

OPTICAL REMOTE SENSING OF ICE IN CLOUDS

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Abstract. A practical method of inferring cloud phase with potential applications in weather modification is presented. The instrument used is a ground-based passive spectroradiometer which measures cloud spectral reflectance or transmittance; differences between ice and liquid water near-infrared absorption spectra are the basis for phase discrimination. Results show that reflectance spectra from cumulus towers abruptly change on the order of minutes due to the conversion of liquid water to ice. Transmission measurements from overlying stratiform clouds are presented and shown to also be useful in predicting cloud phase.

1. INTRODUCTION

Variability in the state (phase) of subfreezing clouds can be very great and is recognized as a notorious nuisance factor in weather modification studies: large amounts of supercooled water may be present in a cloud at one time or place while an apparently similar cloud at another time or place may be almost completely glaciated. Furthermore, appearance is a very poor and unreliable guide for distinguishing the solid state from the liquid (Bigg, 1971). Since cloud seeding by ice nuclei presupposes the existence of liquid water in supercooled parts of treated clouds, any technique which can help to distinguish liquid water from ice has obvious applications. If the technique is remote (i.e., can be applied without actually entering the cloud), more extensive, easier, and safer surveys become possible.

The purpose of this note is to describe and illustrate an optical remote sensing method which has been tested satisfactorily in a number of situations since 1985, is based on straightforward physical principles, and could be operated from the ground, from aircraft, or from a satellite.

2. OPTICAL REFLECTION BY CLOUDS

The sometimes dazzling brightness of clouds attests to their high reflectance, which may approach unity for thicker clouds. Most reflected photons emerge after many scatterings by cloud droplets and after traversing hundreds of meters inside the cloud. Apart from the glory rings (to which only once-scattered photons contribute) light reflected from clouds is colorless and its angular distribution featureless, so that visually one can infer nothing about the composition of the scatterers - clouds of salt, sugar, water, or alcohol drops, for example, would all look similar. Even the interparticle separation is irrelevant - snow and white powders look very much the same as clouds.

Because of the many scatterings which photons experience, a small amount of absorption becomes, in effect, greatly magnified. For example, if the probability of a photon being absorbed while inside a drop were one percent, the probability of absorption after twenty scatterings would be almost 20 percent because $(1-0.01)^{20} \approx .81$. Mixing a very small amount of soot or

dirt with common white powders (salt, sugar, flour) produces a very distinct darkening and it is this mechanism, multiple scattering, that is responsible. The whiteness shared by clouds, fine glass beads, flour, snow, white paint, etc., arises in every case from high order multiple scattering by particles with negligible absorption at visible wavelengths. When absorption is present, its effect is dictated primarily by the probability of absorption vis-a-vis scattering. In radiative transfer theory, a parameter ω_0 (the "single scattering albedo") is favored for historical reasons; this, however, is defined as the probability of scattering vis-a-vis scattering plus absorption, so $(1-\sigma_0)/\omega_0$ gives the first-mentioned probability. Under the circumstances to which we confine ourselves here, $(1-\sigma_0) \ll 1$, so $(1-\sigma_0)/\omega_0$ can be replaced by just $(1-\sigma_0)$ without serious error. For weak absorption in particles or drops appreciably larger than the wavelength, $(1-\sigma_0)$ is proportional to the bulk absorption coefficient k and the ratio of the volume of a scatterer to its cross section (i.e., the droplet radius r); for liquid water droplets, the proportionality factor is 0.85. The relationship is approximate, but it tracks very closely the results of more detailed exact computations, provided that $r \gg \lambda$ (wavelength) and $kr \ll 1$ (Twomey and Bohren, 1980).

Many common substances do not absorb visible light, but as we move just beyond the visible into the near-infrared, absorption bands begin to occur; these bands tend to be fairly regularly spaced and become progressively stronger with increasing wavelength. Both in position and strength, these bands are different for different materials, so, in the near-infrared, many substances become, in a loose sense, "colored." Water vapor has, of course, been extensively studied; between 1 and 2 μm , there are three groups of absorption bands centered at 1.14 μm , 1.35 μm , and 1.88 μm . Between those bands exist "windows" where the transmission through water vapor is virtually complete. Turning now to water in its other forms - solid and liquid - the qualitative behavior is similar, since the same molecule is involved, but now molecules are close enough to their neighbors to become influenced by them; because of that, absorption is a little weaker around the maxima, but it spills over into spectral regions where the vapor is transparent. This fact is crucial to the present discussion: in those windows, one can see through the water vapor

and still detect absorption due to liquid water or ice clouds. A second crucial fact is that because of differences between the molecular structures of liquid water and ice, absorption features (i.e., spectral maxima and minima) in ice are displaced roughly $0.05 \mu\text{m}$ towards longer wavelengths.

Figure 1 shows the spectral variation of absorption coefficient k for liquid water (unbroken curve) and ice (broken curve), from data in Irvine and Pollack (1968), Hale and Querry (1973), and Warren (1984). Transmission by water vapor is shown in figure 2, where shaded regions indicate virtual transparency: absorption by liquid and solid water in these regions, rather than around the regions of absorption maxima, is of primary interest to this study, since it is here that clouds can be observed free from the influence of water vapor (both inside the cloud and in the intervening path). As the figure shows, there is appreciable absorption (~ 1 to 100 cm^{-1}) in liquid water and ice in these regions. Relating back to the quantity σ_0 through the approximate relationship given earlier, one finds that for $10 \mu\text{m}$ drops $(1-\sigma_0) \approx 10^{-3}$ to 10^{-1} , which multiple scattering translates into a reduction in reflectance of several tens percent; furthermore, figure 1 shows large differences between ice and liquid water - especially around $1.65 \mu\text{m}$ and $2.2 \mu\text{m}$. Those differences enable the remote discrimination of ice from water by optical means. Incoming solar energy contains roughly 50% (about 700 watt m^{-2}) in the near-infrared: between 1.0 - $1.1 \mu\text{m}$ there is approximately 67 watt m^{-2} available at the top of the atmosphere, falling to 22 watt m^{-2} between 1.6 - $1.7 \mu\text{m}$ and 7 watt m^{-2} between 2.2 - $2.3 \mu\text{m}$. Before reaching a detector at the surface, that energy is reduced due to scattering and absorption by the cloud (keep in mind that these are "window" regions so that gaseous absorption is not a factor in these bands; typical aerosol extinction could perhaps reduce transmission by a few percent at most). This still leaves a sufficient amount of energy for most standard photodetectors, and, as figure 1 attests, the wavelength resolution required is modest. At the wavelengths which will be discussed here, emission of radiation by the clouds or the atmosphere is entirely negligible; for the same reason cryogenic detectors are unnecessary.

3. OBSERVATIONS

3.1 Reflected near-infrared from cumuli

The spectral behavior of absorption coefficient (figure 1) is seen when reflected sunlight from, say, a growing cumulus turret is spectrally analyzed by a suitable spectrometer, although maxima and minima are reversed since increasing signal means decreasing absorption. Such spectra are illustrated in figure 3; signal maxima for cloud "A" coincide accurately with absorption coefficient minima in the unbroken (liquid) curve of figure 1. (We can, in fact, reproduce figure 3 computationally using the data of figures 1 and 2 while adopting plausible values for the mean droplet size and therefore, not surprisingly, conclude that cloud "A" was composed of liquid water droplets.)

When we observe spectra from growing cumuli congesti in southern Arizona (usually summer monsoon phenomena), a sudden change is often observed which is especially marked around the 1.65 and $2.2 \mu\text{m}$ reflectance maxima. Cloud "B" in figure 3 illustrates such a transition of the $1.65 \mu\text{m}$ feature: the spectrum from the upper part of a growing cloud had varied little from "A" over some 20 minutes, but in the space of less than 4 minutes, changed to "B" and thereafter remained close to "B"; "A" and "B" are actually the same cloud.

In view of the discussion earlier, the relevant quantity is the product kr , and one might propose that

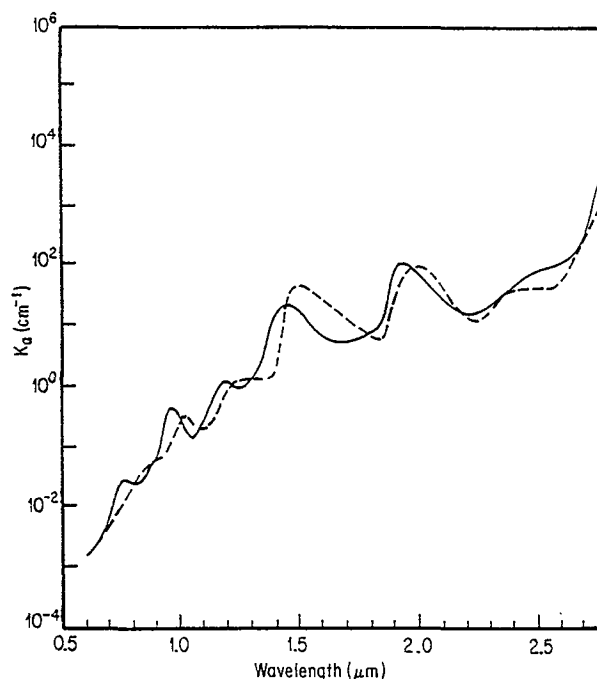


Figure 1. Bulk absorption coefficient for liquid water (solid curve) and ice (dashed curve). (Pilewskie and Twomey, 1987)

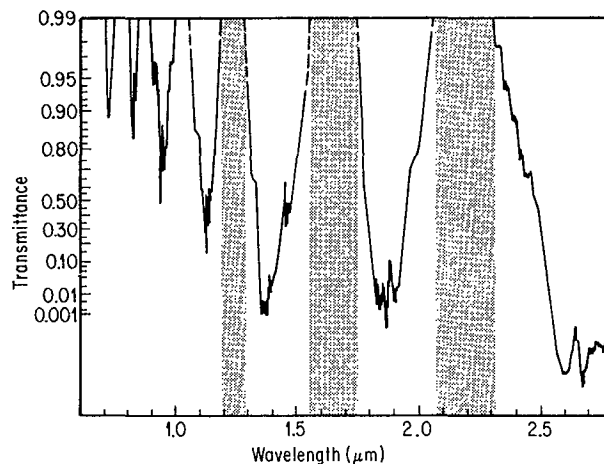


Figure 2. Water vapor (1 g cm^{-2}) transmittance in the visible and near-infrared. Window regions are shaded. (Pilewskie and Twomey, 1987)

the change from "A" to "B" was caused by a sudden increase in average droplet size, or by a change in absorption coefficient (i.e., change of water phase, from liquid to ice), or perhaps some combination of both.

Radiative transfer theory shows that the depression of diffuse reflectance below its value for conservative ($\sigma_0 = 1$, i.e., no absorption) conditions follows closely a simple proportionality in $\sqrt{1 - \sigma_0}$, i.e., \sqrt{kr} , under conditions of weak absorption. (Even though $k \approx 10 \text{ cm}^{-1}$ around $1.65 \mu\text{m}$, absorption is still "weak" since, for cloud drops with $r \sim 10 \mu\text{m}$, $kr \ll 1$.) The decrease in signal near $1.65 \mu\text{m}$ from "A" to "B" could be

accounted for if kr increased by a factor of about 4. From figure 1, near the $1.65 \mu\text{m}$ region, k for ice is seen to be about 3 to 5 times its value for liquid water, so most of the observed change can be accounted for by a liquid - solid transition, i.e., glaciation. If, on the other hand, k did not change, r would have to increase almost fourfold in a couple of minutes - a possibility that most cloud physicists would reject, especially in view of the steadiness of spectrum "A"-type behavior before the abrupt change.

An improved version of our instrument was operated during summer 1986. We consider its results to show quite unambiguously that larger water drops could no longer be proposed as an explanation of gross spectral changes like those occurring in the previous example, since, with improved spectral resolution and extended spectral range, we could observe, concurrent with an abrupt decrease in signal near $1.65 \mu\text{m}$, the wavelength shift alluded to earlier; furthermore, the marked decrease in signal near $1.65 \mu\text{m}$ was accompanied by a reduction in signal between $2.1 \mu\text{m}$ and $2.2 \mu\text{m}$ and a corresponding narrowing of the $2.2 \mu\text{m}$ peak reflectance band. That is just what the measured spectra show (figure 4, where the transition from a liquid water to ice cloud is represented by solid and broken curves): a distinct shift towards longer wavelength; a sharp reduction in signal around $1.65 \mu\text{m}$, and a smaller reduction between 2.1 and $2.2 \mu\text{m}$. (Note that, while figure 1 might suggest that, upon glaciation, there should be a distinct increase in signal at wavelengths less than $2 \mu\text{m}$, that would obtain only if no change in mean size accompanied the liquid-to-solid change. From a cloud physics viewpoint, that would appear unlikely.) Qualitative agreement of predictions based on the spectral data with what was measured therefore quite conclusively requires a phase change to have caused the "A" to "B" spectrum transition. (One might, in addition, infer an increase in the size of the scatterers, but that is another aspect which will not be pursued here.) The time between observation of a type-A spectrum to the first occurrence of a type-B spectrum was approximately 4 minutes.

3.2 Transmission measurements

The radiation transmitted diffusely through a cloud is influenced by weak absorption even more strongly than that which is reflected. Absorption bands, such as that near $1 \mu\text{m}$, are too weak to be useful in reflection, but they can produce sizable reductions in transmission, especially when the cloud is thick. While most of our recent work has concentrated on reflectance, we had the opportunity to participate in the final deployment of the Sierra Cooperative Pilot Project (SCCP) in winter 1986-1987, and there used our equipment in a vertically looking mode under stratiform cloud over the Sierra Nevada. The results were encouraging and lend some promise to the idea of fairly simple ground-based instruments for monitoring overlying clouds. Figure 5 shows transmission spectra recorded under stratocumulus clouds on 18 December 1986. The same spectral characteristics introduced in the previous section are used to categorize the clouds by phase.

Unlike reflected radiation, which has been scattered only by the outer few hundred meters of cloud, diffusely transmitted radiation has been scattered at all levels in the cloud, and a photon may encounter both liquid and solid scatters en route. A mixture of ice and liquid water in appreciable proportions obviously cannot be distinguished from a layer of ice particles overlying a layer of liquid drops, but such mixtures can be distinguished from a homogeneous cloud since the combined absorptions give a spectrum different from that of either purely liquid or purely ice. Spectral transmission

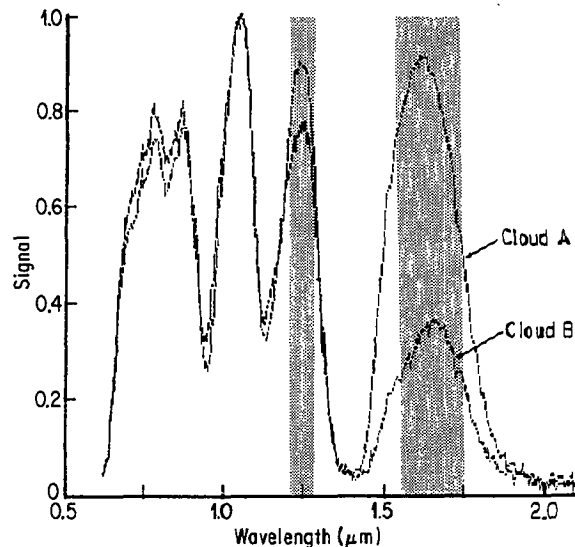


Figure 3. Reflectance spectra from an evolving cumulonimbus turret. The transition from "A" and "B" is typical of the change in reflectance signature following glaciation. The shaded areas indicate water vapor transparency. (Pilewskie and Twomey, 1987)

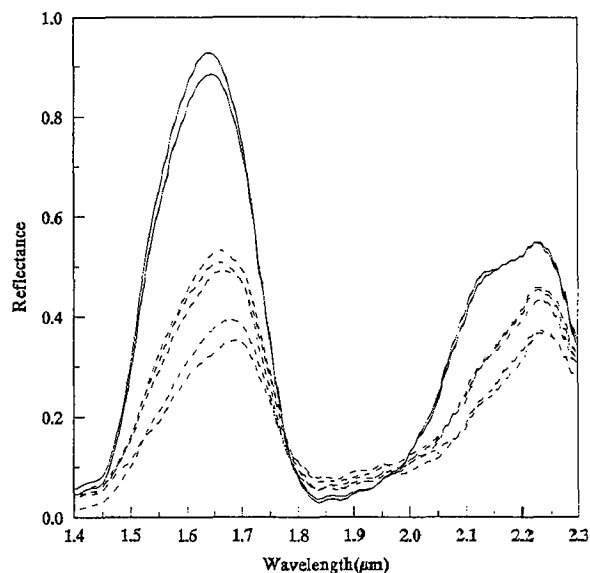


Figure 4. Reflectance from a rapidly evolving cumulonimbus with the extended spectral response out to $2.3 \mu\text{m}$. The reflectance fell with each successive spectrum. Spectrum acquisition time was 90 seconds; the entire set of data represents 18 minutes. Type "A" spectra (liquid) and type "B" (ice) are here represented by solid and dashed curves, respectively.

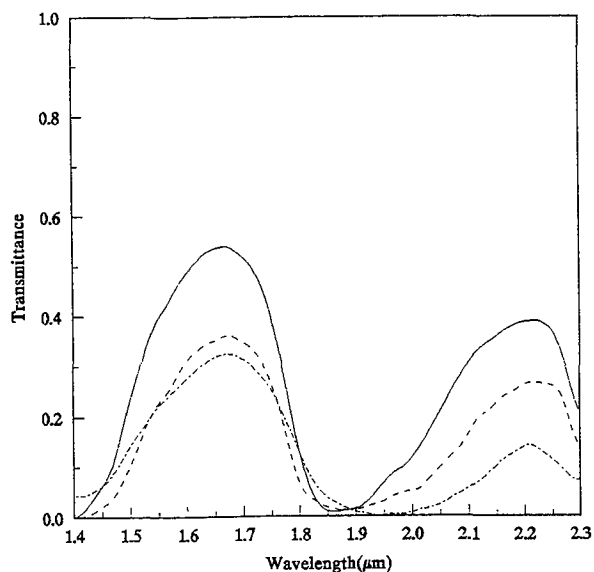


Figure 5. Transmission from stratocumulus. While the solid curve is indicative of a water cloud signature and the dash-dot curve (lowest) indicative of an ice cloud signature, the spectrum denoted by the dashed curve is at an intermediate level (in both signal level and spectral shape) and most likely results from a mixed phase cloud.

measurements at the ground could recognize glaciation of a portion of an overlying cloud, but could not determine where the ice was located; furthermore, the presence of ice particles in low concentrations would be difficult to detect under this method - roughly speaking, each phase contributes as the summed cross-sections of particles of that phase in the entire cloud, and, until it became appreciable in that sense, ice (or water, as the case may be) would not be noticed. We have labelled the spectra in figure 5 to indicate composition and the labelling should be interpreted in that spirit - "mixed", for example, would mean comparable amounts (summed cross-sections) of ice and liquid overhead. These could have existed at separate levels, not necessarily combined.

4. CONCLUDING REMARKS

The instruments employed for the measurements just described were research devices, but they were easily

moved, fairly robust, and relatively inexpensive. A reflecting telescope (amateur astronomer's type) was used for light collection and pointing. The detector was a germanium photodiode in the first version of the instrument and a cooled lead-sulfide photoconductor in the second. Since this method of phase discrimination required only a relative measurement of cloud reflectance or transmittance, only a wavelength response calibration was necessary. A more complete description of the instrument is given in Pilewski and Twomey (1987), and Pilewski (1989), or can be obtained from the authors.

It should be remarked that silicon detectors are of little value for reflectance measurements since the most interesting and informative wavelengths are beyond the response limits for silicon. However, in transmission, the weak absorption bands near $1 \mu\text{m}$ may be useful since their effect is magnified as previously stated. It also should be noted that no discriminating features near $1 \mu\text{m}$ were present in the transmission spectra acquired during SCCP. A similar ground-based instrument with improved spectral resolution will be deployed for cloud phase discrimination during the First International Satellite Cloud Climatology Project Regional Experiment (FIRE), Phase II, in Kansas in November 1991.

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REFERENCES

- Bigg, E.K., and R.T. Meade, 1971: High-level stratocumulus clouds. *Weather*, **26**, 55-57.
- Hale, M.H., and M.R. Querry, 1973: Optical constants of water and ice in the 200 nm to 200 μm region. *Appl. Opt.*, **12**, 2523-2549.
- Irvine, M.I., and J. Pollack, 1968: Infrared optical properties of water and ice spheres. *Icarus*, **8**, 324-360.
- Pilewski, P., 1989: Cloud phase discrimination by near-infrared remote sensing. Ph.D. dissertation. Univ. of Arizona. 110 pp.
- Pilewski, P., and S. Twomey, 1987: Discrimination of ice from water in clouds by optical remote sensing. *Atmos. Res.*, **21**, 113-122.
- Twomey, S., and C.F. Bohren, 1980: Simple approximations for calculations of absorption in clouds. *J. Atmos. Sci.*, **37**, 2086-2094.
- Warren, S.G., 1984: Optical constants of ice from the ultraviolet to the microwave. *Appl. Opt.*, **23**, 1206-1224.