Wide-Area Determination of Cloud Microphysical Properties from NOAA AVHRR Measurements for FIRE and ASTEX Regions

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ABSTRACT

A method for satellite remote sensing of cloud optical thickness and effective particle radius has been developed to apply to NOAA AVHRR multispectral radiance data. Undesirable radiation components such as ground-reflected solar radiation and thermal radiation are guessed from satellite-received radiances in channels 1, 3, and 4 of AVHRR and subtracted from radiances in channels 1 and 3 to derive the reflected solar radiation of a cloud layer that includes information about cloud microphysical properties. This method can be applied to a broad range of water clouds from semitransparent to thick clouds.

This method was applied to AVHRR data acquired over oceans during the First ISCCP Regional Experiment and the Atlantic Stratocumulus Transition Experiment. The authors found good agreement between satellite-derived and in situ microphysical quantities. The presence of drizzle droplets in optically thin clouds was also confirmed from the satellite observation. Furthermore, the results show that marine stratocumulus clouds were drastically modified by ship track effluents and dust-contaminated airflow from the continent.

1. Introduction

Cloud feedback mechanisms are recognized as a major source of uncertainty in the current assessment of global climate change. Planetary boundary layer clouds of the marine subtropics affect the climate through coupling of radiative, microphysical, and convective processes with largely different timescales (Betts and Ridgway 1989). Somerville and Remer (1984) have used a radiative-convective equilibrium model to study cloud optical thickness feedback and concluded that the sign of the feedback is negative; hence, clouds can act as a thermostat and change in such a way as to reduce the surface and tropospheric warming caused by the CO₂ increase. Using recent general circulation models (GCMs), Roeckner et al. (1987) and Mitchell et al. (1989) also concluded that a change in cloud optical properties may result in a negative feedback comparable in size to the positive feedback associated with a change in the cloud cover. On the other hand, Wetherald and Manabe (1988) were led to the conclusion that the contribution of the negative feedback process of clouds is much smaller than the effect of the positive feedback process induced by reduced cloud amount in the upper troposphere and increased cloudiness around the tropopause. In order to improve the treatment of clouds in GCM simulations of climate change, observations of global cloud radiative and microphysical properties are very important. The International Satellite Cloud Climatology Project (ISCCP) was initiated to develop and coordinate basic research on techniques for measuring the physical properties of clouds by satellite remote sensing and to apply the resulting techniques to derive a global cloud climatology for improving the parameterization of clouds in climate models (Schiffer and Rossow 1983).

An important possibility that increasing anthropogenic aerosols may change the cloud radiative properties has been pointed out recently by many researchers (e.g., Coakley et al. 1987; Radke et al. 1989). Twomey et al. (1984) studied the effect of anthropogenic aerosols of continental origin on the cloud albedo and concluded that the loss of sunlight may compensate for the expected warming by CO₂. An interesting example of the effect of anthropogenic cloud condensation nuclei (CCN) on cloud reflectivity is the so-called ship track phenomenon caused by seeding of CCN from ships. Ship tracks appear frequently in satellite imagery as a signature of modification of low-lying stratus and stratocumulus clouds (Radke et al. 1989). Durkee et al. (1986) and Kaufman and Nakajima (1993) noted that the simultaneous study of both cloud and aerosol properties is important for better understanding of cloud–aerosol interaction processes.

Stimulated by the importance of cloud microphysical change problems as discussed above, there has been an
increasing number of studies pertaining to the retrieval
theories and methods for obtaining cloud optical thickness
and particle size from multispectral radiometers on
aircraft (Hansen and Pollack 1970; Twomey and
Cocks 1982; King 1987; Foot 1988; Rawlins and Foot
1990; Nakajima et al. 1991) and on satellites (Curran
and Wu 1982; Arking and Childs 1985; Rossow et al.
1989; Kaufman and Nakajima 1993; Platnick and
Twomey 1994; Han et al. 1994). The soundness of
aircraft remote sensing retrievals of cloud microphysical
properties has been extensively studied by
comparisons between in situ and remote-sensing-derived
values. We need more effort, on the other hand, to es-
tablish the satellite remote sensing method since there
are many sources of uncertainty in the method, such as
removal of undesirable radiation components in the sat-
ellite-received spectral radiance and methods of assum-
ing the model atmosphere, which we need to simulate
the satellite-received radiances. There are not many
comparisons of satellite remote sensing results with in
situ measurements to study these problems. There also
is room for developing techniques to improve the per-
formance of the method. Han et al. (1994), for ex-
ample, have surveyed the effective radius for a near-
global area by using a nadir remote sensing method, in
which the reflected radiation is independent of azi-
muthal angle and the effect of non-plane-parallel fea-
tures of clouds may be small, making use of the ISCCP
CX dataset that prepares ISCCP-derived cloud optical
thickness. It will be useful to study merit of off-nadir
analyses.

The principal intent of this paper, therefore, is to
suggest an efficient method for satellite remote sensing
retrievals of cloud optical thickness and effective par-
ticle radius and to show some applications. The present
method can be applied to an area as large in size as the
field of view of the NOAA polar orbital satellite. We
will show results of the wide-area determination of mi-
crophysical properties over the west coast of California
in the period of the First ISCCP Regional Experiment
(FIRE) and over the area of the Madeira Islands during
the period of the Atlantic Stratocumulus Transition Ex-
periment (ASTEX). The possibility of significant mod-
ification of clouds not only by artificial aerosols but
also nonartificial aerosols (e.g., continental dust) will
be discussed.

2. Retrieval method of cloud microphysical
properties

a. Concept

The solar reflectance method uses nonabsorbing vis-
ible and water-absorbing near-infrared wavelengths,
such as 1.6, 2.2, and 3.7 \( \mu \)m, for simultaneous retrievals
of cloud optical thickness and effective particle radius.
In this paper we mainly discuss the solar reflectance
method making use of channels 1 (0.64 \( \mu \)m), 3 (3.75
\( \mu \)m), and 4 (11 \( \mu \)m) of the Advanced Very High Res-
olution Radiometer (AVHRR) aboard NOAA polar or-
biter.

\[
r_c = \frac{\int_0^\infty r^2 n(r)dr}{\int_0^\infty r^2 n(r)dr},
\]

where \( n(r) \) is the number size distribution as a function
of particle radius \( r \). We used a lognormal size distri-
bution in the calculation:

\[
n(r) = \frac{N}{\sqrt{2\pi} \sigma} \exp \left[ -\frac{(\ln r - \ln r_0)^2}{2\sigma^2} \right],
\]

where \( r_0 \) is the mode radius, related to the effective
radius as \( r_c = r_0 e^{3.5\sigma^2} \), and \( \sigma \) is the log standar-
d deviation of the size distribution; \( \sigma = 0.35 \) was
assumed for marine stra-
tocumulus clouds in our analyses. For the satellite signal
simulation we used an accurate and efficient radiative
transfer scheme (Nakajima and Tanaka 1986, 1988;
Nakajima and King 1992) extended to include thermal ra-
diative transfer as proposed by Stamnes et al. (1988).
In the radiative transfer calculation we used the \( k \) dis-
tribution of absorption coefficient from LOWTRAN-7
(Kneizys et al. 1988) for a gas absorption model. The mid-
latitude summer atmosphere model was divided into four plane-

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FIG. 1. Simulation of reflected solar radiances in AVHRR channels
1 and 3 as a function of cloud optical thickness \( (r_c = 1, 2, 4, 8, 16,
32, \text{ and } 64) \) and effective radius \( (r_c = 2, 4, 8, 16, \text{ and } 32) \) with
the condition of \( \theta = 40^\circ, \theta_s = 60^\circ, \phi = 50^\circ \). Near-vertical and near-
horizontal lines illustrate iso-optical thickness and isoeffective radius
radiances, respectively. Ground-reflected and thermal radiances are
not taken into account in this result. LOWTRAN-7 and the U.S.
standard model were assumed to model gaseous absorption.
parallel layers with interfaces at 1, 2, and 12 km and the top at 120 km in this example. The cloud is inserted into the third layer from the top of the model atmosphere with saturated humidity in the cloud layer. As for the underlying surface, we assumed a Lambert surface. This assumption will not introduce a significant error in the analyses if we use an equivalent flux albedo as suggested by Nakajima et al. (1991) for cloudy atmospheres.

We recognize in Fig. 1 that the reflected solar radiation in channel 1 depends primarily on the cloud optical thickness whereas the reflected radiance in channel 3 depends primarily on the effective radius. This can be explained by the difference in the magnitude of the imaginary index of refraction of liquid water in channel 1 ($\sim 10^{-8}$) and channel 3 ($\sim 10^{-4}$). Since the solar radiation in channel 1 penetrates more deeply into the cloud layer than that in channel 3, larger optical thicknesses ($\sim 100$) can be retrieved from channel 1 radiance than optical thicknesses ($\sim 10$) from channel 3. The moderately large imaginary index of refraction in channel 3 makes the radiance sensitive to particle size. The sensitivity is larger than those of 1.6 or 2.2 µm at which measured radiances do not include thermal radiation. In channel 3, instead, we have to take into account the effect of thermal radiation, as discussed in the following subsection.

As $\tau_e$ and $r_e$ decrease, multiple solutions of $\tau_e$ and $r_e$ are possible, as noted by Nakajima and King (1990). The reason for this phenomenon is that the optical thickness in channel 3 takes the maximum value at $r_e \sim 4$ µm when the optical thickness in channel 1 is fixed constant. When the cloud is optically thin, the ratio of channel 3 radiance to channel 1 radiance (i.e., the slope of isoeffective radius lines in Fig. 1) becomes a good index of particle size. Wielicki et al. (1990) used those slopes at 1.6-µm and 2.2-µm bands of Landsat TM for cirrus particle sizing.

**b. Removal of undesirable radiation components**

Although the concept of retrieval is simple, some difficulties will occur when we try to determine the cloud microphysical properties from measured spectral radiances of AVHRR. We have to decouple the radiation reflected by the cloud layer, which depends on cloud optical thickness and particle size, from other undesirable radiation components, such as solar radiation reflected by the ground surface, especially for optically thin clouds, and thermal radiation emitted from the cloud layer and ground surface in channel 3. Figures 2 and 3 illustrate simulated radiances ($\text{W m}^{-2} \text{sr}^{-1} \mu\text{m}^{-1}$) in channels 1 and 3 as a function of cloud optical thickness $\tau_e$ and ground surface albedo $A_g$. Similar figures were presented by King (1987) at wavelengths of 0.75 and 1.63 µm. In these figures we see that ground reflectance largely contributes to the satellite signal in the full range of $\tau_e$ in channel 1, whereas it becomes important only for $\tau_e$ less than 10 in channel 3.

Figure 4 illustrates simulated radiances in channel 3 as a function of $\tau_e$ in cases containing 1) no thermal radiation, 2) ground thermal radiation, 3) cloud thermal radiation, and 4) both ground and cloud thermal radiation. In this simulation the surface albedo $A_g$, ground temperature $T_g$, and cloud temperature $T_e$ are set to $A_g = 0.0$, $T_g = 288.2$ K, and $T_e = 275.2$ K. This figure indicates that the radiance in channel 3 is influenced by the ground thermal radiation in the range of $\tau_e < 10$, whereas it is affected by the cloud thermal radiation in the range of $\tau_e > 2$. Even for optically thin clouds ($\tau_e < 2$), we can see an effect of thermal radiation.

These undesirable radiation components have to be removed from the measured radiances. Rossow (1989) noted that the success of this kind of analysis depends on the fidelity of the radiative transfer model used in the data analyses and the accuracy of specifying other properties of the atmosphere and ground surface that affect the measured radiance. Kaufman and Nakajima

![Fig. 2](image-url)  
**Fig. 2.** Simulated radiances in channel 1 (0.64 µm) as a function of cloud optical thickness $\tau_e$ and ground surface albedo $A_g$ under the same conditions as in Fig. 1. The effective particle radius is fixed at $r_e = 8$ µm.

![Fig. 3](image-url)  
**Fig. 3.** As in Fig. 2 except for channel 3 (3.75 µm).
(1993) have adopted a simple method of subtracting thermal radiation by using an effective temperature derived from channel 4 and the optical thickness guessed from channel 1, by which we can calculate the cloud reflectivity in channel 3 without thermal radiation. Of course, this approximation is adequate for optically thick clouds because their neglected small transmissivity prevents ground-reflected radiation and ground thermal radiation transmitting to the space. King and Harshvardhan (1986) noted that the transmissivity of cloud layers can be neglected for $\tau_c > 10$. As an alternative method, thermal radiation in channel 3 has been determined from radiance in channel 4 that is fitted as a function of radiance in channel 3 using nighttime measurements (Coakley and Davies 1986). A ground-reflected radiation and ground thermal radiation, however, are not so small as to be neglected for optically thin clouds. Also, the fitting method does not handle well dual-source thermal radiation that is frequently observed when clouds are optically thin.

Thus, it will be very useful to find an effective method to infer the cloud reflectivity by decoupling it from the undesirable radiation components by making explicit use of radiative transfer theory and all of the information of channels 1, 3, and 4. Ou et al. (1993) developed such a method for their cirrus cloud retrievals but they used the thermal radiation information for their purpose by subtracting the reflected radiation from the satellite-received radiances.

According to the radiative transfer theory for plane-parallel layers with an underlying Lambert surface, we can decouple the radiation reflected by the cloud layer, $L$, from the satellite-received radiance $L_{obs}$ by removing the undesirable radiance components as follows: for visible wavelength,

$$L(\tau, r_e; \mu, \mu_0, \phi) = L_{obs}(\tau, r_e; \mu, \mu_0, \phi) - t(\tau, r_e; \mu) \frac{A_g}{1 - r(\tau, r_e) A_g} t(\tau, r_e; \mu_0) \frac{\mu_0 F_0}{\pi}, \quad (3)$$

and for near-infrared wavelength,

$$L(\tau, r_e; \mu, \mu_0, \phi) = L_{obs}(\tau, r_e; \mu, \mu_0, \phi) - t(\tau, r_e; \mu) \frac{1 - A_g}{1 - r(\tau, r_e) A_g} B(T_e) - t(\tau, r_e; \mu) \frac{A_g}{1 - r(\tau, r_e) A_g} t(\tau, r_e; \mu_0) \frac{\mu_0 F_0}{\pi}, \quad (4)$$

where $F_0$ is the extraterrestrial solar flux and $B$ is the Planck function; $\tau, \tau_e$, and $\tau_0$ are, respectively, the optical thicknesses of atmosphere, cloud layer, and the atmosphere above the cloud layer; $\mu_0$ and $\mu$ are respectively the cosines of solar and satellite zenith angles, and $\phi$ is the azimuthal angle of the satellite relative to the sun. The transmissivity $t$, plane albedo $r$, and spherical albedo $\bar{r}$ are given as

$$t(\tau, r_e; \mu_0) = \frac{1}{\pi} \int_{0}^{2\pi} \int_{0}^{\pi} T(\tau, r_e; \mu, \mu_0, \phi) \mu d\mu d\phi + e^{-\tau_0 \mu_0} \frac{1}{\pi} \int_{0}^{2\pi} \int_{0}^{\pi} R(\tau, r_e; \mu', \mu, \phi) \mu' d\mu' d\phi, \quad (5)$$

and

$$\bar{r}(\tau, r_e) = 2 \int_{0}^{\tau} r(\tau, r_e; \mu) d\mu. \quad (7)$$

where $T(\tau, r_e; \mu, \mu_0, \phi)$ and $R(\tau, r_e; \mu', \mu, \phi)$ are bidirectional transmission and reflection functions respectively. The second term of Eq. (3) and the fourth term of Eq. (4) are ground-reflected radiation components, and the second and third terms of Eq. (4) are cloud and ground thermal radiation components, respectively. Multiple reflection between ground surface and atmospheric layers must be considered in these radiative transfer integrals.

<table>
<thead>
<tr>
<th>Quantities</th>
<th>Gridpoint values</th>
</tr>
</thead>
<tbody>
<tr>
<td>$Z$ (km)</td>
<td>1.0 1.5 2.0 2.5 3.0 3.5 4.0</td>
</tr>
<tr>
<td>$D$ (km)</td>
<td>0.1 0.2 0.5 1.0 2.0</td>
</tr>
<tr>
<td>$\theta$ (deg)</td>
<td>0 5 10 20 30 35 40 45 50 55 60</td>
</tr>
<tr>
<td>$\theta_0$ (deg)</td>
<td>0 5 10 20 30 35 40 45 50 55 60 65 70</td>
</tr>
<tr>
<td>$\phi$ (deg)</td>
<td>0 $-$ 180 divided into 18 (every 10°)</td>
</tr>
<tr>
<td>$\tau_c$</td>
<td>1 2 4 6 9 12 15 20 30 50 70</td>
</tr>
<tr>
<td>$r_e$ ($\mu$m)</td>
<td>4 6 9 12 15 20 25 30</td>
</tr>
</tbody>
</table>
Fig. 5. The flow chart of the retrieval method. Initial values ($r_c = 35, r_e = 10 \mu m,$ and $Z = 2 km$) are input to initialize the flow.

and the upper layer is taken into consideration in Eqs. (3) and (4). This effect is very small, however, enough so to regard $F(\tau_c, r_c)A_e$ as almost zero, especially for an optically thin cloud and for a ground surface of low reflectance. On the contrary, with optically thick cloud and large ground albedo, this effect is relatively large at visible wavelength because the large cloud spherical albedo reflects radiation from the ground surface and the relatively large transmissivity allows this radiation component transmitting to space.

These formulations are exact, apart from the neglected thermal radiation from the atmosphere other than the cloud layer, when we treat monochromatic radiances. We further introduce a wavelength averaging of variables in the formulations. For example, $t$ is averaged with subchannel response function of AVHRR as

$$t = \frac{\sum_{n=1}^{N} \frac{\varphi_n \left( \sum_{k=1}^{M} (w_{n,k} \times f_{n,k}) \right)}{N}}{\sum_{n=1}^{N} \varphi_n},$$

where $\varphi_n$ is the response function of the $n$th subchannel wavelength for each AVHRR channel, $w_{n,k}$ is the weight.
of the \( k \)th \( k \)-distribution, and \( t_{n,k} \) is transmissivity for the \( k \)th \( k \)-distribution at \( n \)th wavelength. This averaging, applied to Eqs. (3) and (4), brings a nonnegligible error in the case of thin cloud layers in which spectral variation of \( t_{n,k} \) becomes large. However, for most cases of our application the error remains small, and we can estimate undesirable radiation components in Eqs. (3) and (4) with variables spectrally averaged for each channel.

In Eqs. (3) and (4) \( T_c \) and \( T_k \) are determined by the brightness temperature of channel 4 for the cloudy target pixel as

\[
T_c = B^{-1} \left( \frac{L_{\text{obs,cloudy}} - t_c(1 - A_k)B(T_c)}{1 - t_c} \right) 
\]

and for a cloud-free pixel adjacent to the target cloudy pixel as

\[
T_k = B^{-1} \left( \frac{L_{\text{obs,clear}}}{1 - A_k} \right), \tag{10}
\]

where \( t_c \) is the transmissivity of the cloud layer in channel 4. Similar formulations were suggested by Rossow et al. (1989).

c. Flow of analysis

Cloud optical thickness and effective particle radius are determined in our method from radiances in channels 1, 3, and 4 of AVHRR by using an inverse method making use of lookup tables to maintain accuracy of the analysis and save computation time. Three tables are prepared for this purpose, that is, tables of cloud-reflected radiances in channels 1 and 3 (Table A), transmissivities and reflectivity in channels 1 and 3 (Table B), and channel 4 transmissivity (Table C). Table 1 summarizes the grid system of the tables, and Fig. 5 illustrates the flow of the analysis.

Beginning with initial values \( \tau_c = 35, r_c = 10 \mu m, \) and \( Z = 2 km, \) where \( Z \) is the cloud-top height, we calculate the cloud geometrical thickness \( D \) from the relation \( D = W/w, \) where \( w \) and \( W \) are respectively the liquid water content and the liquid water path calculated as

\[
W \approx \frac{3\tau_c r_c}{2} \tag{11}
\]

We have used the values of \( w \) proposed by Liou (1976) as shown in Table 2 for several classified cloud types. Once \( \tau_c, r_c, \) and \( D \) and \( Z \) are known, \( t_c \) is obtained from table C. Channel 4 radiances for cloudy and clear sky pixels are put in Eqs. (9) and (10) together with the estimated \( t_c \) to derive \( T_c \) and \( T_k \). The updated value of \( Z \) is thus determined by assuming a constant lapse rate as \( Z = (T_k - T_c)/\gamma, \) with \( \gamma = 6.5 K/km. \)

Theoretical values of cloud-reflected radiances \( L \) can be estimated from Table A and estimated values of \( \tau_c, \)

\( r_c, \)

\( Z, \) and \( D, \) and angular variables \( (\theta, \theta_o, \phi). \) We also remove the undesirable radiation components from measured radiances \( L_{\text{obs}} \) by Eqs. (3) and (4). Then the value of \( \tau_c \) is sought so that the difference of theoretical and retrieved cloud-reflected radiances in channel 1 becomes less than 0.1%. Subsequently, we fix \( \tau_c \) at this value and change \( r_c \) until the difference of radiances in channel 3 becomes less than 0.1%. We continue those two iteration procedures until both the differences in channels 1 and 3 become less than 0.1%. The iteration converges after three or four times. If the cloud-reflected radiances in channel 1 becomes less than the lowest model radiances, the analysis is canceled. On the contrary, if the cloud-reflected radiances in channel 1 becomes greater than the largest model radiances, \( \tau_c \) is set as the maximum value (=70) and the loop for determining \( r_c \) is continued until the solution for \( r_c \) is found. We found that the iteration does not converge in some cases of optically thin clouds when the removed radiation significantly dominates over the signal. In this case we cancel the analysis.

d. Error analyses

In our retrieval method the largest error is caused by the thermal radiation emitted from the atmosphere other than the cloud layer, which we neglected from Eqs. (3) and (4). In addition to this major error there is a small systematic error arising from the averaging of transmissivities and reflectivities as in Eq. (8). Table 3 summarizes retrieval errors in \( \tau_c \) and \( r_c \). The upper and lower rows are the percent error involved in the cloud optical thickness \( E_{\tau_c} \) and effective radius \( E_{r_c}. \) This result was obtained, under the same conditions, in Fig. 1, with \( A_k = 0.05 \) by inputting simulated signals of AVHRR, calculated with a full radiative transfer code, to our retrieval code illustrated in section 2. We did not include observation errors in the simulated data so that the error is mainly produced from the approximation error involved in Eqs. (3) and (4). In a real situation, however, there are many other error sources, as discussed by Han et al. (1994), including radiance uncertainties raised by cloud contamination and a multilayered cloud system.

We find that \( E_{\tau_c} \) is always less than 1%, indicating the approximation of Eq. (3) is good enough for retrieving \( \tau_c. \) Here \( E_{r_c} \) is always larger than \( E_{\tau_c} \) and decreases rapidly with increasing \( \tau_c \) and decreasing \( r_c. \) This error is produced primarily from thermal radiation.
emitted by the atmospheric layer between the cloud bottom and the surface, which we did not account for in Eqs. (3) and (4), so that it decreases rapidly as the transmissivity of the cloud layer decreases. The maximum value of $E_r$, which occurs at the smallest $\tau_r$, and largest $r$, in the table, increases slightly with increasing $\theta_0$ and decreasing $\theta$ because of relatively small solar radiance and large transmitted thermal radiation in such cases. Also $E_r$ slowly increases with increasing $A_r$ because of the increasing contribution of surface reflection. Taking all simulation results into account, we would say that $r$ can be retrieved with errors less than 60%, which occurs for large ground reflectivity and low solar elevation. However, for most cases of realistic clouds ($\tau_r > 5, r \sim 10 \mu m$) under realistic conditions ($\theta_0 < 50^\circ, A_r < 0.3$), $\tau_r$ and $r$ will be determined with errors less than 10%.

As for the retrieval error arising from an observation error, we found similar results as in Nakajima and King (1990), with slightly better accuracy in estimating the effective particle radius than the case of 2.2 $\mu m$ due to the larger imaginary index of refraction at 3.7 $\mu m$.

### 3. Data analyses

To show the utility of the present method, we have analyzed AVHRR image data taken in the FIRE and ASTEX regions and compared the results with in situ measurement data. Table 4 summarizes the time and location of AVHRR and in situ observations used in the analyses.

#### a. Satellite data description

We have used NOAA-10 AVHRR Local Area Coverage (LAC) data at 1538 UTC 10 July 1987 for FIRE data analyses and NOAA-11 AVHRR LAC data at 1638 UTC 13 June 1992 for ASTEX data analyses. The spatial resolution of AVHRR LAC data at a subsatellite point and at the edge of scan lines is about 1.1 km and 6.0 km, respectively. Since the visible channel sensor has been degrading from the prelaunch condition, we applied the corrected calibration constants to get the radiance from satellite-recorded digital counts. For the FIRE period we adopted $a = 1.62$ and $b = 36.4$ for channel 1 of NOAA-10 AVHRR as proposed by Teillet et al. (1990) in the formula $C = a \times L + b$, where $L$ is the radiance (W m$^{-2}$ sr$^{-1}$ $\mu m^{-1}$) and $C$ is the digital counts. For the ASTEX period we adopted $a = 1.69$ and $b = 40.0$ for channel 1 of NOAA-11 AVHRR as proposed by Kaufman and Holben (1993).

#### b. Analyses of FIRE data

From 29 June to 18 July 1987 as a part of the FIRE program airborne measurements of marine stratocumulus clouds were carried out off the coast of southern California. On nearly every flight during the FIRE, the area of interest was in the field of view of the NOAA-10 AVHRR. During this period, a large area of the eastern Pacific was covered by stratocumulus clouds as a result of subsiding warm air together with a strong high pressure system over the cold oceanic surface (Kloesel et al. 1988).

We applied the present method illustrated in section 2 to the entire AVHRR LAC scene of 10 July for obtaining the cloud optical thickness and effective particle radius, as shown in Figs. 6 and 7. We assumed the midlatitude summer atmospheric model of LOWTRAN-7 and set $w = 1.28 \times 10^{-6}$ (g cm$^{-3}$) assuming stratocumulus clouds in Table 2 for our analyses. Since we an-

| Date       | Platform | Time (UTC) | Location  
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<th></th>
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<td>(29, 128) (27, 116)</td>
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<tr>
<td>13 June 1992</td>
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<td>1638</td>
<td>(38, 31) (40, 14)</td>
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<td>(32, 29) (33, 14)</td>
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<td>C-131A</td>
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<td>End (33.65, 24.36)</td>
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* The satellite nadir track is along (34, 115)–(27, 116).

* The satellite nadir track is along (38, 31)–(32, 29).
analyzed oceanic clouds, we adopted $A_r = 0.05$ to simulate
sea surface reflection. Pixels over the continent and opti-

cally thin area ($r_c < 1$) were opaqued.

One of the most striking features of these figures is
that the cloud properties were sharply different between
the left and right sides of longitude 122°W. This large-

scale difference of cloud properties may be due to the
difference in the airmass properties. The optically thick
clouds (the right side of 122°W) characterized by small
effective particle radius ($r_e = 5 \sim 15 \mu m$) might be
enhanced as a consequence of cloud modification by
dust-contaminated warm air blowing from the North
American continent as seen in synoptic chart analyses
(e.g., Albrecht et al. 1988). Clouds forming over the
cold oceanic surface can thicken if drizzle is cut off by
a reduction of particle size as suggested by Albrecht
(1989). On the other hand, optically thin clouds (the
left side of 122°W) with relatively large effective
particle radius ($r_e \sim 20 \mu m$) form in clean maritime air.
In this region it is normal for clouds to have a drizzle mode
in their particle size distribution (Radke et al. 1989).

On 10 July, the NASA ER-2 and the University of
Washington C-131A aircraft flew tightly coordinated in
space and time along the flight leg shown by the solid
line in Figs. 6 and 7. The ER-2 aircraft, which flew at
an altitude of 18 km, was equipped with the Multi-
spectral Cloud Radiometer (MCR) for measuring multispec-

tral cloud-reflected radiances. The C-131A aircraft,
which flew within clouds, was equipped with the Particu-
late Measuring System (PMS) for measuring cloud particle
size distribution and Johnson-Williams (JW) hot wire
probe for measuring the cloud liquid content. Nakajima
et al. (1991) used channel 1 (0.75 \mu m) and channel 6
(2.16 \mu m) of the MCR to determine the cloud optical
thickness and effective radius. They adjusted remote
sensing values of the effective radius $r_{remote}$ to the ex-
pected values at the geometric center of the cloud layer
$r_{center}$ using the method described in Nakajima and King
(1990). To compare our results with their results we
sampled pixels along the flight line in Fig. 6 and cor-
rected them by the same method except for the expres-
sion of the diffusion exponent $k$ at 3.7 \mu m as follows:

$$k_{1.7} = 3.48 \times 10^{-2} + 9.62 \times 10^{-2} \ln(r_e).$$

In Figs. 8 and 9 we show the effective radius and optical
thickness estimated by AVHRR, MCR, and in situ mea-
surements with PMS probes, as a function of distance
along the nadir track of the ER-2. MCR and PMS values
were taken from Nakajima et al. (1991). All data were
smoothed by a 5-point moving average. First, we note in
Fig. 8 a very good spatial correlation between values of
AVHRR-derived effective radius $r_{avhrr, center}$ and in situ
values $r_{in situ}$ except for the portion of 0 \sim 40 km. There is a
slight overestimation of about 1 \mu m by AVHRR remote
sensing as compared with in situ values, in contrast with
the large systematic overestimation by MCR remote sen-
ing. Such an overestimation tendency of remote sensing
has been known as *anomalous cloud absorption* in the
near-infrared spectral region (Twomey and Cocks 1989;
Stephens and Tsay 1990). In the FIRE region similar
overestimation by remote sensing was also reported by
Rawlins and Foot (1990). We may attribute the better
retrievals from AVHRR to several reasons. We used 3.7-

\mu m radiance for retrievals, whereas most other studies
used 1.6- or 2.2-\mu m radiances. The 3.7-\mu m channel can
be well calibrated by a blackbody standard, whereas the
other channels need a calibration with a standard lamp
and a integrating sphere, which are less stable. Also the
light absorption coefficient of water vapor and the imagi-

nary index of refraction of liquid water, which are needed
for theoretical simulations, have larger values at 3.7 \mu m
so that they might have been more accurately measured
at that wavelength. The other reason is that we used
LOWTRAN-7 water vapor absorption, whereas most
other studies used LOWTRAN-5 values. LOWTRAN-7
has stronger continuum absorption of water vapor than
LOWTRAN-5 in the 2.2 and 3.7 \mu m windows. Taylor
(1992) got a similar good result from his MCR radiomi-
er using LOWTRAN-7. Han et al. (1994) have com-
pared their results of effective radius and optical thickness
with those of Nakajima et al. (1991) and pointed out that
their AVHRR-retrieved effective radii are slightly larger
(1-2 \mu m) than the FIRE in situ values and slightly
smaller (about 0.5 \mu m) than the MCR-retrieved values.
This finding is consistent with our result in Fig. 8, which
shows a slight overestimation by AVHRR of about 1 \mu m.

As for the systematic difference between AVHRR and
PMS values for the first 40-km leg, it is difficult to iden-
tify the reason. There might be mismatching of time and
location between aircraft and satellite observations. But it
will be noteworthy that in the first 40-km portion there
existed optically thin clouds with drizzle droplets as seen
by the large in situ effective radius. Since the 3.7-\mu m
channel is sensitive to the particle size near cloud top as
compared with the 2.2-\mu m channel, there might be
smaller particles near cloud top than cloud bottom where
drizzle droplets were expected. This guess is supported
by Fig. 10, which illustrates the volume size distributions
at flight distances of 32 km, 40 km, and 55 km. We find
that the portion of cloud at 40 km had a significant drizzle
mode of the size distribution around 100 \mu m, even at the
cloud center where the C-131A flew.

Figure 9 compares the AVHRR-derived cloud optical
thickness $\tau_{avhrr}$, in situ values $\tau_{in situ}$, and the MCR-
derived values $\tau_{mcr}$. Using Eq. (11), $\tau_{in situ}$ was cal-
culated with $r_{in situ}$ and the liquid water path, which was
estimated from the liquid water content measured by the
JW probe. A very good spatial correlation among
$\tau_{avhrr}$, $\tau_{mcr}$, and $\tau_{in situ}$ is found in Fig. 9. If we investi-
gate the difference in detail, however, we find a portion
where $\tau_{avhrr}$ is closer to $\tau_{in situ}$ than $\tau_{mcr}$ and another
portion where $\tau_{mcr}$ is closer to $\tau_{in situ}$ than $\tau_{avhrr}$. There
may be a significant effect of the large field of view of
AVHRR ($\sim 1$ km) as compared with that of MCR
($\sim 120$ m). This effect is clearly seen in the sharp dip
of $\tau_{in situ}$ at $\sim 40$ km of the leg.
Fig. 6. AVHRR-derived effective particle radius on 10 July 1987 in the FIRE region. The solid line labeled "flight line" is the flight leg of the ER-2 and C-131A aircraft used in the comparison.

Fig. 7. As in Fig. 6 except for the cloud optical thickness.
Two lines of optically thick clouds can be seen around 32.5ºN, 121ºW just above the flight line in Fig. 7. These were identified as ship tracks and were penetrated by the C-131A aircraft at 1556–1609 UTC for in situ observation (Radke et al. 1989). Figure 11 shows a scatter plot of \( r_{avhr, center} \) and \( T_{avhr} \) over the ship track region. The contour lines correspond to 30%, 50%, 70%, and 90% occurrence levels of joint probability density function. Two peaks of the joint probability density function, that is, one with small optical thickness and large effective radius (left peak) and the other with large optical thickness and small effective radius (right peak), show that two kinds of cloud existed in the ship track region. This observation indicates that cloud particles were influenced by artificial CCN seeded by ship stack emissions under the clouds, and the cloud properties transformed from the left peak of the joint probability density to the right peak, producing a negative correlation between effective radius and optical thickness. Figures 6 and 7 show that the typical effective radius of cloud top decreases from 12 to 10 \( \mu m \) (83%) in track 1 (left hand, stronger) and from 12 to 11 \( \mu m \) (92%) in track 2 (right hand, weaker). This result is very similar to what Radke et al. (1989) obtained from direct measurements by the C-131A.

![Graph showing effective radius as a function of distance for AVHRR, MCR, and PMS adjusted to cloud center](image1)

**Fig. 8.** Comparison of the effective radii as a function of flight distance along the nadir track of the ER-2 aircraft. Values are shown for AVHRR remote sensing (solid line), MCR remote sensing (dashed line), and PMS in situ (solid circles) measurements. The results from the AVHRR and MCR are adjusted to the cloud center.

![Particle size distributions](image2)

**Fig. 10.** Particle size distributions obtained by the PMS probes at 32, 40, and 55 km of the flight leg.

![Graph showing particle radius as a function of particle size](image3)

**Fig. 9.** As in Fig. 8 except for cloud optical thicknesses.

![Combined distribution of optical thickness and effective particle radius](image4)

**Fig. 11.** Combined distribution of optical thickness and effective particle radius retrieved from AVHRR for ship track clouds. Values of the effective particle radius are adjusted to the cloud center. The four contour lines correspond to 30%, 50%, 70%, and 90% occurrence levels of joint probability density function.
Fig. 12. AVHRR-derived effective particle radius on 13 June 1992 in the ASTEX region.

Fig. 13. As in Fig. 12 except for cloud optical thickness.
c. Analyses of ASTEX data

From 1 to 28 June 1992 as a part of the ASTEX experiment airborne measurements were carried out in the area of the Azores and the Madeira Islands. On 13 June 1992 the region of the experiment was covered by stratuscumulus clouds, of which the estimated thickness was about 1000 ft under a high pressure region (Bretherton and Pincus 1994; Bretherton et al. 1994). On this day the University of Washington C-131A and NCAR Electra were dedicated to airborne measurements, and the area of interest was in the field of view of the NOAA-11 AVHRR. It was reported by the C-131A that the region was characterized by large-particle clouds and extensive drizzle.

Figures 12 and 13 show AVHRR retrievals over the ASTEX region. Clouds, in this case, were optically thin with relatively large particle radii, which is much different, for example, from the 10 July case of the FIRE experiment, when thin clouds were associated with small particles, and is rather similar to the drizzling thin cloud portion of the FIRE 10 July case (Nakajima et al. 1991). It is thus inferred that the clouds had a drizzle-mode size distribution as found by the in situ airborne measurements. There is, however, a tendency of positive correlation between \( r_{\text{avhr}} \) and \( r_{\text{avhr}} \), which is much different from the 10 July FIRE case. In general, as shown by Figs. 6 and 12, the particle size field is more homogeneous than the optical thickness field, but at the same time it can be said that mature and thicker clouds tend to have a less homogeneous effective particle radius field.

Figure 14 compares the retrieved effective radius \( r_{\text{avhr}} \) with the corresponding in situ values \( r_{\text{in situ}} \) obtained by Gerber-probe measurements. We did not adjust \( r_{\text{avhr}} \) to the cloud center in this case because it is difficult to find the aircraft height relative to the cloud top and bottom in this flight leg. The AVHRR retrievals seem to be in good agreement with Gerber-probe results, although the two hour difference between NOAA satellite and aircraft operation time may be too long to confirm this observation. Figures 15 and 16 compare \( r_{\text{avhr}} \) with \( r_{\text{in situ}} \), obtained by FSSP measurements along two sounding flight legs of the Electra as a function of the flight distance. The altitude of the FSSP is also shown in the figures with labels to show the location of the cloud top and bottom. The Electra flew in the cloud, descending from the cloud top to the bottom (Fig. 15) and ascending from the bottom to the top.
(Fig. 16), to measure the vertical profile of cloud particle size. These figures show that FSSP-derived $r_{\text{m situ}}$ increased with increasing altitude, and around the cloud top AVHRR-derived $r_{\text{avhyr}}$ agree quite well with $r_{\text{m situ}}$. The effective depth estimated by the method of Nakajima and King (1990) is also shown in the figures, from which we confirm that our remote sensing retrievals are in good agreement with the in situ values.

4. Discussion and concluding remarks

In the preceding sections we have studied the performance of our retrieval method of the optical thickness and effective particle radius from the AVHRR radiometer by theoretical analyses and by data analyses. The expected error in our analyses is less than 25%, as summarized in Table 3, and less than 10% for most cases of realistic clouds ($\tau_c > 5$, $r_e - 10 \mu m$) under realistic conditions ($\theta_o < 50^\circ$, $A_v < 0.3$). There might, however, be other error sources that we did not take into account, such as adequacy of the model atmosphere we assumed and the effect of neglected thermal radiation emitted from the atmosphere other than the cloud layer. Effects of non-plane-parallel features, ice crystals, and aerosols are other important issues for further quantitative investigation. In spite of such insufficiency, characteristic features of wide-area cloud microphysical properties in the FIRE and ASTEX regions seem to be retrieved in a consistent manner as compared with in situ observation. For example, the left side of Fig. 7 is the wing of the satellite image with view angles as large as $40^\circ$–$50^\circ$ (see the footnote to Table 4). Even with such large view angles we still see clearly the drastic change in optical thickness and particle radius for studying the cloud airmass transition. We do not see noticeable systematic biases in the retrievals along the edge of cloud decks where the non-plane-parallel effect is expected to be large.

As shown by Figs. 8 and 9, we had a unique opportunity to compare the remote sensing results from the 2.2 and 3.7 $\mu m$ channels. It is interesting to note in Fig. 8 that the MCR 2.16-$\mu m$ result falls between the AVHRR 3.75-$\mu m$ result and in situ values, apart from the systematic bias. This may be caused by a different effective depth of remote sensing between the 2.2 and 3.7 $\mu m$ channels. For example, the effective particle radius is in agreement with the in situ values in the smaller particle region, whereas they start deviating from in situ values for the drizzle-rich portion at 0 to 40 km of the flight leg. Since drizzle droplets are expected to exist near the cloud bottom, the difference in the effective depth consequently causes this deviation. This observation suggests the possibility of a vertical sounding of effective particle radius profile using 1.6, 2.2 and 3.7 $\mu m$ radiances simultaneously. These channels will be available in the future Moderate Resolution Imaging Spectrometer on EOS-AM platform (King et al. 1992); these channels can be used for retrieving the stratification of the effective particle radius.

As summarized above, we found AVHRR-derived values of the effective radius were in good agreement with values from the PMS probes. In spite of this finding we should have more measurements by various instruments for particle sizing to improve the validity of in situ measurements. In this study we had an opportunity to compare our AVHRR retrievals with those from a new instrument: the Gerber probe. The averaged values of effective particle radius were very close, although we did not have good spatial correlation be-
cause of the two hour difference in Gerber probe and AVHRR observation time. We need more analyses to confirm the agreement with the Gerber probe. Use of a "glory" feature of cloud-reflected radiance (Spinhrne and Nakajima 1994) will be another new promising technique for detecting effective radius since this technique relies on the single scattering theory, which is different from other remote sensing techniques that utilize multiple scattering of radiation by cloud particles.

One of highlights of our results is the distinct contrast of cloud microphysics shown in Figs. 6 and 7, which may be explained by the large-scale modification of cloud microphysics by the intrusion of continental airflow. It is guessed that maritime cloud particles ($r_c \sim 20 \mu m$) reduce their size ($r_c = 5 \sim 15 \mu m$) as a result of cloud—aerosol interaction. We also found another example of cloud—aerosol interaction in the pair of ship tracks. Clouds were drastically enhanced by CCN seeded by ship stack emissions under the clouds. For these clouds we found a negative correlation between $\tau$ and $r_c$, as shown in Fig. 11. The large optical thickness change, which is much stronger than the con-
stant liquid water path assumption $\tau_c \propto r_c^{-1}$, suggests drizzle quenching occurred in this area.

Figures 17 and 18 show scatterplots of $\tau_c$ and $r_c$ dividing the corresponding AVHRR scenes (Figs. 6, 7, 12, and 13) into 16 sections. We see many features in the distributions, but there are two distinct patterns, that is, positive and negative correlations, similar to the cloud microphysical statistics found by Nakajima et al. (1991). The negative correlation is very similar to the ship track cases shown in Fig. 11. This observation suggests that, whatever the CCN sources, there is a similar transition of microphysical states of clouds in these regions (Nakajima 1993). If we compare Figs. 17 and 18, it is found that the FIRE region has more negative correlation patterns than the ASTEX region, indicating that more drizzle-rich clouds existed in the FIRE region. Since it has such a local tendency in microphysical state transition, it will be important to obtain similar statistics for different cloud types and locations to improve our knowledge in assessment of the cooling effect of SO$_2$ emissions (e.g., Charlson et al. 1992; Kaufman and Chou 1993). In this context, it is
very interesting to see that such correlations also have been found in the global statistics given by Han et al. (1994). According to their analyses, optically thin clouds tend to have a positive correlation, whereas thicker clouds tend to have a negative correlation. Although this tendency is very similar to what we found in this study, we have to undertake more investigations to understand the mechanism for this tendency on a global scale. This is because on a global scale the mechanism mentioned above may not be a dominant factor in determining cloud microphysics; rather, dynamical factors may become more important.

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