Studies with a flexible new radiation code. II: Comparisons with aircraft short-wave observations

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SUMMARY

Calculated irradiances from a new radiation code are compared with in situ observations of short-wave irradiances from the UK Meteorological Office’s C-130 aircraft. Three cases of clear skies are studied and four where a liquid-water boundary-layer cloud was present. Under clear-sky conditions the modelled and in situ observations agree to within 3%, which is the estimated accuracy of the observations. In the cloudy-sky cases the albedo and transmittance agree to within ±0.1 but the absorption in the model is higher than that observed, sometimes by a factor of two; there is no evidence of anomalous absorption in the observations. The observed absorptions do not exceed 6% for the stratuscumulus cases considered. The results clearly identify the problems of representing inhomogeneous clouds as plane parallel layers in radiation models. Analysis of the variability of the cloud microphysics provides some insight into the importance of regions of low optical depth within the clouds.

KEYWORDS: Airborne observations Cloud radiative properties Radiation parametrization

1. INTRODUCTION

The development of a new radiation code which uses the same code for both long-wave and short-wave calculations is described in the companion paper (Edwards and Slingo 1996), hereafter Part I. Since the code described in Part I represents the most up to date two-stream radiative-transfer formulation it is essential to compare it with in situ observations. In this manner one can identify the major issues which should be of concern to the community.

In this paper short-wave calculations with this radiation code are compared with observational data from the UK Meteorological Office’s C-130 aircraft. Given that the new code is intended for a variety of applications, including operational and research roles, it is important to test its utility and accuracy under a range of conditions. For this purpose, a number of flights have been selected to include both cloud-free conditions with a range of water vapour and aerosol amounts, and single-layer marine boundary-layer clouds. The code is initialized with thermodynamic and microphysical profiles from these flights and the results discussed. The code is run in four configurations: (a) with a 220-band resolution, (b) with a 95-band resolution, (c) using the 24-band configuration of Slingo and Schrecker (1982, hereafter SS) and, finally, (d) in a 4-band configuration designed for use in the UK Meteorological Office Unified Forecast/Climate Model (UM) (Cullen 1993). In this way, the code’s performance at full resolution can be compared with that used in the UM and the results can be traced back to previous work using the Slingo–Schrecker model.

This paper presents broad-band short-wave (0.3–3.0 μm) irradiances for seven flights. In section 2 the aircraft instrumentation and the errors associated with the irradiance measurements are discussed. In section 3 we describe briefly the radiative-transfer code; a thorough description of the code is given in Part I. In section 4 we describe the flights and the initial conditions input to the code and discuss the comparison of modelled and observed irradiances for clear and cloudy atmospheres. Section 5 deals with the important issue of cloud absorption and includes an analysis of the importance of accounting for the inhomogeneity of the clouds. A sensitivity analysis of the new code is presented in

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section 6, and in section 7 the new code is compared run using differing spectral data. A discussion of the results and their implications are given in section 8.

2. Aircraft instrumentation

The Meteorological Research Flight (MRF), a part of the UK Meteorological Office, operates a Royal Air Force C-130 Hercules aircraft which has been modified extensively to make it suited to a wide range of atmospheric research work.

A brief description of the instruments used to measure the basic meteorological parameters required for input into a radiation code is given here. A more complete description of the standard meteorological instrumentation is given by Nicholls (1978), of the cloud physics instrumentation by Martin et al. (1994) and Brown (1993), and the radiation instrumentation by Kilsby et al. (1992) and Saunders et al. (1992).

(a) Temperature and humidity measurements

Two Rosemount platinum-resistance temperature sensors are mounted on the aircraft, one de-iced and the other not. Corrections are applied for kinetic heating and for the heat used in the de-iced probe. The accuracy of both temperatures is approximately ±0.5 degC. For measurements in cloud a radiometric temperature sensor, which measures the emission in the 4.3 \( \mu \text{m} \) carbon dioxide band, is also flown. The accuracy of this instrument is ±0.1 degC at +30 °C reducing to ±1 degC at −50 °C; its resolution is 0.025 degC.

The main humidity measuring instrument is a thermo-electrically cooled dew-point hygrometer, formerly a Cambridge Instruments device and latterly a General Eastern. These instruments are accurate to ±0.5 degC for temperatures above freezing, and to ±1.0 degC for temperatures below. Ström et al. (1994) give a summary of the performance of the MRF measurements compared with those of other aircraft.

(b) Liquid-water content

A Johnson–Williams meter is used to measure cloud liquid-water content. Cloud droplets are deposited on a heated wire exposed perpendicular to the airstream and then evaporated. The wire is cooled by an amount which depends upon the latent heat of evaporation supplied to the droplets, which depends upon the liquid-water content of the sample. A compensating wire oriented parallel to the airstream, on which it is assumed no droplets are deposited, is used to monitor the heat loss to the air. The two wires form part of a Wheatstone-bridge circuit. From the change in resistance of the wires the water content incident on the wire perpendicular to the airflow is deduced. The meter is insensitive to droplets larger than about 30 \( \mu \text{m} \) radius, and its accuracy is ±10%.

(c) Cloud particle size distribution

A Knollenberg FSSP (Forward Scattering Spectrometer Probe) is fitted to the aircraft and is mounted on the port wing maxi-pod. Light from a 5 mW He–Ne laser (at 0.63 \( \mu \text{m} \)) is focused to a diameter of approximately 200 \( \mu \text{m} \) at the centre of the sampling aperture through which air passes. The collection optics accept light scattered only between 4° and 15°. In this range, small water droplets of radius \( r < 50 \mu \text{m} \) scatter light with an intensity which is approximately proportional to \( r^2 \). If the sampling size is made so small that it contains only one drop at any time, measurements of the scattered light intensity enable water drops to be counted and sized automatically.

In practice, the scattered intensity depends upon the particle size and shape, the optical properties of the material composing the particle, the location of the particle within
the beam, and the optical properties of the instrument (beam intensity, uniformity etc.). Particles are assumed to be perfectly spherical and composed of pure water.

Measurements of the rate-of-sizing events and the true air speed allow inference of droplet number densities. The instrument records the number of droplets detected in 15 size categories in one of three selected ranges: 1 to 15 μm, 2 to 30 μm or 3 to 45 μm, where all the sizes are droplet radii. The effective radius, _r_\_e, of a droplet distribution is defined as the ratio of the third to the second moment of the size spectrum, that is

\[
_r_e = \frac{\int_{r=0}^{\infty} N(r)r^3 \, dr}{\int_{r=0}^{\infty} N(r)r^2 \, dr}
\]

where _r_ is the cloud droplet radius and _N(r)_ the droplet concentration. Together with the liquid-water content, _r_\_e can be used to determine the single-scattering properties of the cloud drops (Slingo and Schrecker 1982).

The effective radius is measured by the FSSP using the finite form

\[
_r_e = \frac{\sum_{n=1}^{M} r_n^3 N_n}{\sum_{n=1}^{M} r_n^2 N_n}
\]

where _M_ is the number of size bins resolved by the FSSP (_M_ = 15), _r_n_ is the middle radius value for that size bin and _N_n_ is the concentration of droplets in that size bin.

The FSSP is calibrated by sampling glass beads of known sizes, allowances having to be made for the differences between the refractive indices of glass and water. With careful calibration the accuracy of the FSSP sizing is ±1 μm.

(d) Aerosol concentration

A Knollenberg PCASP (Passive Cavity Aerosol Spectrometer Probe) is fitted to the aircraft. This instrument counts and sizes aerosol in 15 size intervals covering the size range from 0.1 to 3.0 μm diameter. The PCASP measures the intensity of light scattered from a particle as it intercepts a laser beam; this intensity is a function of the particle size, shape and refractive index. The large collection angle of > 2π steradians means that there is little sensitivity to the real part of the refractive index of the aerosol, which is assumed to be that for pure water at the wavelength of the He–Ne laser (0.635 μm). As the chemical composition of the aerosol is not determined, if the scattering properties of the aerosol are different from those of water there will be an error in the PCASP derived size and concentration. Mie calculations carried out assuming spherical particles have shown that, for aerosol below 1 μm diameter, the errors associated with assuming the refractive index is that of water are negligible. Heaters in the tip of the inlet pipe to the PCASP and the sheath of dry air used to ‘focus’ the sample into the path of the laser beam will both act to dry the aerosol. We are unable to measure the humidity of the aerosol as it passes through the laser beam, and hence any effects due to residual moist aerosol or any changes in the aerosol chemistry that have resulted from its being dried are not accounted for in the processing. In cloud, water droplets can shatter on the intake nozzle of the PCASP, resulting in spurious measurements. For this reason measurements made by the PCASP are used only out of cloud.

e) Broad-band irradiance

Broad-band hemispherical irradiances are measured with Eppley pyranometers, two on the top of the aircraft and two on the underside. Each pair consists of one instrument fitted with a clear dome (Schott filter WG295, passband ≈ 0.3 to 3.0 μm) and another
fitted with a red dome (Schott filter WG715, passband \( \approx 0.7 \) to \( 3.0 \; \mu \text{m} \)). The red-dome filter therefore separates the near infrared part of the solar spectrum, where absorption by water vapour and condensed water are important, from the visible part, where molecular scattering is significant.

The pyranometers are calibrated on the ground, once a year, during the summer months by exposing them outside, alongside a national standard pyranometer, for a number of days of both clear (direct beam and diffuse radiation) and cloudy (diffuse radiation) skies. The diffuse-only component can also be monitored using a pyranometer with a motor-driven occulting disk. The national standard pyranometer is itself calibrated outside against a primary standard cavity radiometer which is in turn calibrated once every five years at the World Meteorological Organization (WMO) intercomparison at Davos, Switzerland. The objectives of the annual calibration are threefold; firstly to determine a diffuse sensitivity under cloudy skies by comparison with the standard as a check against changes since the previous calibration. Secondly, to detect any gross changes in the zenith-angle dependence under clear skies relative to the standard and, finally, to measure any significant azimuthal dependence of the sensitivity by rotating the pyranometer under clear skies. There is no evidence of any long-term drift in the MRF pyranometers between annual calibrations.

The calibration of the red-dome pyranometers is more elaborate. Initially the red dome is replaced by a clear one and the instrument calibrated as before. However, the instrument sensitivity derived in this way assumes that, integrated over the passbands of the red dome, the product of the solar spectrum and the transmission function with the clear and red domes is identical. On the whole this is unlikely to be true. Furthermore, because of the strong water absorption bands in the near infrared this difference will vary according to the effective absorber amount, as the spectral shape changes. This current study is unusual in exploiting the variation of absorber amounts over most of the depth of the troposphere to achieve the widest range of conditions under which to compare observations and model calculations.

Transmission functions, as measured by Semour and McHaffie (1993), were used to calculate correction factors for a number of model atmospheres. The factor to be applied to the measured irradiance could be reduced to a simple dependence on altitude, or alternatively pressure. The magnitude of this correction can be substantial, up to \( 8\% \) of the solar irradiance, and must, in a critical application such as this, be taken fully into account.

Under conditions of predominantly direct irradiance the upper instruments have to be corrected for deviations from the local horizontal and a non-ideal cosine response. A thorough description of the techniques used is given by Saunders et al. (1992), whose results suggested that diffuse irradiances were the most accurate measurements, approaching \( 2\% \). For clear-sky downwelling irradiances the levelling and non-cosine corrections act to increase the uncertainty in the measurements to \( \approx 3\% \).

3. The Radiation Code and Preparation of Data

The radiation code employed in subsequent calculations is described in detail in Part I. Here, only a few key points will be mentioned.

Radiative fluxes are calculated using the \( \delta \)-Eddington form of the two-stream equations. The spectral region is chosen to match the dome fitted to the pyranometer and extends from \( 0.295 \; \mu \text{m} \) to \( 2.77 \; \mu \text{m} \) in the case of the clear dome and from \( 0.707 \; \mu \text{m} \) to \( 2.77 \; \mu \text{m} \) in the case of the red dome. The cut-off for the red dome is determined from the measurements of Seymour and McHaffie (1993). The sensitivity to the representation of the spectral response of the filter as a sharp cut-off has been investigated. Changing from a true representation of the filter response to a sharp cut-off had an insignificant effect.
on the irradiances. All the red domes used by the MRF were consistent and showed a cut-off wavelength (defined as the 50% transmission point) of 0.707 μm at 23 °C. The slight temperature dependence of the cut-off wavelength has a negligible effect over the range encountered.

The spectral region considered is divided into a number of bands, within each of which gaseous transmission is represented by exponential sum-fitting of transmissivities (ESFTs). The ESFTs are obtained from tables of gaseous transmissions generated using the line-by-line model GENLN2 (Edwards 1988) and the HITRAN92 database (Rothman et al. 1992) or, in the case of ozone, using LOWTRAN7 (Kneizys et al. 1988). An important feature of the radiation code is that its spectral resolution is not fixed; this freedom is achieved by storing spectral information in a separate file generated by a preprocessor. An appropriate decomposition of the spectrum for a particular application may thus be chosen. Three spectra have been used for the calculations in later sections. Most calculations were performed using bands 33 to 194 (0.294–2.78 μm) of the 220-band spectrum described in Part I. Bands 40 to 83 (0.708–2.77 μm) of a 95-band version were also used; in general this 95-band spectrum has a coarser spectral resolution than the former but is more highly resolved in the region from 0.69 to 0.72 μm and, in particular, is appropriate for representing the spectral response of the red filter. Also, for three of the cases presented, the 4-band version of the code developed for use in the UM has been used. The 4-band version covers the range 0.2–5.0 μm; a range slightly wider than that of the other two versions of the code presented. Matching the exact wavelength range of the instruments, which has been carried out with the 220- and 95-band versions, is not possible without splitting one of the four bands. For this reason the irradiances produced by the 4-band version of the code are expected to be 1–2% larger than those from the other versions of the code. Full details of the spectral bands of the code are given in Part I.

The optical properties of water droplets in each band are parametrized in terms of the effective radius and liquid-water content using the functional form suggested by Ackerman and Stephens (1987). To derive the coefficients in this parametrization a Mie code was run to generate single-scattering properties at a range of wavelengths for the size distributions given by Rockel et al. (1991), omitting those distributions with \( r_e \) smaller than 3 μm; for more details refer to Part I. These properties were then averaged across the spectral bands and fitted as functions of \( r_e \).

Observations from the aircraft alone are not sufficient to specify the atmospheric state completely and must be combined with data from other sources to allow the radiation code to be run. Typically, measurements of the height, pressure, temperature and dew-point are made from the aircraft at many atmospheric levels up to a ceiling of about 500 hPa. From these measurements the profiles of specific humidity and temperature in the lower atmosphere may be determined. Above this level the profiles are completed by inserting values from either the mid-latitude summer or tropical McClatchey atmosphere (McClatchey et al. 1972), depending on the flight. The mixing ratio of ozone is taken from the corresponding McClatchey profile. The volume mixing ratio of oxygen is taken as 0.20946, following the value given in Goody and Yung (1989). The mixing ratio of carbon dioxide is calculated for the year of the flight from the information given in the Intergovernmental Panel on Climate Change report of 1990 (Houghton et al. 1990). For flights in cloudy conditions the liquid-water content and \( r_e \) must also be determined.

A thorough quantitative analysis of the errors in the calculated fluxes is not possible, but three main sources of error may be identified. There are, first of all, uncertainties in the atmospheric state specified to the code, arising both from the limitations on the accuracy of the measurements discussed above and from the use of standard profiles above the lower troposphere. This point is addressed by a sensitivity study below. Another
source of error is uncertainties in the databases used to derive gaseous-transmission data, but we are not aware of any sensitivity studies pertaining to this question. There are also errors inherent in the approximations used within the radiation code. These approximations comprise chiefly the two-stream approximation for the diffuse irradiances, and the use of scaling functions for the pressure and temperature dependence of gaseous absorption (see Part I). King and Harshvardhan (1986) have investigated the accuracy of a number of two-stream approximations. Whilst they caution against drawing general conclusions about the validity of a specific approximation from restricted intercomparisons, for the cases which they consider, restricting the cosine of the solar zenith angle to the range 0.5−1, the absolute error in the total transmission is less than 5%. Under clear skies, where the diffuse irradiances amount to less than 10% of the total, errors from this cause would not be expected to exceed 1%. A recent study of scaling functions in the infrared (Chou et al. 1993) suggested that the errors in the calculated fluxes arising from this cause were around 1%. A partial assessment of such modelling errors can be made from comparisons with results from line-by-line models, such as those obtained by Ramaswamy and Freidenreich (1992) (see Part I). In cloudy conditions, where the diffuse irradiances are greater, larger errors may be expected, but in such cases there are also considerable uncertainties in the specification of the physical properties of the cloud.

4. AERIAL OBSERVATIONS COMPARED WITH MODEL RESULTS

Data from seven flights are used to compare with model irradiances; three are in cloudless sky conditions and four are flights where a single layer of liquid-water stratocumulus cloud was observed.

(a) Cloudless sky results

The flights in cloudless skies were chosen to obtain a range of column water vapour amounts. The radiation code was run for each of the cloudless sky cases at the mean solar zenith angle for each of the level flight runs where in situ irradiance measurements were made. In these comparisons the radiation code has been initialized with the profile of temperature and humidity measured by the aircraft, supplemented at high level by a suitable model atmosphere or radiosonde data, where possible. The solar constant of 1365 W m$^{-2}$ was used and this has been corrected for the effects of eccentricity in the earth’s orbit.

The surface albedo was determined for the clear-sky case from the equation:

$$A = \frac{0.037}{1.1 \mu_0^{1.4} + 0.15}$$

where $\mu_0$ is the cosine of the zenith angle. This equation was derived from an ensemble of observations made by the MRF C-130 over several years in a variety of conditions. Further work is under way to characterize more accurately the dependence of the sea-surface albedo on solar zenith angle and sea state. The code has been run in its 220- and 95-band modes and for two of the flights in its UM configuration of four bands as described in section 3. Most emphasis in the discussion of the results is given to the downwelling irradiances. The variability in the upwelling irradiances will reflect mainly those of the downwelling irradiances combined with any errors associated with the determination of the surface albedo.

(i) Flight A345. Flight A345 took place off the coast of California during the Monterey Area Ship Tracks (MAST) experiment. On 23 June 1994 a sortie was flown where radiometric observations were made at nine levels between 30 m and 7300 m. The atmosphere
above the boundary layer was exceptionally dry, with dew-point depressions of 40 K, resulting in a low total water column of 0.91 cm. The results are presented in Fig. 1 as comparisons between the 220-band and 4-band versions of the code with the \textit{in situ} observations of downwelling clear-dome irradiances; the range of zenith angles covered was 12° to 46°. Figure 1(b) shows the percentage difference between flight and model irradiances as a function of pressure. The 220-band version of the code results and aircraft results are nearly all within ±2%. The model irradiances are mostly higher than the aircraft irradiances and there is a tendency for the percentage difference in the irradiances to increase with the length of the atmospheric path, i.e. with increasing pressure. The 4-band results are in systematically poorer agreement with the observations with an increase in the difference of 1–2%. This is to be expected because of the wider spectral range of the 4-band model.

The 95-band results are compared with the observed red-dome irradiances in Fig. 2. The agreement between the observed and modelled irradiances is very good, with most of the points lying within the ±2% lines.

(ii) \textit{Flight A143}. Flight A143, which took place in the tropical South Atlantic in November 1991, had a total water vapour column amount of 3.96 cm. This period was close to the peak of the Pinatubo aerosol loading. Saunders (1993) made measurements of the aerosol optical depth and the direct and diffuse components of the short-wave downwelling irradiances using the C-130 during the period which bounded flight A143. Saunders found that the aerosol loading led to 20% of the direct solar irradiance being scattered into the diffuse irradiance component but the total downward irradiance was not measurably different from the normal aerosol-free case. The effects of the Pinatubo aerosol have not been accounted for here in the radiation modelling. Radiometric observations were made at seven levels during this flight between 89 m and 7900 m. Figure 3 shows the aircraft measured clear-dome downwelling irradiance versus the model irradiance, evaluated at the same solar zenith angle as the observations. The agreement between the model and flight irradiances for the 220-band spectrum is generally within 3%, with the exception of one point where the errors are of the order of 6% for a run where the solar zenith angle at the time of the observations was 53°. At large solar zenith angles the non-cosine response of the pyranometers reduces the absolute accuracy attainable.

The 4-band version of the code gives a higher irradiance at all levels as expected from the differences between the spectral range of the versions of the code. The agreement between the observed irradiances and the 4-band spectrum results is within 4%.

The red-dome downwelling irradiances, observed with the pyranometer and modelled using the 95-band version of the new code, are shown in Fig. 4(a) with the percentage difference between model and \textit{in situ} results in Fig. 4(b). The model and observed irradiance, with the exception of two points, agree to within ±3%. The largest error of 25% occurs for the observation made at a solar zenith angle of 53° as noted before. There is no apparent reason that explains the magnitude of the error at 670 hPa.

(iii) \textit{Flight A200}. Flight A200 took place over the North Sea to the east of the United Kingdom in May 1992. There was a strong southerly flow and the air mass had originated over continental Europe. Aircraft runs at ten levels between 70 m and 9700 m were flown. Data from the PCASP showed significant aerosol with peak concentrations in the boundary layer of 5000 cm\(^{-3}\). An attempt was therefore made to account for this in the model calculations. Particle size distributions from a profile between 15 m and 7900 m were used. However, lacking any information on chemical composition, the code was initialized with the refractive indices of several aerosol types. The refractive index data of Deepak and Gerber (1983) were used, except for ammonium sulphate where the data of Toon \textit{et al.}
Figure 1. Flight A345: Clear-dome (0.3–3.0 μm) downwelling irradiance. 220-band model spectrum = C, 4-band = 0. (a) Observed irradiance versus model irradiance; solid line is the 1:1 fit and the dashed lines are ±2%. (b) Percentage difference in irradiances (observed – model)/(observed) as a function of pressure.

Figure 2. Flight A345: Red-dome (0.7–3.0 μm) downwelling irradiance. 95-band model spectrum. (a) Observed irradiance versus model irradiance; solid line is the 1:1 fit and the dashed lines are ±2%. (b) Percentage difference in irradiances (observed – model)/(observed) as a function of pressure.
Figure 3. As Fig. 1 but for Flight A143.

Figure 4. As Fig. 2 but for Flight A143.
Figure 5. Flight A200: Clear-dome (0.3–3.0 μm) downwelling irradiance. 220-band model spectrum. (a) Observed irradiance versus model irradiance; solid line is the 1:1 fit and the dashed lines are ±2%. (b) Percentage difference in irradiances (observed – model)/(observed) as a function of pressure. The chemistry of the aerosols is represented by the various symbols explained in the key.

(1976) were used. Figure 5 shows the model and aircraft clear-dome downwelling irradiances in the same format as before. In this figure the model results using a range of aerosol types are shown along with the no aerosol case; details are given in the figure key. The squares are the model irradiances without any aerosol and they clearly show an increasing difference between observed and modelled irradiances that reaches 6% at the lowest level. The inclusions of aerosol has the largest effect at low level where the highest aerosol concentrations were observed. The use of a dust, ammonium sulphate or water soluble aerosol, has a similar effect on the model irradiances, the percentage difference being reduced to less than 2% at low level whilst the soot aerosol absorbs considerably more radiation and brings all the model and aircraft irradiances to within ±2%. These simulations show the significant effects that aerosol can have on the downwelling irradiance, ≈ 40 W m⁻² at the surface. Although the composition of the aerosol is not known, the use of assumed aerosols consistent with a continental airmass (e.g. ammonium sulphate) does bring all but one of the model irradiances to within 3% of the observations.

The 95-band model downwelling irradiances over the wavelengths of the red-dome pyranometer are shown in Fig. 6. The inclusion of a soot aerosol, using the refractive indices from Deepak and Gerber (1983), brings the agreement between the modelled and observed irradiances to within ±2%.
The radiation code was initialized with temperature, humidity and cloud microphysical profiles for a range of stratocumulus capped boundary layers. Flight H806 took place during the First ISCCP (International Satellite Cloud Climatology Project) Regional Experiment (FIRE) in 1987 off the west coast of California. During this flight, runs were made vertically stacked below, within and above a relatively uniform stratocumulus cloud sheet at levels between 23 m and 650 m. Flights A140 and A146 took place in the South Atlantic; here the effective radius of the cloud droplets near cloud tops was larger than that observed during FIRE. Runs were made at flight levels between 673 m and 1350 m. Flight H941 was flown to the south-west of the British Isles on 3 October 1989 over a relatively homogeneous stratocumulus sheet. Runs were made at levels between 364 m and 1638 m. Between them these cases represent a variety of cloud microphysical and dynamical conditions under a range of different atmospheric columns.

As before, the aircraft profiles of temperature and humidity were supplemented at high level with either a local radiosonde ascent or a standard atmosphere. The surface albedo in these simulations was set at that for a Lambertian surface of 0.045, as all the radiation incident on the ocean surface is assumed to be diffuse.

Details of the cloud structure and microphysical properties for the four flights studied here are given in Table 1. A major problem in the modelling of the radiative properties of cloud fields is their inherent horizontal and vertical inhomogeneity. In this work the profiles of cloud droplet $r_c$, as measured by the FSSP, and liquid-water content, LWC, as measured by the Johnson–Williams probe, have been combined with the data obtained during straight and level runs within cloud to determine the range of the measurements within each cloud. In each case LWC and $r_c$ were observed to increase with height through the cloud. The
TABLE 1. CLOUD STRUCTURE AND MICROPHYSICAL PROPERTIES FOR THE CLOUD CASES MODELLLED

<table>
<thead>
<tr>
<th>Flight</th>
<th>Date</th>
<th>Location</th>
<th>Range of zenith angle (°)</th>
<th>Cloud depth (m)</th>
<th>Cloud-top effective radius (µm)</th>
<th>Cloud-top liquid-water content (g Kg⁻¹)</th>
</tr>
</thead>
<tbody>
<tr>
<td>A140</td>
<td>2 November 1991</td>
<td>South Atlantic</td>
<td>21–32</td>
<td>516</td>
<td>14.69 11.08</td>
<td>0.68 0.27</td>
</tr>
<tr>
<td>A146</td>
<td>11 November 1991</td>
<td>South Atlantic</td>
<td>25–30</td>
<td>570</td>
<td>11.53 9.38</td>
<td>1.13 0.52</td>
</tr>
<tr>
<td>H806</td>
<td>8 July 1987</td>
<td>FIRE</td>
<td>9–18</td>
<td>450</td>
<td>10.76 7.65</td>
<td>0.69 0.23</td>
</tr>
<tr>
<td>H941</td>
<td>3 October 1989</td>
<td>UK</td>
<td>55–62</td>
<td>365</td>
<td>8.53 5.77</td>
<td>0.99 0.43</td>
</tr>
</tbody>
</table>

radiation code has been initialized with three separate cloud microphysical profiles for each flight which represent the maximum, mean and minimum values of LWC/r_c (proportional to the visible optical depth) from the observations.

Figure 7 shows the ratio of the observed to model downwelling irradiances (0.3–3.0 µm), from the 220-band version of the code, for all four cloud cases, plotted as a function of the scaled height \( z_s \) where:

\[
z_s = \frac{z - z_b}{z_t - z_b}.
\]  

(4)

Here \( z_t \) and \( z_b \) are the heights of the cloud top and base respectively and \( z \) is the height of the observation. This scaled height is such that cloud base for all cases is at \( z_s = 0 \) and cloud top is at \( z_s = 1 \). A value of \( z_s = -1 \) represents a distance below cloud base equal to the cloud thickness. The key in Fig. 7 describes the symbols used. The error bar on each point represents the range of the in situ observations.

Above cloud, \( z_s > 1 \), the modelled and observed irradiances are in very good agreement similar to the results shown for the clear-sky cases, the ratio being 1.0 ± 0.04. As one moves to larger distances below cloud top the difference between modelled and observed irradiances increases markedly. This is true except for the lowest run below cloud where the differences between modelled and observations are smaller than the differences just below cloud. This observation may be due to the increase in the amount of cloud seen by the pyranometer at the lower level averaging out some of the cloud inhomogeneities. The results are closest for the minimum case, and the divergence from perfect agreement is greatest for the model runs initialized with the maximum cloud case.

Figure 8 is in the same format as Fig. 7 but shows the upwelling irradiances (0.3–3.0 µm). The large error bars on the observations represent the variability of the observed irradiances. These results are strongly influenced by the code’s ability to predict the downwelling irradiance correctly and also the accuracy of the assumption about the sea-surface albedo which is fixed at 0.045.

The 4-band version of the code has been initialized with the cloud properties of just one flight (H941) for comparison with the 220-band version. Figure 9 shows the results of this comparison with the modelled data plotted against the in situ aircraft observations. For the run skimming cloud base at 885 hPa and the run below cloud at 981 hPa the absolute magnitude of the irradiances is small and hence the percentage errors tend to be large. The vertical error bars show that although the cloud appeared uniform there was considerable variability in the observed irradiances due to the internal inhomogeneity of the cloud field. There is no systematic difference between the results using the 220- and 4-band versions of the code, which gives confidence in the use of the 4-band version in the UM where its speed of operation makes it preferable to the full 220-band version. Note there are no near
Figure 7. All cloudy flights: observed clear-dome downwelling irradiances/modelled 220-band downwelling irradiances for the max, mean and minimum LWC/$r_e$ profiles (see text). The error bars for the maximum LWC/$r_e$ cases have been displaced upwards and those for the minima cases displaced downwards.

Figure 8. As Fig. 7 but for observed clear-dome upwelling irradiances/modelled 220-band upwelling irradiances.
infrared data presented for this case as the 4-band version of the code cannot perfectly match the filter response of the red domes.

5. CLOUD ABSORPTION

The absorption, albedo and transmittance of the four cloud cases considered can be computed using both the modelled and measured irradiances above and below the cloud. Ackerman and Cox (1981) proposed a method for correcting errors in aircraft observations of radiative properties of clouds. Their method computes the true cloud absorptance by subtracting the apparent absorptance in the visible region (0.3–0.7 μm) from that in the total region (0.3–3.0 μm) under two assumptions as follows. Firstly, the absorption of solar radiation by cloud droplets and water vapour is negligible in the visible region and secondly the scattering properties in the near-infrared are similar to those in the visible.

The first assumption is reasonable for pure cloud droplets with no contaminants (Stephens and Tsay 1990). The second assumption was based on the theoretical work of Welch et al. (1980) who, using a Monte Carlo model, showed that the percentage of the incident total solar energy exiting through the sides of finite clouds is well represented by the percentage of incident energy exiting the sides in the visible (λ < 0.7 μm) spectral region. This relationship was developed assuming zero absorption in the visible spectrum.
The amount of energy ‘leaking’ out of the inhomogeneous cloud in the 0.3–3.0 \( \mu \text{m} \) interval is therefore approximated from the measured 0.3–0.7 \( \mu \text{m} \) convergence.

Rawlins (1989) describes in some detail the mathematics of this correction scheme and his working will be presented here for clarity. If one considers a horizontal slab of inhomogeneous clouds then the radiative properties of the slab can be characterized by:

\[
A + R + T + E = 1. \tag{5}
\]

Where \( A \) is the cloud absorptance, \( R \) the reflectance, \( T \) the transmittance and \( E \) is a term describing the net energy gain or loss through the cloud sides. The absorptance of the inhomogeneous cloud layer is given by:

\[
A = \tilde{A} - E \tag{6}
\]

where

\[
\tilde{A} = 1 - (F_{\text{top}} \uparrow + F_{\text{base}} \downarrow + F_{\text{top}} \downarrow) / F_{\text{top}} \downarrow \tag{7}
\]

is the directly measured absorptance if cloud edge effects are ignored and \( F_{\text{top}} \downarrow \) is the downward irradiance at cloud top, \( F_{\text{top}} \uparrow \) the upward irradiance at cloud top, \( F_{\text{base}} \downarrow \) the downward irradiance below the cloud layer, and \( F_{\text{base}} \uparrow \) the upward irradiance below the cloud layer.

In any practical application of Eq. (6), the term \( E \) will include effects of both cloud inhomogeneities and sampling errors. Ackerman and Cox’s assumptions, discussed above, are a consequence of the scattering properties of clouds varying only gradually with wavelength in the solar region. Using the technique of Ackerman and Cox one can equate the term \( E \) to the apparent absorption at visible wavelengths \( A_{\text{vis}} \), if the true absorptance in the visible, \( A_{\text{vis}} \), is assumed to be negligible. Strictly, \( A_{\text{vis}} \) is a slight overestimate of \( E \). If one uses the subscript ‘vis’ to denote the measurement of derived quantities in the visible band (0.3–0.7 \( \mu \text{m} \)) and non-subscripted symbols refer to quantities in the total solar band (0.3–3.0 \( \mu \text{m} \)) then the solar absorptance of an inhomogeneous cloud layer can be found from:

\[
A - A_{\text{vis}} = \tilde{A} - \tilde{A}_{\text{vis}}. \tag{8}
\]

On the MRF C-130 the visible irradiance is not measured directly but is instead computed from the difference between the total irradiance (0.3–3.0 \( \mu \text{m} \)) and the near-infrared irradiance (0.7–3.0 \( \mu \text{m} \)).

In these calculations it is assumed that there has been no major cloud evolution during the time taken to make the above and below cloud measurements, which was a maximum of 60 minutes. By calculating the absorptance of a cloud layer using Eq. (8) it is therefore possible partly to remove sampling errors associated with inhomogeneous cloud fields, and this allows the direct comparison of the absorptance of different cloud sheets with those predicted by models. The aircraft observations in and around the cloud were made such that the aircraft was advected with the mean boundary-layer wind and hence remained in the same cloud layer for the duration of the observations. This removes some of the problems of having different cloud advect into the measurement region during the course of the observations; however, there is still a residual error due to the evolution of the cloud field, but this is assumed negligible.

Newiger and Bähnke (1981) performed radiative-transfer calculations at a wavelength of 0.55 \( \mu \text{m} \) to assess the magnitude of the \( E \) term. Their results showed that the magnitude of the \( E \) term varies with the horizontal extent of the cloud but could amount to 20–30\%.

Rawlins (1989) presented an error analysis for the technique of applying a correction to the measured absorptance, using apparent absorption in the visible region of the spectrum. The
magnitude of the error in absorption is likely to be different for each case as it is dependent on the cloud structure. Rawlins, in his Table 6, estimated the error in the absorption, due to the assumption that $E$ can be determined from the divergence in the visible region of the spectrum, ranged between 0.01 and 0.02, based on the Monte Carlo simulations of Welch et al. (1980). When combined with instrument errors Rawlins found that, for his cases, the possible systematic error in the absorption ranged from about 0.02 for the smallest absorption to around 0.04 for the largest.

More recently, Hayasaka et al. (1995) have returned to this problem; they also used a Monte Carlo model to check the validity of the second assumption. Hayasaka et al. point out that since the difference between visible and near-infrared flux divergence depends on the cloud geometry and the cloud microphysics it cannot be evaluated quantitatively in advance. Their conclusions were that such a technique is valid as a means of correcting aircraft measurements of cloud absorption.

If one takes a systematic calibration error of 2% for measurements under diffuse illumination and 3% for measurements under direct illumination, then if one uses the irradiances from the cloudy flight A146, the error in the measurement of the true absorptance is ±0.05. This calculation has assumed that since the same pyranometers are used above and below cloud the measurements are not truly independent. If we take this error as an indicator of the upper range of possible errors in all the observations then, given the estimate by Rawlins (1989) of 0.01–0.02 for assuming that $E$ is equated to the visible divergence, one arrives at a maximum possible error in our calculation of the true absorptance of between 0.06 and 0.07.

Figure 10 shows the modelled (220-band spectrum) and observed absorptance, albedo and transmittance for the four cloud cases studied using both the maximum and minimum profiles of LWC/r_c in the model initialization. Figure 10(a) shows the model absorptance plotted against the apparent absorptance, $A$ in Eq. (8). The apparent absorptance ranges from −0.13 to +0.17. Figure 10(b) shows the true absorptance, $A$ in Eq. (8), this ranges from 0.025 to 0.057. The difference between these two measurements of absorptance clearly indicates the problems of the inhomogeneity of the cloud field and the importance of the term $E$. The model runs with the minimum profiles of LWC/r_c show the better agreement with the observations. The albedo and transmittance calculations generally agree to within ±0.1 when the minimum LWC/r_c profile results are considered. The calculation of absorption is inherently more sensitive as it is the residual of several larger numbers; the scatter of values is therefore quite large. Of note is the fact that the model gives larger absorptances than the true absorptance, Fig. 10(b), for all initializations. Also the true absorptance is never greater than 0.06 and the modelled absorptance, even using the maximum profiles of LWC/r_c, does not exceed 0.1. If one takes our estimate of the largest possible error in the calculation of the true absorptance then the largest absorptance in this data set could be as much as 0.13.

The inhomogeneity and sampling-error term, $E$, described above, for the four flights was −0.18 for flight A140, 0.02 for A146, 0.14 for H806 and 0.06 for H941. The variability and magnitude of this term shows the importance of accounting for the inhomogeneity when calculating absorptance from observations.

6. Sensitivity tests

Using the aircraft profiles of temperature and humidity from flight A200 the sensitivity of the code to changes in the profile, minor constituents and model spectrum has been studied. Figure 11 shows the model downwelling irradiances for a range of test cases. The line labelled CLR is that using bands 3 to 83 (0.295−2.77 μm) of the 95-band spectral
Figure 10. Scatter plots of the modelled (220-band spectrum) and (a) observed absorptance, (b) true absorptance, (c) observed albedo, and (d) observed transmittance for the four cloud cases. The model results are those initialized with the minimum and maximum profiles of LWC/r_c as explained in the key.

Figure 11. A comparison of the model downwelling irradiance, initialized using the profile of flight A200. Details of model runs are given in the text.
file and CLRH that using bands 33 to 194 (0.294–2.78 μm) of the 220-band spectral file. MAXH and MINH are the same as CLRH but the specific humidity inferred from the aircraft profile of dew-point has been modified by increasing and decreasing, respectively, the dew-point measurements by 1.0 K. The line labelled NTH is the model profile of downwelling irradiance as for CLRH but without CO₂, CH₄, N₂O or O₃. It should be noted that the MINH line is overlaid with the CLR line and hence is not visible on the figure. The results of changes to the dew-point data and separate changes of the number of bands used have had no appreciable effect on the model irradiances. The inclusion of minor constituents, particularly CO₂, and O₂ is, however, shown to be important as they combine to give a further 10 W m⁻² absorption in the atmosphere. Removing the O₂ results in an increase in the irradiance of 4.97 W m⁻², then removing CO₂ gives a further increase of 3.54 W m⁻² and finally removing CH₄ gives a further increase of 1.01 W m⁻².

7. Comparisons with other absorption data

The radiation code of SS has been widely used in the atmospheric science community to model the short-wave radiative properties of the atmosphere. This code used the absorption data of LOWTRAN 3B (Selby et al. 1976). It is, therefore, instructive to compare results from a model run using LOWTRAN 3B spectral data with both the new code and aircraft observations.

Figure 12 shows a comparison of the new code run in three modes for flight A200 (with no aerosol in the profile). The solid line is the irradiance profile obtained using bands 33 to 194 of the 220-band resolution new radiation code as in Fig. 11. The gaseous transmission data in this run comes from HITRAN92 (Rothman et al. 1992) and LOWTRAN 7 (Kneizys et al. 1988) with the continuum of Clough et al. (1989).
The dashed line is that obtained with bands 33 to 194 of the 220-band resolution new radiation code but with the spectral data used in the SS code, namely that of LOWTRAN 3B with no continuum. To make this spectral data the ESFTs and Rayleigh scattering coefficients were extracted from the original SS code and re-arranged in a form suitable for the new code.

The dotted line uses a spectrum of 24 bands in the new code, with limits identical to those of the original SS code, but with the newly defined spectral data from HITRAN92 and LOWTRAN 7 which includes the continuum of Clough et al. (1989). The code run to produce the dotted line also differs from that used to produce the dashed line in its specification of the solar spectrum (based on Labs and Neckel (1970) for the dotted line but Thekaekara and Drummond (1971) for the dashed line).

In the calculations the total irradiance at the top of the atmosphere is fixed, but because the revised solar spectrum has a larger fraction of its irradiance outside the pass-band of the clear dome, the predicted downwelling irradiance at the top of the atmosphere is smaller. The causes of differences lower in the atmosphere are harder to apportion, and reflect the differences in both the solar spectrum and the gaseous data. The differences between the 220-band results and the revised 24-band results arise mainly from the neglect of absorption by CO₂ and O₂ in the 24-band version of the code, as can be seen from comparing Fig. 11 with Fig. 12.

8. DISCUSSION AND CONCLUSIONS

(a) Summary of clear-air comparisons

Figure 13 shows all the results of the clear-dome downwelling irradiances compared with the 220-band version of the new code, and Fig. 14 shows the results for the downwelling red-dome irradiances compared with results from the 95-band version of the code.

With the exception of a few points, some of which can be explained by observational errors at large solar zenith angles, the model results lie within, or very close to, the estimated accuracy of the observations. The accuracy of the measurements of diffuse irradiances is around 2% whereas for direct-beam irradiance of the pyranometers this is increased to 3% due to the non-cosine response of the domes and the necessary levelling corrections.

There is some evidence in Fig. 13 of a trend in the differences between model and observations giving larger differences at longer path lengths. However, this trend is not apparent for the red-dome irradiances in Fig. 14; it may be that there is some residual absorption by aerosol that has been unaccounted for in the visible but the effect is too small to distinguish unambiguously.

The limited range of model results using the 4-band version of the code, designed for use in the UM, has shown that there is a small systematic difference which amounts to 1–2%, such that the model and observations agree to within 4% (Fig. 1 and Fig. 3). This systematic difference between the 4-band and the other model results is due to its spectral range which extends to 5 µm, leading to a 1–2% increase in irradiances. This accuracy is probably sufficient for the forecast mode of the UM, where the importance of the increased speed of operation of the radiation code of the 4-band version outweighs the slight decrease in accuracy.

(b) Summary of boundary-layer cloud results

All the boundary-layer-cloud results are shown in Figs. 7 and 8. The downwelling irradiances above the cloud from the model and observations agree to within 4%. However, once in and below the cloud the differences become larger. The cloud simulations using
Figure 13. All clear-air flights: Clear-dome (0.3–3.0 μm) downwelling irradiance, 220-band model spectrum. (a) Observed irradiance versus model irradiance; solid line is the 1:1 fit and the dashed lines are ±2%. (b) Percentage difference in irradiances (observed − model)/observed) as a function of pressure.

Figure 14. As Fig. 13 but for red-dome (0.7–3.0 μm) downwelling irradiance and 95-band model spectrum.
the minimum LWC/\(r_e\) profiles result in the best agreement, generally to within ±20%. This suggests that even in a stratiform cloud field which appears uniform to the eye, as was the case for most of these observations, regions within the cloud where the ratio LWC/\(r_e\) is low (i.e. regions of low optical depth) have a major impact on the cloud's radiative properties, more so than regions where the ratio is above the mean. This may be due to the nature of the changes in albedo with optical depth such that, for higher optical depths, the albedo is much less sensitive to further increases in optical depth.

Another possible explanation for this observation is that the funnelling of radiation through regions of low optical-depth cloud is an efficient way in which the albedo can be reduced and the transmittance increased. This problem of the internal inhomogeneity of clouds and their representation in models is an area of further study and beyond the scope of this paper. Nevertheless, these results show that there can be serious deficiencies in treating a cloud as a plane parallel layer.

The transmittance and albedo calculations have shown that the 220-band version of the code and the observations agree to within ±0.1. For both maximum and minimum LWC/\(r_e\) profiles the model albedo tends to be higher than that observed, in common with previous measurements (Stephens and Tsay 1990). However, the calculations of absorption show differences of a factor of two. The calculation of the error term, \(E\), clearly shows the problems of cloud inhomogeneity, and casts doubts over the validity of treating clouds as plane parallel layers in radiation models.

(c) Conclusions

These results have clearly demonstrated the new radiation code's ability to model the downwelling solar irradiance, both over the entire solar spectrum (0.3–3.0 \(\mu m\)) and over the near-infrared region (0.7–3.0 \(\mu m\)), in clear skies, for a range of atmospheric paths, to within 3% of the observed irradiances. It is important to note that the calculation of flux divergence is inherently less accurate owing to its calculation from the difference between several irradiance measurements. It is, therefore, much more difficult to validate this aspect of radiation codes to this level of accuracy. The importance of aerosol in the energy balance at the surface has also been demonstrated in flight A200.

The modelling of the upwelling solar irradiance has shown larger differences compared with observations. The upwelling irradiances in clear skies include any errors in the downwelling irradiances plus an error term associated with the modelling of the sea-surface albedo. There is still considerable scope for improving the modelling of sea-surface albedo, and this is a region of further study being undertaken by the authors.

The cloudy-sky results have further highlighted the problems of representing an internally inhomogeneous cloud field as a plane parallel layer within a radiation model. Clearly further study is required to address this problem. All of the cases presented have been for single-layer water clouds. Further study is required to test the ability of the new code to model the radiative transfer associated with a mixed phase or ice cloud. Here the problems of inhomogeneity will be further complicated with the problems of representing irregular crystals in the scattering and absorption calculations.

The measurements of cloud absorption have shown maximum values of around 6% for the stratocumulus cases studied. Our estimate of the absolute error in measuring cloud absorbance was between 0.06 and 0.07. Even after taking these possible errors into account, and assuming they act to increase the cloud absorption, these results are consistent with previous measurements in boundary-layer cloud (Stephens and Tsay 1990) and do not support the enhanced absorption suggested by Cess et al. (1995). An analysis of inhomogeneities and associated sampling problems has shown that up to 18% of the solar
irradiance incident on the cloud top can be affected, and if this radiation were not attributed to inhomogeneity effects but included in the absorption term then considerably higher cloud absorptions could be inferred.

This work has addressed the modelling of broad-band irradiances and compared them with aircraft observations. There is still a need for further narrow-band spectral measurements to be compared with modern radiation codes to enable the study of the details of gaseous absorption.

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REFERENCES


Houghton, J. T., Jenkins, G. J. and Ephraums, J. J. (Eds) 1990 *Climate change. The IPCC scientific assessment*. Cambridge University Press


Labs, D. and Neckle, H.


Martin, G. M., Johnson, D. W. and Spicu, A.


Newiger, M. and Bahnke, K.

Nicholls, S.


1994 The measurement and parameterisation of the ejective radius of warm stratocumulus clouds. J. Atmos. Sci., 51, 1823–1842

1972 ‘Optical properties of the atmosphere (third edition)’. Air Force Cambridge Research Laboratories


Ramawat, V. and Freidenrich, S. M.

Rawlins, F.

Rockel, B., Raschke, E. and Weyer, B.


Saunders, R. W.

Saunders, R. W., Brogniez, G., Buriez, J. C., Meeköetter, R. and Wendling, P.

Selby, J. E. A., Shettle, E. P. and McClatchey, R. A.

Seymour, J. H. and McHaffie, A.


1992 A comparison of measured and modelled broad band fluxes from aircraft data during the ICE’89 field experiment. J. Atmos. Oceanic Technol., 9, 391–406


1982 On the short wave radiative properties of stratiform water clouds.


Slingo A. and Schrecker, H. M.

Stephens, G. L. and Tsay, S. C.

Ströml, J., Busen, R., Quante, M., Guillemet, B., Brown, P. R. A. and Heinzeinberg, J.

Thekaekara, M. P. and Drummond, A. J.

Toon, O. B., Pollack, J. B. and Khare, B. N.

Wetzel, R. M., Cox, S. K. and Davis, M.


1992 A comparison of measured and modelled broad band fluxes from aircraft data during the ICE’89 field experiment. J. Atmos. Oceanic Technol., 9, 391–406


1976 The optical constants of several aerosol species: ammonium sulphate, aluminium oxide and sodium chloride. J. Geophys. Res., 81, 5733–5748