unknown, but does not vary with wavelength (see section 2.1). Determining absolute absorption of an ice cloud would prove difficult, but particle shape should have negligible effect on relative spectral features.
3.2 Aerosol and Molecular Scattering

Along with water vapor absorption, aerosol scattering contributes to the near-infrared transmittance of the atmosphere but we can easily show that it can be neglected. Visible range $s$ is defined by

$$s = 3.912/a,$$

where $a$ is the aerosol scattering coefficient at $0.5 \, \mu m$. For a visible range of 50 km ($a = 0.08 \, \text{km}^{-1}$, considered very "hazy" for Tucson where our cloud reflectance measurements were made; see Young and Herman, 1988), Zuev (1970) shows that through a 10 km horizontal path, transmittance due to aerosol scattering between 1 \, \mu m and 2 \, \mu m is about 0.95. More importantly, however, it varies slowly with wavelength, especially compared to absorption by water. Since we are interested in making relative measurements, aerosol scattering can be ignored.

Extinction due to molecular scattering at near-infrared wavelengths (Rayleigh scattering) is very small and therefore, can be neglected. Transmittance due to molecular scattering is about 0.99 at 1.65 \, \mu m (from LOWTRAN).
CHAPTER 4
INSTRUMENTATION

A spectroradiometer was built and used to measure the near-infrared reflectance spectra from clouds. In the early phase of this project our aim had been to assess the feasibility of making these measurements from the ground and to ascertain if any information could be gleaned from the resulting spectra.

The first design, operating during the summer of 1984, used a germanium photodiode as the radiometer detector and it was found to be unstable over a period of a few hours, resulting in excessively noisy spectra. A thermoelectrically cooled Ge photodiode with substantially lower noise equivalent power was installed the following summer. Measured spectra were considerably smoother than those acquired in 1984 and it became apparent that certain spectral features showed marked differences depending upon what portion of a cloud had been observed and at what stage of a cloud's evolution the measurement had been made. Over 100 spectra obtained during a two month period were consistent with what we had perceived to be differences between liquid water and ice clouds, based upon the measured
signal in the 1.55 μm – 1.75 μm atmospheric window. Unfortunately, the long-wavelength limit of the detector was 1.9 μm which excluded the potentially useful 2.1μm – 2.3μm window. Furthermore, data transfer and storage was rather sluggish, requiring 4 minutes to obtain a single spectrum. Many of the observed cumulonimbi developed, matured, and dissipated over periods as short as an hour; a higher temporal frequency of measuring reflectance spectra was desired to track cloud evolution.

The initial success of the spectroradiometer subsequently led to modifications for the measurement period of summer, 1986. A new detector, a "self-cooled" lead-sulfide photoconductor, extended the long-wavelength cutoff; an overhauled data transfer and storage system allowed for spectra to be acquired as rapidly as every 90 seconds. Most of the discussion in this chapter pertains to this most recent state of the instrument; brief reference is given to the Ge photodiode and the data transfer device prior to 1986. Treatment of DC offset signal and signal normalization is reserved for the next chapter where measured reflectance spectra are presented.
4.1 The Spectroradiometer

A spectroradiometer consists of three primary components: collecting and focusing optics; a dispersing element; a detector. Our design comprised the following:

a) Collecting and focusing optics

The front-end optical system was a Newtonian-type reflecting telescope (Celestron) that had a 125 mm diameter, 725 mm focal length parabolic mirror and a plane mirror to direct light out the side of the telescope tube. The telescope was mounted in an altazimuth configuration to allow independent motion about horizontal and vertical axes. A finder scope attached to the main optical tube and aligned with the primary mirror enabled aiming at specific cloud features.

b) Dispersing element

A two-piece circular variable interference filter (CVF) was used as the dispersing element (Optical Coating Laboratory, Inc.). Spectral transmission for the first filter half ranged from .61 μm to 1.41 μm with an average percent half-bandwidth of 2%, or roughly 15 nm to 30 nm. For the second filter half it ranged from 1.16 μm to 2.30 μm with an average percent half-bandwidth of 1.85%, or
20 nm to 44 nm. Transmission curves, derived from laboratory specifications, are shown in figure 4.1 for both halves of the CVF. The bandpass center wavelengths were highly linear with angle about the disk, as determined with a Cary spectrometer, the results from which are listed in table 4.1. Here $\theta$ is the angular position on the filter referenced, for each half, from the short-wavelength cutoff, $\lambda$ is the center bandpass wavelength, and $\lambda'$ is wavelength determined by a least-squares linear fit; percent deviation is defined by $100\times|\lambda-\lambda'|/\lambda$.

c) Detector

Two types of infrared photodetectors (Hammamatsu) were used; a self-cooled Germanium photodiode with a peak response at 1.55 $\mu$m and long-end cutoff at 1.90 $\mu$m; a self-cooled lead-sulfide photoconductor with peak response at 2.2 $\mu$m and long-end cutoff at 3.0 $\mu$m. The sensitive area of the Ge detector was 1 mm$^2$ and that of the PbS detector was 20 mm$^2$. Each detector was cooled to near 0°C to ensure stability and increase detectivity ($D^*$, shown in figure 4.2) to about $1.5\times10^{11}$ cm-Hz$^{1/2}$/W.

The CVF and photodetector were housed in a box, hereafter referred to as the radiometer, mounted onto the telescope and positioned with its 1/4 inch diameter entrance aperture at the telescopic focus. By imaging an object at
Figure 4.1. Circular Variable Filter transmission.
Table 4.1. Linearity of Circular Variable Filter.

<table>
<thead>
<tr>
<th>θ</th>
<th>λ(μm)</th>
<th>λ'(μm)</th>
<th>%Deviation</th>
<th>λ(μm)</th>
<th>λ'(μm)</th>
<th>%Deviation</th>
</tr>
</thead>
<tbody>
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<td>5</td>
<td>0.652</td>
<td>0.639</td>
<td>2.04</td>
<td>1.198</td>
<td>1.198</td>
<td>0.04</td>
</tr>
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<td>1.261</td>
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<tr>
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<td>0.749</td>
<td>0.24</td>
<td>1.353</td>
<td>1.356</td>
<td>0.22</td>
</tr>
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<td>45</td>
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<td>0.816</td>
<td>0.29</td>
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<td>1.451</td>
<td>0.13</td>
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<td>0.57</td>
<td>1.544</td>
<td>1.546</td>
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<td>0.43</td>
<td>1.736</td>
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<td>0.16</td>
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<td>1.928</td>
<td>1.926</td>
<td>0.10</td>
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<td>0.06</td>
<td>2.023</td>
<td>2.021</td>
<td>0.09</td>
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<tr>
<td>150</td>
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<td>1.283</td>
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<td>2.116</td>
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<tr>
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<td>1.353</td>
<td>1.349</td>
<td>0.26</td>
<td>2.212</td>
<td>2.211</td>
<td>0.03</td>
</tr>
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<td>175</td>
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<td>1.394</td>
<td>0.43</td>
<td>2.270</td>
<td>2.275</td>
<td>0.20</td>
</tr>
</tbody>
</table>

Plate 1: \( \lambda = (0.616 \pm 0.003) + \theta (4.44 \pm 0.03) \times 10^{-3} \)

Plate 2: \( \lambda = (1.165 \pm 0.001) + \theta (6.35 \pm 0.01) \times 10^{-3} \)

(\( \lambda \) is in μm, \( \theta \) in degrees)
Figure 4.2. Detectivity of the two photodetectors used.
100 feet, the field of view of the composite system was estimated to be .007 radians. The radiometer box also housed preamplification electronics, consisting primarily of an operational amplifier operated in reverse bias mode as a current to voltage converter. A 12-volt dc motor rotated the CVF along with an accompanying disk that was attached to the drive shaft and through the edge of which a 2mm diameter hole had been pierced. An infrared emitter/detector switch (Monsanto) enveloped a segment of that disk and, as the hole passed through the switch, would generate a 2 volt pulse used to trigger a signal averager (Princeton Applied Research) to begin sampling. The configuration is illustrated in figure 4.3a.

The signal averager served the duel purpose of providing further signal amplification and enhanced signal-to-noise ratio (S/N). In summation mode, if \( n \) was the number of summations, or "sweeps," S/N would be increased by a factor of \( n^{1/2} \). A spectrum was divided into either 1024 or 2048 channels, each of which was represented by a 29-bit word. A dwell-time (DT) setting determined the sampling and 8-bit A/D conversion rate for each channel and therefore, along with the number of sweeps \( n \), also determined the time to acquire a spectrum from the product \( nxDT \times 1024 \) (or 2048).

When 1024 channels were used, the drive motor speed was set at approximately 10 rotations per second and when 2048
Figure 4.3a. Radiometer dispersing element and detector.

Figure 4.3b. Schematic diagram of the complete system.
channels were used, the speed was set at 5 rotations per second. Thus, for either case, a dwell time setting of 100 \( \mu s \) enabled one sweep to capture the spectrum from a complete revolution of the CVF (i.e., \( 100\mu s \times 1024 \approx 1\text{ s} \) and \( 100\mu s \times 2048 \approx 2\text{ s} \)). The analog output from the signal averager was sent to an x vs y oscilloscope for real-time monitoring of the spectrum being captured. The total number of sweeps to acquire a spectrum was determined from experiment; a minimum number of sweeps was necessary to build the signal to noise level above an acceptable threshold but the transient nature of cumuliform clouds dictated a maximum number of sweeps such that cloud microphysics would not be expected to vary significantly over the measurement period. 300 to 700 sweeps were found to satisfy these criteria, enabling an integrated spectrum to be acquired in 1 to 2 minutes.

For the measurements made during summer, 1985, when the Ge detector was used, a Rockwell Aim 65 microcomputer transferred digital output from the signal averager onto cassette tape. The time to complete the measurement and storage of one spectrum took approximately 4 minutes. The following summer, 1986, the PbS detector was used and a Commodore-64 computer transferred the data onto disk, during which time a running mean average over 7 consecutive data words reduced data storage with negligible reduction in
spectral resolution. The time required to obtain and store a spectrum was reduced to 90 seconds, thereby increasing the temporal frequency with which spectra from evolving cumulonimbi could be measured. Figure 4.3b is a schematic of the composite system, and a photograph of the instrument is shown in figure 4.4.

4.2 Wavelength Calibration

A set of narrow-band (30 nm half-bandwidth) interference filters (Corion) was used to provide wavelength markers for calibration. 0.8 μm and 1.0 μm filters supplied markers for the first half-plate of the CVF while 1.6 μm and 1.98 μm supplied markers for the second half-plate. The filters were positioned at the entrance aperture of the radiometer and the radiometer was pointed at an incandescent lamp source. Since the CVF wavelength was a linear function of angular position, \( \theta \), and the CVF was spun at a constant angular speed, wavelength was a linear function of channel number, \( c_n \); that is,

\[
\lambda = a_o + a_1 \theta = a_o' + a_1' c_n
\]

Since there were two markers for each half of the spectrum, each half could be calibrated independently by determining the coefficients \( a_o' \) and \( a_1' \). Repeated calibration indicated a precision of ± 2 channels or roughly ± .01 μm. The rotation speed of the motor drifted slightly
Figure 4.4. The spectroradiometer.
with time, by approximately 1% over several hours, but this
could be remedied by scaling since sharp markers on recorded
spectra indicating the beginning and end of each spectrum
half were provided by the seam where the two semicircular
segments of the CVF were cemented together.
CHAPTER 5
MEASUREMENTS

The theoretical treatment in chapters 2 and 3 demonstrates that the phase of the condensed water in clouds can be distinguished on the basis of differences between liquid water and ice absorption in the parts of their near-infrared spectral bands where water vapor absorption is negligible. The instrument used to measure relative cloud reflectance was described in chapter 4 and those results are to be presented in this chapter.

We first give a brief description of the Arizona monsoon season and the conditions under which measurements were made, followed by a discussion of the strategy involved in collecting the data. Two spectra obtained during summer, 1985, with the older version of the spectroradiometer are presented first, followed by two sets of spectra, each relating to single cumulonimbus during the 1986 monsoon, after the instrument was modified.
5.1 The Arizona Monsoon

The climate of southern Arizona, while on average very dry, provides an ideal setting for observing convective storms during the summer months of July, August, and September. In Tucson, nearly half of the annual average 11 inches of precipitation falls during this period. The source of this moisture is a maritime tropical airmass which makes several incursions into Arizona during the summer, temporarily replacing a continental airmass. This exchange is brought about by the northward migration of the surface subtropical ridge located in the southwest U.S. Many of the features associated with the boundary separating the moist and dry air masses resemble typical, albeit weak, mid-latitude frontal systems (Adang and Gall, 1988).

Near Tucson, clouds first develop when southeasterly currents, carrying moisture from the Gulf of Mexico, rise over mountains to the east and north of the city. As the surface heats, by mid-afternoon local convection gives rise to scattered cumuli throughout the valley. The relative isolation of individual convective pockets and absence of obstructing low clouds allows a clear view to the sides and tops of thundercells, unlike the organized line of storms associated with a front. During the monsoon season upper-air winds tend to be relatively light (about 10 to 20 knots
between the 500 mb and 300 mb pressure levels). For this reason, storms generally show little horizontal motion, developing, maturing, and dissipating in the same locale.

5.2 Measurement Procedures

Cloud reflectances were measured in the following manner: The spectroradiometer was aimed at a particular feature of a cloud, for instance a rising turret, by an observer using the spotting telescope. During acquisition time (approximately 90 seconds) the observer manually tracked the cloud, adjusting the instrument accordingly. Because most clouds showed little horizontal motion, tracking usually involved only slight vertical adjustments: a cloud at a distance of 20 km and rising 5 ms\(^{-1}\) would change in elevation by only about 1° over the acquisition period.

There were two stages involved in gathering a collection of spectra from a single cloud. While a cloud was in early stages of growth we aimed the instrument at upper regions of a rising turret (since this would be the most likely part of a non-glaciated cloud to glaciate), recording a number of spectra from the same turret as it evolved. After vertical growth had ceased, or after phase change had been inferred to have occurred on the basis of gross changes in the measured spectra, the cloud was scanned at several
vertical levels. In this manner we attempted not only to determine the onset of glaciation but to compare spectra from the same cloud at various altitudes as well.

Each time a spectrum was recorded a photograph of the cloud was taken and later marked to indicate the approximate region of the cloud that was scanned. The purpose was not to precisely map the targeted area of cloud but rather to indicate very broad regions, for example, top, bottom, left, right, etc. Furthermore, the photograph served as a record of the visible features of the cloud at the time its near-infrared spectrum was measured, so that it could be determined whether certain attributes, such as a "fuzzy" or "diffuse" appearance, were correlated with phase.

5.3 Signal Normalization

As figures 3.3-3.5 indicate, relative reflectance should suffice in discriminating ice from liquid water in clouds. This is fortunate because absolute measurements would be virtually impossible from the ground. The path from sun-to-cloud and cloud-to-observer would produce a significantly different atmospheric transmittance from that along the path from sun to observer, where a reflectance standard would be situated. Even if measurements were restricted to water vapor "window" regions, for low solar and observation angles, one could not account for the small,
but finite, vapor absorption or aerosol extinction. Measured signals were also dependent upon photodetector response and circular variable filter transmission. However, since only relative reflectances were desired, data were normalized with respect to the signal value at a wavelength where absorption by water (vapor, liquid, and solid) is negligible. The .68 μm signal was chosen because it also coincided (for most of the measured spectra) with the peak signal value. (When the earlier version of the instrument was used, signals were also normalized with respect to its peak signal level, which occurred near 1.1 μm.)

To account for dc detector gain and possible light leakage into the radiometer, a signal was recorded with the telescopic entrance aperture covered. This signal was subtracted from measured spectra in one of two ways: the signal averager could be operated in an "inverse" mode where successive sweeps were subtracted instead of summed, allowing for offset to be eliminated immediately following the acquisition of a spectrum. Often, however, a rapid succession of spectra were desired from an evolving cloud, so that immediately following the capture of one spectrum, the instrument was positioned to record another. In these instances, the offset signal was recorded when time permitted and subtracted via software at a later time.
5.3 Measurements: 1985

The example presented in this section depicts a representative case from the summer of 1985. As stated in the previous chapter, data transfer and storage did not allow spectra to be recorded faster than one per 4 minutes. Rather than follow the evolution of a single turret (which might change dramatically in that time), the strategy was to obtain spectra from various clouds in different stages of development.

Figures 5.1 shows photographs of a towering cumulonimbus to the east of Tucson at 15:30 and 15:34 on August 27, 1985. Reflectance was measured first from the top section of a finely structured, newly developing cloud in the foreground, indicated by the circle in the photograph labeled A. (Note that the circle only depicts the general region where the cloud was viewed by the spectroradiometer, not the instrument's field view). The next target was the upper portion of the main cell (indicated in photograph B) which formed a visible anvil 5 minutes later. Accurately determining the range of clouds is very difficult, but the proximity of the mountains to the east of the cloud in these pictures suggests a range of 15 to 20 km. At that distance, the spectroradiometer, with a field of view of .007 radians, would "see" an area on the order of 100 m wide.
Figure 5.1. Clouds A and B. Time of day is indicated above each photograph; the solar elevation angle was approximately 39°. (Circles depict region of cloud viewed by the instrument.)
The resulting signals are shown in figure 5.2 where shaded regions delineate two water vapor transmission windows. Clearly the greatest difference between the spectra is found in the window centered near 1.65 μm where the signal from cloud B was slightly more than one-third that of cloud A at the same wavelength. Between 1.6 and 1.7 μm absorption in ice is about 14 to 30 cm⁻¹ while in water it is 5 to 6 cm⁻¹, so these measurements strongly suggest that cloud A was composed of liquid water, while cloud B was composed of ice. The peak signal of the cloud B spectrum near 1.65 μm appears to be slightly shifted (in wavelength) from that same peak in the cloud A spectrum; a transition from water to ice would produce such a shift, but in figure 5.2 the shift is not convincing. The results from the following summer, however, proved capable of detecting that shift unambiguously.

5.4 Measurements: 1986

Measurements made during the summer of 1986 allowed an extended spectral range to include the water vapor window centered at 2.2 μm, lending support to evidence of phase changes deduced from signal levels near 1.65 μm. Also, the number of sampled points per spectrum was increased from 1024 to 2048 with a 7 point weighted running mean performed at every fourth point. (The previous method was to average
Figure 5.2 Signals from clouds A and B of figure 5.1. Shaded areas indicate regions where water vapor absorption is slight.
over 2 points). Although the ultimate spectral resolution was limited by the bandwidth of the circular variable filter (about 30 nm at 1.6 μm), this should not have affected the estimation of locating shifts in the peak position of the relatively broad spectral feature between 1.55 and 1.75 μm. The combination of higher sampling resolution and increased signal to noise ratio (S/N, determined by the magnitude of "grass", improved from about 20:1 to 200:1) of the PbS detector enabled the detection of the shift.

Figure 5.3 shows a series of 9 photographs (labeled C through K) of a cumulonimbus cloud on August 8, 1986. Again, circles represent the region of cloud scanned at the time indicated above each picture. Time of day is not in sequence because each of the three rows were grouped according to similarity of targeted area. Spectra are plotted in figures 5.4-5.6 in groups of three, labeled according to their corresponding letters in figure 5.3. Figures 5.4a-5.6a show signal measured from the short-wavelength segment of the circular variable filter. Notice that between .7 and .9 μm, where absorption by liquid water and ice is very weak, there are only minor differences in signal level among the spectra, but reduced signals in some of the spectra are evident at 1.05 μm and 1.25 μm where absorption by liquid water and ice becomes appreciable (see figure 2.1).
Figure 5.3. Cloud of August 8, 1986. Circles depict region being viewed. Time of day is indicated above each photograph; solar elevation angle varied from $26^\circ$ (cloud C) to $20^\circ$ (cloud K).
Figure 5.4a. Short-wavelength reflectance from clouds C, D, and E.
Figure 5.4b. Long-wavelength reflectance from clouds C, D, and E.
Figure 5.5a. Short-wavelength reflectance from clouds F, G, and H.
Figure 5.5b. Long-wavelength reflectance from clouds F, G, and H.
Figure 5.6a. Short-wavelength reflectance from clouds I, J, and K.
Figure 5.6b. Long-wavelength reflectance from clouds I, J, and K.
The long-wavelength spectral features are shown in figures 5.4b-5.6b with the extended spectral range out to 2.3 \( \mu m \). Small differences between measured signals at 1.05 \( \mu m \) and 1.25 \( \mu m \) become magnified at these longer wavelengths, as evidenced by substantial reductions in reflectance in both the 1.55\( \mu m \) - 1.75\( \mu m \) and 2.1\( \mu m \) - 2.3\( \mu m \) water vapor transmission windows. However, before detailing characteristics of specific spectra, we present measurements from another cloud, shown in figure 5.7. This series of photographs (labeled L through T) was taken on the afternoon of August 14, 1986 at the times indicated above each picture. The resulting spectra are displayed in figures 5.8a-5.10a (short-wavelength end) and 5.8b-5.10b (long-wavelength end).

We will focus on three distinct aspects of the spectral differences between absorption in ice and absorption in liquid water; each of these influence cloud reflectance spectra in ways which can be predicted qualitatively and used to determine cloud phase:

1) Reduced signal level at 1.65 \( \mu m \)

Based upon the results in chapters 2 and 3, and owing to greater absorption in ice than in liquid water over the spectral region 1.6\( \mu m \)-1.7\( \mu m \), one would predict a cloud
Figure 5.7. Cloud of August 14, 1986. Circles again depict region being viewed. Time of day is indicated above each photograph; solar elevation angle varied from $13^\circ$ (cloud T) to $19^\circ$ (cloud L).
Figure 5.8a. Short-wavelength reflectance from clouds L, M, and N.
Figure 5.8b. Long-wavelength reflectance from clouds L, M, and N.
Figure 5.9a. Short-wavelength reflectance from clouds O, P, and Q.
Figure 5.9b. Long-wavelength reflectance from clouds O, P, and Q.
Figure 5.10a. Short-wavelength reflectance from clouds R, S, and T.
Figure 5.10b. Long-wavelength reflectance from clouds R, S, and T.
composed of ice particles to show weaker reflectance in this spectral band than a similar cloud composed of liquid water. Furthermore, for an ice cloud, the ratio of reflectance at 1.65 \( \mu m \) to that at 2.2 \( \mu m \) would be near unity (neglecting detector response) because absorption in ice is very similar at these two wavelengths. Liquid water, on the other hand, absorbs about three times more at 2.2 \( \mu m \) than it does at 1.65 \( \mu m \). Thus, for a liquid water cloud, the ratio of reflectance at 1.65 \( \mu m \) to 2.2 \( \mu m \) would be significantly greater than that ratio for an ice cloud; this behavior would not be affected by detector response since detector response would enter only as a constant factor (the ratio of response at 1.65 \( \mu m \) to that at 2.2 \( \mu m \)).

2) Shift of wavelength of peak signal between 1.6\( \mu m \)-1.7\( \mu m \).

The peak signal occurs near the wavelength at which absorption attains a minimum. The optical constant data (figure 2.1) shows the absorption minima for ice to be shifted towards longer wavelengths than the absorption minima for liquid water. The extent of this shift is 0.03 \( \mu m \) to 0.07 \( \mu m \) throughout the near-infrared. (Absorption maxima are simultaneously displaced, but it is the minima that are relevant to this discussion.)
3) Narrowing of the peak in the 2.1\(\mu m\)-2.3\(\mu m\) region.

Theoretical reflectances in figures 3.3-3.5, show ice cloud spectra which have sharper and narrower peaks between 2.1 and 2.3 \(\mu m\) than the corresponding features for liquid water. Of course water vapor absorption ultimately defines the width of this feature but the measurements were all from the same cloud with only small differences in observation angle and over a short period of time, so the water vapor path would have varied negligibly.

Table 5.1 lists, for cloud spectra C-K and L-T, the following: a) position of maximum signal in the 1.6\(\mu m\)-1.7\(\mu m\) band; b) ratio of reflectance at 1.65 \(\mu m\) to that at 2.2 \(\mu m\) \((R_{1.65/2.2})\) - the values shown represent reflectances integrated over 1.6 to 1.7 \(\mu m\) and 2.15 to 2.25 \(\mu m\), respectively; c) Half-bandwidth of the 2.1\(\mu m\)-2.3\(\mu m\) reflectance band - this is defined as the width at the midpoint between the peak value and the value at 2.3 \(\mu m\); d) elevation angle of the measured part of the cloud, given by an inclinometer attached to the telescope mount. (Angle was read to within \(\pm 0.3\) of a degree.)

There is a great deal of consistency among these spectra. Most could be classified in one of just two categories. For those spectra where the signal peaked near
Table 5.1. Near-infrared spectral features of clouds shown in figures 5.3 and 5.7.

<table>
<thead>
<tr>
<th>Cloud</th>
<th>Elev. Angle (degrees)</th>
<th>1.6μm-1.7μm Max (μm)</th>
<th>$R_{1.65/2.2}$</th>
<th>Half-BW of 2.2 μm Band (μm)</th>
<th>Phase</th>
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<td>7.7</td>
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<td>1.69</td>
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<tr>
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<td>liquid</td>
</tr>
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<td>1.66</td>
<td>1.24</td>
<td>.09</td>
<td>ice</td>
</tr>
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<td>1.65</td>
<td>1.62</td>
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</tr>
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<td>1.66</td>
<td>1.25</td>
<td>.09</td>
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<tr>
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<tr>
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<td>1.15</td>
<td>.08</td>
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</tr>
<tr>
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<td>1.03</td>
<td>.07</td>
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<td>liquid</td>
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<td>O</td>
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<td>1.35</td>
<td>.12</td>
<td>?</td>
</tr>
<tr>
<td>P</td>
<td>10.2</td>
<td>1.67</td>
<td>1.24</td>
<td>.11</td>
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</tr>
<tr>
<td>Q</td>
<td>8.8</td>
<td>1.66</td>
<td>1.34</td>
<td>.11</td>
<td>?</td>
</tr>
<tr>
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<td>1.65</td>
<td>1.87</td>
<td>.15</td>
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<tr>
<td>S</td>
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<td>1.21</td>
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</tr>
<tr>
<td>T</td>
<td>11.3</td>
<td>1.67</td>
<td>1.29</td>
<td>.09</td>
<td>ice</td>
</tr>
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</table>
1.65 μm, the ratio $R_{1.65/2.2}$ tended to be around 1.5 or greater (characteristic of liquid water clouds). When that peak shifted to 1.66 μm or beyond, that ratio dropped to about 1.2 or lower. Similarly, those spectra in the former category showed an approximate 50% increase in half-bandwidth of the 2.2 μm band over spectra in the latter category (.15 μm to .1 μm). Based upon these results, cloud phase is indicated for each spectrum in table 5.1. Two of the spectra, O and Q, have characteristics that lie somewhere between expected liquid water and ice cloud signatures. There is only a slight shift of the maximum in the 1.6μ-1.7μm band (1.655 μm) and $R_{1.65/2.2}$ (1.35) is near the middle of the extremes that are seen (1.0-1.9). Such behavior could indicate a mixture of water and ice. It will be discussed further in the following chapter.

5.5 Discussion

In the case of the August 8th cloud, spectra C, D, and E were measured from the same portion of turret (see the photographs in figure 5.3). In 6 minutes the elevation of the cloud rose 2°. Judged by visible features, the cloud had appeared to change very little. However, the spectra in figure 5.4b show a substantial change and indicate that the cloud had glaciated sometime between 17:10 and 17:14. Likewise, the spectra in 5.5b show that the turret targeted
in the series FGH changed phase over the time period indicated. Spectrum G has characteristics of water cloud absorption even though signal level had been reduced compared to spectrum F. This could be due to change in droplet size or optical depth (see figures 2.4-2.6) and will be discussed further in the next chapter. In the final sequence from this cloud, I-J, spectra were measured from upper regions of the cloud as it reached maturity (I) and after an anvil had formed (J and K). These spectra, in figure 5.6b undoubtedly indicate glaciation.

The cloud on August 14th (figure 5.7) provided 6 consecutive spectra, L-Q, in roughly 2 minute intervals. As indicated in the previous section, it is difficult to determine phase from spectra O and Q. Unfortunately, at the time the measurements were made, scanning of this turret had been terminated because the oscilloscope display showed an ice-like signature. However, the cloud had clearly glaciated by the time spectra S and T were obtained. Spectrum R, incidently, shows that the lower regions of this cloud still gave an unambiguous liquid water signal.

The conclusions drawn from these measurements simply imply that young, immature liquid cumuli rapidly glaciate as they grow to higher (cooler) levels. This is not startling, but the phase inferred from some of the spectra contradicts a standard meteorological method of determining cloud phase.
from visible features. The Observer's Handbook (1982) states "The change from large cumulus with domed tops and hard outline (produced by water clouds) to a top with a softer fibrous outline (produced by ice crystals) marks the change from cumulus to cumulonimbus." Our measurements showed several clouds with "domed tops" and "hard outlines" to have glaciated. These include clouds B (from 1985), E, H, and I. Investigators measuring cloud microphysics in-situ have frequently found ice in clouds which had features similar to those associated with the definition of a water cloud stated above. However, the notion that one can infer the phase of a cloud by its visible features persists (see Scorer, 1972, for another example).
CHAPTER 6
LIMITATIONS OF TECHNIQUE

The limitations of using near-infrared cloud reflectance to discriminate ice from liquid water will be discussed in this chapter. Three factors are considered as they apply to the result presented in chapter 5: heterophase clouds; variation of optical depth; absolute verification of cloud phase. Further applications, specifically, satellite and transmission measurements, are suggested.

6.1 Heterophase Clouds

Most of the measured spectra presented in chapter 5 bore unambiguous liquid water or ice cloud signatures, making phase simple to identify. However, two of the spectra, those labeled O and Q in figure 5.9b, exhibit a spectral behavior between the expected behavior for liquid water and for ice. Apart from measurement error, the ambiguity could have arisen if the part of the cloud being viewed by the instrument contained a mixture of liquid water and ice particles. Certainly a heterophase cloud could confuse matters if the total absorption cross-section of each condensed species was comparable to one another.
Mixed-phase cloud reflectances were computed and are shown in figure 6.1; the curves are normalized reflectance (allowing for absorption by water vapor between cloud and ground) from infinitely thick clouds composed of: 5 μm liquid water droplets (curve A); 25 μm ice particles (curve B); a mixture of 5 μm liquid and 25 μm ice particles in concentrations which gave equal total geometric cross-sections (curve C); and equal number concentrations (curve D). Reflectance was computed in the manner described in chapter 2. For the clouds C and D, average single scattering albedos and asymmetry parameters were computed by weighing extinction and scattering cross-sections according to the assumed number concentration. These averaged quantities have no meaning in single particle scattering but are appropriate for multiple scattering calculations where photons interact with several particles.

The curves C and D in figure 6.1 demonstrate the effect heterophase clouds have on near-infrared reflectance. As the concentration of ice increases, so does the shift of the peak near 1.65 μm, while the magnitude of that same peak decreases. That is, the signal becomes more ice-like. Both the number concentration and average size of the water droplets and ice crystals ultimately determine the cloud's near-infrared spectral features. We have shown only two examples out of an infinite number of possible combinations;
Figure 6.1. Reflectance from clouds composed of: A) 5 μm liquid water drops; B) 25 μm ice particles; C) liquid/ice mixture, equal geometric cross-section; D) liquid/ice mixture, equal number concentration.
the task of determining the relative abundance of each phase would be hopeless. If, however, the presence of a mixed-phase cloud could be detected, it could prove beneficial to weather modification studies.

Clouds composed of both super-cooled liquid water droplets and ice crystals are colloidally unstable (Wegener, 1911), the consequence of which produces precipitation-size particles in many types of clouds (Bergeron, 1935). The diffusion of vapor from liquid to solid particles can lead to rapid depletion of the liquid. If ice crystals were present in concentrations comparable to that of ice nuclei (on the order of 1 to 10 per liter between -15°C and -20°C; see, for example, Bigg and Hopwood, 1963, and Mossop et al., 1970), one would expect a water-like reflectance signal from a cloud with smaller, but more abundant, water drops until the cloud was almost completely converted to ice. (The measurements in liquid water cumuli by Warner, 1969, suggest several hundred to a thousand droplets per cubic centimeter, with average radii from 5 μm near cloud base to 10 μm near cloud top.) Observations have indicated that the concentration of ice crystals often exceed nuclei concentration, as measured by ice nucleus counters, by several orders of magnitude (see chapter 1) and the nucleation of large numbers of particles can occur quite rapidly. Koenig (1963) reported several hundred crystals per
liter in a mixed phase cumulonimbus turret within 5 minutes of the onset of glaciation. It does not seem unreasonable to suggest that under these conditions a mixed-phase reflectance signal could be detected.

6.2 Variation of Optical Depth

We have focussed most attention on reflectance from infinitely thick (optically) clouds since it was shown in section 2.3 that a cloud of optical depth 64 (geometric depth of 1 km) would be virtually indistinguishable from a semi-infinite cloud at wavelengths where absorption in liquid water and ice is appreciable. However, measured spectra were normalized to the signal at a non-absorbing wavelength (.68 μm). At this wavelength, for a cloud of optical thickness 64, reflectance is about 90% of that for an infinitely thick cloud. Thus, the greatest uncertainty when comparing the magnitude of normalized reflectance from several clouds (or from several spectra from the same cloud) may be due to variations in optical depth.

Figure 6.2 shows for average droplet radius 5 μm, the variation of reflectance with optical depth at .75 μm where scattering is conservative (ω₀=1, as it is at .68 μm), and at 1.65 μm where absorption in liquid water is appreciable (ω₀=.997). Note that as optical depth increases, reflectance increases but at a decreasing rate. Also, beyond an optical
Figure 6.2. Reflectance versus $\tau$; $\bar{F} = 5 \, \mu m$
depth of about 10, an increase in $\tau$ leads to a larger increase in reflectance at $0.75 \, \mu m$ than at $1.65 \, \mu m$. For example, the reflectance at $0.75 \, \mu m$ for $\tau=50$ is about $0.85$, and at $\tau=55$, $0.88$, an increase of $3.5\%$; at $1.65 \, \mu m$, reflectance increases by only $1\%$ percent for the same increase in optical depth. (Of course, a cloud's optical depth at $1.65 \, \mu m$ is not identical to its optical depth at $0.75 \, \mu m$). However, the ratio of reflectance at $1.65 \, \mu m$ to that at $0.75 \, \mu m$ increases by $2.5\%$ when $\tau$ is increased from 50 to 55.

Variations in optical depth of this magnitude would not be uncommon for, say, a growing turret whose dimensions and droplet size spectrum are changing. The differences in signal at $1.65 \, \mu m$ between spectra C and D in figure 5.4a might be due to a slight change in optical depth; it is unlikely that an increase in average droplet size produced the decreased signal in spectrum D because an even greater reduction in signal should have occurred at $2.2 \, \mu m$.

Although the normalized reflectance signal at a particular wavelength would vary between clouds with different optical depths, inferences of cloud phase by the methods we have adopted are not critically dependent on differences in optical depth. The qualitative dependence of optical depth on reflectance is illustrated in figures 2.4-2.6. Thin clouds show weaker reflectance than their
infinitely thick counterparts. With increasing \( \tau \), the contrast between non-absorbing and absorbing spectral regions is enhanced. However, because near-infrared absorption in liquid water and ice is weak, reflectance in a spectral band would be affected by changes in optical depth similarly to that in any other spectral band. Hence, the ratio of the reflectance between 1.55-1.75 \( \mu m \) to that between 2.1-2.3 \( \mu m \), as well as the position of the peak signal near 1.65 \( \mu m \), and the width of the 2.1-2.3 \( \mu m \) feature show negligible variation with optical depth, but considerable variation with phase. Therefore, changes in the optical depth over the course of measurements from a cloud should not affect phase discrimination by the methods which we adopted.

6.3 Verification of Results

Complete verification of results from any remote sensing can only be achieved through corroborative "in-situ" measurements - in this case, aircraft penetration of the observed clouds, unavailable at the time of our measurements.

During the 1986 monsoon season, a thermal-infrared radiometer was aligned with the spectroradiometer in attempt to measure the temperature of a cloud at the position it was scanned. The device converted the integrated thermal
emission from 8\(\mu\text{m}\)-14\(\mu\text{m}\) to temperature; that spectral region, however, while it roughly coincides with the atmospheric window, does include several absorption (and, therefore, emission) bands, resulting in large errors in temperature. Attempts to reduce the data to useful quantities by considering water vapor transmission have been unsuccessful.

Similarly, no attempt has been made to assign cloud-top temperatures from measured elevation angles by estimating the range because of the large uncertainties involved. (If range were known accurately, temperatures could be computed from the inferred height and temperature soundings by radiosonde.) For example, the range of the cloud in figure 5.7 was estimated to be 24 miles (40 km) from pilot reports. At that range, the temperature at an elevation angle of 10° would have been around \(-10^\circ\text{C}\). Assuming (optimistically) a 5 mile uncertainty in range, the uncertainty in temperature would have been \(\pm 10^\circ\text{C}\).

More importantly, however, between 0°C and \(-40^\circ\text{C}\) temperature is not indicative of phase. Observations have shown that, typically, at \(-10^\circ\text{C}\) supercooled liquid water clouds occur about as often as ice clouds (Prupacher and Klett, 1978). At warmer temperatures liquid water clouds are favored, while at colder temperatures, ice clouds are favored. Again, the only "ground-truth" verification of the
near-infrared remote measurements would be in-situ observations. Nevertheless, it would be difficult to attribute the large differences seen in the measured reflectance features in chapter 5 to anything other than differences in cloud phase.

6.4 Other Applications

The most attractive application of near-infrared cloud reflectance measurements would be on board satellites. The Thematic Mapper on Landsat satellites has broadband (approximately .2 μm) channels centered at 1.65 μm and 2.1 μm. Bradley (1989) has analyzed cloud reflectance from Landsat data by considering various ratios formed using those channels and a visible channel. Curren and Wu (1982) used Skylab near-infrared measurements in an attempt to infer cloud phase (see chapter 1). The proposed EOS satellites will have appropriate sensors to make similar measurements. The major drawback in measuring reflectance from above a cloud would be determining phase in clouds consisting of ice in upper regions and liquid water below (similar to the problem discussed in section 6.1), or in more complicated stratifications with clouds of different composition at different levels - as, for example, a thin cirrus sheet above a liquid water stratiform layer. The scattering from mixed-phased clouds needs to be examined in
greater detail before useful inferences from satellite measurements could be made.

Transmission measurements made at the surface could also indicate cloud phase but these measurements are somewhat more difficult than reflectance measurements because of weaker signals from thick clouds and ground reflectance. However, there would be no constraint on pointing a sensor so that wide-field device could be employed. Again, the information content of such measurements would be limited by potential mixed-phase clouds: radiation reaching the surface would traverse all levels of a cloud which may contain particles of varying phase and size.
CHAPTER 7
CONCLUSIONS

We have demonstrated the feasibility of detecting gross changes in near-infrared reflectance signals from cumulus clouds as they mature and evolve into cumulonimbi. From measured spectra, phase was inferred on the basis of the expected behavior of reflectance in the 1.55-1.65μm and 2.1-2.3μm spectral bands where water vapor absorption is negligible, while absorption in ice and liquid water is still appreciable and distinctly different from one another.

The predicted and observed differences between liquid water and ice clouds are attributable solely to differences between the respective spectral absorptions. This spectral information is well documented and there is close agreement between different tabulations and measurements. (For example, they all show the wavelength shift discussed in section 3.1)

The measurements are simple to make, requiring neither high resolution nor absolute response calibration. The examples of spectra presented in chapter 5 are representative of approximately 500 spectra* acquired over

*This data can be acquired by contacting the author through The University of Arizona, Dept. of Atmospheric Sciences.

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the 1985 and 1986 southern Arizona monsoon seasons, and inferred results are consistent with young cumuli becoming glaciated as they reach colder levels; this glaciation is found well before any anvil formation or development of visible fibrous structure takes place.

Finally, it should be noted that most of the reflected radiation from a cloud which reaches an observer at the ground could not have penetrated deeply - say, several kilometers - into the cloud (see section 2.4). Therefore, reflectance signals, and inferences drawn from them, relate to the few hundred meters near the cloud boundary.


LUDLAM, F.H., 1980: Clouds and Storms, the Pennsylvania State University, University Park, Pa., 405 pp.


PALTRIDGE, G.W., 1988: "Spectral and total albedo to solar radiation of ice and water clouds- experimental results from ASPIRE," Atmosfera, 1, pp. 5-16.


TWOMEY, S., 1985: Unpublished manuscript.

TWOMEY, S., 1985: Personal communication.


WEGENER, A., 1911: Themodynamik der Atmosphäre, J.A. Barth, Leipzig, p.331
