Some observations of the optical properties of clouds. 
I: Stratocumulus

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SUMMARY

Aircraft measurements of broad-band solar irradiances together with radiance measurements in the 1.2 to 2.1 μm region are presented for a stratocumulus sheet. Measurements of the solar albedo and cloud physics parameters indicate that the cloud was very uniform and optically thick. Broad-band observations of albedo agreed well with calculations but observed absorption was about double the model values. Narrow-band observations of vertical reflectance at three wavelengths suggested a marked increase in effective cloud droplet radius along the cloud, which was not observed. Some evidence that the discrepancy might be due to cloud top unevenness is presented.

1. INTRODUCTION

The atmospheric radiation budget is significantly affected by clouds and it is important to undertake more observational programmes to understand the processes involved. There have been a number of studies in which aircraft observations of broad-band solar irradiances have been compared with theoretical calculations but the results are inconclusive. For example, Herman (1977) and Stephens et al. (1978) found that the observed absorption of solar radiation exceeded predicted values, but in other instances (Slingo et al. 1982) satisfactory agreement was obtained. The accuracy of measurements of this type is degraded by instrumental limitations as well as by space and time fluctuation in the structure of even apparently uniform layer cloud. For many instances the magnitude of the absorption is comparable to uncertainty in its measurement. Hignett (1987) has measured albedo in two broad bands (0.3–3.0 μm and 0.7–3.0 μm). In his case study he found that although the albedo for the whole solar spectrum was in satisfactory agreement with calculations, the observed near-infrared albedo was less than the predicted values. Narrower band radiance measurements (Rozenberg et al. 1974; Twomey and Cocks 1982) have also concluded that the reflection in the near-infrared is lower than calculated indicating stronger absorption. Apart from experimental errors certain assumptions used in the calculations have been questioned, in particular the plane-parallel approximation (Welch et al. 1980), the omission of the effects of aerosol (Twomey 1976) and of large droplets (Wiscombe et al. 1984) from the droplet spectra.

In the present study airborne radiometers were used to measure solar irradiances above and below a stratocumulus sheet and irradiances in the 1.2 to 2.1 μm wavelength region above the cloud. This spectral region is of great interest because water and ice exhibit a variable and significant absorption of solar radiation, while cloud thermal emission is negligible. The optical depth of the cloud in this study was sufficient that in the near-infrared the cloud behaved as an infinitely thick cloud. This is an ideal situation for comparing observations and calculations of the reflectance of a cloud. Calculations were performed to study the sensitivity of the observations to the size of the cloud particles, their phase, and also variations in cloud top structure. The influence of an interstitial carbon aerosol was also considered.
2. Instrumentation

(a) General

The instrumental fit of the Canberra and Hercules aircraft of the Meteorological Research Flight is described by James and Nicholls (1976) and Readings (1985). Two particular key systems are explained in some detail in the next two subsections.

(b) Multi-channel radiometers (MCRs)

Details of the MCRs fitted to the Canberra aircraft are given in Doherty (1980) and summarized in Doherty and Houghton (1984). Essentially there were two separate radiometers viewing vertically upwards and downwards. The upward looking radiometer had an angular field of view of 10°, the lower one 1½°. The upper instrument had a single detector (denoted by E), the lower instrument had three detectors fitted during this period (denoted by A, B and D). Each detector had four filters which were selected every second and the radiances were integrated over 0·8 s. The filters were denoted by A1, A2, A3, A4, B1, etc. and channels on the downward facing instrument that have the same number (for example B3 and D3) made simultaneous measurements.

The channel wavelengths used are shown in Table 1 and are the same as in Doherty (1980) with the addition of the A detector and filter A2. Table 1 also shows the absorption characteristics in relative terms for the gases in a typical atmosphere. The refractive indices (real and imaginary, \(n_r\) and \(n_i\)) for water and ice are also shown; these values come from interpolating the values given by Irvine and Pollack (1968), for these wavelengths they are broadly consistent with more recent observations, for example Hale and Querry (1973) and Warren (1984). For A2 however, reference should be made to Warren and Shettle (1986). The absolute accuracy of \(n_r\) is probably of order 20%—there is a problem in producing an average value for \(n_r\) for the D2, E2 channels for water because of the strong absorption feature at 1·4 \(\mu\)m; however, this is not critical in this work.

<table>
<thead>
<tr>
<th>Channel description</th>
<th>Filter bandwidth ((\mu)m)</th>
<th>Refractive index of L (liquid water) (n_r)</th>
<th>Characteristics of gaseous absorption</th>
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<tr>
<td></td>
<td></td>
<td>(n_i)</td>
<td>(n_r)</td>
</tr>
<tr>
<td>D1, E1</td>
<td>1·23–1·26</td>
<td>L 1·322</td>
<td>8·9 (\times) 10^{-6}</td>
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<tr>
<td></td>
<td></td>
<td>I 1·297</td>
<td>1·4 (\times) 10^{-5}</td>
</tr>
<tr>
<td>D2, E2</td>
<td>1·35–1·38</td>
<td>L 1·320</td>
<td>7 (\times) 10^{-5}</td>
</tr>
<tr>
<td></td>
<td></td>
<td>I 1·296</td>
<td>1·5 (\times) 10^{-5}</td>
</tr>
<tr>
<td>D3, E3</td>
<td>1·53–1·57</td>
<td>L 1·317</td>
<td>1·2 (\times) 10^{-4}</td>
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<td></td>
<td></td>
<td>I 1·294</td>
<td>5·8 (\times) 10^{-4}</td>
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<td>D4, E4</td>
<td>1·83–1·87</td>
<td>L 1·311</td>
<td>1·4 (\times) 10^{-4}</td>
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<tr>
<td></td>
<td></td>
<td>I 1·292</td>
<td>6·3 (\times) 10^{-5}</td>
</tr>
<tr>
<td>B4</td>
<td>1·32–1·35</td>
<td>L 1·320</td>
<td>2·0 (\times) 10^{-5}</td>
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<tr>
<td></td>
<td></td>
<td>I 1·296</td>
<td>1·5 (\times) 10^{-5}</td>
</tr>
<tr>
<td>B2</td>
<td>1·62–1·66</td>
<td>L 1·316</td>
<td>6·7 (\times) 10^{-5}</td>
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<tr>
<td></td>
<td></td>
<td>I 1·293</td>
<td>3·0 (\times) 10^{-4}</td>
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<tr>
<td>B3</td>
<td>1·99–2·04</td>
<td>L 1·304</td>
<td>1·1 (\times) 10^{-3}</td>
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<tr>
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<td>10·6–12·8</td>
<td>L 1·148</td>
<td>0·16</td>
</tr>
<tr>
<td></td>
<td></td>
<td>I 1·410</td>
<td>0·12</td>
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The upward looking radiometer had a pot opal glass shutter that could be interposed to allow a direct calibration using the sun as a source. Details of the properties of this diffuser and the calibration technique are given by Doherty (1980). Doherty used a calibrated filament lamp to calibrate the downward facing instrument. In this study a large hot steel oven plate was used to compare signals between the upper and lower instruments by allowing them, in turn, to view this source. This provided a direct comparison of the gains of the D channels relative to the E channels. By assuming that the emissivity of the plate varied smoothly as a function of wavelength a calibration of the three B channels was also obtained. The absolute accuracy of these calibrations was probably of order 10% for the directly calibrated E channels and 20% for the indirectly calibrated B and D channels. Calibration of A2 was performed with external black bodies. Although the absolute accuracies were not high the results of this work depend on a high relative accuracy of two channels (i.e. D1 and D3) and the variability of the radiances. This is discussed later in this section.

Reflectance, \( R \), and transmittance, \( T \), of a single cloud or surface are all normalized to the value for a perfectly Lambertian diffuse reflector or transmitter. If \( I \) is the extraterrestrial intensity for a particular wavelength of the solar beam at a zenith angle \( \theta \) and \( L_R \) is the radiance measured by the downward facing radiometer, then

\[
R = \frac{\pi L_R}{I t \cos \theta}
\]

and for the upward facing radiometer measuring radiance \( L_T \)

\[
T = \frac{\pi L_T}{I t \cos \theta}
\]

where \( t \) is transmission of the atmospheric path, taken to be unity for all the window channels. Measurements of the solar beam transmitted through the diffusing shutter made in cloud-free conditions from the surface to an altitude of 13 km indicated negligible change with altitude at 1.25 and 1.55 \( \mu \)m. Rayleigh scattering is not significant at these wavelengths and was ignored in the interpretation of the results. These normalized values of \( R \) and \( T \) are hereafter denoted by the channel designators, i.e. \( D1 \) is the normalized reflectance of the layer beneath the aircraft at a wavelength of 1.25 \( \mu \)m and \( E3 \) the normalized transmittance of the layer above the aircraft at a wavelength of 1.55 \( \mu \)m. All these values are vertically measured, unless they are in parentheses with an angle equal to the angle of bank of the aircraft.

The transmission, \( t \), for the non-window channels has been calculated using the results of Doherty (1980), the path length through or reflected by the layer under investigation being assumed to be the minimum possible. Results where measurements have been made with different atmospheric paths have indicated satisfactory agreement although the absolute accuracy is probably further reduced by 5% for these channels. The resolution of the equipment is such that where it is viewing a perfect diffusing surface there are typically 300 counts in the absence of any attenuation. The strength of the water vapour absorption for D2, E2 and D4, E4 is such that these channels can be used only for high or medium level clouds.

A high relative accuracy between D1 and D3 channels has been proven by observing the reflection from the sea surface as the aircraft circled while banking (an orbit) in clear sky conditions. Figure 1 shows results for channel D3 on two orbits with 30° and 50° of bank. The reflectance is plotted against the angle between the viewing and the specular reflection directions. Comparison between the observed values of the ratio \( D3/D1 \) with values calculated from the Fresnel equation for salt water gave agreement to within 2%. Other pairs of channels could not be established to the same accuracy because the resolution of the data was not sufficient.
Figure 1. Reflectance of sea surface observed on 30° and 50° banked orbits at a wavelength of 1-55 μm ((D3)$_{30^\circ}$ and (D3)$_{50^\circ}$) as a function of the angle between the viewing and the specular reflection direction. Solar zenith angle of 50°.

(c) The ASSP cloud probe

The cloud droplet size probe used on this flight was the ASSP (Axially Scattering Spectrometer Probe manufactured by Particle Measuring Systems Inc.) and has been reported in detail elsewhere (Brown 1982). There was a fault experienced with the instrument during this flight, later traced to an amplifier. This resulted in the number of particles recognized to have passed centrally through the system, and which could therefore be sized, being unusually low compared with the total number of particles counted. The data could, however, be used in two ways. Firstly the particle spectra information where it existed was used to provide two average sizes; $r_m$, the mass average, and $r_e$, the effective radius, are defined by

$$r_m = \left[ \frac{\int n(r) r^3 \, dr}{\int n(r) \, dr} \right]^{1/3}$$

(3)
\[ r_e = \int n(r) r^3 \, dr / \int n(r) r^2 \, dr. \] (4)

This could be done only for limited periods particularly during HP2 (see later). Secondly the total number of particles counted could be used with the Johnson–Williams liquid water content device to derive a separate estimate of \( r_m \), this could be achieved all the time. In the second case the total concentration was increased by a factor 2, found necessary over a series of flights to obtain the best agreement in liquid water content. In both cases the value of \( r_m \) near the top of the cloud was 6·5 \( \mu m \) to within 0·5 \( \mu m \).

3. Experimental details

On 20 February 1981 a ridge of high pressure extended from Scandinavia south-westwards to Cornwall. Stratocumulus persisted throughout the day in the south-west approaches. Figure 2 shows the extent of the low cloud identified from the morning and evening NOAA-6 satellite passes, by the evening the cloud had dissipated near the coastline and did not extend as far to the west. There were some bands of cirrus also visible on the images, these were oriented north–south and associated with an old front over western Ireland.

![Figure 2. Sketch of flight area and high and low cloud detected by polar-orbiting satellite. ABC is position of L pattern flown by both aircraft. HP2 is the second profile flown by the Hercules, the first profile HP1 being close to C.](image)

The deployment of the two aircraft is schematically shown in Fig. 3. Both aircraft flew on a fixed ground pattern, ABC (shown in Fig. 2), the Hercules flying below the cloud, near the cloud top and just above cloud top. The Canberra could not fly through the cloud because of icing and performed slightly longer straight runs, which included runs AB and BC, above cloud top. Only the track BC has been used in this study because the cloud along AB was not optically thick: its solar albedo gradually increased from 60% to 80% from A to B. There was also thin cirrus evident in this area which was not
observed from B to C. The two Hercules runs (HR2 and 3) were unfortunately interrupted; the in-cloud run had to be aborted prematurely because of icing and the above-cloud run because the cloud top height was elevated at the point where a sharp cloud edge was reached; beyond this edge broken cloud existed. The Hercules run beneath cloud, HR1, was within a convective layer with small cumulus clouds. A profile descent (HP1) was performed around point C at the commencement of the experiment. At mid-cloud level the aircraft flew out beyond the cloud edge. The profile was halted and continued on a reciprocal heading with the result that the aircraft re-entered cloud near cloud base. After run HR3 the Hercules flew to another area of cloud to investigate the evolution of the cloud as part of a separate experiment. A very rapid profile HP2 to the north of C was carried out later in the day—the high rate of climb means that the liquid water content instrument would not have produced very accurate results but the vertical structure observed was valid.

Profiles of temperature, liquid water content (l.w.c.) and an average particle radius (effective radius, $r_e$, see Eq. (4)) are presented in Fig. 4 for both the profiles, HP1 and HP2, and for the level run HR2. The increase of liquid water content and particle size with height is a common observation in stratocumulus, for example Slingo et al. (1982). Particle concentrations (not shown) were also observed to be fairly constant with height. The dashed lines shown in Figs. 4(b) and (c) represent an idealized single cloud layer with liquid water content increasing linearly with height and the size of the particles also increasing within four layers consistent with the increase in l.w.c. and maintaining a constant concentration of droplets. The maximum in l.w.c. at about 1.5 km (HP2), the increased thermal stability below 1.5 km, and the fact that $r_e$ does not continue to decrease below this level suggest the existence of two distinct layers decoupled at about 1.5 km. The lowest layer had a l.w.c. very close to the adiabatic value given the uncertainties in deciding upon cloud base, and the upper layer a value just below adiabatic value. Although this complicates the description of the cloud, the dashed lines shown on these graphs are a realistic description of the observed cloud properties for carrying out numerical prediction of radiative transfer, particularly as the total optical depth was large.
4. Models

To calculate broad-band solar irradiances a two-stream model based on the Delta–Eddington approximation and developed by Slingo and Schrecker (1982) was used. Two vertical distributions of effective radius and liquid water content have been considered. Firstly, the values representing the observed profiles shown as the dashed lines in Figs. 4(b) and (c) were used. The approximate vertical optical depth, $\tau$, of the cloud is 48, evaluated from

$$\tau = \frac{3}{4} \int \left( \frac{\text{lwc}}{r_e} \right) dh$$  \hspace{1cm} (5)

where $h$ is height. The second vertical distribution used constant values of $r_e$ and lwc. With fixed $\tau$, negligible differences in cloud albedo and absorption were found in model calculations using the vertical variation in $r_e$ compared with a constant value of 7-7 $\mu$m—the value observed near cloud top. To simulate the sensitivity of the calculations to variations in $\tau$ and $r_e$ a constant vertical distribution has been used, values of $r_e$ were 7-7, 6-2, 9-6 and 15-4 $\mu$m. The first three radii represent the observed and its likely minimum and maximum values and the last value is included to demonstrate the change if the cloud were composed of much larger droplets. The parametrization of cloud properties using $r_e$ and lwc was used, Slingo and Schrecker (1982). To validate the use of the parametrization in this study, model calculations were performed using Mie calculations of the cloud properties based upon the observed size distribution of droplets at cloud top for each of the 24 bands of the model. This yielded cloud albedo of 80-6% and absorption of 7-1% compared with 82-4% and 7-0% respectively for the parametrization; the differences are smaller than those resulting from the likely errors in $\tau$ and $r_e$.

The model was run with observed humidity and temperature profiles in much the same manner as Slingo et al. (1982). The variation of zenith angle over the period was
small and the mean angle was considered to be appropriate. This model has been compared with other schemes, for example Slingo et al. (1982) and Hignett (1987), and found to be reasonably consistent.

An interstitial carbon aerosol was included in the model cloud layer. This was achieved using the absorption coefficients of the soot model defined by the Radiation Commission (1986) for each of the spectral bands of the two-stream model and adjusting the cloud single-scattering albedo to take account of the additional absorption. Scattering by the aerosol has been ignored as scattering will be predominantly from cloud droplets. Different volume mixing ratios of soot within the cloud were used. Chylek et al. (1984) have shown that carbon within the cloud droplets is about twice as efficient at absorbing radiation compared with the same quantity of interstitial carbon aerosol. Thus the results also represent the behaviour of carbon within the droplets at about half the volume mixing ratio of the interstitial aerosol.

Radiances above the cloud have been computed using a Monte Carlo model developed by Ley (1983). Mie theory was used to calculate single-scattering albedos and phase functions from the observed drop size distribution for three wavelengths. As with the two-stream model this model is not sensitive to drop size distribution, which can be varied considerably provided the effective radius remains constant. Statistical errors of the results associated with the technique are less than 1%.

The vertical optical depth $\tau$ was taken to be constant with wavelength, with a value of 48. The parametrization used in the two-stream model gave marginally higher optical depths for these wavelengths. Vertical structure was introduced in the manner already described. There was again no significant difference obtained when comparing modelled vertical radiances derived from the vertically varying $r_v$ with those derived using a constant value of 7.7 $\mu$m, so a constant profile was used in other simulations. Interstitial carbon aerosol was introduced in the same manner by adjusting the single-scattering albedo.

Some horizontal structure was also introduced. This amounted to having a two-dimensional ridge and trough structure similar to Wendling (1977). Ridges and troughs were set to be of equal length and results could either be averaged over a whole number of features or be just from ridges or troughs. The maximum vertical optical depth was kept at 48 and the solar azimuth was randomized in relation to the pattern direction.

5. Comparisons of observations and calculations

(a) Solar irradiances

The albedo of the cloud averaged in 60-second intervals from B to the point when the aircraft went into cloud on HR3 gave six values with a mean of 0.82 and standard deviation of 0.02. The Barnes radiometer gave a mean and standard deviation of $-12.3 \pm 0.1^\circ\text{C}$ for the cloud-emitting temperature. Data from the A2 channel on the Canberra also indicated very little change in cloud top temperature from the cloud edge to B. With the Hercules travelling at 100 m s$^{-1}$ the cloud over this 60 km length was very uniform on the scale of kilometres. The net flux into the cloud at cloud top (calculated from the three 60 s means closest to point B) was in the range from 98 to 110 W m$^{-2}$. The data beneath the cloud on HR1 were difficult to interpret because for the first few minutes the lower pyranometer’s temperature had not stabilized and its output was clearly drifting; the run was further complicated because it intersected a low cumulus cloud. The best estimate for the downward solar irradiance in regions away from the cumulus cloud and the stratocumulus edge, i.e. near B, was 35 to 47 W m$^{-2}$ (from the three 60 s samples closest to B). Using a sea surface albedo of 10.6 $\pm 1.0\%$ measured on the other run (AB) beneath cloud, which did not suffer from these problems, the net flux beneath the
stratocumulus was in the range 31 to 42 Wm⁻². This gives a cloud absorption between 56 and 79 Wm⁻², or in terms of the available solar irradiance at cloud top between 10 and 15%. The range includes temporal variations which would be removed if measurements above and below the cloud could be made simultaneously. The spatial variabilities of 60s means of the net flux both above and below cloud are both around 10 Wm⁻², hence giving a range of about 20 Wm⁻² in absorption; however, one would anticipate these two being correlated so that the range of values would probably be reduced around the mean absorption of 13%.

The main contribution to instrumental inaccuracies is the relative calibration of the upper and lower pyranometers. These are routinely calibrated in an integrating sphere and their relative response is known to about 1%. However, there were differences in the experiment compared with the calibration because above cloud the upper instrument was primarily measuring the direct beam whereas the lower instrument was measuring a diffuse one. Differences in sensitivity because of non-ideal cosine responses could be as large as ±2% or ±10 Wm⁻² in the net flux into cloud top. This would also alter the albedo by ±2%.

The two-stream model predictions for albedo and absorption are given in Table 2 and compared with the observed range. Some of the model predictions of absorption are presented to 0.1% merely to indicate the lack of sensitivity to ±25% changes in optical depth and effective radius. The observed albedo is well described by the model and its variability could be due to a not unrealistic variation (±25%) in optical depth. The predicted absorption of 7% is about half the mean observed value, and even doubling the effective radius will only just bring the model values in line with the likely minimum value observed. Unless there was considerable change in the cloud between making measurements above and below it, there is evidence that the absorption in the cloud was higher than predicted by this model.

Interstitial carbon aerosol with a volume mixing ratio of 5 x 10⁻¹² could also account for the high absorption, see Table 2. Patterson et al. (1984) have made measurements in the north-eastern U.S.A. on both interstitial aerosol and cloud water and concluded that in their samples interstitial aerosol was more important than contamination of the droplets in increasing cloud absorption. However, the typical aerosol absorption coefficient was about one order of magnitude smaller than those required to reproduce the observed total absorption found in this work.

| TABLE 2. COMPARISON OF OBSERVED AND MODELLLED CLOUD ALBEDO AND ABSORPTION |
|--------------------------------|-----------------|----------------|
| Observed range for 18 km closest to B | Albedo % | Absorption % |
| 80-82 | 10-15 |

<table>
<thead>
<tr>
<th>Model values</th>
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<tr>
<td>r</td>
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Figure 5. Upper panel: Variation of means observed from cloud edge (near C) to B measured above the cloud of: Barnes cloud top temperature; solar albedo of cloud; vertical reflectance, D1; Ratios of vertical reflectances D3/D1, B3/D3 (● CR1; × CR2). Lower panel: Measured within cloud top: mass average radius, $r_m$; liquid water content.

(b) Reflectance in narrow spectral bands

Figure 5 shows the means and the standard deviations of 60 1-second observations of albedo and cloud top temperature measured on HR3 as a function of the distance from cloud edge to B. Also shown are the one-minute averages of the reflectance, D1, of the cloud and the ratio of mean reflectances D3/D1 and B3/D3 for the two Canberra runs, CR1 and CR2. The reflectance at 1.25 μm (D1) was very constant apart from near the cloud edge. The standard deviation of the 15 observations making up these ten means was between 4 and 12% of the mean. The ratios D3/D1 and B3/D3 show a consistent pattern between the two runs and indicate real variability of the radiative properties along this 60 km length. The variability of the mass mean radius, $r_m$, and liquid water content on HR2 is also shown. Regrettably they do not extend over the whole run, but the fractional change in the average microphysical properties within 30 km of the cloud edge was relatively small compared with the corresponding changes in D3/D1 and B3/D3. Similar variability of near-infrared reflectances has been observed by Twomey and Cocks (1982).

The Monte Carlo model has been used to simulate these results. Figure 6 shows the two reflectance ratios plotted against a reflectance and demonstrates the sensitivity of
these ratios to the effective radius and to the difference between liquid water and ice spheres. The main reason for the changes is different single-scattering albedos rather than changes in the scattering phase function. There is no change in these graphs if the optical depth is doubled—i.e. the transmission of the cloud at these wavelengths is negligible. The model results are for plane-parallel clouds and clearly demonstrate the ability to infer effective radius and optical properties of the droplets. The assumption that ice particles are spherical is not valid; this point is discussed in part II of this paper (Foot 1988); the calculations indicate, however, the sensitivity to changes in the refractive index. The results for an effective radius of 7.7 $\mu$m should fit the observation most closely.

Figure 7 repeats the water sphere calculations shown in Fig. 6 and includes the
1-minute-mean observations shown in Fig. 5, excluding the two values near cloud edge. The data have been shifted within the systematic errors discussed to provide the best fit to the calculation with \( r_e = 7.7 \, \mu m \). Figures 7(a) and (b) both indicate that the variability of the data can be matched to the calculations provided \( r_e \) changes from 7.7 to 15 \( \mu m \) along the run. The fact that both graphs show a similar trend compared with the model indicates that the refractive index changes from wavelength to wavelength in a manner very similar to liquid water and, with reference to Fig. 6, very unlike ice. Also studied was the variation of \( B2/D1 \) which was smaller and is not shown here but was also consistent with these conclusions.

6. DISCUSSION OF THE COMPATIBILITY BETWEEN THE OBSERVED AND PREDICTED REFLECTANCES

(a) Particle size

The comparisons shown in Fig. 7 indicate that the observations are compatible with a plane-parallel water cloud with a horizontal variation of \( r_e \) from 7.7 \( \mu m \) to 15 \( \mu m \). Such a variation was not seen, though it is possible that there was an area of larger droplets not sampled by the aircraft. The author is not aware of any data showing large horizontal changes of \( r_e \) in stratocumulus.

Wiscombe et al. (1984) have discussed the possibility of \( r_e \) being significantly altered by the existence of large droplets (radius \( \geq 50 \, \mu m \)) not detected by cloud probes. A two-dimensional Particle Measuring Systems Inc. probe was also run during HR2. Some large spherically shaped droplets were detected, but their concentration was 2 to 3 orders of magnitude too low to account for \( r_e \) increasing to 15 \( \mu m \). Results presented by Nicholls (1984) of size spectra up to precipitation droplets for a number of stratocumulus clouds also indicated that the concentration of large droplets was not sufficient to alter \( r_e \) significantly, determined by integrating the droplet spectra up to about 30 \( \mu m \). The results presented by Hegg (1986) of quite large \( r_e \) values seem very unrealistic.

In the next subsections other possible explanations are explored.

(b) Ice crystals and pollutants

No ice-shaped particles were detected by the 2D PMS probe. If glaciation had started in one area of cloud then it would be expected to convert most of the water to ice very rapidly with the result that the Johnson–Williams liquid water device would have detected a dramatic reduction of liquid water in that part of the cloud, which was not observed. As already pointed out the spectral signatures at the various wavelengths are also consistent with the refractive indices of water. The possibility of ice being the cause of these results therefore seems remote, even though the cloud top temperature was about \(-10^\circ C\).

When a carbon aerosol was introduced into the model at the mixing ratio necessary to account for the broad-band absorption it made very little difference to these predictions because water at these wavelengths is a significantly strong absorber. If twenty times this amount of carbon aerosol were introduced into a plane-parallel cloud with \( r_e \) equal to 7.7 \( \mu m \) the ‘W’ points in Figs. 7(a) and (b) were predicted. This quantity of carbon would produce a broad-band solar absorption of 43\%, see Table 2, clearly not consistent with section 5(a) or with experience. The predictions using the model aerosol were for higher values of \( D3/D1 \) and \( B3/D3 \) whereas the observations gave lower values compared with the model without aerosol. Thus although carbon aerosol may play a role in the total solar absorption it does not account for these near-infrared results. Similarly the results from a stratus sheet presented by Hignett (1987) suggested that the observed albedo in
the near-infrared was lower than predicted by the two-stream model, a result not consistent with carbon contamination, which would have had the greatest effect at visible wavelengths.

(c) **Cirrus overhead**

The model was used to simulate the effects of a thin cirrus layer with an optical depth of 0.5 composed of ice spheres of effective radius 15 μm overlying the stratocumulus sheet with effective radius 7.7 μm. The results are marked as points 'X' on Fig. 7. They indicate a trend consistent with the data although the magnitudes of the variations observed are much higher and would require a far thicker cirrus sheet, which would have been detected by the pyranometers on the Hercules and the E channels on the Canberra’s upper radiometer. Also the area where cirrus was actually noted, i.e. near to B, gave the best agreement with the model plane-parallel cloud with \( r_e \) equal to 7.7 μm. The treatment of the cirrus cloud particles as spheres with effective radius 15 μm is very crude but in terms of transmittance of such a layer is not without some reality, see part II (Foot 1988).

The existence of thin cirrus in the area is not believed to explain the reflectance results.

(d) **Irregularities in cloud top**

The striated cloud top addition to the Monte Carlo model described in section 4 was used to introduce horizontal structure at cloud top. Patterns of wavelength 200 m and depth 100 m representing typical scales of undulations averaged over whole wavelengths reduced \( D3/D1 \) and \( B3/D3 \) by only 3%. Smaller scale patterns, wavelength 25 m with depth of 100 m, gave larger reductions in these ratios, for \( r_e \) equal to 7.7 μm the points

![Figure 8. Comparison of radiance calculations with individual observations. +: As Fig. 6 for B3/D3 and water sphere only; bars show variations for \( r_e = 7.7 \) μm between peak and troughs for striated clouds (peaks have higher values of D3); ——: striation wavelength 25 m, depth 12.5 m; -----: striation wavelength 100 m, depth 50 m; ||||: striation wavelength 25 m, depth 100 m; ● individual observations.
(a): Results close to B. (b): Results midway between B and C. Note that at cloud top 12.5 m is equivalent to an optical depth of unity.](image-url)
'Y' in Fig. 7 were calculated. These small-scale features might represent the entrainment of dry air from above the cloud giving rise to vertical columns of varying optical depth near cloud top. If this process was taking place then it is conceivable that there were variations of $r_c$ on the same scale. If $r_c$ was increased to 9.6 $\mu$m and combined with this small-scale structure then the points 'Z' were obtained, which closely match the departures from a cloud with a plane top and $r_c = 7.7$ $\mu$m.

Further evidence for cloud top structure being important was obtained by studying the individual 0.8 s observation of reflectance at B3 and D3 wavelengths which were made simultaneously. At the Canberra speed this corresponds to a distance of 130 m along-track, and the across-track sampling width was only 10 m. Figures 8(a) and (b) present data from two typical 1-minute periods in the run. Figure 8(a) shows a highly variable ratio found near ground position B where the average $B3/D3$ ratio was in good agreement with the plane-parallel cloud with $r_c = 7.7$ $\mu$m. Figure 8(b) shows results with very little variability in the area where $B3/D3$ was lowest. On Fig. 8(a) are marked bars showing the values predicted from just peaks or just troughs for three different striations. For the results near B (Fig. 8(a)) the large variability might be associated with large wavelength features which would imply that individual observations could be sampling just peaks or troughs. For the result in Fig. 8(b) any structure must be small scale in terms of the 0.8 s sampling time or otherwise the data would be variable. Such a structure, if it is as deep as 100 m, would result in a reduction of $B3/D3$, shown as points 'Y' in Fig. 7, which is the average of the peaks and troughs given in Fig. 8(a).

7. Conclusions

The stratocumulus sheet studied in this work has provided a particularly interesting set of data because its large optical depth has meant that the reflecting properties in the near-infrared are insensitive to changes in optical depth.

On scales of order 10 km, away from the cloud edge, the mean near-infrared reflectance values showed little variation at wavelengths with weak water absorption but considerable variations with stronger absorption. Under the plane-parallel assumption it is possible to model this variation by allowing the effective radius to vary from its observed value to a value almost twice as large; such a variation was not consistent with the observations. Because the reflectances at the various channel wavelengths changed in a manner consistent with the optical properties of water, pollution within the cloud is not believed to explain the results. A simulation of carbon aerosol within the cloud increased the discrepancy between observation and calculations. There was evidence that the cloud top structure played a role. In a region where agreement between the observations and the model was best, individual (130 m x 10 m) reflectances showed considerable variation, whereas in the region where the agreement was worst the variability of the individual values was least. It was possible with the model to simulate lower vertical reflectances by introducing small-scale (~25 m wavelength) and deep (~100 m) voids in the cloud top. In a real cloud entrainment at its top will give rise to vertical columns of denser and less dense cloud. If these features are on horizontal scales <100 m then they will affect near-infrared vertical reflectances. Future studies should include the ability to study microphysical data on scales less than 100 m.

Allowing for the uncertainties introduced by the time delay in making measurements above and below cloud, there was some indication that the total absorption within cloud was higher than predicted. Interestingly the cloud top structure invoked to explain the vertical reflectances has only a small effect on the net irradiances at cloud top. This is because this structure is vertically oriented and so the vertical reflectances are changed
whereas the albedo is not. Other models of horizontal structure may be more efficient at increasing total absorption within a cloud.

The results of this work demonstrate the usefulness of multi-spectral measurements, particularly in the near-infrared, to test fully our understanding of radiative transfer within clouds. This understanding is not complete and the role that horizontal non-uniformity may play requires further study.

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