Microphysical and short-wave radiative structure of wintertime stratocumulus clouds over the Southern Ocean

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SUMMARY

Results are presented from an aircraft measurement campaign carried out over the Southern Ocean near the north-west coast of Tasmania, Australia. The microphysical and radiative characteristics of marine stratocumulus cloud sheets were sampled on four days, three of which are considered baseline days when air parcels had traversed long distances over the ocean without having been exposed to anthropogenic sources of pollution. Clouds were depleted by drizzle, with cloud liquid-water often smaller than 50% of the expected adiabatic value. Horizontal variability in liquid-water content associated with non-drizzle droplets was primarily caused by variations in the droplet number concentration, which was found to be among the lowest ever recorded (10 to 40 cm⁻³ for clouds of up to 300 m deep). Cloud albedos integrated over the solar spectrum varied from 40 to 60%, which roughly agreed with radiative-transfer computations. Data from one non-baseline day were also obtained and compared with the data obtained during baseline conditions. The functional dependence of cloud optical depth on liquid-water path was found to be much stronger on this day than the others, due to the smaller size of the cloud droplets present. Calculations show that on the days when drizzle was intense, the cloud optical depth could have been reduced by as much as 50% by shifts to larger values of the cloud droplet effective radius.

KEYWORDS: Cloud albedo, Cloud microphysics, Marine stratocumulus, Radiative transfer

1. INTRODUCTION

A large portion of the oceans in the temperate climate regimes is covered by sheets of stratocumulus clouds. They are formed near the top of the marine boundary layer and derive their characteristics from boundary-layer processes which link the lower troposphere to the ocean surface. Observations and modelling of the marine boundary layer over the last decade have uncovered some of the basic principles controlling the existence of stratocumulus clouds. The fundamental reason for the interest of the scientific community in stratocumulus clouds derives from their impact on climate. Being one of the lowest water clouds in the troposphere, they emit long-wave radiation at a high temperature, but also have a high albedo. Thus, they exert a cooling influence on climate.

The albedo of stratocumulus clouds is governed by the size of the particles, which is usually expressed as the effective radius (i.e. the ratio of the third and the second moment of the size distribution), and the depth of the cloud. The effective radius, \( r_e \), is an artificial size parameter linking the liquid-water content with the extinction coefficient, and is controlled by the number of condensation nuclei existing in the sub-cloud layer which act as cloud condensation nuclei (CCN) near cloud base. The more CCN present, the smaller \( r_e \) will be, and vice versa. Thus, the mechanisms responsible for the creation of CCN exert a powerful influence on cloud albedo.

The theory describing this process has been developed by Twomey (1977) in the context of the influence of anthropogenic pollution on the earth’s climate. There is evidence in the form of ship track measurements (Radke et al. 1989) to suggest that the general principles of this theory are correct.

Over the oceans, in the absence of anthropogenic influences, Charlson et al. (1987) have suggested that the emission of dimethylsulphide (DMS) from phytoplankton sources in the ocean provides a natural source of CCN. DMS is oxidized in the lower atmosphere to sulphates which can act as CCN. Since the production of DMS is controlled (amongst

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others) by temperature, it was postulated that global warming would increase DMS, CCN production, and hence increase cloud albedo, thus providing a natural negative feedback to global warming. Boers et al. (1994) have shown that there appears to be a link between surface-based observations of CCN and satellite-retrieved cloud optical depth by examining the seasonal cycle in these measurements. The cloud optical depth and CCN follow coherent seasonal patterns in phase with the cycles in DMS production, suggesting that the Charlson et al. (1987) hypothesis is probably sound as well.

It is clear that it will be hard to discern from the results of most stratocumulus field experiments conducted in the last four decades whether the observations are representative of natural conditions, or contaminated by anthropogenic influences. Most experiments took place near the continent of Europe (Slingo et al. 1982) or the west coast of the United States (Brost et al. 1982a, b; Randall 1984; Boers and Betts 1988) in the vicinity of major pollution sources so that some form of anthropogenic ‘contamination’ probably did occur. However, it is essential to know the natural background condition because it provides the reference against which any (perturbed) conditions can be checked. It is with this concept in mind that a cloud study programme was formulated to investigate the structure of marine stratocumulus clouds in the unperturbed conditions over the Southern Ocean.

The location of the experiment is the Southern Ocean in the immediate vicinity of the north-western tip of Tasmania (Cape Grim) where the Cape Grim Baseline Air Pollution Station (CGBAPS) has recorded the baseline atmospheric chemistry components over a period of 15 years. Under baseline conditions the marine boundary-layer air arriving at this station has a history of 4000 to 5000 km of travel over the world’s most remote ocean, and can be viewed as almost completely devoid of anthropogenic influences. Therefore, the clouds that are formed under these conditions probably provide the best ‘laboratory’ to study the microphysical and radiative structure of clouds in a pristine environment.

Since it is known that DMS and CCN at this site go through well described seasonal cycles (Gras 1989) the Southern Ocean Cloud EXperiment (SOCEX) was designed to consist of two phases, one in the austral winter (July) when the source of oceanic DMS is almost non-existent, the other in the austral summer (January) when the DMS activity is at its peak. This paper describes microphysical and radiative results from the first (winter) phase of this experiment which was conducted in July 1993.

Cloud droplet, radiation and CCN observations around the Australian continent have been collected for well over 40 years now. As early as 1956 it was found (Squires 1956) that maritime clouds generally produced larger and fewer droplets than continental clouds. Over the years, as more and more experimental evidence was gathered, a link was found between the cloud droplet concentration and available CCN (Squires 1958a, b, c; Warner and Squires 1958; Twomey 1959; Twomey and Squires 1959; Twomey and Warner 1967), as well as a difference between available CCN in maritime and continental atmospheres (Twomey 1963). The microphysical structure of cumulus cloud from mostly maritime origin was comprehensively dealt with by a series of articles by Warner (Warner 1969a, b; 1970; 1979).

While this early work mostly focused on the microphysical aspects of cloud development, the importance of the radiative structure and its link to the microphysical structure was mostly recognized in the early 1970s from which time several aircraft studies focused on the radiative structure of clouds east and south-east of the continent (Platt 1971, 1976; Paltridge 1974, 1976; Stephens 1978; Stephens et al. 1978; Stephens and Platt 1987; Twomey and Seton 1980; Twomey and Cocks 1982).

Interestingly, very few of these measurements actually took place near Tasmania. In our area of interest it seems that the only measurements taken were by Stephens (1978, south of Hobart), Twomey and Cocks (1982, north-west Tasmania), Mossop (1985, west
Tasmania), and Mossop et al. (1968, 1970, west Tasmania), although from some papers it is unclear where measurements took place. Also, in some papers (in particular Mossop's work) the emphasis was mostly on ice phase clouds, and the collection of droplet spectra was mostly justified on the basis of their importance in the initiation of ice multiplication processes.

Most research focused on regions south-east, east, or north-east of the main continent and, apparently, the region west of Tasmania has rarely been sampled. In the light of the new evidence linking DMS to CCN to cloud microphysics, it is of considerable interest to study the cloud structure in the unpolluted maritime regions of the globe; hence the choice of the ocean near the north-western tip of Tasmania.

2. INSTRUMENTS

(a) Aircraft

The aircraft platform used in the experiment is the Fokker F-27 formerly owned by the Commonwealth Scientific and Industrial Research Organisation (CSIRO) (now the property of the Australian Flight Test Services in Adelaide). It has a heavy lift capability, and a medium-range distance. During SOCEX, the maximum range, fully loaded and fully manned, was 4 hours. Instruments used to study the microphysical and radiative structure consisted of sets of radiometers (upward- and downward-looking retractable Eppley pyranometers (0.3–3 µm), and pyrgeometers (> 4 µm)); four cloud probes: a Particle Measuring Systems (PMS) Forward Scattering Spectrometer Probe (FSSP-100), a Particle Volume Monitor (PVM) cloud probe (Gerber et al. 1994), a King probe (King et al. 1978), and a PMS Optical Array (2D-C) particle probe; a Barnes PRT-5 narrow field-of-view radiometer; and instruments to recover the state variables (pressure, temperature and humidity). The aircraft has been used over many years and its capabilities are well described (Stephens and Platt 1987).

(b) Measurements

(i) Liquid water. Because of the importance of accurate liquid-water measurements for the experiment, four liquid-water probes were on board the F-27. These instruments work on different principles with the result that each yields different values of liquid water.

The FSSP instrument sizes and counts individual droplets. Its problems are more or less known from evaluations by Baumgardner and Spowart (1990) and include non-uniformity of the laser beam, response time, and coincidence and dead-time losses. The FSSP observations from the F-27 were corrected for coincidence and dead-time losses using the iterative procedure of Baumgardner et al. (1985). The data have not been corrected for laser beam non-uniformity. For sizing we have used the PMS nominal size bins; i.e. 15 size bins of 3 µm width up to the maximum diameter of 47 µm. Baumgardner and Spowart's (1990) analysis of the laser beam non-uniformity shows that the sizing is not absolutely accurate. Some droplets are sized larger than their actual size, some smaller. This is the case over the entire sizing range including the nominal maximum size of 47 µm.

The PVM probe measures light diffracted in the forward direction by cloud droplets. Variable transmission filters are used in order to derive the liquid-water content and the particle surface area from the flux of scattered light incident on large-area photosensors. The area illuminated by the laser source is 1.25 cm², so that the probe observes a volume of particles rather than individual droplets; hence its name. Because the liquid-water content and the particle surface area can be obtained directly, a measurement of the effective radius is available because this parameter is proportional to the fraction of these two quantities.

The PVM probe was designed to overcome the well-known problems with the FSSP. The response curve for the PVM-100 probe shows a roll-off for droplets larger than 40 µm,
so that at 70 μm the liquid-water content is underestimated by 50% (Gerber 1991; Gerber et al. 1994). The FSSP provides droplet spectra, but these are based on small droplet volumes. The PVM uses volume scattering and therefore provides a better statistical representation. None of the two probes is designed to provide accurate measurements of droplets above 50 μm.

Since wintertime clouds over the Southern Ocean produce large droplets with a significant fraction of droplets of diameter larger than 40 to 50 μm, the two probes are operated near the limits of their respective response ranges and are likely to give somewhat inaccurate values of liquid water and effective radius. In seeking to understand the differences between the two probes, a third liquid-water probe (the King probe) was mounted on the aircraft. The King probe is a calibrated hotwire probe. Power necessary to evaporate droplets is extracted from the probe. The maximum diameter of droplets that stick to the probe without breaking up in the process is unknown, but probably larger than 47 μm. As a result, this probe is likely to measure more liquid water than either the PVM probe or FSSP as they reach their response limits.

Standard alignment and sizing calibrations of the FSSP were performed within two weeks of the beginning and end of the experiment. Small shifts in sizing were observed, but no changes in depth of field. The PVM probe was calibrated before the experiment and checked several times during the experiment. As expected, significant differences were found between the liquid-water measurements of the three probes, which we will discuss later (section 4).

One further particle probe was mounted on the F-27, namely the PMS 2D-C probe. The probe measures droplet sizes well into the range of precipitation particles (up to 1000 μm radius), and thus can be used to improve on the values of the liquid water and effective radius obtained from the FSSP and PVM probe. The 2D-C probe proved to be essential in the interpretation of the observed microphysical and radiative structure, as on some occasions up to 40% of observed liquid water was distributed in the spectral range of the 2D-C probe.

(ii) Short-wave radiometers. Pyranometers were calibrated against standard pyranometers before the experiment, and against absolute standard sources at a calibration facility at the Bureau of Meteorology in Melbourne after the experiment. Data collected above clouds were corrected for roll, pitch and yaw of the aircraft. However, analysis of the data from the upward-pointing radiometer showed inconsistent comparisons between the measured down-welling solar flux and computations based on a 24-band solar-transfer model through an atmosphere with local temperature and humidity profiles. Further analysis revealed the inconsistency to be due to a 3° tilt of the upward-looking radiometer. With high sun angles (larger than 60°) typical of wintertime conditions over the Southern Ocean, the tilt gives rise to large errors in the observed downward solar flux, very little of which is diffuse. The procedure by which the tilt angle could be removed from the data is briefly described in the appendix.

Results show that after the correction procedure had been applied, measured and computed solar flux agreed to within 17 W m⁻², or 4% of the observed flux. This agreement is approximately equal to the 5% error typically assigned to this instrument. Any tilt of the downward-pointing radiometers is unknown, but of no consequence to the result.

(c) Flight plans

Figure 1 shows the geographic location of the experiment. All five flights (on four days) described in this study were conducted in the immediate vicinity of the west coast of
Tasmania. There were two principal conditions to determine whether flights would be conducted: (1) the existence of stratocumulus clouds as observed from real-time satellite data (NOAA\textsuperscript{t}-11, -12, and GMS\textsuperscript{t} data), and (2) the occurrence of ‘baseline’ conditions. The atmospheric state was considered to be baseline when at CGBAPS the wind direction was between 190 and 280\degree. The baseline condition is probably somewhat over-restrictive when applied for regions over the open ocean. However, the CGBAPS records taken over many years indicate that this condition is a reasonable guarantee that the sampled air is uncontaminated by anthropogenic influences, which was confirmed by back-trajectory analysis (courtesy of the National Institute of Water and Atmosphere, New Zealand (Kristaoten 1994)).

Flight plans called for direct transits to predetermined positions over the ocean where several stacked flight legs were flown. Flight legs consisted of one radiation leg at 200 to 500 m altitudes above the clouds, followed by a descent to near the ocean surface. After a surface leg, several flight legs were conducted starting just below cloud base, followed by two or three legs at various altitudes above cloud base. A stack was completed by a second radiation leg above cloud top. If time permitted, a stack was repeated. A stack was oriented perpendicular to the mean boundary-layer wind and was flown ‘true heading’, i.e. the aircraft was permitted to drift with the mean wind in order to maximize the possibility of sampling the same airmass a number of times, although this can never be guaranteed. The positions changed from day to day depending on the clouds. Except for one stack (11 July) all were executed within 150 km from the coast. Each leg was approximately 40 km in horizontal distance.

\textsuperscript{t} National Oceanic and Atmospheric Administration.
\textsuperscript{t} Geostationary Meteorological Satellite (Japan).
3. Radiative-transfer model

A significant part of the analysis has focused on the analysis of cloud albedo measurements. These measurements were compared against radiative-transfer calculations with local conditions as inputs. The radiative-transfer model has been developed along the lines outlined by Slingo and Schrecker (1982). It is a 24-band two-stream model using the delta-Eddington method to compute multiple scatter in clouds. Some small modifications were introduced primarily with respect to the parametrization of the cloud extinction in terms of the liquid-water content and effective radius (Boers and Mitchell 1994). Furthermore, the band water-vapour transmissivities were computed using the LOWTRAN-7 transfer model (Kneizys et al. 1988).

A special issue of the Journal of Geophysical Research was published in 1991 on the intercomparisons of radiation codes. One particular study in this issue (Fouquart and Bonnel 1991) details a comparison of line-by-line calculations of various standard atmospheric profiles with high- and low-resolution short-wave radiative parametrizations. We submitted our code to 30 of the 57 benchmark tests reported in this study, including six cloudy test cases. Accuracy of our computations was generally better than 2%.

4. Results

(a) Weather condition

North-west Tasmania is situated in the ‘roaring forties’. In the wintertime the weather is characterized by frequent passages of cold fronts. Ahead of those cold fronts, Cape Grim is exposed to continental air often arriving directly from Melbourne which is located 300 km north of Cape Grim on the mainland of Australia. After the cold front passage, the Cape is usually exposed to several days of baseline conditions during which the weather rapidly stabilizes and stratocumulus clouds are prevalent.

In the time period between 11 July and 20 July 1993, the weather in northern Tasmania was primarily governed by two high-pressure systems which (except for 11 July) limited the exposure of this region to air from the mid- to high-latitude regions of the Southern Ocean (Fig. 2). On the 16th a high-pressure system (1028 hPa) was situated over South Australia and another (1033 hPa) west of Perth over the Indian Ocean. The airflow in the vicinity of Cape Grim was moderate from the west-north-west ahead of a weak cold front in the South Australian Bight. This front crossed Tasmania on the 17th, after which two high-pressure systems merged into one on the 18th with the centre (1040 hPa) in the Southern Bight. On the 19th winds were weak from the west, with the region situated close to the axis of a ridge extending from the high into the Tasman Sea. Early on the 20th a cold front crossed the area.

The continuous CCN record at Cape Grim shows that the period from 15 to 20 July 1993 was characterized by the lowest CCN concentrations measured during the one month of experimental measurements. Two flights were carried out on the 16th, one on the 19th and one on the 20th. Except for the 19th, the observed cloud decks covered the ocean near Cape Grim more or less uniformly. On the 19th, the cloud deck was broken and consisted of mostly ‘popcorn’ type structures. Clouds were observed to precipitate on all days, especially on the 16th and 20th. The sampled clouds consisted exclusively of water droplets, except for one flight leg on the 20th, where occasional ice crystals were recorded by the 2D-C probe. One further day of baseline conditions was sampled. However, significant ice formation occurred on this day (26 July), and consequently the results of the radiative analysis cannot easily be interpreted. Figure 3 shows an AVHRR* visible image

* Advanced Very High Resolution Radiometer.
taken on the afternoon of 16 July. It shows the experimental area more or less uniformly covered by stratocumulus clouds. As the clouds approach the coastline there appears to be a large reduction in cloudiness over land. This is a typical occurrence near the west coast of Tasmania which acts as a large obstacle in the predominant westerly flows and even shows up in the ISCCP* data set on fractional cloudiness.

Both early on and late in the time period of the experiment (between 7 July and 26 July), deep cut-off low-pressure systems developed in the South Australian Bight region; these slowly drifted over the experimental area giving rise to long periods of unsettled weather, with non-baseline conditions prevalent. On 11 July a flight was carried out near a stratocumulus cloud bank under non-baseline conditions with the wind coming from continental directions. Results from this flight were compared with the other data and were found to be distinctly different from the data obtained during baseline conditions.

(b) Microphysical structure

(i) Probe comparison. Some effort was spent in comparing liquid-water-probe data characteristics. The primary result of this effort has been the conclusion that the characteristics of each individual probe are consistent from day to day, but when compared with each other they are not. The reason for this inconsistency is the uncertain response of the probes in the presence of large cloud droplets. It was found that on most occasions the King probe yielded the largest liquid-water values, and that the difference between the King probe and the other two probes increased when the droplet number concentration decreased. We will show just one example of a comparison between the probes.

Figure 4(a) shows a comparison between the King probe and the FSSP on 16 July 1993. Although the King probe measures more liquid water than the FSSP, the difference is less than 15% when the liquid water is less than 0.2 g kg\(^{-1}\), and increases nonlinearly to

about 40% for larger values of the liquid water, presumably due to the presence of droplets larger than 47 μm. Figure 4(b) shows a comparison for the PVM probe and the FSSP. The character of the point cluster is quite similar except that the two probes tend to agree better. Similar differences were obtained by Gerber (1995). Figure 4(c) shows the comparison between the PVM and King probes, which shows the best agreement, although the King probe gives slightly larger liquid water than the PVM probe, in particular for large values of the liquid water. Unfortunately, such comparisons yield different results on different days, the reason of which is unclear. The lack of change between the pre- and post-flight calibrations of the FSSP, and the continuous updated calibrations of the PVM probe, suggest that sudden calibration shifts were probably absent, although they cannot be ruled out entirely.

The difference between the probes is somewhat disconcerting. However, it should be kept in mind that the comparisons were done under testing conditions, when the droplet number concentrations were small, and the size of the droplets was near the limit of the
Figure 4. Comparison of liquid-water content (LWC) on 16 July 1993 between (a) the King probe and the FSSP, (b) the PVM probe and the FSSP and (c) the PVM and King probes.
response capability of both the FSSP and the PVM probe. In most of the results to be shown, we will focus on the FSSP data because of the spectral information that is available from this probe.

(ii) **Horizontal structure.** Figure 5 shows the horizontal structure of liquid-water content, effective radius and droplet number concentration, during one in-cloud flight leg on 16 July 1993 (first flight), while Fig. 6 shows a similar plot for an in-cloud flight leg on 20 July. Two curves are shown on Figs. 5(a), 5(b), 6(a) and 6(b) corresponding to results based on the FSSP data, and those based on the addition of FSSP and 2D-C data. What is particularly interesting in Fig. 5(b) is the lack of horizontal variation in effective radius based on the FSSP measurements only. The variability in liquid water is almost completely due to variations in the droplet number concentration. This is true for the 20 July data as well, although to a lesser degree. An explanation for this phenomenon is difficult to give.
If we examine the usual definition of effective radius and droplet number concentration:

\[
\begin{align*}
    r_e &= \frac{\int_0^\infty r^3 n(r) \, dr}{\int_0^\infty r^2 n(r) \, dr} \\
    N &= \int_0^\infty n(r) \, dr
\end{align*}
\]

where \(n(r)\) is the droplet density distribution as a function of droplet radius \(r\), and \(N\) is the droplet number concentration, it seems that the only possible explanation is that along the line of flight the droplet density distribution varies by a constant multiplication factor over the complete spectral range of non-drizzle droplets, i.e. a constant fraction of the density distribution is affected by drizzle and/or mixing processes, although it is not intuitively obvious why this should be so. Both figures also indicate the importance of the spectral ranges exceeding the response function of the FSSP in shaping the value of the effective radius and the liquid-water content. There is a large variability in liquid water
and effective radius when the 2D-C data are included in the computations. Although some of this variability must be due to natural variations in drizzle droplets along the line of flight, some of it may have been caused by restrictions in sampling due to the fact that in 1-second averaged data few drizzle droplets are found. However, it is important to note that, despite the presence of drizzle, the FSSP effective-radius data are not affected when drizzle droplets are present. This result will be used later when we address the influence of drizzle droplets on cloud albedo (see section 4 (c)).

(iii) *Average.* All microphysical data collected in the region where the stacks were flown
were averaged in 10 hPa pressure bin widths. Averages were determined as follows:

\[
\text{LWC} = \frac{4}{3} \rho \pi \frac{\sum_i \sum_j r^3 n(r)}{K} \tag{3}
\]

\[
N = \frac{\sum_i \sum_j n(r)}{K} \tag{4}
\]

\[
r_e = \frac{\sum_i \sum_j r^3 n(r)}{\sum_i \sum_j r^2 n(r)} \tag{5}
\]

\[
\sigma = \frac{\sum_i \sum_j Q_{ext} r^2 \pi n(r)}{K} \tag{6}
\]

where LWC is the average liquid-water content, \(\sigma\) is the extinction coefficient, and the summation of \(j\) refers to the spectral bins of both the FSSP and the 2D-C probe. The extinction factor \(Q_{ext}\) was obtained from Mie calculations for liquid water at 0.7 \(\mu\)m. The summation over \(i\) (number of samples is \(K\)) can have two different meanings. In the first instance it can refer to all samples in the 10 hPa bin, in the second instance it can refer to cloudy samples only. From the standpoint of radiative transfer, the first average is the most important one for both the liquid-water content and the extinction coefficient, as a radiometer receives radiation from cloudy and non-cloudy parcels alike. However, the second type of average can be equally relevant. In particular, this average could shed light on the departure of the cloud from the adiabatic lapse rates, and the value of the droplet number concentration in cloudy parcels. For the effective radius both types of averages result in the same value, as no division by \(K\) is necessary.

The averages are presented as a set of panels, starting with the first three case studies of uniform stratocumulus clouds, namely: 16 July, morning flight (Fig. 7); 16 July, afternoon flight (Fig. 8); 20 July (Fig. 9); the one flight with broken clouds, 19 July (Fig. 10); and the one non-baseline flight, 11 July (Fig. 11). The panels show the effective radius, the liquid-water content according to the first type of averaging, the liquid-water content and droplet number concentration according to the second type of averaging, the extinction coefficient which was generated by summing over clear and cloudy parcels alike (i.e. the
Figure 8. As Fig. 7 but for 16 July, afternoon flight.
Figure 9. As Fig. 7 but for 20 July.
Figure 10. As Fig. 7 but for 19 July.
Figure 11. As Fig. 7 but for 11 July.
first type of averaging), and finally the height integral over the extinction from the top of the cloud down to the actual level \((\tau(z) = \int_0^{\text{cloud top}} \sigma dz)\). \(\sigma\) indicates which levels in the cloud contribute most significantly to the cloud optical depth, while the last parameter near cloud base yields the cloud optical depth itself. When plotted as a function of height, it is the optical depth between the observing height and the top of the cloud, so that the parameter gives an indication for each level inside the cloud of how much the cloud above the level contributes to the total optical depth.

There are two lines in each graph, one line corresponding to the calculation based on the FSSP data only, the other based on the addition of the FSSP results and the 2D-C results. The second line clearly represents the true liquid-water content much more accurately than the first. However, the inclusion of the first line in the plot is instructive as it demonstrates the limits of the FSSP in measuring the total liquid-water content, which is particularly relevant for the first three case studies (see also Martin et al. (1994) for similar results).

The effective-radius calculations of the first three case studies demonstrate that the FSSP by itself is not a good probe to represent this important radiative parameter adequately. If we were to believe the results from this probe, the effective radius increases with height in a manner to be expected of non-precipitating clouds. However, if the observations of the 2D-C probe are included, the effective radius goes up by several microns in the top part of the clouds, and by several hundred microns near the bottom of the cloud. The inset in the figures is required as the scale of the large graph is insufficient to show the detailed variations in effective radius based on both probes. Similarly, the liquid-water content calculated with the observations of the 2D-C probe included show that up to 40% of liquid water is contained in droplets exceeding 47 \(\mu m\) diameter in the middle and top part of the cloud, and up to 100% near cloud base. These are very large numbers, and would cast doubt on any measurement of liquid water that relies solely on FSSP or similar probes. As will be shown later, the spectra are bimodal which shifts the effective radius to large values.

Interestingly, although a large proportion of liquid water is contained in the large droplets, they hardly contribute to the extinction coefficient, and thus they neither contribute much to the optical depth. The largest contribution was measured on the 20 July case study, but even in this case it was no more than 5%. The reason is that the larger droplets only make up a small proportion of the total number of droplets, and thus contribute to the projected cross-sectional area in only a minor way.

When considering Figs. 7(a), 8(a), 9(a) with Figs. 7(f), 8(f), 9(f) it can be concluded that the levels where the effective radius exceeds 25 \(\mu m\) (in the lower part of the cloud) only have a small contribution to the cloud optical depth.

Both on 19 July and 11 July, the contribution of the large droplets to all microphysical parameters is small. On 19 July this is due to the fact that the clouds were very broken so that not enough drizzle patches may have been sampled, and on 11 July to the fact that droplet number concentrations were high and non-baseline air was sampled.

On the 20th the massive depletion of liquid-water cloud droplets (Fig. 9(d)) in the upper part of the cloud is at least in part due to the conversion of droplets to drizzle particles. However, we suspect that this depletion is also partly due to mixing of the cloud with dry air situated in the overlying inversion. On this day the flight was conducted behind a cold front, with strong southerly winds in the lower boundary layer, which shifted towards the west inside and above the cloud and increased in strength. A similar type of vertical variation in droplet number concentration was found on the 11 July case study.

The unusual distribution of droplet concentration on these two days has been observed before in this region by Stephens (1978) and Stephens and Platt (1987). Both of these older
data sets were collected over the ocean south of the Australian continent, under conditions of very low droplet count, although it is unclear from both accounts whether or not their measurements truly represented 'baseline' conditions; i.e. advection of air over unpolluted regions.

From the liquid-water averages using cloudy samples only (Figs. 7(c)–11(c)) it appears that on all days the in-cloud liquid water is about 30 to 60% of the adiabatic value. Since the deviation is much larger than can be explained by errors in the measurements of any of the microphysical probes used, this result is robust.

From the measurements of the droplet number concentrations (Figs. 7(d)–11(d)) it appears that for baseline conditions the maximum 10 hPa averaged value does not exceed 50 cm⁻³. The maximum 10 hPa averages are summarized in Table 1. The difference between the non-baseline and baseline case studies is very pronounced. Several summaries exist of global droplet concentration measurements (Heymsfield 1993; Seidl 1994). When compared with these summaries it appears that our measurements rank among the lowest.

Although our only non-baseline case (July 11) is a maritime case study, the continental influence on this day more than triples the droplet number concentration when compared with the maritime baseline cases. It indicates the danger of simply classifying clouds as maritime and continental and may be the reason for several 'maritime' case studies mentioned by Seidl (1994) exhibiting large droplet number concentrations (200–350 cm⁻³). Despite the 'continental' contamination the concentrations recorded on this day would still have to be considered as low when compared with the data from Seidl's (1994) summary.

<table>
<thead>
<tr>
<th>Date</th>
<th>Condition</th>
<th>( N ) (cm⁻³)</th>
</tr>
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<tbody>
<tr>
<td>11 July 1993</td>
<td>non-baseline</td>
<td>165</td>
</tr>
<tr>
<td>16 July 1993(1)</td>
<td>baseline</td>
<td>36</td>
</tr>
<tr>
<td>16 July 1993(2)</td>
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<tr>
<td>20 July 1993</td>
<td>baseline</td>
<td>34</td>
</tr>
</tbody>
</table>

Note: (1) and (2) after the date refer to the first and second case study of 16 July.

(c) **Optical depth and liquid-water path**

For the purpose of parametrizing the optical properties of a cloud in large-scale models it is of some interest to quantify the relation between cloud optical depth and liquid-water path. The idea behind such quantification is that most numerical models predict the liquid-water path of clouds as a result of dynamical and thermodynamical considerations. If there would be a precise relation between the optical depth and liquid-water path, then the specification of the optical properties in those models would be greatly facilitated. However, an added advantage is that this type of relation can be used to compare individual case studies, and eventually different experiments. It can be shown that for a homogeneous cloud (i.e. a cloud with uniform liquid water and particle size) the relation between optical depth and liquid-water path can be approximated as

\[
\tau \approx \frac{3 \ LWP}{2 \ \rho r_e}
\]
Figure 12. Liquid-water path (LWP) versus cloud optical depth (τ). The crosses refer to data based on FSSP observations only, the stars to data based on the FSSP and 2D-C probe. For each case study the points are connected to show the effect on the LWP–τ relation of including the 2D-C data. The cross–star combination between the Stephens (1978) line and effective radius, \( r_e = 10 \, \mu m \) (\( \sim 8 \, \mu m \)) corresponds to the 11 July case study.

where \( \tau \) is the cloud optical depth, LWP is the liquid-water path, \( \rho \) is the density of liquid-water, and the geometric optics limit has been used. Both LWP and \( \tau \) can be found by integration over the third and second moment of the size distribution so that (7) can be derived in a straightforward manner from (1). This equation shows that the optical depth is inversely proportional to the effective radius, and linearly dependent on the liquid-water path.

For all the stacks that were completed during the five case studies presented above, the average cloud optical depth and liquid-water paths were calculated and are plotted in Fig. 12. The calculations were done by simply summing the liquid-water content and extinction values from Eq. (3) and Eq. (6) with height. Also plotted in Fig. 12 are relations of the type of Eq. (7), for different values of the effective radius, plus a separate relation based on Stephens (1978). Optical-depth values are moderate, and generally ranging from 5 to 8, with the exception of the 19 July case study, where the optical depth was close to 1.5.

It appears from the plot that there is a substantial shift in the relation between optical depth and liquid-water path when the 2D-C probe data are included in the microphysical observations. One can interpret each data point as being representative of a homogeneous cloud with the measured liquid-water path and representative effective radius found from Eq. (7). While most data points fall into a range of effective radii between 10 and 14 \( \mu m \) when only the FSSP data are included, the representative effective radius is increased to 16-18 \( \mu m \) when the 2D-C data are included. Interestingly, this result applies to all case studies where the effective radii near the bottom of the clouds were several hundred \( \mu m \) (16 and 20 July). The reason is that much of the multiple scatter of sunlight takes place in the top and middle part of the cloud, so that when the reflective capability of the cloud is considered the bottom part of the cloud does not contribute very much to the optical depth. This can be verified by examining the panels in Figs. 7 (e),(f)–11(e),(f).
The point representing the 11 July case study (near 8 \(\mu m\)) is the only significant departure from the point clustering between 16 and 18 \(\mu m\). This is not surprising given the fact that the droplets were on the whole a lot smaller and more numerous than on the other days.

The plot shows that no uniform relation exists between the liquid-water path and the cloud optical depth. The occurrence of drizzle increases the effective radius to values beyond those expected from a unimodal distribution. As a result, significant divergent relations are found dependent on whether the cloud contains drizzle droplets, or not, and probably dependent on the amount of drizzle droplets as well. Clearly, for wintertime baseline conditions over the Southern Ocean, it is possible to group the data in a limited range of effective radii (16 to 18 \(\mu m\) in this case). However, this relation is unlikely to hold for other areas of the globe. In view of the very large departure of our data from the relation found by Stephens (1978), which roughly corresponds to an isopleth of \(r_c = 6 \mu m\), it would be useful to review all microphysical data in this manner to assess the range of naturally occurring variability in the link between cloud optical depth and liquid-water path.

\((d)\) Albedo measurements and model comparisons

Figures 13 and 14 show cloud albedo as measured by the pyranometers, and the cloud-top temperature as measured by the Barnes PRT-5 infrared radiometer, as functions of time for two case studies, namely the uniform stratocumulus cloud on 16 July (first flight), and the scattered cloud on 19 July. In Fig. 13 cloud albedo ranges between 0.40 and 0.65, although from the rather constant cloud-top temperature it appears that there is probably little variation in cloud-top height with distance. There are small breaks in the cloud, as visible by the rapid increase in cloud-top temperature. It appears that the albedo
of the cloud is affected well away from the breaks on either side. It can be shown that the variations in cloud albedo near the cloud break cannot be caused by the large field of view of the radiometer, but are probably caused by a natural decrease in liquid water near the cloud break, most likely due to entrainment events.

Figure 14 shows the albedo signature for broken clouds. Although a relatively large clear section is observed near 1354 Australian Eastern Standard Time (EST), the cloud albedo seems slow to adjust to the typically observed 'clear' values of 0.05 to 0.06. Also, there is a sudden increase to about 0.10 well before the next 'cloudy' path is encountered. These signatures are attributed only partly to the response characteristics of the radiometer to very rapid changing conditions, but more importantly to the occurrence of scattered clouds outside the view of the Barnes radiometer, but inside the view of the downward-pointing pyranometer. Photographs taken during the flights revealed a very wide range of cloud sizes and spatial distributions.

For all flights, including the 19 July case study with the broken clouds, average albedo values were calculated. Tests were performed with the two-stream radiative-transfer model to investigate whether these albedo values could be adequately modelled. As explained above, model computations are to a large extent dependent on an adequate description of the microphysical structure, and in light of the discrepancy between the various probes the results shown in Fig. 15 should be viewed with some caution. Most of the albedo calculations (except for 19 July) range from 0.45 to 0.52, while the range of measurements is from 0.35 to 0.57. Based on these results it is clear that for this set of experiments the albedo could not be calculated with an accuracy better than about 10 to 15%.

The reasonable agreement between theory and observations found for the 19 July low-albedo case study must be seen as fortuitous. The assumption used in the radiative-transfer calculations, namely plane-parallel clouds, clearly does not hold for this case.
study. However, it is instructive to include this result for comparison with the other case studies, for which this assumption is more realistic. However, the horizontal variability of the albedos on 16 July for those regions where the PRT-5 indicates overcast conditions suggests that even in the case of uniform cloud height there must be considerable horizontal variation in liquid-water path in order to obtain albedo variations of 10–15%.

To investigate the importance of the liquid-water measurements in calculating albedos, the question may be asked by how much either the liquid-water path, or the effective radius, must be changed in order to obtain the actually measured cloud albedos. Cloud albedos were recalculated by multiplying the observed cloud liquid-water and effective-radius profiles by variable fractions and recording those fractions for which the calculated albedos matched the observed ones. The results are summarized in Table 2. This table also contains the measured, and modelled albedos, the liquid-water paths and optical depths derived from the combination of FSSP and 2D-C measurements. Table 2 indicates that the required changes in either liquid-water path or effective radius are quite large, and exceed the typical values for differences between the particle probes.

Although this seems a somewhat unsatisfactory result, it highlights the difficulty of accurately comparing measured and modelled cloud-top albedo. A comparison of radiation-model results with data is only as good as the specification of the local microphysical conditions. Furthermore, there are significant variations in the three-dimensional structure of the cloud which may not be adequately sampled by a single aircraft. This statement holds true for measurements of cloud absorption as well. As reviewed by Stephens and Tsay (1990) there appears to be a persistent anomaly between measurements and calculations of cloud absorptivity. Although it is unclear what is responsible for the discrepancy, it seems that inaccuracies in the microphysical measurements and imperfect sampling are likely to play an important role. Based on the results presented here, an evaluation of cloud absorption from our measurements and model computations was abandoned.
### TABLE 2. SUMMARY OF MICROPHYSICAL AND RADIATIVE OBSERVATIONS AND MODELLING FOR ALL CASE STUDIES

<table>
<thead>
<tr>
<th>Date</th>
<th>Liquid-water path (g m(^{-2}))</th>
<th>Optical depth</th>
<th>Sun angle (°)</th>
<th>Albedo</th>
<th>Liquid-water path (%)</th>
<th>Effective radius (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>11 July 1993</td>
<td>34</td>
<td>6.47</td>
<td>70</td>
<td>0.57</td>
<td>0.52</td>
<td>+30</td>
</tr>
<tr>
<td>16 July 1993(1)*</td>
<td>52</td>
<td>6.02</td>
<td>64</td>
<td>0.48</td>
<td>0.46</td>
<td>+10</td>
</tr>
<tr>
<td>16 July 1993(2)*</td>
<td>87</td>
<td>8.46</td>
<td>65</td>
<td>0.58</td>
<td>0.54</td>
<td>+25</td>
</tr>
<tr>
<td>16 July 1993(3)*</td>
<td>51</td>
<td>4.87</td>
<td>77</td>
<td>0.58</td>
<td>0.50</td>
<td>+60</td>
</tr>
<tr>
<td>19 July 1993</td>
<td>14</td>
<td>1.34</td>
<td>64</td>
<td>0.18</td>
<td>0.20</td>
<td>-15</td>
</tr>
<tr>
<td>20 July 1993(1)*</td>
<td>101</td>
<td>7.68</td>
<td>66</td>
<td>0.37</td>
<td>0.49</td>
<td>-50</td>
</tr>
<tr>
<td>20 July 1993(2)*</td>
<td>110</td>
<td>8.56</td>
<td>65</td>
<td>0.44</td>
<td>0.50</td>
<td>-25</td>
</tr>
</tbody>
</table>

The percentages recorded under the headings liquid-water path (%) and effective radius (%) refer to the amount by which the observed liquid-water path or effective radius need to be changed in order to calculate a cloud albedo that matched the observed albedo.

* The numbers (1) and (2) after the date refer to the first and second stack of the first case study of 16 and 20 July. (3) corresponds to the only stack on the second case study of 16 July.
(e) Drizzle and cloud albedo

The influence of drizzle on cloud optical depth and cloud albedo is a topic which has received considerable attention in the last few years. There appear to be at least three mechanisms associated with drizzle that can affect global cloud optical depth. The first one is that perturbations in CCN and drizzle affect fractional cloudiness (Albrecht 1989). The second is that drizzle can affect the CCN concentration by washout (Ackerman et al. 1993) which will invoke cloud collapse. The third one is that perturbations in CCN through perturbations in drizzle will affect cloud depth directly (Pincus and Baker 1994; Boers 1995).

In the data presented so far there is ample evidence of drizzle. It is clear from our results that the influence of drizzle on the microphysical and radiative properties is indeed very important. In this section, we will suggest that beyond the three mechanisms already mentioned there is probably a fourth mechanism by which drizzle can affect cloud optical depth. This mechanism is associated with modifications to the cloud droplet spectra due to drizzle formation.

The starting point of the arguments presented here is that the autoconversion and collision/coalescence process generally will modify the liquid-water path as well as the effective radius of the cloud droplets. Given Eq. (7), the evolution of the cloud optical depth in response to any perturbations (such as the collision/coalescence process) can be written as

$$\frac{d \tau}{\tau} = \frac{d \text{LWP}}{\text{LWP}} - \frac{dr_e}{r_e}.$$  \hspace{1cm} (8)

From an observational standpoint is is impossible to know the initial conditions for a cloud consisting entirely of non-drizzle droplets. Hence, the complete evolution of the cloud optical depth due to the initiation of the drizzle process can only be known from a detailed modelling effort. However, it is possible to estimate from observations how much of the liquid water is associated with the drizzle process. From this observation the question can then be asked: What would happen if all liquid-water associated with drizzle droplets would instead be associated with non-drizzle droplets? This question is equivalent to asking how the effective radius changes in response to the initiation of a drizzle process, under the condition that the liquid-water content remains fixed. In essence, this question can be reduced to solving Eq. (8) while neglecting the first term on the right-hand side.

The procedure to calculate the effect of drizzle on cloud optical depth from the observations during SOCEX will be demonstrated from one case study, namely the 20 July case, when the drizzle process was most intense. The computations from all case studies will then be summarized in Table 3. Figure 16 shows the average droplet spectra for both of the stacks from this case study, at three levels within the cloud. The graphs show that there is an increase in droplets towards the tail end of the spectrum for lower levels in the cloud. This is hardly surprising as larger droplets are formed by collection of smaller droplets as they fall down towards cloud base. The spectra can be used to compute the contribution of each spectral bin to the downward liquid-water flux associated with drizzle. To this end the liquid water is multiplied by the Stokes velocity for small droplets and an empirical formula for larger droplets, as in Brost et al. (1982b), and integrated over the full spectrum.

$$F_{\text{drizzle}}(r) = \frac{4}{3} \pi r^3 n(r) w$$  \hspace{1cm} (9)

where $F_{\text{drizzle}}(r)$ is the vertical liquid-water flux contribution due to drizzle in each spectral range, and $w$ is defined as

$$w = (1.19 \times 10^8 \text{ m}^{-1}\text{s}^{-1})r^2 \text{ for } r < 40 \mu\text{m}$$  \hspace{1cm} (10)
Figure 16. Mean droplet spectra for three levels inside the cloud on 20 July 1993. The average cloud is divided into three layers of equal pressure thickness. The vertical line indicates the separation between spectral intervals recorded by the FSSP and the 2D-C probe. See text for further explanation.

and

\[ w = (8 \times 10^3 \text{ s}^{-1})r \quad \text{for } r > 40 \mu\text{m}. \]  \hspace{1cm} (11)

Figure 17 shows the calculation of the second and third moment of the size distributions displayed in Fig. 16, and a calculation of \( F \) according to (9). The measurements and calculations are shown on a log–log scale to emphasize the FSSP in comparison with the 2D-C data, and because of the very large range of values of the parameters. In Fig. 17 the values of the second and third moment and the drizzle flux are multiplied by the value of \( r \) in each bin (i.e. in Fig. 17(a) we show \( r \times r^2 n(r) \) as a function of \( r \), so that the second moment is defined as \( \int_{0}^{\infty} r \times r^2 n(r) \text{d}(\ln(r)) \)). This multiplication facilitates a visual comparison between the contributions of the FSSP and the 2D-C probe. Figures 17(a) and (b) represent the contribution of each of the spectral intervals to the extinction and total water mass. There is clear evidence that the spectra are bimodal, a result also found for the case studies on 16 July. In particular, near cloud base (thick line, marked as low) the contribution to the liquid-water mass of the drizzle mode is very important. However, at the same time, the contribution of extinction (proportional to the second moment of the size distribution) towards total optical depth is small at these levels when compared with the mid and high region in the cloud. From this figure it can also be deduced that the contribution to the drizzle process is mainly due to the droplets measured by the 2D-C probe. This is particularly true for the middle and lower cloud levels, where the contribution of 2D-C measured liquid water to drizzle is such that it increases the drizzle flux by 1.5 and 3 orders of magnitude above what it would have been if only the FSSP measurements would have been taken into account. It is, therefore, not entirely unrealistic to make the assumption, as will be done now, that all drizzle droplets are associated with the 2D-C probe. Using this assumption, Fig. 12 can be used to calculate what the optical depth would have been for the case that liquid water would have been associated with the non-drizzle mode (i.e. associated with the FSSP spectral bins).
Figure 17. Mean contribution to (a) extinction from the data shown in Fig. 16, (b) contribution to the liquid-water mass, and (c) contribution to drizzle flux. The average cloud is divided into three layers of equal pressure thickness. The vertical line indicates the separation between spectral intervals recorded by the FSSP and the 2D-C probe. See text for further explanation.

In Fig. 18 the point A corresponds to the liquid-water path and optical depth associated with the 20 July case study, when only the FSSP data are considered. As shown earlier in Fig. 12, when the data from the 2D-C probe are included, the representation of the case study as point A is no longer correct. In that case the appropriate representation is point B (large increase in liquid-water path, small increase in cloud optical depth). The manner in which the spectral ranges contribute to the liquid-water paths and extinction are shown in the panels of Fig. 17. Clearly, the contribution to liquid-water path of drizzle droplets is quite large while for the extinction it is much smaller. The spectra are bimodal in the sense that there is a peak in the spectral range of the FSSP, but also one in the spectral range of the 2D-C probe.
Figure 18. Calculation of the effect of drizzle on cloud optical depth ($\tau$). Point A refers to data from the FSSP; point B refers to data from the FSSP and 2D-C probe; while point C refers to the liquid-water path based on the FSSP and 2D-C probe, but the effective radius based on the FSSP. The distance C-B indicates the optical-depth reduction due to drizzle conversion, assuming that the liquid-water path (LWP) does not change at the same time.

If it is now assumed that all liquid water associated with the drizzle process is converted back to small non-drizzle droplets, point B is translated to point C. This translation is equivalent to modifying the optical depth of the cloud while leaving the liquid-water path constant. Essentially, it modifies the cloud in such a way that the effective radius representative of the cloud is reduced to the value of the effective radius corresponding to the FSSP data only. In Figs. 17(a) and (b) this would mean the removal of the second (drizzle) mode and distributing all liquid water from this second mode back into the first (non-drizzle) mode. Consequently, a line can be drawn from point A to point C along a line of equal effective radius. There is some support for this assumption when Figs. 5 and 6 are re-examined. These figures show that despite the presence of drizzle droplets in the cloud samples the effective radius of the non-drizzle droplets is hardly affected.

Figure 18 shows that drizzle formation has a profound effect on the optical depth of clouds. For the case study shown the optical depth of over 12 (point C) for a cloud of non-drizzle droplets is reduced to about 8.2 (point B) when the collision/coalescence process modifies the spectrum and increases the effective radius. These are very large changes and highlight the importance of the drizzle process in modifying cloud albedo. The drizzle process demonstrated in Fig. 18 is only approximately correct as no account can be given of any changes in liquid-water path when the drizzle process is initiated. It is now possible to compute the fractional change in albedo for all case studies when clouds with properties of B are transformed to C (or vice versa). To this end the short-wave radiative-transfer model described earlier was employed. The results are tabulated in Table 3. The perturbation, $\delta \alpha$, in Table 3 is defined as

$$\delta \alpha = 100 \times \frac{\alpha_B - \alpha_C}{\alpha_B}. \quad (12)$$

The calculated perturbations in cloud albedo are very large, in particular for the case studies of 16 and 20 July. Not surprisingly, the 11 July case study with few drizzle droplets has the smallest perturbation.
TABLE 3. **Albedo Perturbation (see text) due to Drizzle**

<table>
<thead>
<tr>
<th>Date</th>
<th>Stack number</th>
<th>Perturbation δα (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>11 July 1993</td>
<td>1</td>
<td>-2</td>
</tr>
<tr>
<td>16 July 1993(1)*</td>
<td>1</td>
<td>-7</td>
</tr>
<tr>
<td>16 July 1993(2)*</td>
<td>2</td>
<td>-14</td>
</tr>
<tr>
<td>16 July 1993(3)*</td>
<td>1</td>
<td>-17</td>
</tr>
<tr>
<td>19 July 1993</td>
<td>1</td>
<td>-3</td>
</tr>
<tr>
<td>20 July 1993</td>
<td>1</td>
<td>-18</td>
</tr>
<tr>
<td>20 July 1993</td>
<td>2</td>
<td>-15</td>
</tr>
</tbody>
</table>

* The numbers (1) and (2) after the date refer to the first and second stack of the first case study of 16 July. (3) corresponds to the only stack on the second case study of 16 July.

The mechanism described here is reasonably realistic, in that it adequately accounts for the conversion of the radiative properties of a hypothetical cloud without drizzle droplets to an observed cloud with drizzle droplets. Clearly, the complete description of the physical mechanism including the depletion of liquid water awaits further study using high resolution modelling of the cloudy boundary layer with explicit microphysics. Furthermore, modelling studies of the optical properties of clouds with drizzle need to be undertaken.

5. **Discussion and Conclusions**

The data set reported on here represents the microphysical and radiative structure of the marine boundary layer devoid of anthropogenic influences. The Southern Ocean probably is the best ‘natural laboratory’ to obtain an inventory of the natural variability of background conditions.

The cloud droplet concentrations reported in this study are among the lowest reported in the literature; this is due to a general dearth of CCN in the sub-cloud atmosphere. There are at least three important consequences: (1) the cloud droplet effective radii recorded were among the highest ever reported; (2) cloud optical depth was quite small (5–10) despite the moderate depth of these clouds (200–300 m); and (3) drizzle formation impacts cloud reflectance.

By computing both the liquid-water path and the cloud optical depth from the microphysical probes, it is possible to arrive at a representation of each case study in a single point on the optical-depth/liquid-water-path plot. Thus, each case study has also a corresponding representative effective radius, computed by solving Eq. (5) for $r_c$. This representation facilitates comparison between the different case studies and furthermore elucidates several microphysical aspects of individual case studies.

One aspect pertained to the effect of drizzle droplets on the representative effective radius of the case studies. It was found that despite the fact that the actually observed effective radius of droplets near cloud base was several hundreds of microns, the representative effective radius for each case study only increased by several microns at the most. This result can be interpreted by considering that for retrievals of optical depth and effective radii from satellite observations, the cloud is considered to be homogeneous and consisting of uniform particles with the representative effective radius as its size parameter. Although near cloud base very large effective radii were observed, their radiative effect is not so large as to increase the representative effective radius to hundreds of microns.
The other aspect pertained to the influence of drizzle on cloud albedo. An approximate process was outlined whereby a hypothetical cloud consisting entirely of non-drizzle droplets is converted into an observed cloud consisting of drizzle and non-drizzle droplets alike. It was shown that this conversion accounts for a large reduction in cloud albedo. This important result highlights yet another potential impact of drizzle on clouds and climate, namely the reduction in projected cross-sectional area of the droplets when non-drizzle droplets are converted to drizzle droplets.

Liquid-water content was found to be only about 30 to 60% of the expected adiabatic value. The likely reasons were mixing events with overlying dry air, but also the generally low droplet number concentration, which invokes drizzle, which in turn inhibits a compensating upward flux of moisture due to a weak stabilization of the boundary layer.

One non-baseline case study was compared with the baseline case studies, from which it was clear that there were large differences between the two types of case studies, both in the size of the particles and in their ability to affect the radiative structure of the cloud.

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APPENDIX

Tilt corrections to the downward solar-flux measurements

After the aircraft (upward-pointing) radiometer data were analysed and corrected for roll, pitch and yaw, inconsistent results were found when compared with a two-stream model of solar flux impinging on the radiometers. Although the response of both radiometers is somewhat sensitive to the azimuth angle of the incoming direct radiation, the large inconsistencies could not be attributed to this effect. From the available data it became apparent that an unknown tilt angle of the radiometer, either in the radiometer itself, or through a small unknown angle of the pod holding the radiometer with the horizontal of the aircraft, contaminated the data.

The following procedure was applied to the data in order to correct for this tilt. First, flux-data portions from flight legs above cloud top were averaged, divided by the cosine of the solar zenith angle, and plotted as functions of the difference between the aircraft heading and solar azimuth angle. If the response of the instrument is independent of orientation, then to a rough first-order approximation this data should be constant with aircraft orientation. The results are shown in Fig. A.1. The resulting functional dependence of the flux with orientation can be well fitted by the sinusoidal curve as indicated by the solid line through the data. The phase of the sinusoid indicates the orientation of the tilt, while the amplitude indicates the tilt angle. Once the tilt angle is known, all data can be corrected by introducing an apparent roll and pitch into the original data stream and recalculating the fluxes. Figure A.2 shows the comparison of the corrected flux data with the results from the two-stream radiative-transfer model which was adopted for this work.
to include only the spectral region where the radiometer is sensitive. The results show excellent agreement.

Figure A.1. Observed cosine-corrected downward solar flux as a function of the difference between the sun and aircraft orientation.

Figure A.2. Corrected observed versus modelled downward solar flux.
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