Microphysical and Radiative Properties of Marine Stratocumulus from Tethered Balloon Measurements

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ABSTRACT

Vertical profiles of cloud microphysical data and longwave and shortwave radiation measurements through the marine boundary layer were obtained using an instrument package on the NASA tethered balloon during the FIRE Marine Stratocumulus Experiment. The radiation observations were analyzed to determine heating rates inside the stratocumulus clouds during several tethered balloon flights. The radiation fields in the cloud layer were also simulated by a two-stream radiative transfer model, which used cloud optical properties derived from microphysical measurements and Mie scattering theory.

The vertical profiles of the observed longwave cooling rates were similar in structure and magnitude not only to previous measurements of marine stratocumulus, but also to the cooling rates computed by the two-stream radiative transfer model. The solar heating rates measured in the clouds, however, were systematically much larger than the rates calculated in the model.

Solar albedo measurements showed that the visible spectrum tended to be reflected by the clouds more than the near IR spectrum. This is similar to the results reported by Hignett, although the discrepancies between the observed and calculated near IR to visible albedo ratios were generally much smaller. The results from the flights on 10 and 13 July 1987, however, suggest that the effects of heterogeneities on the radiative transfer through the cloud may be more important in the visible than in the near IR.

1. Introduction

Each summer, large areas of stratocumulus clouds form under the eastern branch of subtropical oceanic high pressure systems. Due to the different radiative properties of the stratocumulus compared to those of the ocean, these cloud sheets greatly affect the flow of solar radiation into the atmosphere-ocean system and therefore influence the global atmospheric circulation and climate. Since changes in the size, composition, and location of these cloudy areas may significantly alter the global climate, an understanding of the processes that maintain these cloud systems is required.

In the past two decades, marine stratocumulus clouds have been studied extensively with numerical models of varying complexity (e.g., Driedonks and Duyzerke 1989) in order to simulate some of the physical processes in these clouds. Despite the large amount of theoretical work on marine stratocumulus, many physical processes in the marine boundary layer including entrainment and the influence of radiative heating on the heat and moisture budgets are still not understood due to a lack of detailed observations of these clouds. One of the specific goals of Project FIRE (First ISCCP Regional Experiment) was to seek a better understanding of the roles played by these processes in determining the life cycles of marine stratocumulus systems (Albrecht et al. 1988). Toward this goal, FIRE was designed to produce an improved set of observations of the clouds by coordinating several different instrument systems during an intensive field operation (IFO) on and near San Nicholas Island (SNI).

A NASA tethered balloon was included with the surface-based instruments on SNI. Colorado State University participated in the FIRE stratocumulus IFO by adding an instrument platform to the NASA balloon. The instrument platform was attached to the tethered balloon cable and carried several instruments to measure the thermodynamic, radiative, and microphysical characteristics of the cloud layer.

Marine stratocumulus clouds are maintained by a balance between several processes including entrainment and radiative cooling/heating at cloud top, and fluxes of heat and moisture from the surface. Tethered balloon platforms have been used in the past few years (Slingo et al. 1982; Gerber 1986) to study maritime stratocumulus, and may be one of the best ways to study these clouds because they offer several advantages over other instrument systems in the measurement of the energy balance components. The tethered balloon can probe the entire depth of the marine boundary layer more easily than aircraft and thus can provide complete vertical profiles of several meteorological variables. Unlike aircraft, the balloon platform can re-
main in one location over a long period of time and therefore can measure local time series of cloud properties. The instrument platform can also obtain high vertical resolution data, which are essential for the study of radiative fluxes since they can vary dramatically through the first 10–50 m below cloud top.

The CSU instrumentation aboard the balloon platform was primarily designed to study the radiative contributions to the energy budget of the cloud. A separate instrument package deployed by the United Kingdom Meteorology Office measured momentum, heat, and moisture fluxes. This paper provides a description of analyses of the radiation and cloud microphysical data collected from measurements made by the tethered balloon. These measurements were made with the following scientific issues in mind:

- What is the depth and intensity of the longwave cooling and shortwave heating at cloud top?
- How do these fields vary over time inside the cloud?
- How do inhomogeneities in the cloud and changes in the microphysical properties affect the transfer of radiation through the boundary layer, especially in the near infrared and visible regions of the spectrum?

The following section contains a description of the details of the experimental program, including the flight patterns of the balloon; the radiation and microphysics instruments carried aboard the tethered balloon platform; and the procedures used to reduce the raw data from the instruments into a usable form for each flight. Section 3 gives a brief overview of the general characteristics of three flights. Section 4 presents some results from the flights, which include measurements of the observed fluxes and solar and infrared heating rates inside the clouds. Since the tethered balloon was located in a region where both continental and maritime air masses are often present, the influence of different air masses on the radiative properties of the cloud via changes in the microphysical characteristics of the stratocumulus is also examined. Simulations of the radiative transfer through the stratocumulus clouds by a two-stream radiative transfer model are presented in section 5 in order to compare current understanding of the radiative transfer through these clouds with the tethered balloon observations. The final section gives a summary and a discussion of the results.

2. Details of the experimental program

a. The NASA/GSFC tethered balloon flight strategy

The tethered balloon used to carry the CSU instrument package was developed by the Wallops Flight Facility of the NASA/Goddard Space Flight Center. The balloon is 31.6 m long with a volume of 1275 m³ and is capable of lifting a 400-kg payload up to an altitude of 1 km. The platform was attached to the tether line approximately 60 m below the balloon and measured radiation, microphysical, and thermodynamic properties in the cloud.

A total of nine research flights were made using the CSU package. Results from three of the flights launched on 7 July (flight 2), 8 July (flight 3), and 10 July 1987 (flight 5) are presented in sections 3 and 4. Information on the vertical structure of the boundary layer was collected in two ways. The first was a continuous profile through the boundary layer (the rise rate of the balloon was approximately 20 m min⁻¹). The second method involved making a sounding in a series of steps; the balloon would be raised or lowered a set distance (usually 100 m) and remained at a constant level 5–20 min. This stepping method allows one to compute the mean and the range of the radiative and microphysical variables at several levels, while the continuous profile provides a vertical profile of data over a short time interval.

b. Radiation instrumentation

The tethered balloon instrument platform carried several different types of radiation instruments to measure both solar and infrared radiation. Two sets of Epingle pyranometers (upward and downward facing) were used to obtain measurements of the visible and near IR parts of the shortwave spectrum within an accuracy of 5 W m⁻². One set of pyranometers measured the total shortwave spectrum (from 0.3–2.8 µm). The other set, which were filtered by dark red Schott glass domes, measured the near IR spectrum (0.7–2.8 µm).

Epingle pygeometers were used to measure the upward and downward infrared fluxes in the range from approximately 4–50 µm. Although both the measured shortwave and longwave irradiances were converted directly from the radiation instruments and logged into a data platform system, the output voltages from the pygeometers were also recorded so corrections to the longwave fluxes could be made. Errors in the measured fluxes due to differential heating of the instrument were corrected using the procedure described by Albrecht and Cox (1977).

The absolute accuracy of the corrected longwave fluxes was 10 W m⁻², but for the flux divergence measurements the observations agree with model calculations by an average of 0.03 W m⁻² mb⁻¹ (Albrecht et al. 1974). Both the pyranometers and the pygeometers were calibrated before the start of the balloon flights. The radiation data from each instrument were logged at an interval from 4–8 s.

c. Microphysics instrumentation and data reduction

1) Analysis procedure

The instrument on the platform used to obtain microphysical information was a forward scattering spectrometer probe (FSSP), developed by Particle Measuring Systems, Inc. (Knollenberg 1981). From the
raw FSSP data, droplet-size distributions were calculated for each constant-level leg of the flight. Additional variables, such as the liquid-water content (LWC), effective droplet radius \( r_e \), mean droplet radius \( r_m \) and total number density \( N_0 \) were computed from the droplet-size distributions. These quantities provide information on the microphysical structure of the clouds, and with the use of Mie scattering theory, an estimate of the optical properties of the clouds. The data collected by the FSSP probe were compared to estimates of the cloud liquid-water path (LWP). The LWP is the vertically integrated liquid-water content that was derived from observations by a NOAA/WPL microwave radiometer stationed on SNI (Hogg et al. 1983; Snider 1988).

The droplet-size distributions were found by computing the droplet concentrations for each size interval of the FSSP. The droplet concentration is the number of droplets accepted during the sampling interval divided by the sampling volume. The sampling volume is obtained as the product of the effective cross-sectional area \( A_e \), the particle velocity \( v \), and the sampling interval \( \delta t \) (5 s).

The effective cross-sectional area of the laser beam used to sample cloud droplets was determined by multiplying the measured cross section of the beam by the ratio between the accepted counts and the total counts:

\[
A_e = A \frac{A_C}{T_C},
\]

where \( A \) is the measured cross section \( (4.6 \times 10^{-7} \text{ m}^2) \), \( A_C \) is the total number of particles that are accepted by the probe as passing through the center of the laser beam, and \( T_C \) the total number of particles detected by the beam.

Since the wind in the boundary layer was generally strong throughout the experiment \( (5-10 \text{ m s}^{-1}) \), it was the only agent used to blow particles through the probe. Although the balloon was designed so it would constantly point into the wind, the drag on the instrument platform was such that it would twist the tether cable and the probe would not face directly into the wind. The angle between the wind and the probe was measured and recorded along with the other meteorological variables. An estimate of the speed of the particles through the tube was then made by multiplying the wind speed by the cosine of this angle and the sampling volume was determined to be

\[
V = A_e v \cos \theta \delta t,
\]

where \( v \) is the airspeed and \( \theta \) is the angle of the wind direction. The concentration of particles in one channel is then

\[
n_j = C_j / V,
\]

where \( C_j \) is the number of droplets counted in the \( j \)th channel. In order to prevent the estimates of the particle concentrations from being erroneously large, the calculations were not made when \( \theta \) became large \( (>60^\circ) \) or the particle speed became small \( (<4 \text{ m s}^{-1}) \). During flights 1-4, the least significant bit was missing from the output of the pulse height analyzer, which resulted in a loss of some resolution \( (6-\mu \text{m} \text{ intervals instead of } 3-\mu \text{m intervals}) \) in the bin size data. An example of a measured size distribution is shown in Fig. 1.

Several other microphysical variables were derived from the measured droplet-size distributions. The average droplet radius in each channel depends on the size range set for the instrument. The radius of the droplets in each channel was assumed to be the midpoint value of each channel. Since the instrument was always set on range 0, the average droplet radius in each channel was

\[
r_j = (3j + 0.5)/2,
\]

where \( r_j \) is expressed in microns. The mean radius of the particles was computed from

\[
r_m = \frac{1}{N} \sum_{j=1}^{15} n_j r_j,
\]

and the effective radius was computed from

\[
e = \frac{1}{15} \sum_{j=1}^{15} n_j r_j^2
\]

and

\[
e = \frac{1}{15} \sum_{j=1}^{15} n_j r_j^3
\]

2) Sensitivity analysis of the FSSP Measurements

As stated earlier, the droplet number densities measured by the FSSP probe are a function of the wind

![Fig. 1. Mean normalized droplet-size distributions for the indicated flights.](image-url)
speed and relative direction between the probe and the wind according to
\[ n_j = \frac{C_j T_C}{AA_C v \cos \theta \beta t}, \tag{7} \]
where variables are as defined above.

The measured cross-sectional area \( A \) and the sampling time \( \beta t \) were fixed throughout the flights, and thus may be substituted by a constant. The other variables, however, change throughout the flight. If it is assumed that the probe accurately counts the number of particles flowing through it, then only errors in wind speed \( v \) and the relative direction between the wind and the probe \( \cos \theta \) will affect \( n_j \).

The change produced in the number density per unit error (radian) in the wind direction is
\[ \frac{\partial n_j}{\partial \theta} = \frac{C_j T_C}{KA_C v} \tan \theta \sec \theta. \tag{8} \]

From this equation it is evident that
- An increase in \( \theta \) will lead to an increase in \( n_j \).
- The sensitivity of \( n_j \) to changes in \( \theta \) is less for large wind speeds, but increases greatly for large values of \( \theta \).

A measure of the relative sensitivity of \( n_j \) to observed wind direction then results as
\[ \frac{1}{n_j} \frac{\partial n_j}{\partial \theta} = \tan \theta \text{ in units of rad}^{-1}. \tag{9} \]

Thus for the reported 5° uncertainty in the wind direction measurements, we expect a relative error of 1.5% in \( n_j \) when \( \theta = 10^\circ \) and an increase to a 15.1% error when \( \theta = 60^\circ \). For most of the flights, the average value of \( \theta \) is below 40°, and thus the expected error in \( n_j \) is below 7%.

The sensitivity of \( n_j \) to an error in wind velocity \( v \) can be estimated by differentiating (7) with respect to \( v \):
\[ \frac{\partial n_j}{\partial v} = -\frac{C_j T_C}{KA_C v^2 \cos \theta}. \tag{10} \]

According to this expression, the sensitivity decreases as \( v \) increases. A measure of the relative sensitivity of \( n_j \) to wind speed, expressed as \( n_j^{-1} (\partial n_j/\partial v) \) suggests that a 1 m s\(^{-1}\) error in the measurement of \( v \) results in a 25% error in \( n_j \) when \( v = 4 \) m s\(^{-1}\), and decreases to a 10% error when \( v = 10 \) m s\(^{-1}\). Since the wind speeds recorded at the time of measurement were typically 9 m s\(^{-1}\), the expected error in \( n_j \) owing to imprecision in the measurement of \( v \) is likely to be less than 11%.

3. Overview of case flights

Table 1 briefly describes the three flights considered in the analyses reported in this paper, while Fig. 2 shows a typical profiling procedure used during one of the flights. From the 850-mb and surface pressure analyses and the measurements of wind direction by the platform, the direction of the air flow over SNI was generally from the north during flights 2 and 3 and from the south during flight 5. The offshore flow over SNI during flights 2 and 3 affected the droplet-size distributions and gave the clouds over the island a more continental character, with a predominance of small droplets. During flight 5 however, southerly flow prevailed resulting in a more marine-like air mass over SNI.

Accurate retrieval of cloud LWC was one of the major problems of the FSSP dataset. Due to the variable nature of the wind speed and direction (which control the sampling volume of the instrument), estimates of the microphysical variables also fluctuated greatly during the flights. For example, during most of the afternoon legs taken on 8 July, the standard deviation of the radiometer-derived LWP was generally 20% of the mean, while the standard deviation of the LWC at each level was often as large as 75%. The standard deviation in the LWC was largest when the relative angle between the wind and the platform was large (>40°). Derivations of the cloud LWP using the FSSP data may be unreliable, given the nature of this data.

The data collected from the ascending profile taken
during the morning of 7 July show that the cloud extended from 980 mb (250 m above sea level) to 945 mb (550 m). The height of the cloud top decreased during the afternoon, and was near 950 mb (510 m) at 2300 UTC. A time series of the radiometer-derived LWP measured during flight 2 shows that while the instrument platform was measuring the radiative fluxes above cloud top (1600–2030 UTC) the LWP ranged from 70 to 150 g m⁻². However, when the balloon entered the cloud in the afternoon, the LWP dropped significantly and remained around 30 g m⁻² when the microphysics measurements were made. The albedo data collected during flight 2 are from a totally different type of cloud than the microphysical data.

The cloud layer extended from roughly 970 mb (300 m) to 935 mb (605 m) at the start of flight 3, and the cloud top rose above 930 mb (650 m) during the day. During the morning the LWP was large and variable, ranging from 60 to over 300 g m⁻². Like flight 2, the LWP in the afternoon (after 1900 UTC) was much lower (near 35 g m⁻²) and less variable.

The cloud layer observed during flight 5 ranged from roughly 915 mb (800 m) to 895 mb (1000 m). The radiometer data and the radiative flux data indicate that the clouds in the boundary layer varied considerably during the day and the cloud cover probably was not homogeneous. The LWP measured at the time the instrument platform was in the cloud varied from 7 to 195 g m⁻².

4. Results

a. Microphysics measurements

From the raw dataset, all of the microphysical variables mentioned in section 2 were calculated for each 5-s data interval, and the average and standard deviation of each variable were computed for every horizontal leg of the flight. The microphysical variables measured during flights 2, 3, and 5 are presented below. Table 2 summarizes the microphysical observations obtained from the flights.

Figure 1 shows the mean normalized droplet-size distributions from all three flights. The droplet-size distributions from the flights can be divided into two regimes. The distributions from flights 2 and 3 are dominated by particles in the two smallest FSSP bins from 2–8 μm in diameter, while flight 5 has a distribution that is more marine-like in appearance, with a peak in the 14–20 μm range. The high concentrations of smaller droplets in flights 2 and 3 suggest the clouds during those flights were formed in a boundary layer influenced by continental air, which is also consistent with the synoptic conditions observed during these days.

Profiles of the average LWC computed from each leg of the three flights are shown in Fig. 3. In flights 2 and 5, the LWC generally increased from cloud base to cloud top, while in flight 3 the LWC peaked in the middle of the cloud. The time series of the radiometer-
TABLE 2. Cloud microphysical data.

<table>
<thead>
<tr>
<th>Variable</th>
<th>Flight 2</th>
<th>Flight 3</th>
<th>Flight 5</th>
</tr>
</thead>
<tbody>
<tr>
<td>Cloud top</td>
<td>945–950 mb</td>
<td>930–935 mb</td>
<td>895 mb</td>
</tr>
<tr>
<td>Cloud base</td>
<td>980 mb</td>
<td>970 mb</td>
<td>915 mb</td>
</tr>
<tr>
<td>Maximum LWC</td>
<td>0.14 g m⁻³</td>
<td>0.19 g m⁻³</td>
<td>0.50 g m⁻³</td>
</tr>
<tr>
<td>Std. dev. LWC</td>
<td>36% of mean LWC</td>
<td>30%–75% of mean LWC</td>
<td>45% of mean LWC</td>
</tr>
<tr>
<td>LWP (FSSP)</td>
<td>35–40 g m⁻²</td>
<td>52–60 g m⁻²</td>
<td>65 g m⁻²</td>
</tr>
<tr>
<td>LWP (radiometer)</td>
<td>90 g m⁻² (before 2000 UTC)</td>
<td>135 g m⁻²</td>
<td>80 g m⁻²</td>
</tr>
<tr>
<td>Range of $N_0$</td>
<td>40–50 cm⁻¹</td>
<td>25–55 cm⁻¹</td>
<td>50–100 cm⁻¹</td>
</tr>
<tr>
<td>$r_e$</td>
<td>10–14 μm</td>
<td>10–13 μm</td>
<td>10–13 μm</td>
</tr>
<tr>
<td>$r_m$</td>
<td>5–7 μm</td>
<td>7.5–10 μm</td>
<td>7.5–10 μm</td>
</tr>
</tbody>
</table>

Derived LWP shows that the cloud layer had the largest liquid-water paths when the platform was in the middle of the cloud. Thus, most of the variability in the LWC during flight 3 is due to changes in the entire cloud rather than the differences between the top and bottom of the cloud.

The measured LWC from flight 2 was much less than the adiabatic liquid-water content, and this is also reflected in the radiometer-derived LWP. The peak LWC measured during flight 2 was 0.136 g m⁻³. Assuming the LWC increased linearly from cloud base to cloud top, the mean liquid-water path would be 35–40 g m⁻². The average LWP measured during the same time was 30 g m⁻². By contrast, the LWC measured during flight 3 was derived from a cloud layer with a widely varying structure. Most of the microphysical measurements were taken before the LWP decreased sharply in the afternoon, but the measurements taken at cloud top seem to be a mixture of clear-and-cloudy-air observations, and cannot be used to estimate the cloud-top LWC. The measurements in the middle of the cloud, average near 0.17 g m⁻³. Assuming again a linearly increasing LWC profile, the cloud liquid-water paths for flight 3 (before the afternoon minima) would be from 52–60 g m⁻². However, the average LWP obtained from the microwave radiometer during the same time was 135 g m⁻². It is noted that during the time when the radiometer retrieved the largest LWP, the FSSP could not accurately compute microphysical data since the angle between the wind and the platform was larger than 60°. The loss of data may explain some of the discrepancy between the FSSP and radiometer-derived LWP, but it appears that in flight 3 the FSSP underestimated the LWC.

The marked decrease in the cloud LWP during the afternoon has been observed in several studies of the stratocumulus-topped boundary layer, and the afternoon minimum in the cloud LWP seems to be the result of the decoupling of the cloud layer from the subcloud layer (Duynkerke 1989). Once the layers are decoupled, the transport of moisture from the ocean is drastically reduced and the cloud begins to dissipate due to a combination of the turbulent transport of moisture from the cloud layer and shortwave radiative heating. The partial evaporation of the cloud droplets may explain part of the predominance of small droplets measured during flights 2 and 3. Most of the microphysical measurements during flight 3 were made in the morning, however, and the origin of the small droplets on 8 July is not certain. The droplets may also have been the result of the activation of new small droplets or the remnants of a distribution of larger droplets that had evaporated.

The LWC measured in flight 5 is more nearly adiabatic throughout the cloud layer, and the FSSP-estimated LWP is only 20% less than the microwave radiometer-derived LWP. Since the cloud layer in flight 5 was not homogeneous, a filter was added to screen the microphysical data in order to compensate for the possibility of any bias produced by sampling clear air. All data points with liquid-water contents less than 0.04 g m⁻³ and wet-bulb temperature depressions greater than 0.1 K were considered to be taken “outside” of the cloud. Although the filter reduced the standard de-
viation of the LWC measurements, the average LWC was not significantly affected.

Since the droplet-size distributions did not vary much inside a cloud during each flight, the shape and the magnitude of the liquid-water content profiles were greatly influenced by the shape and magnitude of the total number density profile. The mean particle concentration measured during flight 2 generally increased with height from 15 cm\(^{-3}\) near cloud base to approximately 50 cm\(^{-3}\) near the top of the cloud. The mean total number density from flight 3 was variable through the cloud layer, with a peak value of 55 cm\(^{-3}\) at 947 mb. As with the LWC data, the standard deviation of the total number density data was extremely variable at each level and often was nearly as large as the mean values. The total number density from flight 5, which was obtained between 1950 and 2345 UTC on 10 July, ranged from 50 to 100 cm\(^{-3}\). The larger mean particle concentrations inside the cloud reflect the larger LWC observed by the FSSP. Measurements of total droplet number density were also made by the University of Washington’s C-131A research aircraft on 10 July approximately 140 km SW of SNI (Radke et al. 1989). The total number density of droplets measured in the middle of a 500-m deep cloud at 1600 UTC was near 50 cm\(^{-3}\).

In each flight, both the effective and mean radii slowly increased from the cloud base to cloud top. Although the shape of the droplet-size distribution from flights 2 and 3 differed from the distribution recorded during flight 5, the profiles of effective radius from those flights were nearly identical. The effective radius profile from flight 2 ranged from approximately 10 at cloud base to over 14 \(\mu m\) at cloud top, while the profiles from flights 3 and 5 increased from 10 to near 13 \(\mu m\). In comparison, the Multispectral Cloud Radiometer on board the ER-2 aircraft retrieved effective radii between 7 and 15 \(\mu m\) over the stratocumulus IFO region on 10 July (Nakajima et al. 1990). The mean radii measured from flights 2 and 3 were smaller than those measured from flight 5. The mean radii from the first two flights range from 5 to near 7 \(\mu m\), while the mean radii from flight 5 varied from 7.5 to 10 microns.

b. Radiation measurements

1) LONGWAVE RADIATIVE HEATING RATES

After the upward and downward infrared fluxes were corrected for the differential heating of the instrument, the upward fluxes were subtracted from the downward fluxes to find the net longwave flux throughout the boundary layer. Figure 4 shows a smoothed profile of the net infrared fluxes measured in the morning (1442–1554 UTC) during flight 2. A least squares polynomial fit was made through both the longwave and shortwave profiles that was later used to calculate the heating rates. The measurements show that the net flux increased to near zero as the platform entered the cloud, since the contribution from the downward flux increased with depth in the cloud. The net flux stayed near zero throughout the middle of the cloud, and the upward and downward fluxes were within 5 W m\(^{-2}\) of their blackbody values. Inside the cloud, the standard deviation of the measured fluxes for all flights was 2 W m\(^{-2}\). The observed downward flux decreased sharply above the cloud top, and a layer of strong net flux divergence was measured in the top 50 m of the cloud. The radiative heating rates determined from the flux profile is presented on the right-hand side of Fig. 4. The heating rates were calculated from the flux gradients over 2-mb intervals. A small amount of radiative warming was found near cloud base (0.2 K h\(^{-1}\)), while large radiative cooling rates were found in the top 4 mb (approximately 40 m) of the cloud on the order of 3 to 4 K h\(^{-1}\). The depth and the magnitude of the cooling compares favorably to the rates found in previous studies of marine stratocumulus by Stephens et al. (1978) and by Slingo et al. (1982).

A profile of the longwave radiative heating rates measured in the afternoon (2320–0030 UTC) is shown in Fig. 5. The profile is similar to the morning profile, with a region of radiative warming near cloud base, little warming or cooling in the middle of the cloud, and the largest flux divergence in the top 4–6 mb of the cloud layer. The maximum cooling rates in the profiles was 5 K h\(^{-1}\). However, the net flux profile in Fig. 5 shows that the depth of the cooling at cloud top during the afternoon increased from the profile measured during the morning. The change in the net flux profiles from morning to afternoon suggests the optical thickness of the cloud diminished during the afternoon. This change is consistent with the microwave radiometer estimation of LWP from flight 2, which indicated a sharp decrease in the cloud LWP after 1900 UTC.

The profiles of the net infrared fluxes measured during flight 3 are similar in shape to the profiles from flight 2, and the maximum cooling rate near cloud top was estimated to be about 4 K h\(^{-1}\). Since no rapid profile was made through the cloud layer on flight 5, and the cloud appeared to be broken, no clear profile could be made of the net longwave radiative fluxes. However, from the measurements made during the stepped levels toward the end of the flight the net flux above the cloud was consistently near −70 W m\(^{-2}\), and the net flux inside the cloud was approximately −10 W m\(^{-2}\). Assuming a profile similar to the other flights, the maximum longwave cooling rates at cloud top would be 5 K h\(^{-1}\).

2) CLOUD-TOP SOLAR FLUXES

The Eppley pyranometers measured the upward and downward total solar and near infrared fluxes throughout each of the flights. The downward solar fluxes in
the clear sky above the boundary layer measured during the long constant-level runs were normalized to local noon by

$$F_{\text{norm}}^4 = F_{\text{obs}}^4 \left( \frac{\mu_{\text{noon}}}{\mu_{\text{obs}}} \right)$$  \hspace{1cm} (11)

where $F_{\text{norm}}^4$ is the normalized flux, $F_{\text{obs}}^4$ the observed flux, $\mu_{\text{noon}}$ the cosine of the zenith angle at local noon and $\mu_{\text{obs}}$ the cosine of the zenith angle at the time of the observation.

The average normalized downward flux measured at the top of the boundary layer during flight 2 was 1135 W m$^{-2}$, and varied by an average of ±5 W m$^{-2}$ over each leg. However, the mean normalized downward flux measured above the cloud during flight 3 was only 1105 W m$^{-2}$. The normalized flux measured during flight 5 was 1131 W m$^{-2}$.

Computations of the downward solar fluxes at cloud top were made using the model described in section 5. The downward flux through the atmosphere was simulated for flights 3 and 5 using the temperature and moisture profiles from the CLASS soundings (Schubert et al. 1987) that were launched during each flight and the radiative properties of the clouds derived from microphysics measurements and Mie scattering theory. A solar zenith angle of 20° and a surface albedo of 25% were used in the calculations. The calculated normalized downward flux for flight 3 was 1108 W m$^{-2}$, while the calculated flux for flight 5 was 1152 W m$^{-2}$.

The differences in the moisture profiles above the cloud produced most of the differences in the downward fluxes between flights 3 and 5.

3) SHORTWAVE RADIATIVE HEATING RATES AND ABSORPTION

Figure 6 shows the net solar fluxes measured throughout the cloud during the profile run taken in the afternoon during flight 2. The strong flux convergence near cloud top shows the solar heating rate to be as large as the infrared cooling rate. Although the possibility of large solar forcing in stratocumulus has previously been speculated by Twomey and Cocks (1982) and Twomey (1983), the measured solar heat-
Fig. 5. Same as Fig. 4, except the profile was taken from 2320 to 0030 UTC.

ing rates appear to be unreasonably large. When visible fluxes were calculated by subtracting the measured near-IR fluxes from the measured total solar fluxes, both the visible and near IR parts of the spectrum contributed almost equally to the total solar heating rate. The large amounts of heating due to the visible light are also unreasonable given that nearly all absorption of solar radiation by water occurs in the near-infrared region.

One method used to measure solar absorption in cloud fields has been described by Rawlins (1989). This technique makes a correction for cloud-edge effects by taking the difference of measured solar absorption and the measured visible absorption, assuming absorption in the visible region in the cloud is negligible compared with that in the near-infrared. The broadband solar radiation properties of a cloud layer (absorptance $A$, transmission $T$, and reflectance $R$) are defined by this method such that

$$A + R + T + E = 1,$$  

(12)

where $E$ is an error term that describes the net energy exchange through the sides of the cloud layer. This term accounts for errors in sampling and for extinction in the visible wavelengths. The absorptance of the cloud layer is thus

$$A = \bar{A} - E,$$  

(13)

where $\bar{A}$ is the measured absorption and is defined as

$$\bar{A} = 1 - \frac{F_{top}^+ + F_{bot}^+ - F_{bot}^-}{F_{top}^+}.$$  

(14)

Assuming the visible absorption to be zero, the solar absorptance of a cloud layer is

$$A = \frac{F_{top}^\text{vis} + F_{bot}^\text{vis} - F_{bot}^\text{vis}}{F_{top}^\text{vis}} - \frac{F_{top}^+ + F_{bot}^+ - F_{bot}^-}{F_{top}^+}.$$  

(15)

Table 3 presents values for $\bar{A}$, $A$, and $E$ for flights 3 and 5 and the corresponding model simulations. The properties measured during the balloon flights show a wide range in magnitude and a large ratio between $E$ and $A$, which confirms the large uncertainties in the shortwave flux measurements. The values of $E$ calculated from the model runs are due only to visible absorption and are near zero.
Herman and Curry (1984) also report large values of cloud absorption in the visible from aircraft measurements of Arctic stratus. Herman and Curry note that a substantial observational uncertainty is present in bulk absorption measurements since the absorption is the small residual of four large fluxes. The uncertainty is especially apparent in the visible because the fluxes are computed as the difference between the total and near IR values. The problem of measuring solar fluxes in the cloud was also complicated by the inhomogeneity of the cloud layer.

4) SOLAR ALBEDO MEASUREMENTS

Along with the radiative heating rates, solar albedo data were also analyzed for the three flights. Since the solar albedo is easier to measure and more reliable than other observed cloud radiative properties, it is a useful quantity to compare with the results from radiative transfer models and other observations. Figure 7 shows a plot of the total solar, visible, and near-infrared albedos measured while the instrument platform was above the cloud during flight 2. The albedos were defined as the ratio of the flux measured by the downward facing pyranometer divided by the flux measured by the upward facing pyranometer at cloud top. The time series of the albedos were compared to a time series of the radiometer-derived LWP (Fig. 8). During the first two hours above the cloud, the instruments measured a total solar albedo that varied between 60% and 65%. The LWP during the same period generally increased from 70 to 100 g m⁻². From approximately 1815 UTC to 1915 UTC the albedo rose above 70%, while the LWP jumped to 150 g m⁻². The measured albedos
then decreased along with the liquid water path before the platform re-entered the cloud at 2042 UTC. The relationship between the total solar albedo and the LWP is presented in Fig. 9. As suggested by the above time series, albedo increases with LWP. Due to differences in the solar zenith angle at the time of measurement, two clusters of points appear at low LWP values. The measurements along the low LWP end of the larger cluster were obtained when the zenith angle was between $11^\circ$ and $25^\circ$, while nearly all of the points in the cluster of higher albedo values were measured when the solar zenith angle was between $35^\circ$ and $50^\circ$. The points along the rest of the larger cluster were measured when the zenith angle was between $11^\circ$ and $60^\circ$.

The visible component of the solar albedo was generally larger than the near infrared component while the platform was above the cloud. This is consistent with the optical properties of liquid water as more light is absorbed in the near infrared wavelengths by the cloud droplets. Figure 10 presents a time series of the
Table 4. Mean shortwave fluxes (in W m⁻²) from constant-level legs measured during flights on 7 and 10 July 1987.

<table>
<thead>
<tr>
<th></th>
<th>Tot</th>
<th>NIR</th>
<th>VIS</th>
<th></th>
<th>Tot</th>
<th>NIR</th>
<th>VIS</th>
</tr>
</thead>
<tbody>
<tr>
<td>7 July</td>
<td>F₄</td>
<td>1119.4</td>
<td>561.4</td>
<td>558.0</td>
<td>Leg 13</td>
<td>F₄</td>
<td>1131.3</td>
</tr>
<tr>
<td></td>
<td>F₃</td>
<td>711.4</td>
<td>337.4</td>
<td>374.0</td>
<td></td>
<td>F₄</td>
<td>657.2</td>
</tr>
<tr>
<td>albedo</td>
<td></td>
<td>.6355</td>
<td>.6010</td>
<td>.6703</td>
<td>albedo</td>
<td></td>
<td>.5809</td>
</tr>
<tr>
<td>NIR/VIS albedo</td>
<td>.8966</td>
<td></td>
<td></td>
<td></td>
<td>NIR/VIS albedo</td>
<td>.9727</td>
<td></td>
</tr>
</tbody>
</table>

The solar albedos measured during flight 5 were much more variable during this flight than the other cases. The average albedo was 64%, although it ranged between 30% to 75% before the platform re-entered the cloud. The near IR to visible albedo ratio was also more variable than the ratio measured from other flights (the albedo ratio ranged from 92% to 102%), and averaged over 96%. The higher value of the ratio arose from the fact that the visible portion of the spectrum tended to reflect less in this flight, rather than the near-IR part reflecting more. In order to illustrate this, Table 4 shows the mean downward and upward fluxes in watts per square meter measured during two legs taken above the cloud layer on 7 July and 10 July. The standard deviation of the LWP during the leg taken on 7 July was 17% of the mean, while it was 34% of the mean during the 10 July leg. Although the albedos were lower on 10 July, note that the difference in the upward visible fluxes between the flights was 46 W m⁻², while the difference in the upward near-IR fluxes was only 8 W m⁻². This suggests that the effects of heterogeneities on the radiative transfer through the cloud may be more important in the visible than in the near-IR.

The relationship between the near-IR to visible albedo ratio and the cloud LWP is presented in Fig. 11. The albedo ratio was averaged over 1-min intervals and matched with the LWP data for the times when the instrument platform was above the cloud. Data from the three flights and 13 July are presented and show that clouds with low LWP have higher albedo ratios than high LWP clouds. The albedo ratio decreases with increasing LWP until it reaches an asymptotic limit near 0.9. Unlike Fig. 9, the albedo ratio/ LWP relationship does not seem to have a strong de-

5. A comparison between observation and theory

a. Two-stream radiative transfer model

The two-stream radiative transfer model used in the radiative flux comparisons is a version of the model developed by Stackhouse and Stephens (1990). The model includes the effects of gaseous absorption, absorption and multiple scattering by cloud particles, Rayleigh scattering in the solar wavelengths, and e-type absorption in the longwave region. Absorptions by H₂O, CO₂, and O₃ are treated by the use of the K-distribution method (Chou and Arking 1980; Chou and Arking 1981), which includes high spectral resolution in the shortwave (50 cm⁻¹ wavebands) and longwave (20 cm⁻¹ wavebands) regions of the spectrum. Both Rayleigh scattering (Paltridge and Platt 1976) and e-type absorption (Roberts et al. 1976) are parameterized in the model by using empirical formulas to calculate the optical depth resulting from each

![Fig. 11. Scatter plot of the near-IR/visible albedo ratio and the microwave radiometer-derived LWP. The solid curve represents a least squares fit of the observations.](image-url)
process. Multiple layers in the model are vertically integrated using the adding method.

The composite droplet-size distributions obtained from each flight leg were incorporated into Mie theory to compute the total extinction coefficient ($\sigma_{\text{ext}}$), single scatter albedo ($\omega_0$), and asymmetry parameter ($g$) for each waveband at several levels inside the cloud. Two different model cases were used to illustrate the effects of differing microphysics. The first case was derived from the microphysics obtained during flight 5, which contained maritime clouds with a moderate LWP (77 g m$^{-2}$). The second case included the microphysics from flight 3, which measured clouds influenced by continental air. The droplet-size distribution was predominantly composed of small droplets and the LWP was 34 g m$^{-2}$.

The atmospheric profile assumed in case 1 was obtained from a CLASS sounding (Schubert et al. 1987) launched from SNI at 2006 UTC 10 July, while the instrument platform was inside the cloud. Above 16 km ($\approx 100$ mb) for the temperature profile and above 14 km ($\approx 150$ mb) for the moisture profile, the sounding was modified to merge with the midlatitude summer sounding of McClatchey et al. (1972). In case 2, the data from a CLASS sounding launched at 1211 UTC 8 July were also merged with the midlatitude summer sounding data for the temperature (above 16 km) and moisture (above 10 km) profiles. The ozone concentrations in both cases were taken from the McClatchey sounding. The atmosphere was divided into 60 layers in case 1 and 64 layers in case 2, and a solar zenith angle of 20° was used in the calculations. The surface albedo used in the shortwave section of the model was 25%, which was the mean albedo measured below the cloud layer during the flights.

**Fig. 13.** Calculated longwave heating rates for case 1.

**b. Longwave fluxes and cooling rates**

The net longwave irradiances were calculated for the wavenumber bands in the model that corresponded to the spectral range of the Eppley pyrgeometers. For both cases, the greatest flux divergence occurred in the 5-mb layer below cloud top. The flux divergence was greater for case 1 (shown in Fig. 12), with a difference in net flux of nearly 120 W m$^{-2}$ near cloud top, while the corresponding net differential flux in case 2 was 75 W m$^{-2}$. The calculated flux divergence tended to be larger than the divergence observed during the flights, which varied from 40 to 70 W m$^{-2}$ in the top 4 mb of the cloud. As in the observations, the calculated net flux approached 0 W m$^{-2}$ in the middle of the cloud and a layer of weak flux convergence was produced near cloud base by the contribution to the upward flux by the relatively warmer surface.

The calculated longwave heating rates for case 1 are presented in Fig. 13. The peak cooling occurred in the top 5 mb of the cloud, with a value near $-10^\circ$K h$^{-1}$. The calculated rates shown in Fig. 14 for case 2 indicate strong cooling extending through the top 20 mb of the cloud, with a peak value of $-1.5^\circ$K h$^{-1}$. The peak cooling rates measured inside the clouds varied from $-3$ to $-5^\circ$K h$^{-1}$.

The heating rate in case 2 is much different in depth and intensity than case 1. A comparison of the heating rates in each case suggests that the depth of the cooling inside the cloud is influenced mainly by the LWC structure inside of the cloud, rather than by the droplet-size distribution. This result has been confirmed by another model case (not shown) using a similar LWC structure to case 1, but with different microphysics. Stephens (1978) and Davies and Alves (1989) have
noted the dependence of heating rate profiles on the LWC structure in low-level clouds. Although some differences in the depth of the longwave cooling appear in the profiles taken during flight 2, the observations could not be used to confirm the relationship between the longwave heating rates and the vertical distribution of the LWC. Since the sampling interval of the FSSP was 5 s, it is believed the instrument could not make enough measurements during the profiles for an accurate retrieval of the LWC structure inside of the cloud. The LWC could only be obtained during the constant-level legs.

c. Shortwave fluxes and absorption

The profile of the net shortwave irradiances through the model cloud for case 1 is shown in Fig. 15. The upward and downward fluxes were calculated by matching the wavenumber bands with the spectral ranges of the filtered and unfiltered Eppley pyranometers. Unlike the observations, the flux convergence through the cloud is produced almost entirely in the near-infrared wavelengths. The net visible flux is nearly constant through the cloud, as expected from the theory of conservative scattering. The calculated change in net flux through the cloud in case 1 is 44 W m⁻², while it is approximately 28 W m⁻² in case 2. These changes are much smaller than the flux convergence measured during the early afternoon, which decreased over 200 W m⁻² through the cloud.

Figures 16 and 17 present the calculated shortwave heating rates for cases 1 and 2 respectively. The maximum heating in case 1 occurs in the top of the cloud and decreases roughly linearly through the cloud. The peak heating rate of 1°K h⁻¹ is much less than the heating rates observed by the tethered balloon instru-
d. Solar albedo

The analysis of observed albedos described above in section 4 indicated that the ratio between the measured cloud top near-IR and visible albedos remained relatively constant except for low LWP clouds and the inhomogeneous cloud observed during flight 5. In both cases, the measured visible albedo decreased relative to the near IR albedo. The differing albedo ratios may have been caused by either the heterogeneities in the cloud or by the different microphysics. The radiative transfer model was used to investigate some of the effects produced by the differences in the clouds between flights 2 and 3 and those from flight 5. Since the radiative transfer model assumes a horizontally homogeneous cloud, a comparison between the model and observed results may provide some information on the importance of heterogeneities on the cloud-top albedo.

A comparison of the near-IR to visible albedo ratios for the model results and the tethered balloon observations is presented in Table 5. The difference between the albedo ratios from case 1 and flight 5 were within the uncertainties in the albedo ratio measurements and the albedo ratio from flight 3 was only 4% lower than the ratio calculated from case 2. The differences in the albedo ratios were much less than the discrepancies between the computed and measured albedo ratios presented by Hignett (1987). Using aircraft observations of marine stratus clouds off the coast of Great Britain, Hignett found the difference between model-derived and observed albedo ratios to be as large as 15%. The clouds in that study had a LWP between 33.6 and 56.2 g m⁻² and the mean albedo ratio ranged between 0.80 and 0.85. By contrast, clouds with similar LWP measured during the flights had albedo ratios between 0.95 and 1.05.

Figure 18a illustrates the relationship between the total solar albedo and the radiometer-derived LWP for the observations and the model results. The small crosses represent the observations taken when the solar zenith angle was less than 30°. In the model calculations, the cloud LWP was modified for the 10 July case (case 1) by changing one of three variables. Line A presents the results from changing the total number density of the cloud droplets. Line B shows the changes in albedo due to differences in cloud depth. Line C represents the differences resulting from increasing and decreasing the effective radii of the cloud droplets. The model results match well with the observations. Figure 18b presents the relationship between the albedo ratio and the radiometer-derived LWP for the observations and the model calculations. The solid curve represents a least squares fit of the measurements. Unlike the observations, the calculated albedo ratio tended to slowly decrease or remain constant as the LWP decreased. Hignett (1987) reported a similar weak dependence of the near-infrared to visible albedo ratios on LWP in both model results and observations.

It was mentioned in section 4 that the observed albedo ratio increased in flight 5 since the visible portion of the spectrum reflected less during this flight. This result was also confirmed from the radiation measurements obtained on 13 July, a case with an inhomogeneous, low LWP cloud. (The standard deviation of the LWP was 36% of the mean during the cloud-top measurements.) This behavior was not observed in the model calculations, which show more equal decreases between the reflected visible and near-IR portions of the spectrum. The model results support the conclusion from section 4 that the effects of heterogeneities on the radiative transfer through the cloud are more important in the visible than in the near IR.

6. Summary and conclusions

During the FIRE marine stratocumulus IFO, an instrument platform was attached to a tethered balloon in order to obtain in situ measurements of the microphysical and radiative properties of stratocumulus. A sensitivity analysis of FSSP data was made to estimate the errors in the droplet number densities due to errors in measuring the sampling volume of the instrument. Errors in the number densities can be large in some situations and a fan or some device should be added to the tethered balloon FSSP in order to regulate the air flow through the instrument and improve the microphysical measurements.

| Table 5. Near-IR to visible albedo ratios for the model results and the tethered balloon observations. |
|---------------------------------|---------|
| Flight 3, leg 2                 | 0.92    |
| Case 2                         | 0.958   |
| Flight 5, leg 13                | 0.97    |
| Case 1                         | 0.944   |
The data from the balloon were analyzed for three flights. The results of the analysis showed that the shape of the droplet-size distributions remained constant throughout each flight and through all levels of the cloud. Flights 2 and 3 revealed clouds predominated by small droplets (2–8 μm in diameter), while flight 5 had a more typical marine stratocumulus distribution. The observed values of cloud LWC from flights 2 and 3 during the afternoon were much less than the adiabatic LWC, and may be due to the partial evaporation of the cloud droplets due to solar heating and the turbulent transport of moisture from the cloud layer. The difference between the FSSP-derived liquid-water path and the LWP measured by the NOAA/WPL microwave radiometer on SNI during flights 2 and 5 was only 20%, but the agreement in LWP between the two instruments during flight 3 was poor. The maximum cloud-droplet densities for flights 2 and 3 were near 50 cm\(^{-3}\) and near 100 cm\(^{-3}\) for flight 5. The effective and mean radius slowly increased with height inside the cloud. Although the effective radius did not vary greatly between flights, the mean radii observed during flight 5 were 45% larger than those measured during flights 2 and 3.

The observed longwave cooling rates exhibited a sharp peak in the top 40 m of the cloud in each case with maximum cooling rates up to 5°K h\(^{-1}\), which compare favorably with other tethered balloon measurements and aircraft measurements of marine stratuscumulus. The measured solar heating rates, however, were unreasonably large, and mainly due to the uncertainties in the measurements. It is suggested that the computation of the cloud absorptivity may be improved by attaching two radiation packages on the tethered balloon cable that could make simultaneous net flux measurements above cloud top and below cloud base.

The observed solar albedos showed that the visible radiation tended to be reflected by the clouds more than the near-IR radiation. The ratio of reflected-visible to reflected near-IR remained fairly uniform except for the nonhomogeneous and low LWP clouds. For these cases, the visible albedo decreased relative to the near IR albedo.

Two runs of the radiative transfer model were made using the optical properties derived from three different droplet-size distributions. The comparison between the calculated and observed radiative profiles is summarized below:

1) The calculated and observed net longwave irradiance profiles agree closely in shape and in magnitude to the cloud-top flux divergence. The maximum cooling rates from all three model cases compare well with the cooling rates observed by the instrument platform.

2) The distribution of the calculated longwave cooling inside the cloud appears to be influenced by the LWC structure inside of the cloud.

3) The calculated shortwave heating rates are considerably smaller than the rates observed during the flights.

4) For small zenith angles, the computed solar albedos generally matched the observations.

5) Unlike the results from Hignett (1987), the near-IR to visible albedo ratios measured during the flights compared well with the ratios computed from the corresponding numerical simulations. For liquid-water
paths less than 50 g m$^{-2}$, however, the observed albedo ratio was significantly greater than the calculated values.

From the analysis of the observations and the radiative transfer model calculations, the following conclusions have been drawn:

1) The observations of the visible and near-infrared albedos suggest that visible light is more strongly influenced by cloud heterogeneities than light at near IR wavelengths. Inhomogeneous clouds may allow for some of the visible light to “escape” through the sides and bottom of the cloud.

2) The observations of solar heating rates in stratumulus clouds were unreasonably large, especially in the visible part of the spectrum. More observations of stratumulus clouds are necessary, preferably by instruments with finer spectral resolution in the visible and near-IR wavelengths, in order to clarify this problem.

3) Measurements of the cloud microphysics show that both the drop-size distributions in marine stratumulus and the albedo of the cloud were affected by the presence of maritime or continental air masses. More measurements of marine stratumulus are needed to examine more thoroughly the variability of the microphysical structure in the stratumulus cloud fields, especially when they are influenced by large-scale, offshore airflows.

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