Properties of PSCs and cirrus determined from AVHRR data

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Abstract. Polar stratospheric clouds (PSCs) and cirrus have been investigated using thermal emission measurements at 10.8 and 12 μm wavelength (channels 4 and 5) from the Advanced Very High Resolution Radiometer (AVHRR). The AVHRR signal was evaluated from a theoretical basis to understand the emission from clear and cloudy skies, and models were developed to simulate the AVHRR signal. Signal simulations revealed that nitric acid PSCs are invisible to AVHRR, while ice PSCs and cirrus are readily detectable. Methods were developed to retrieve cloud optical depths, average temperatures, average effective radii, and ice water paths, from AVHRR channels 4 and 5. Properties of ice PSCs retrieved from AVHRR were compared to values derived from coincident radiosondes and from the Polar Ozone and Aerosol Measurement II instrument, showing good agreement.

1. Introduction

This work explores the detection of polar stratospheric clouds (PSCs) and cirrus using measurements from the Advanced Very High Resolution Radiometer (AVHRR). Interests in cirrus include their role in radiative forcing of the climate system and their role in the chemical balance of the atmosphere [e.g., Sassen et al., 1989; Borrmann et al., 1996]. Because cirrus usually occur below the spontaneous freezing temperature of water, they are typically composed of ice crystals [Jensen et al., 1994]. PSCs have been implicated as key factors in the loss of ozone under cold polar conditions. Observational and theoretical evidence suggests that PSCs can be composed of either liquid ternary H₂SO₄–H₂O–HNO₃ aerosols (LTA), nitric acid trihydrate (NAT), or ice [e.g., Tolbert, 1996; Carslaw et al., 1994].
AVHRR measures reflected and/or emitted radiation at nadir (±50°) in five spectral bands between 0.58 and 12.5 μm wavelength (λ). A series of AVHRR sensors on board polar–orbiting satellites have been collecting data since 1979, and provide excellent temporal and spatial coverage of the polar regions. Poleward of 60°, each point on the earth’s surface is sampled in two or more consecutive passes twice a day for each of the two polar–orbiting satellites. These images can be combined to provide daily coverage of each polar region with horizontal resolution of 1.1 km at nadir. The spatial overlap of consecutive images allows for some tracking of cloud movement, although for a relatively short time period when compared to geostationary satellite data. AVHRR measurement characteristics, including the ability to measure with or without sunlight, offer unique advantages for PSC detection. Garcia, et al. [1995] and Pagan [1996] have shown that AVHRR measurements of 12 μm brightness temperature can be used to detect optically thick PSCs when underlying Earth surface temperatures are well above PSC temperatures. Most PSCs, however, are optically thin and occur over a variety of underlying surface conditions, and cannot be detected using a single AVHRR channel.

Cirrus and ice PSCs can have similar characteristics, so it is reasonable to expect that techniques established for cirrus characterization could be applied to PSC measurements. Inoue [1985, 1987] found that cirrus clouds can be identified over tropical oceans using the difference between channel 4 and 5 brightness temperatures (T4 and T5). This brightness temperature difference (BTD = T4 - T5) was then used by Yamanouchi et al. [1987] to detect thin cirrus over Antarctica. The radiative contrast between channels 4 and 5 has been investigated as a tool for inferring physical cloud properties such as temperature and particle effective radius [e.g., Wu, 1987; Parol et al., 1991; Lubin and Harper, 1996]. Parol et al. concluded that while cloud effective radii cannot be accurately determined, that useful limits can be placed on its range. Lubin and Harper inferred the temperature, ice water path, and effective radius of Antarctic cirrus, and suggest that these properties can be determined with some certainty for conditions relevant to cirrus and PSCs.

For this work, the AVHRR channel 4 and 5 radiances were evaluated from a theoretical basis to understand the emission from clear and cloudy skies. Methods were developed for retrieving
cloud extinction optical depths, average temperature \(T_c\), average effective radius \(r_e\), and ice water path (IWP), from the channel 4 and 5 radiances. These retrievals were applied to AVHRR cloud measurements for comparison to values derived from radiosonde measurements and to measurements from the Polar Ozone and Aerosol Measurement (POAM) II instrument.

2. Radiative Transfer and Signal Simulation

The nadir emission from a column of atmosphere containing cloud, \(L_m(\lambda)\), can be expressed in terms of net radiances \(L\) and transmissivities \(\tau\)

\[
L_m(\lambda) = \{ [L_E(\lambda) \tau_E(\lambda) + L_{At}(\lambda)] \tau_c(\lambda) + L_c(\lambda) \} \tau_u(\lambda) + L_Au(\lambda) \quad (1)
\]

where \(L_E(\lambda)\) is due to the Earth’s surface, \(\tau_E(\lambda)\) and \(L_{At}(\lambda)\) are due to clear air below the cloud layer, \(\tau_c(\lambda)\) and \(L_c(\lambda)\) are due to the cloud layer, and \(\tau_u(\lambda)\) and \(L_Au(\lambda)\) are due to clear air above the cloud. Each \(L\) and \(\tau\) is an integral in height and wavelength, and the expanded relationship is tedious compared to (1). The terms in (1) are now explored.

2.1. Earth’s Surface

Radiance emitted from the Earth’s surface can be expressed in terms of the surface emissivity \(\varepsilon_0(\lambda)\) and the blackbody radiance \(L^*(\lambda)\) at surface temperature \(T_E\), \(L_E(\lambda) = L^*(T_E, \lambda) \varepsilon_0(\lambda)\). For typical surface cover underlying PSCs (sea water, snow, and ice), values of \(\varepsilon_0(\lambda)\) are close to unity and are roughly constant in the 10 to 12 \(\mu\)m region (Table 1). Thus, the spectral content introduced to \(L_m(\lambda)\) by \(L_E(\lambda)\) is primarily that of the Planck function.

2.2 Clear Skies

Atmospheric absorption and emission were determined using rigorous line–by–line calculations [Gordley et al., 1994] for interfering gases in the channel 4 and 5 bandpasses. In order of radiative importance, the species considered were \(H_2O\), \(CO_2\), \(O_3\), \(HNO_3\), \(CCl_3F\), and \(CCl_2F_2\). Frequency integrals were convolved over the AVHRR spectral response functions to more accurately simulate an AVHRR measurement. The vertical integrals considered transfer from the Earth’s surface to altitudes where gaseous contributions become negligible (about 50 km). Clear
sky radiances were simulated for conditions ranging from polar winter to tropical. For this purpose, vertical profiles of temperature, pressure, and gas mixing ratios were taken from climatologies. These calculations show that the atmospheric contribution is determined mainly by H₂O and CO₂ in the lower troposphere, and that transfer above the tropopause can be safely neglected (i.e., τₜ(λ) = 1 and Lₐₜ(λ) = 0). Thus, the clear sky emission, Lₖₛ(λ), is determined by the Earth’s surface and radiative transfer in the troposphere, Lₖₛ(λ) = Lₑ(λ) τₑ(λ) + Lₐₑ(λ).

The clear sky results allow some simplifications of (1) which lead to a more general expression for cloudy sky emission. Lₖₛ(λ) could be measured from clear sky adjacent to a cloud, allowing the cloudy sky radiance to be expressed in terms of two measurements and two cloud terms. In a more general sense, the radiance illuminating a cloud from beneath may be affected by uncharacterized cloud layers at lower altitude. Considering that this background radiance, L₉(λ), can be characterized by AVHRR measurements away from a cloud of interest, the cloudy sky emission (1) is reduced to

\[ Lₘ(λ) = L₉(λ) τₑ(λ) + Lₑ(λ) \]

where L₉(λ) can be from a clear background, L₉(λ)=Lₑ(λ), or cloudy background. The form of (2) is similar to expressions derived by Parol et al. [1991] and Lubin and Harper [1996] for cloudy sky emission. Figure 1 depicts the concept of clear and cloudy sky signals adopted for this work. Potential limitations to this approach come quickly to mind. However, (2) eliminates the need to accurately characterize the Earth’s surface and atmosphere for input to rigorous radiative transfer calculations.

The dependence of clear sky radiance on telescope scan angle should be understood since values of Lₖₛ(λ) and Lₘ(λ) could be taken from measurements at different scan angles. Clear sky calculations for a range of scan angles considering tropical and polar conditions show that sensitivity to scan angle increases for optically thick (e.g., tropical) atmospheres. Transmissivity is noticeably reduced at large scan angles. However, emission increases in proportion to τₑ(λ) so that net radiance is relatively insensitive to scan angle (Figure 2).
Figure 1. A conceptual model of thermal emission from clear sky and one or two cloud layers.

Background: cirrus
\[ L_{ci}(\lambda) = L_{cs}(\lambda) \tau_{ci}(\lambda) + L_{ci}(\lambda) \]

PSC measured radiance:
\[ L_{m}(\lambda) = L_{bk}(\lambda) \tau_{c}(\lambda) + L_{c}(\lambda) \]

Background: clear sky
\[ L_{cs}(\lambda) = L_{E}(\lambda) \tau_{c}(\lambda) + L_{At}(\lambda) \]

Upper atmosphere:
transmissivity = 1
emission = 0

Cirrus, transmissivity = \( \tau_{ci}(\lambda) \)
emission = \( L_{ci}(\lambda) \)

Earth surface, emission: \( L_{E}(\lambda) = L^{*}(T_{E}, \lambda) \epsilon_{0}(\lambda) \)

Troposphere:
transmissivity = \( \tau_{t}(\lambda) \)
emission = \( L_{t}(\lambda) \)

Figure 2. The fraction of nadir (scan angle = 0°) radiance versus scan angle and transmissivity versus scan angle. The results are for channel 4 from line-by-line calculations considering O\(_3\), CO\(_2\), and H\(_2\)O, for clear skies. The state parameter and mixing ratio profiles were from climatologies.
Table 1. Various Properties Relevant to AVHRR Channels 4 and 5.

<table>
<thead>
<tr>
<th>Channel</th>
<th>Bandpass Wavelengths (μm): Lower Limit, Center, Upper Limit</th>
<th>Cloud Particle: Complex Refractive Index</th>
<th>Earth Surface: Emissivity</th>
</tr>
</thead>
<tbody>
<tr>
<td>4</td>
<td>10.3, 10.8, 11.3†</td>
<td>ice: 1.09 + 0.18i (W)</td>
<td>snow: 0.993 (M)</td>
</tr>
<tr>
<td></td>
<td></td>
<td>NAT: 1.63 + 0.13i (T)</td>
<td>sea ice: 0.985 (G)</td>
</tr>
<tr>
<td></td>
<td></td>
<td>LTA: 1.58 + 0.24i (H)*</td>
<td>sea water: 0.991 (G)</td>
</tr>
<tr>
<td>5</td>
<td>11.5, 12.0, 12.5†</td>
<td>ice: 1.28 + 0.41i (W)</td>
<td>snow: 0.972 (M)</td>
</tr>
<tr>
<td></td>
<td></td>
<td>NAT: 1.53 + 0.27i (T)</td>
<td>sea ice: 0.970 (G)</td>
</tr>
<tr>
<td></td>
<td></td>
<td>LTA: 1.59 + 0.27i (H)*</td>
<td>sea water: 0.986 (G)</td>
</tr>
</tbody>
</table>

The information was taken from Gao et al. [1994] (G), MOSDIS UCSB Emissivity Library, www.icess.ucsb.edu (M), Warren [1984] (W), Hervig et al. [1997] (H), and Toon et al. [1994] (T). Refractive indices and emissivities are at the band center wavelengths.

* LTA at 192 K composed of 10 wt.% H2SO4 – 35 wt.% HNO3.
† Values for the NOAA–12 sensor.

2.3. Cloud Optical Properties

The extinction coefficient ($\beta_e$) for a population of particles is calculated by integrating the extinction efficiency ($Q_e = \text{optical cross section} / \text{geometric cross section}$) over the size distribution:

$$\beta_e(\lambda) = \int_{r_1}^{r_2} \pi r^2 Q_e(r, \lambda, m) n(r) \, dr$$  \hspace{1cm} (3)

where $m$ is particle refractive index (Table 1), $n(r)$ is the particle size distribution, and $r_1$ and $r_2$ are the smallest and largest radii in the size distribution. The scattering coefficient ($\beta_s$) is calculated by substituting the scattering efficiency ($Q_s$) into (3), and the absorption coefficient ($\beta_a$) is the difference $\beta_e - \beta_s$.

The cloud particle size distribution can be described using a lognormal distribution for PSCs [Deshler et al., 1994] or for cirrus [Jensen et al., 1994]. While real cirrus or PSC size distributions can be quite different from lognormal, the lognormal form is useful for considering a broad range of conditions in modeling cloud radiative properties. The lognormal distribution is described by

$$n(r)dr = \sum_{i=1}^{k} N_i \frac{\exp[-\ln^2(r/\tilde{r}_i)/(2\ln^2 w_i)]}{r (2\pi)^{1/2} \ln(w_i)}$$  \hspace{1cm} (4)

where $n(r)$ is the differential concentration in the radius interval $dr$, $k$ is the number of modes in the distribution, $N_i$ is the number concentration, $w_i$ is the distribution width, and $\tilde{r}_i$ is the median radius,
for each of i modes. Cloud volume density \( (V, \text{ particle volume per unit volume of air}) \) and effective radius \( (r_e, \text{ surface area weighted mean radius}) \) can be derived from the size distribution. These important quantities are introduced here for convenience.

\[
    r_e = \bar{r} \exp\left[\frac{5}{2} (\ln 2w)\right] \tag{5}
\]

\[
    V = N \left(\frac{4}{3}\pi \bar{r}^3\right) \exp\left[8 (\ln 2w)\right] \tag{6}
\]

In this work, optical cross sections for (3) were determined using Mie theory, which explains the scattering and absorption by spheres. While cirrus and PSC particles are probably nonspherical, Mie theory can still provide reasonable estimates [e.g., Ebert and Curry, 1992]. To test this assumption, calculations based on Mie theory were compared to results from Fu and Liou [1993] (FL) who used ray tracing and spheroidal solutions to calculate extinction cross sections for randomly oriented hexagonal ice crystals. FL computed integral extinctions, \( \beta_e(\lambda) \), for 11 measured size distributions and fit the ratio \( \beta_e(\lambda)/V \) versus effective radius for numerous broad spectral bands across the visible and infrared. The bands used by FL did not match AVHRR channels 4 or 5, so one of their bands between 10.2 and 12.5 \( \mu \text{m} \) (central \( \lambda = 11.3 \mu \text{m} \)) was chosen for the comparisons. Using the Mie-lognormal model, \( \beta_e(11.3) \) and \( V \) were calculated for a range of lognormal distributions. The resulting volumes encompass the FL results (Figure 3a), so it should be possible to mimic their size distributions by carefully selecting lognormal parameters. Because volume and extinction are roughly proportional, we expect the ratio \( \beta_e(\lambda)/V \) to be somewhat independent of size distribution. To the extent that this is true, the comparison of \( \beta_e(11.3)/V \) is a good measure of the differences between Mie extinction and the more rigorous calculations of FL (Figure 3b). The Mie extinctions were scaled by the ratio of volume from FL over the lognormal volume, to account for size distribution differences. The scaled Mie extinctions are 10 to 30% greater than FL (Figure 3c). However, it is evident in Figure 3c that the Mie-lognormal model could exactly reproduce FL by adjusting the choice of lognormal size distributions. Since FL considered only 11 size distributions, there is no reason to expect that their results are more versatile than results based on a large family of lognormal size distributions. Thus, this work proceeds using the readily available Mie-lognormal model.
Cloud transmissivity is determined by radiative loss due to scattering and absorption (extinction) in the cloud layer,

$$\tau_c(\lambda) = \exp(-\beta_c(\lambda) \Delta z)$$  \hspace{1cm} (7)

where $\Delta z$ is geometric cloud thickness. Cloud emissivity is determined by a cloud’s absorptive properties,

$$\varepsilon_c(\lambda) = 1 - \exp(-\beta_a(\lambda) \Delta z)$$  \hspace{1cm} (8)

and cloud radiance is determined by the cloud temperature ($T_c$) and emissivity,

$$L_c(\lambda) = L^*(T_c, \lambda) \varepsilon_c(\lambda)$$  \hspace{1cm} (9)

where $L^*(T_c, \lambda)$ is the blackbody radiance. Optical depth, $\sigma_c(\lambda) = \beta_c(\lambda) \Delta z$ or $\sigma_a(\lambda) = \beta_a(\lambda) \Delta z$, can be substituted into the above relationships. The cloudy sky emission (2) can be restated in terms of a single optical depth by defining the ratio $R_{ae}(\lambda) = \sigma_a(\lambda)/\sigma_c(\lambda)$ and substituting (7), (8), and (9).

$$L_m(\lambda) = L_{bk}(\lambda) \exp[-\sigma_c(\lambda)] + L^*(T_c, \lambda) \{ 1 - \exp[-\sigma_c(\lambda) R_{ae}(\lambda)] \}$$  \hspace{1cm} (10)

If $R_{ae}(\lambda)$ can be treated as constant, (10) defines a closed system, relating two measurements and two unknowns. Calculated values of $R_{ae}(\lambda)$ are bounded between 0.45 and 1.0 for ice refractive
indices and a range of size distributions representative of PSCs or cirrus (Figure 4a). It can also be shown that $R_{ae}(10.8)$ and $R_{ae}(12)$ are strongly correlated for ice indices and a wide range of size distributions (Figure 4b). While fixing $R_{ae}(\lambda)$ may or may not prove acceptable, its value is bounded and its importance therefore easily understood.

Figure 4. The data shown here were computed using lognormal size distributions (w from 1.5 to 2.5 and f from 0.1 to 87.1 \(\mu m\)) and Mie theory with ice refractive indices. a) The ratio of absorption optical depth to extinction optical depth, $R_{ae}(\lambda)$, versus effective radius for channels 4 and 5. b) The relationship between $R_{ae}(10.8)$ and $R_{ae}(12)$, which was fit according to: $R_{ae}(12) = 0.396 - 0.287 R_{ae}(10.8) + 0.878 R_{ae}(10.8)\times10^{-2}$

The Mie-lognormal model was challenged to reproduce relationships between $T_4$ and BTD measured by AVHRR. For these simulations, it was necessary to vary optical depth to reproduce a measured range of brightness temperature. This can be accomplished by fixing number concentration and cloud thickness ($N\Delta z$) and varying $r_e$ or by fixing $r_e$ and varying $N\Delta z$ (equations 3, 4, and 10). While results from both approaches can be similar, fixing $N\Delta z$ and varying $r_e$ usually gives the most convincing reproduction of the measured $T_4$ versus BTD. This is consistent with the understanding of crystal population growth, where particle concentration becomes stable shortly after cloud nucleation and particle size continues to increase as a cloud matures. While this generalization may not apply to a broad region of cloud, it is probably a good assumption for small cloud segments. In general the AVHRR measurements can be explained using the Mie-lognormal model, and an example comparison is offered in Figure 5.
Figure 5. Channel 4 brightness temperature versus BTD from AVHRR cloud measurements at 5:50 UT on October 7, 1993 over the Weddell Sea (roughly 70°S, 335°E). 855 four km resolution AVHRR pixels were used (an area of roughly $120 \times 120$ km). $T_{bk}$ was taken from clear air adjacent to the cloud and $T_c$ was retrieved from AVHRR, as described below. The AVHRR data are compared to model calculations considering two values of NAz and a range of size distributions ($w=2$ and $r$ from 0.1 to 87.1 μm).

3. Predictions for PSCs at Equilibrium

It is useful to estimate optical depths and AVHRR signals expected for typical clouds. Given a pressure, temperature, and gas mixing ratios of H$_2$O ($Q_w$) and/or HNO$_3$ ($Q_n$), the volume densities of stratospheric ice, NAT, and LTA can be calculated from theory by assuming that vapor in excess of equilibrium condenses onto cloud particles [e.g., Koop et al., 1997]. This treatment is valid because the stability of the stratosphere ensures that vapor loss to cloud growth is localized and therefore limited, on the time scale of a cooling event. This approach is invalid in the troposphere in principal because a cloud's vapor supply can be continuous. Infrared extinctions are strongly correlated to $V$, which allows the estimation of optical depths and therefore AVHRR signals. Figure 6a shows an example of the correlation between $\beta_{c}(\lambda)$ and $V$ determined from theory. These data were fit according to

$$\log_{10}(\beta_{c}(\lambda)) = A + B \log_{10}(V) + C \log_{10}(V)^2$$

where $A$, $B$, and $C$ are constants. Constants for (11) are given in Table 2 for ice and NAT and $\lambda = 10.8$ and 12.0 μm. In a different approach, the ratio $\beta_{c}(\lambda)/V$ can be determined versus effective radius (Figure 6b). Either approach is valid, and allows conversion from $V$ to $\beta_{c}(\lambda)$.
Figure 6. The data shown here were computed using lognormal size distributions (w from 1.5 to 2.5 and r from 0.1 to 87.1 μm) and Mie theory with ice refractive indices. a) Volume versus extinction for λ = 12.0 μm and a fit to these data (equation 11). b) The ratio of extinction over volume versus effective radius for λ = 10.8 and 12.0 μm.

Table 2. Constants for Equation 11.*

<table>
<thead>
<tr>
<th>Wavelength (μm), particle type</th>
<th>A</th>
<th>B</th>
<th>C</th>
</tr>
</thead>
<tbody>
<tr>
<td>10.8, NAT</td>
<td>-3.48</td>
<td>1.11</td>
<td>-0.054</td>
</tr>
<tr>
<td>10.8, ice</td>
<td>-3.56</td>
<td>1.00</td>
<td>-0.035</td>
</tr>
<tr>
<td>12.0, NAT</td>
<td>-3.40</td>
<td>1.04</td>
<td>-0.045</td>
</tr>
<tr>
<td>12.0, ice</td>
<td>-3.30</td>
<td>0.96</td>
<td>-0.035</td>
</tr>
</tbody>
</table>

*For βe(λ) in km⁻¹ and V in μm³ cm⁻³.

Volumes corresponding to a range of stratospheric temperatures and pressures were computed for upper limit mixing ratios of stratospheric H₂O (5 ppmv) [Abbas et al., 1996] and HNO₃ (15 ppbv) [Roche et al., 1993]. Peak NAT volumes (and therefore optical depths) are roughly three orders of magnitude less than peak ice volumes (Figure 7). This difference is due to the fixed stoichiometry of NAT so that HNO₃ becomes the limiting reagent for particle growth. Extinctions were determined from V, and optical depths were determined for fixed cloud thickness (Figure 7b). The optical depths were used with (10) to calculate the change in background sky brightness temperature (T_{bk}(λ) = L*(T_{bk}, λ)) when clouds are introduced, T_{bk} - T_m(λ), and associated cloudy air BTDs (BTD = T_4 - T_5) for assumed values of T_{bk} and R_{ae}(λ). The thickest stratospheric NAT cloud (σ_e(λ) less than about 10⁻²) changes the background brightness temperature by less than 0.1 K and NAT PSCs can therefore be considered invisible to AVHRR. This conclusion holds for a
Figure 7. a) Equilibrium volumes for ice and NAT determined using 5 ppmv H2O and 15 ppbv HNO3 (the V = 1000 μm³ cm⁻³ contour corresponds to IWP = 0.917 g m⁻², for Δz = 1 km). b) Optical depths determined from the volumes in Figure 7a for Δz = 1 km, using the value of βₜ(12)/V from Figure 6b for rₑ = 1 μm for ice. A similar curve was used for NAT. The NAT contours were terminated 3 K below the ice point temperature in Figures 7a and 7b. c) BTDs and the difference Tₑk - Tₑm(12) determined for the ice volumes in Figure 7a using (10) with Δz = 1 km, Tₑk = 250 K, Rₑ(λ) = 0.6, and values of βₑ(λ)/V from Figure 6b for rₑ = 1 μm.

broad range of conditions (Rₑ(λ), Tₑk, and Tₑm), and the same conclusion is drawn for PSCs composed of LTA. Considering that the stratospheric conditions in Figure 7 are generous in terms of pressure and mixing ratio (in reality HNO3 disappears near the tropopause), this conclusion is well founded and the remainder of this study proceeds without consideration of NAT or LTA. The thickest ice PSCs change Tₑk by up to 16 K and produce BTDs of up to 14 K, for the scenario presented in Figure 7c. Tₑk - Tₑm is less sensitive to the shape of the size distribution than BTD, and the estimates of BTD in Figure 7c can change notably for using different size distributions, due to the dependence of Rₑ(λ) on particle size (Figure 4a). Calculations considering tropospheric
conditions produce a broad range of optical depths and BTDs, which include stratospheric values. Thus, a threshold optical depth could be determined which precludes a cloud from being stratospheric. However, a threshold which precludes a cloud from being tropospheric is more elusive.

It is useful to consider the relationship between cloud optical depth, cloud temperature, and brightness temperature. Figure 8 shows example calculations for a single cloud layer in an otherwise clear atmosphere. In order for a cloud to radiate at its physical temperature \( T_m(\lambda) = T_c \), its optical depth must be greater than about one. Optical depths which produce \( T_m(\lambda) = T_c \) decrease with decreasing \( T_{bk} \) (or increasing \( T_c \)), although they are often near the upper limit for ice PSC optical depth (Figure 7b).

![Figure 8](image)

**Figure 8.** Channel 5 cloudy sky brightness temperatures computed from (10) for \( R_{se}(12) = 0.6 \) and two values of \( T_{bk} \), contoured for a range of cloud temperatures and optical depths. The lower threshold in optical depth where cloudy sky brightness temperature equals physical cloud temperature \( T_m(12) = T_c \) is indicated.

4. Cloud Property Retrievals

4.1. Cloud Temperature and Optical Depth

Given cloudy and background sky measurements from both channels, it is possible to retrieve \( T_c \), \( \sigma_c(10.8) \), and \( \sigma_c(12.0) \) simultaneously. This is accomplished by solving (10) for the blackbody cloud radiance \( L^*(T_c, \lambda) \).
and finding $T_c$ through the Planck function. Cloud temperatures can be determined from (12) for a range of synthetic optical depths and fixed $R_{ae}(\lambda)$. This step is performed for each channel and a solution is taken where both channels yield the same cloud temperature. Solutions are not unique, yet the range of possible solutions usually contains the exact solution. It is difficult, however, to discern the correct solution from this family of possibilities. One approach is to find the average temperature, and use the corresponding optical depths. The retrieval approach is demonstrated in Figure 9a, where cloudy sky radiances were simulated for known conditions and (12) was solved for both channels to determine $T_c$, $\sigma_e(10.8)$, and $\sigma_e(12.0)$. While an average solution can be a good approximation in some cases, it can be biased by over 10 K in others. The assumption is now made that horizontal variations in cloud radiance over small distances are due to changes in optical depth, while temperature remains fairly constant. This is an important characteristic because for cloud radiances corresponding to a range of $\sigma_e$ at constant $T_c$ the families of solutions tend to overlap around the exact value of $T_c$. This behavior is demonstrated in Figure 9b, where the potential improvements for considering a range of optical depth are apparent. The errors for retrievals based

\[
L \cdot (T_c, \lambda) = \frac{L_0(\lambda) - L_{th}(\lambda) \exp[-\sigma_r(\lambda)]}{1 - \exp[-\sigma_r(\lambda)R_{ae}(\lambda)]} \tag{12}
\]
on a single measurement are greatest at low to intermediate optical depths, while $T_c$ can be accurately determined for optically thick clouds (Figure 9b). Uncertainties in retrieved $T_c$ and $\sigma_e$ were determined for a range of $T_c$ and optical depth ratio, $R_{54} = \sigma_e(12.0) / \sigma_e(10.8)$. In general, $R_{54}$ ranges from two for $r_e = 0.1 \mu m$ to unity for $r_e = 100 \mu m$. The fundamental uncertainties in retrieved $T_c$ and $\sigma_e$ are systematic (Figure 10) for a chosen scenario of cloud properties. Of course, since the cloud properties are unknown for a real measurement, the uncertainties will take on a more random nature.

**Figure 10.** Cloud temperatures and extinction optical depths were retrieved from simulated radiances based on four cloud optical depths ($\sigma_e(10.8) = 0.5, 1.0, 1.5, 2.0$ and $\sigma_e(12.0) = R_{54} \sigma_e(10.8)$) at constant $T_c$. Differences between the retrieved and exact values are shown for $T_c$ and $\sigma_e(10.8)$, for a range of $T_c$ and optical depth ratio. The differences in $\sigma_e(10.8)$ are an average for the four optical depths. The radiance simulations used $T_{bk} = 250 K$, $R_{ae}(12) = 0.75$, and $R_{ae}(10.8) = 0.82$.

### 4.2. Effective Radius and Ice Water Path

As illustrated in Figure 6, the ratio $\beta_e(\lambda)/V$ can be a unique function in effective radius. Furthermore, it can be shown that the two functions, $f(r_e, 10.8) = V/\beta_e(10.8)$ and $f(r_e, 12.0) = V/\beta_e(12.0)$, are monotonic and non–intersecting for roughly $2 < r_e < 100 \mu m$. Curves for $f(r_e, \lambda)$ can be generated using a broad range of size distributions and thus $r_e$. Scaling $f(r_e, \lambda)$ by a measured optical depth yields a relationship between $r_e$ and $V\Delta z$, $f_s(r_e, \lambda) = f(r_e, \lambda) \sigma_e(\lambda) = V\Delta z$. Because the atmospheric values of $V\Delta z$ and $r_e$ will be unique for a given measurement, a broad range of size distributions can be searched for one that yields the same $r_e$ and $V\Delta z$ from both channels, $f_s(r_e, 10.8) = f_s(r_e, 12.0)$. 

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where the optical depths are from AVHRR. For illustration, \( f(r_e, 10.8) \) and \( f(r_e, 12.0) \) were calculated using Mie theory with lognormal size distributions. These curves were scaled by realistic channel 4 and 5 optical depths, and solutions for \( r_e \) and \( V\Delta z \) are found at their intersection (Figure 11). The ice water path is easily calculated, \( \text{IWP} = V\Delta z q_i \). While size distributions that satisfy (13) can vary, the integral quantities tend to fall in a narrower range. Never the less, the curves describing \( f_s(r_e, \lambda) \) tend to have steep intersections, so that solutions become highly sensitive to errors in optical depth.

\[
\frac{V(r_e)}{\beta(r_e)} \frac{\Delta z}{\Delta z} \sigma_e(12.0) = \frac{V(r_e)}{\beta(10.8)} \frac{\Delta z}{\Delta z} \sigma_e(10.8)
\]

(13)

To arrive at the curves in Figure 11, the size distribution width was fixed and the median radius was varied. This scenario yields one solution. However, a variety of realistic scenarios can yield solutions. Sensitivity of \( r_e \) and \( V\Delta z \) to the choice of size distribution scenario was evaluated by generating optical depths and corresponding values of \( r_e \) and \( V\Delta z \) for a broad range of size distributions. These synthetic optical depths were then used with the curves in Figure 11 to retrieve \( r_e \) and \( V\Delta z \), for comparison to the known values. For each value of \( r_e \), the uncertainty analysis used a Monte Carlo approach considering randomly chosen distributions which satisfy (5). Random and systematic variability about a known solution were then calculated from this sample set, to determine the effects of searching a limited set of size distributions for a solution. The effects of 10% random errors in the retrieved optical depths were then added to the size distribution errors for these calculations. Random uncertainties in \( r_e \) due to uncertainties in the size distribution are less...
than 30% for the range of $r_e$ considered (Figure 12a). When errors in retrieved optical depth are added, random uncertainties in $r_e$ increase to over 50%, and significant retrieval bias appears for $r_e < 3 \mu m$ and $r_e > 30 \mu m$ (Figure 12a). Uncertainties in $r_e$ and $V\Delta z$ due to size distribution and optical depth errors are summarized in Figure 12b, showing nearly 100% uncertainty in both for some size regimes. While $r_e$ and $V\Delta z$ cannot be accurately determined, limits can be placed on their possible range, as concluded by Parol et al. [1991]. These large uncertainties may be tolerable for some applications, and averaging retrieved values over many pixels may provide some reduction in uncertainty.

![Figure 12](image_url)

**Figure 12.** a) Random and systematic uncertainties in $r_e$ determined using Monte Carlo calculations considering errors in the size distribution, $n(r)$ ($w$ varied randomly between 1 and 3 with the corresponding $r$ determined by (5)), and errors in both $n(r)$ and 10% errors in the optical depths. b) The nominal values of $r_e$ and $V\Delta z$ (heavy solid line) and random uncertainties (due to both $n(r)$ and optical depth) indicated as vertical bars and lighter shading for $V\Delta z$ and horizontal bars and darker shading for $r_e$.

5. Results Using AVHRR

Cloud properties retrieved from AVHRR were compared to values derived using measurements from radiosondes and from POAM II. For this purpose, raw AVHRR data were calibrated and geo-registered at 4km horizontal resolution [e.g., Lubin and Harper, 1996].

As previously discussed, equilibrium theory allows stratospheric ice clouds to be characterized from temperature, pressure, and water vapor mixing ratio ($Q_w$). While radiosondes provide reliable temperature and pressure measurements, they cannot measure $Q_w$ in the
stratosphere. Stratospheric water vapor, however, is fairly uniform and has a well known upper limit near ~5 ppmv [e.g., Abbas et al., 1996]. Assuming constant $Q_w$, stratospheric radiosonde profiles were used to calculate $V$ and $\beta_c(\lambda)$ (as in Figure 7), which were integrated vertically to yield $IWP$ and $\sigma_c(\lambda)$. $T_c$ was then computed from a volume weighted mean for all cloud layers. The abundance of radiosonde measurements and the straightforward analysis provide a useful approach for validating retrievals of $T_c$, $\sigma_c(\lambda)$, and $IWP$ from AVHRR ice PSC measurements. An example radiosonde analysis is shown in Figure 13, where two values of constant $Q_w$ were considered. In this example, changing $Q_w$ from 5 to 3 ppmv reduces $\sigma_c(\lambda)$ and $IWP$ by a factor of four, while $T_c$ is relatively unaffected. Despite sensitivity to $Q_w$, radiosonde $IWP$ and $\sigma_c(\lambda)$ provide reliable upper limits, and therefore useful validation data.

AVHRR cloud temperatures were also compared to solar occultation measurements from POAM II [Randall et al., 1996]. Ice PSCs are identified in POAM II aerosol extinction profiles when the signal saturates in the stratosphere, temperatures are below $T_{ice}$, and the termination altitude is above the coldest layer in the profile [Fromm et al., 1999]. When these criteria are met, PSC tops are estimated to occur from 1 to 2 km below the profile termination altitude, and PSC temperature can be estimated from the corresponding temperature profile.
A case study of clouds over Scoresbysund, Greenland beginning on February 19, 1996 was used to test the AVHRR retrievals by comparing them with radiosondes and POAM II. AVHRR imagery on this day show clouds with brightness temperatures ($T_b$) between 189 and 200 K over much of Greenland. A cloudy area within roughly 200 km of Scoresbysund was selected from each AVHRR image for the cloud property retrievals. Each AVHRR cloud area was comprised of about 100 to 300 pixels (an area of roughly 1600 to 4900 km$^2$), and the cloud temperature retrievals used all pixels in an area to ensure variability in optical depth for reduced retrieval errors (e.g. Figure 9b). Values of $r_e$, $\sigma_0(\lambda)$, and IWP are presented as averages of all retrievals in a cloud area. Time series from AVHRR and the radiosondes are compared in Figure 14. These time series depict the passage of various PSCs and cirrus over Scoresbysund, rather than the evolution of a single stationary cloud.

The AVHRR cloud temperature at 3 UT on the 19th is too warm to be stratospheric, even when
considering a possible 10 K overestimation at low optical depth (e.g., Figure 9). The other AVHRR
cloud temperatures are consistent with PSCs, and are in good agreement with coincident POAM
II and radiosonde data. The AVHRR retrievals of $r_e$, $\sigma_e(\lambda)$, and IWP are consistent with PSC
conditions, yet they do not exclude the possibility of cirrus. Although AVHRR $\sigma_e(\lambda)$ and IWP
retrievals tend to be larger than the radiosonde values, the two data sets agree within the mutual error
limits.

Lubin and Harper [1996] suggested that BTDs greater than about 5 K may be characteristic
of ice PSCs. Their idea was tested in this work using the relationship between measured BTD and
retrieved cloud temperature from the AVHRR measurements over Scoresbysund. For these
measurements, stratospheric frost point temperatures ($T_{ice}$) were always less than about 193 K (e.g.
Figure 13). Thus the Scoresbysund AVHRR data suggest that ice PSCs are associated with BTDs
greater than about 5 K (Figure 15), consistent with the suggestion of Lubin and Harper. The use
of BTD as a PSC identifier may involve more than simply defining a threshold value, and this idea
is currently the subject of rigorous evaluations based on model calculations and observations.

![Figure 15. Mean BTD versus retrieved cloud temperature for AVHRR measurements over Scoresbysund on February 19, 1996. Mean BTDs were computed within each cloud area and vertical lines show the range of BTD within an area.](image)

6. Summary

AVHRR channel 4 and 5 radiances from clear and cloudy skies were evaluated from a
theoretical basis and techniques were presented for inferring cloud optical depth at $\lambda=10.8$ and 12
$\mu m$, average temperature, average effective radius, and ice water path, for ice PSCs or cirrus. The
cloud optical depth and temperature retrievals assume that upwelling radiation beneath a cloud layer can be characterized from adjacent measurements, and that radiance variations across small cloud segments are due to variations in optical depth at constant average cloud temperature. The retrieval of ice water path and average effective radius requires signal simulations, which were based on Mie theory and lognormal size distributions. While the retrieval assumptions and signal simulations can have obvious limitations, they are justifiable for a useful portion of the AVHRR data set. Initial assessments show that AVHRR cloudy sky measurements can be explained in terms of the Mie-lognormal model. Additionally, cloud temperatures, ice water paths, and optical depths retrieved from AVHRR are in good agreement with values derived from coincident radiosonde measurements.

While large values of optical depth, effective radius, and ice water path can preclude the possibility of PSCs, small values do not bar the possibility of cirrus. Thus retrieved cloud temperatures should provide the most useful information for separating PSCs and cirrus in AVHRR data. The retrieval methods described in this paper require a considerable amount of image interpretation to determine an accurate underlying background temperature as input to the calculations. While this approach is useful for specific case studies, in its current form it would be difficult to employ on the entire AVHRR data set. There is compelling evidence, however, that BTDs could be a straightforward PSC identifier which can be easily applied to the extensive AVHRR data set, allowing the development of first-ever PSC climatologies in both hemispheres.

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