A study of the short-wave radiative properties of marine stratus: Aircraft measurements and model comparisons

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SUMMARY

Aircraft measurements of the short-wave albedo of a marine stratus cloud are described and compared with results from radiative transfer models of plane–parallel clouds. The albedos are measured in the visible and near-infrared parts of the solar spectrum using clear dome and red dome pyranometers. Using all the available liquid water content data and the measured microphysical properties, two model clouds are constructed to represent the maximum and minimum liquid water paths encountered during the experiment. Model albedo calculations based on these clouds are compared with the range of albedos measured just above cloud top. The natural variation of liquid water path at the cloud edge is exploited as a further qualitative assessment of the models’ behaviour.

1. INTRODUCTION

Clouds are of fundamental importance in the determination of the atmospheric radiation budget. A long term aim of cloud and radiation studies is to develop efficient and accurate methods for the specification of cloud radiative properties in numerical general circulation models. In the study of radiative transfer through clouds considerable simplifications are possible if attention is restricted to plane–parallel layers, thus removing the effects of finite cloud geometry. This idealization introduces an extra degree of uncertainty when attempting to compare the results of such simplified models with measurements of real clouds. Nevertheless the comparison of model results with experimental measurements is a necessary part of the development process. The purpose of this paper is to present aircraft measurements of the short-wave albedo of a single, extensive layer of maritime stratus and to compare these data with the results of model calculations of plane–parallel clouds. Irrespective of the difficulties of making direct quantitative comparisons an attempt is made to assess the qualitative behaviour of the models used.

Ideally a rigorous experimental programme would include measurements made over a wide range of solar zenith angles, cloud liquid water paths and droplet sizes. However, such opportunities are rare. In the present study advantage has been taken of the near-ideal conditions of a single layer of cloud with clear skies above. This lack of any high cirrus cloud in particular simplifies the modelling problem considerably. A simplification has been attempted by making measurements at an almost constant solar zenith angle and exploiting the natural liquid water path variations within the main body of the cloud and at its edge. Furthermore, by measuring the cloud albedo separately in the visible and near-infrared parts of the spectrum, and taking their ratio, it is anticipated that the dependence on the droplet size spectrum can be reduced (DeVault and Katsaros 1983). In this way attention is focused on the effect of liquid water path variations on the cloud albedo. Measurements of cloud albedo in the visible and near infrared have also been made recently by Herman and Curry (1984). This way of splitting the solar spectrum is particularly useful as it separates the active regions of the two major gaseous absorbers of solar radiation: ozone in the visible and water vapour in the near infrared. Also, since pure liquid water absorption is restricted to the near infrared, the effects of droplet
scattering, which occurs in both the visible and near infrared, should be reduced by taking the ratio of the albedos in these two spectral bands.

There has been a number of aircraft studies of cloud radiation properties (e.g. Slingo et al. 1982; Stephens et al. (1978) summarizes several others); all are agreed that cloud absorption is intrinsically the most difficult measurement to make and the results the most controversial. In the experiment described here the aircraft could not get beneath the cloud and although measurements were made in the body of the cloud it was not felt that the absorption, which in principle could be calculated, would constitute a meaningful quantity, and attention was restricted to the albedo. This is a more reliable quantity to measure since downward and reflected fluxes are recorded simultaneously. However, the problem of defining the cloud physical properties at the time of the albedo measurements still remains. In this study a composite description of the cloud liquid water content has been produced for use mainly with the models of Stephens (1978b) and Slingo and Schrecker (1982). The aircraft instrumentation and the cloud physical properties are described in sections 2 and 3. In sections 4 and 5 radiation measurements made over the main body and edge of the cloud are presented along with model calculations. In section 6 an attempt is made to assess the qualitative and quantitative behaviour of the models over the range of liquid water paths encountered.

2. AIRCRAFT INSTRUMENTATION

The experimental data were obtained with the Hercules W Mk2 aircraft of the Meteorological Research Flight (MRF). For the purposes of this study the aircraft was equipped with three Eppley pyranometers (model PSP); a clear-domed instrument (WG295, passband approximately 0.3 to 3 \( \mu \)m) and a red-domed instrument (RG715, passband approximately 0.7 to 3 \( \mu \)m) faced downward and a single clear-domed instrument faced upward. At the time of the experiment no red-domed instrument was available to fill the remaining upward facing position. When irradiated by the direct solar beam the upper pyranometer was corrected for the pitch and roll of the aircraft using data from the aircraft’s inertial navigation system. This correction is normally no more than a few per cent. Each instrument is fitted with ‘obscurers’, small black posts or plates which mask various parts of the aircraft from the field of view. The lower instruments were exposed to essentially diffuse fields, the sea surface or cloud top, and their readings were increased by 0.5% to allow for the obscured fraction of the hemispheric field of view. The upper pyranometer was fitted with a cylindrical pillar which, when the aircraft was flown directly away from the sun, obscured the instrument from the direct solar beam and gave a measurement of the diffuse radiation. This technique is fully described by Foot et al. (1985). In all the runs of interest here the upper pyranometer was either obscured from or irradiated by the direct solar beam. In this latter case the correction required for the presence of the obscurer was negligibly small.

A thermistor buried in the body of each pyranometer was used to monitor the temperature of the thermopile heat sink. The pyranometers will give a spurious reading if they are not in thermal equilibrium, so data were always taken from straight and level runs with sufficient time allowed for equilibrium to be reached following any manoeuvring. The cosine error of the particular upper pyranometer used is known (Foot et al. 1986) to be significantly worse than the typical curve shown by Nast (1983); at the solar zenith angles encountered the instrument reads about 2% low. Such a correction has not been applied to the data used here, but it would not change the conclusions.
reached; the effect of including it would be to reduce the measured albedos by about 2%, although, of course, it does not affect their ratio.

A Rosemount platinum resistance thermometer and Cambridge Instruments dew-point hygrometer provided the air temperature and dew-point while cloud liquid water content (l.w.c.) was measured with a Johnson–Williams meter. Aircraft height (below approximately 1500 m) was from a radar altimeter. Cloud microphysical data were acquired by a Particle Measuring Systems Forward Scattering Spectrometer Probe (FSSP) measuring droplet concentration in 15 channels, each 1 μm wide, from centre radii of 1.5 μm to 15.5 μm. The particular instrument used is known to suffer from a form of noise in the lowest two channels; this has the effect of causing these channels to register too high a count, but this is not thought to have affected the present results significantly. The droplet spectra are produced as 1-second averages.

3. CLOUD PHYSICAL PROPERTIES

During spring and early summer there are frequent occurrences of extensive stratus or sea fog forming in easterly winds over the relatively cool North Sea. On 7 June 1984 low status, that had formed in a north-easterly flow, extended approximately from 54°N to 57°N and from the east coast of Britain to approximately 1.5°E. There was a well-defined edge at the northern and eastern extent of the cloud, which extended north-eastwards during the day. Aircraft runs to measure albedo and cloud physical properties were carried out above and in the main body of the cloud to a minimum height of 150 m. The average cloud top height was approximately 230 m, with the base at or very near the sea surface. Further runs made crossing the cloud edge will be described in section 5. During the course of the experiment there was no other cloud present, except during the latter part of the afternoon when a small amount of cirrus intruded from the north, but not sufficient to have any significant effect.

Cloud top structure, which appeared to take the form of rolls with wavelengths of a few kilometres, extended over the upper 70 to 80 m. These irregularities produce large variations in the vertical liquid water path (l.w.p.) and result in a cloud which is far removed from the plane-parallel cloud on which it is convenient to perform model radiative calculations. To help overcome this problem and give a realistic representation of the cloud structure a composite vertical l.w.c. profile was formed. Figure 1 shows the l.w.c. plotted as averages over 5-metre height bands; the bars indicate the variation between the maximum and minimum values encountered in each band. The data come from level runs and profiles corresponding to horizontal traverses varying between 10 and 50 km. Below approximately 180 m the l.w.c. was very uniform at any particular level with the vertical variation very close to the adiabatic. Above this level the l.w.c. varied between values indicating dry air and those which would be achieved by continued adiabatic ascent of saturated air from below. The maximum values were reached around 230 m, decreasing to the maximum height of the cloud tops at about 260 m.

To permit comparison with the model radiative calculations of section 4 two model clouds were constructed from the data of Fig. 1 to represent the maximum and minimum l.w.p. which could have been encountered. Extrapolation of the measured profile below 180 m on an adiabatic assumption gave zero l.w.c. at approximately 23 m. The maximum l.w.p. was formed by continuing this profile to 230 m and then decreasing linearly to zero at 260 m; this gives a l.w.p. of 56.2 g m\(^{-2}\). The minimum l.w.p. cloud has the same profile below 180 m and then decreases linearly to zero at 210 m to give a l.w.p. of 33.6 g m\(^{-2}\). These profiles effectively form an envelope to the range of l.w.c. values shown in Fig. 1.
Figure 1. Cloud liquid water content against height above the surface. The triangles denote the average in each 5 m height band; the bars are the range of values encountered.

A measure of the droplet spectrum is required as part of the input for the radiative calculations of section 4. Figure 2 shows an example of the droplet spectrum averaged over a 10-minute level run (about 60 km in distance) at 150 m. The FSSP spectrum gives an integrated l.w.c. about 30% lower than the Johnson–Williams meter. The values from the Johnson–Williams are very close to those expected from adiabatic ascent. For

Figure 2. The cloud droplet distribution measured by the FSSP at a height of 150 m. The triangles are the averages in each 1 μm-wide channel.
example, at 200 m, the maximum measured l.w.c. is 6% less than the adiabatic value, and the mean is within 30%. No measurements of droplet concentrations beyond radii of 15-5 μm were available; however, observations of similar clouds by, for example, Herman and Curry (1984), show contributions from drops in the range 15-5 to 23-5 μm to be between zero and 8% of the integrated l.w.c. This is insufficient to account for the present discrepancy, so to give a more representative spectrum the number concentration in each channel has been normalized such that the integrated l.w.c. is equal to the Johnson–Williams l.w.c. (e.g. Nicholls 1984). The slight upturn at the lowest radii probably arises from the noise problem associated with this instrument, but should not have affected subsequent calculations significantly. The effective radius of this spectrum, \( \int n(r)r^2dr/\int n(r)r^2dr \), where \( n(r) \) is the number concentration at droplet radius \( r \), was 6.4 μm at 150 m, increasing to 7.4 μm at 230 m (in cloud).

4. ALBEDO MEASUREMENTS AND MODEL COMPARISONS

(a) Experimental procedure

All measurements were made between 1020 and 1415 GMT. Constant rate ascents and descents carried out at the beginning and end of this period, to determine the atmospheric temperature and humidity up to a maximum altitude of 7.5 km, showed no sign of significant change. A series of five level runs, each of approximately eight to ten minutes duration, was made above and in the main body of the cloud along the sun’s azimuth at approximately 0°5′E and between 56°N and 56°30′N. The first was above cloud top into sun, followed by a reciprocal run in cloud, from which the data of Fig. 2 were taken. This was followed by a run at cloud top level and then two final runs above cloud top; the first of these was away from the sun, to measure the diffuse radiation, and the second into sun. The two runs into sun above cloud top, to measure the albedo, were equally spaced either side of local noon, so giving a minimal variation in solar zenith angle.

The true airspeed was maintained within one or two per cent of 100 ms\(^{-1}\). If a pyranometer has a perfect cosine response then about 75% of the signal from an isotropic reflecting surface comes from a 120° field of view centred on the normal to the plane of the thermopile. Level runs were made above the cloud top at 300 m above the surface, so that readings taken every two seconds (approximately 200 m of travel) give successive fields of view which, although not completely independent, have a relatively small overlap, the degree depending on the cloud top height.

(b) Radiative models

Principally, two models of radiative transfer within plane–parallel clouds were employed, those of Stephens (1978a and b, to be referred to as ST) and Slingo and Schrecker (1982, to be referred to as SS). The former is a parametrization derived by using a two-stream approximation to evaluate reflectance and transmittance in two broad spectral bands, 0-3 to 0.75 μm and 0.75 to 4.0 μm. Values of single-scattering albedo and back-scattering fraction for each of these bands were then tuned to match the results of more precise calculations using a fifteen-band model over a range of model cloud types. These tuned parameters were presented as a set of coefficients to use in a polynomial expansion. An improved set was published recently (Stephens et al. 1984) and it is these revised values which have been used here. An advantage of this scheme is its simplicity since only the vertical l.w.p. and solar zenith angle need be specified. The Slingo–Schrecker model has 24 spectral intervals between 0.25 and 4.0 μm and includes gaseous
absorption by water vapour and ozone. The delta–Eddington routine is used to deal with cloud droplet scattering. There is the option of specifying the droplet extinction parameters in terms of the l.w.p. and effective radius from a parametrization of Mie theory calculations over a range of model cloud types. It is also possible to input directly the Mie scattering properties of the measured droplet spectrum. This model requires the vertical profile of temperature, humidity and ozone amount. Aircraft measurements of temperature and humidity were used up to 7.5 km and then merged with the mid-latitude summer model atmosphere of McClatchey et al. (1972) above this. The ozone column amount measured at Lerwick (60°8'N 1°11'W) was within 3% of the model atmosphere total during the experimental period and so the model profile was retained.

Some results will also be presented from the model of Liou and Wittman (1979, to be referred to as LW). This calculates reflectance, transmittance and absorptance of various cloud types as functions of l.w.p. and solar zenith angle and covers the solar spectrum from 0.2 to 3.5 μm.

(c) Results and comparisons

The total albedo was formed from the ratio of upward and downward fluxes (corrected for pitch and roll) from the clear-domed pyranometers. No downward near-infrared flux was available directly so this was derived by calculating the fraction of the total flux beyond 0.7 μm from the SS model and applying this to the measured total downward flux. This procedure is considered reasonable for two reasons. Firstly there was no cloud above the main layer, and secondly the measured diffuse-to-direct ratio of 0.12 was only slightly higher than the SS model value of 0.10 and hence the effects of aerosol could be neglected. Later flights with an upward facing red-domed pyranometer fitted gave measured ratios of near-infrared to total flux within 3% of the corresponding model value. The visible flux was taken as the difference between the near infrared and total fluxes. The uncertainty in any single measurement of albedo is probably in the range 3 to 5%; however, this is mainly systematic and arises primarily from calibration and zero offset errors. The uncertainties in the water vapour column amount arising from errors in temperature and dew-point are estimated to be about 4%. This translates to an error in the downward flux in the SS model of a few watts per square metre, which is not significant compared with the instrumental errors quoted above.

The measured albedos are those of the cloud and sea surface. The sea surface albedo was measured by a low run over the sea away from cloud and is included explicitly in the SS model. The ST and LW models use multiple reflection formulae to incorporate the effect of a non-zero surface albedo. The values used were 0.1 in the visible and 0.05 in the near infrared. The combination of bands in the SS model was chosen to give the closest match to the pyranometer spectral responses. The wavelength intervals of the ST model are slightly different but the resulting differences in albedo are less than 1% between the models.

Figures 3(a) and (b), with the notation explained in the caption, show the total albedo at a height of 300 m on level runs, each of approximately 50 km and not less than 30 km from the cloud edge. In both cases the solar zenith angle was 34°. The short period variations (≤1 min) in albedo which are apparent derive from the horizontal roll structure, but longer period variations of several minutes are also present. The positions of low albedo occurred when the aircraft was passing through warm, dry air and the high albedos at positions of cooler, moist air. A possible explanation is that there exists an overturning at the cloud top, with low albedo corresponding to the troughs (minimum l.w.p.) and high albedo to the peaks (maximum l.w.p.). Also shown on the figure are the model
values for the maximum and minimum l.w.p. clouds. The two sets of values agree well with each other and cover most of the observed variation in albedo. The corresponding albedos for the LW model stratus cloud are 0.36 and 0.31 respectively. It is not clear why these results should be almost a factor of two lower than both the observations and the other models, but comparisons between the results of the original paper and that of Stephens (1978b) show this to be a consistent feature.

Splitting the albedo into separate contributions from the visible and near infrared results in the curves of Figs. 4(a) and (b). Although the shape of the curves is not substantially different from that of the total albedo, the visible albedo is significantly
greater than that of the near infrared and there is now more variation between the model values, particularly in the visible. Overall the model values tend to overestimate in the infrared but underestimate in the visible. From this it appears that the agreement between model values and measurements seen on Fig. 3 is probably a little fortuitous. This is shown in Figs. 5(a) and (b) by forming the ratio of near-infrared to visible albedo. This ratio, as observed by DeVault and Katsaros (1983), should reduce the influence of the droplet size distribution in determining the albedo. The model values are significantly higher than the observations and, with the exception of the largest changes, there appears to be less correlation with the organized l.w.p. variations evident in Fig. 3. The implication of this is that the albedo ratio is relatively insensitive to l.w.p. variations over this range.

For the measured ratios in Figs. 5(a) and (b) to lie in the range of model values it would be necessary to increase them by about 15%. This is approximately a factor of two larger than the expected error arising from the uncertainty in a single albedo measurement. The albedo ratio can also be systematically increased by reducing the fraction of the total downward solar flux in the near infrared. This quantity was determined by comparison with the SS model; its reduction is therefore equivalent to increasing the water vapour and reducing the ozone column amounts above the aircraft, since the total flux must remain the same. However, the magnitude of the changes required to give a 15% rise in the albedo ratio would be a three-fold increase in water vapour and a reduction of the ozone amount to zero, both of which are widely outside the uncertainties in these quantities.

The calculated albedos presented so far are based on parametrizations of model clouds and can be applied over a wide range of l.w.p. and effective droplet radius. However, it is possible with the SS model to make more specific calculations using Mie scattering parameters derived from the measured droplet spectrum. Using refractive index data for liquid water from Irvine and Pollack (1968), a volume extinction coefficient, an asymmetry parameter and a single-scattering albedo were calculated for the centre wavelengths of the 24 bands and scaled by the l.w.c. profile of the maximum and minimum l.w.p. clouds. The albedo resulting from this procedure and all the albedo estimates
TABLE 1. SUMMARY OF MODEL VALUES FOR MAXIMUM(MINIMUM) L.W.P. CLOUDS

<table>
<thead>
<tr>
<th>Model</th>
<th>Albedo</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Total</td>
</tr>
<tr>
<td>Stephens</td>
<td>0·586(0·474)</td>
</tr>
<tr>
<td>Slingo–Schrecker</td>
<td>0·572(0·470)</td>
</tr>
<tr>
<td>Slingo–Schrecker/</td>
<td></td>
</tr>
<tr>
<td>Mie-scattering parameters</td>
<td>0·551(0·432)</td>
</tr>
</tbody>
</table>

presented so far are summarized in Table 1. The main effect of using the Mie parameters is a general reduction of the various albedos. This is particularly so in the infrared, implying significantly greater liquid water absorption, and results in a lower albedo ratio with almost no variation between the two l.w.p.s, a result more in agreement with Fig. 5.

5. Cloud edge measurements

As an alternative to measuring the albedo over a range of l.w.p. in a variety of different clouds, the large variation in l.w.p. at the cloud edge can be exploited. This has the advantage of maintaining essentially constant conditions above the cloud as the l.w.p. changes. To measure the variation of albedo and cloud physical properties two level runs were made, each of approximately 80 km, crossing perpendicular to the cloud edge. The first was at a height of 150 m above the sea travelling from clear air into cloud. The l.w.c. variation along this run is shown in Fig. 6. Some tenuous broken patches were encountered from 2 to 4 minutes into the run after which, although very variable, the l.w.c. increases for a further 3 minutes (approximately 18 km) after entering the main body of the cloud, before reaching a more uniform value. It is estimated that the l.w.p. varied by a factor of about four over the main adjustment region on Fig. 6. The effective droplet radius over this run increased rapidly once the cloud was entered. By 4·5 minutes into the run the effective radius exceeded 6 μm; the average droplet spectrum over the remainder of the run was essentially identical to that of Fig. 2, so substantial changes in l.w.c., and hence in l.w.p., occurred without correspondingly large changes in the effective droplet radius.

![Figure 6. Liquid water content from the Johnson–Williams meter against time for an aircraft run at 150m crossing from clear air into cloud.](image-url)
Figure 7(a). Total albedo against time for an aircraft run just above cloud top, and crossing the edge.

Figure 7(b). Visible (triangles) and near-infrared (crosses) albedos for the run in (a).

A second run was made above cloud top at a height of 300 m on a reciprocal heading, although it is not possible to follow exactly the same track as the first run. The total, visible and near-infrared albedos are shown in Figs. 7(a) and (b). There is a steady fall in albedo as the edge is approached but a more rapid decrease occurs over the three minutes immediately before the edge is reached. This seems to agree with the more rapid change of l.w.c. near the edge seen in Fig. 6. The more gradual decrease for the first eight minutes may signify changes in the l.w.p. occurring above 150 m and which therefore do not appear in Fig. 6.
Forming the albedo ratio as before results in the plot of Fig. 8. There is some evidence of a slight fall in the ratio over the three minutes before the edge is crossed but no clear trend before this. The overall effect of this lack of sensitivity to the l.w.p. is to produce a quantity which gives a sharper definition of the cloud edge. Furthermore because of the steadiness of the effective droplet radius in a region where the l.w.c. is still changing markedly, and the constant conditions above the cloud, the observed behaviour of the albedos can be ascribed very largely to changes in l.w.p.

6. DISCUSSION

The preceding sections have presented a series of aircraft measurements of cloud albedo and some radiative model comparisons. Previous authors have recognized the problem of the l.w.p. variability by reducing their data to that applicable to an 'average' cloud. Some such procedure is clearly required if use is to be made of models of plane-parallel clouds; such models are readily available and are candidates for inclusion in larger numerical models. The technique adopted here has been to present the aircraft data as a time series of approximately non-overlapping albedo measurements and to compare with model calculations based on the maximum and minimum l.w.p. encountered. It is argued that this procedure is justifiable because the horizontal scale of l.w.p. variations (a few kilometres and more) was much larger than both the depth of cloud top structure (<100 m) and the overall cloud depth (about 230 m). It follows therefore that the cloud structure and albedo were not varying rapidly within the field of view of the downward facing pyranometers. Effects arising from the cloud geometry could be expected very close to the edge, but the majority of the measurements were at positions equivalent to many cloud thicknesses from the edge.

The accuracy and representativeness of aircraft measurements of cloud radiative properties are difficult to quantify, particularly when coupled with the problems of
measuring cloud microphysical data. Even though in the present study the range of l.w.p. has been compared with the variability of the measured albedos, direct, quantitative comparisons remain difficult. Perhaps a more practical approach is to assess the usefulness of the models and their associated parametrization schemes and to examine whether their qualitative behaviour sheds any light on possible shortcomings.

The results of section 4(c) implied that overall the models tended to overestimate the albedo in the near infrared and underestimate in the visible. Wiscombe et al. (1984) performed cloud radiative calculations using a 'large-drop' distribution with a substantial number of drops at radii greater than 50 μm. Their results for this 'large-drop' distribution show enhanced near-infrared absorption and reduced visible reflectance, and hence reduced albedo ratio, relative to their 'normal' drop distribution. For example, from the results for a 2 km-thick cloud the albedo ratio falls from 0.48 to 0.44 for an overhead sun, and from 0.68 to 0.67 for a zenith angle of 78°. However, this cloud has a l.w.p. of 1000 g m⁻², approximately seventeen times that of the maximum observed in the present study. It is unreasonable to suppose that a cloud of 200 m thickness could support a sufficient number of undetected large drops to allow direct comparison with the results of Wiscombe et al. Even for such a large l.w.p. the reduction in albedo ratio when large drops are present is relatively small compared with the differences between the modelled and observed values here.

The behaviour of the near-infrared to visible albedo ratio in sections 4 and 5 implied this quantity had only a weak dependence on l.w.p. over the range encountered. Certainly the equivalent model values (summarized in Table 1) show only a small spread, although their absolute values are significantly higher than the measured ratios. The largest range of l.w.p. occurred at the cloud edge, where it is estimated that it fell to approximately 14 g m⁻² before the cloud became partly broken. To examine this behaviour both the ST and SS models were run for a range of l.w.p. with a solar zenith angle of 34°. For the SS model, uniform clouds with effective droplet radii of 5, 10 and 15 μm were used. For the purposes of this argument, uniform clouds are more convenient and, as far as the albedos are concerned, the results are not significantly different from those that would be achieved using more realistic distributions of l.w.p. and effective droplet radius. The resulting albedo ratios are plotted on Fig. 9 against log(l.w.p.); the arrows indicate a range of 14 to 56 g m⁻². At low values of l.w.p. the sea surface albedo is comparable with that of the cloud, whereas at high l.w.p. the effect of the sea surface is relatively unimportant.

For effective droplet radii typical of observed values, the SS model shows only a very weak variation over the estimated range of l.w.p. This is in contrast to the ST model, which shows a much more sharply peaked curve. Part of this more rapid variation arises because the parametrization of optical depth includes an implicit dependence on effective radius, such that effective radius increases as l.w.p. increases. Hence, qualitatively at

![Figure 9. Near-infrared to visible albedo ratio from Stephens (ST) and Slingo-Schrecker models (SS) against log(liquid water path). The dash-dot curve is the ST model. The other three curves are from the SS model for effective radii of 5 μm (upper), 10 μm (middle) and 15 μm (lower). The arrows denote the estimated range of l.w.p. encountered.](image-url)
least, the SS model appears to produce a more reasonable dependence on l.w.p. Note that in an attempt to obtain better quantitative agreement with the experimental values it is not possible to propose that the effective radius has been seriously underestimated since the shape of the $15 \mu m$ radius curve on Fig. 9 is quite different from the experimental results.

The results discussed above have some implications for the suggestion by DeVault and Katsaros (1983) that the albedo ratio might be used as a remote measure of cloud l.w.p. At low values of l.w.p. the effects of droplet size and the surface albedo are still important; at larger l.w.p., although the influences of droplet size and surface albedo are much less, the dependence on l.w.p. is comparatively weak. Hence the measured albedo ratio is likely to be a poor predictor of l.w.p. A contrary viewpoint can be found in the work of Coakley and Davies (1986), who have calculated cloud albedos at wavelengths of 0.63 and 3.7\mu m. From their results the ratio of these albedos appears to show considerable sensitivity to changes in l.w.p. However, their study requires an increase in l.w.c. to be accompanied by a corresponding increase in droplet sizes; this does not match with the observations at the cloud edge described in section 5, where an initial rapid increase in effective radius is followed by little change in the droplet distributions despite a continued increase in l.w.c.

The overestimate of the albedo ratio by both models is consistent with the observations of Herman and Curry (1984) in their study of arctic stratus. These authors present aircraft measurements and comparisons with the SS model; if the albedo ratio is formed from their data then a similar picture appears of the model consistently overestimating the aircraft measurements. The main discrepancy appears to lie in the near infrared rather than in the visible. This carries the implication that there is insufficient liquid water absorption in the models and may well be due to the behaviour of the parametrization schemes at the low end of the l.w.p. range over which they are applicable. Use of the scattering parameters calculated directly from the droplet spectrum (Table 1) gives a substantial reduction in the near-infrared albedo, and a smaller one in the visible. This progressive change in albedo as a more complex representation of the cloud physical properties is employed is primarily a reflection of the applicability of the parametrization schemes, rather than of the absolute spectral resolution of the models. The difference in resolution between the ST and SS models is less than is at first apparent, as the ST model is a parametrization of results from a fifteen-band model. The implication of the above results is that a parametrization applicable over a more limited range of l.w.p. could give better quantitative results.

An unsurprising conclusion that follows from these results is that as more computational effort is expended then better agreement can be obtained in any particular case. For its simplicity in use the ST model performs very well, within the limitations imposed by its parametrization scheme, which reduces the dependent variables to just the solar zenith angle and the l.w.p. Although more elaborate in use, the SS model gives qualitatively better results; the limits imposed by its parametrization scheme can be circumvented by using Mie scattering parameters calculated from the measured droplet spectrum. However, this increases greatly the computational effort and introduces a further limitation in the accuracy of the liquid water refractive index data and the way these are incorporated into the model.

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