The Pennsylvania State University

The Graduate School

Department of Meteorology

RADIATIVE IMPACT OF CONTINENTAL STRATUS AT THE SOUTHERN GREAT PLAINS:
A CLIMATOLOGICAL PERSPECTIVE

A Thesis in

Meteorology

by

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Submitted in Partial Fulfillment of the Requirements for the Degree of

Doctor of Philosophy

August 2002
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Abstract

A four-year climatology of the radiatively important parameters of warm boundary layer clouds, including cloud liquid water path, cloud base height and normalized cloud forcing, was developed for the Southern Great Plains, Department of Energy Atmospheric Radiation Measurement Program Cloud and Radiation Testbed, site. The climatology shows that cloud liquid water paths were approximately exponentially distributed, ranging primarily from just above 0 to 0.2 mm. Effective radii from one year of retrievals were normally distributed, varying primarily between 5.5 µm and 9.5 µm. A one-year observational study of overcast boundary layer stratus illustrated that surface radiation was primarily sensitive to cloud liquid water path with effective radius having a secondary influence. Radiative transfer calculations using the previously determined natural ranges of effective radius and liquid water path explained the observed sensitivities. Overall, there was a 79% correlation between observed and computed surface fluxes when using a fixed effective radius of 7.5 µm in the calculations. The use of one-dimensional radiative transfer was demonstrated to be appropriate for the study, as the independent column approximation held for the overcast cases.

A cloud liquid water path retrieval using infrared emission measurements was developed, as uncertainties in microwave radiometer retrieved liquid water paths were comparable to the mode of the observations. Independent verification of the results, which are unbiased on some days, but not others, with respect to the microwave radiometer derived values, is not straightforward and was not attempted in the study.
Ultimately, this retrieval is valuable for estimating liquid water paths of clouds with low emissivities, thereby acting as a complement to microwave radiometer retrieved liquid water paths.

A parameterization of normalized cloud forcing based on liquid water path and solar zenith angle was developed and tested against observations, yielding results comparable to the best set of explicit surface flux calculations. The parameterization provides a fast and accurate way of estimating either surface flux or liquid water path given the other. This parameterization is useful for research requiring solar energy estimates in the presence of clouds as well as evaluation of cloud-radiation interactions in weather prediction and climate models.

It is also concluded from this study that measurement of the indirect aerosol effect will be problematic at the site, as variations in cloud liquid water path will most likely mask effects of variations in particle size. Furthermore, the absence of any bias relative to observations in the computed surface broadband fluxes based on effective radii derived from narrowband (415 µm) transmission indicates the absence of anomalous absorption in the clouds analyzed in this study.
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Acknowledgments

This work is the result of contributions and sacrifices by many people. I will always remain indebted to Tom Ackerman for his immense contribution to my life and work. Without his support I could not have reached this point. I would like to thank the incredible Eugene Clothiaux for patiently guiding and encouraging me at every step of my scientific endeavours. A special thanks to my committee members Tim Kane, Dennis Lamb and Johannes Verlinde for providing me the use of their experience and expertise. My wife Nilanjana deserves special mention for the extreme sacrifices she made with a smile. I am also grateful to my daughter Pourna who, while struggling to survive at the dawn of her life, showed me how important it was keep fighting even when the chips were down. My parents and in-laws deserve thanks for continuously supporting me on my venture. Last but not least, I would like to mention my great friends spread over two continents who have helped me and my family in innumerable ways.
Chapter 1

Introduction

Clouds play a major role in the partitioning of energy between the upper and lower atmosphere, as well as the atmosphere and land and ocean, and they have a significant influence on the hydrological cycle of the Earth. The overall effect of clouds on the radiation budget is cooling in the shortwave, as a result of increased albedo, and warming in the infrared, from a reduced effective emitting temperature (Ramanathan et al., 1989). Analyzing observations, Ramanathan et al. (1989) and Ardanuy et al. (1991) demonstrated that shortwave cooling is dominant, though there is uncertainty in the magnitude of the overall cooling effect of clouds.

While the net cooling effect of clouds is valid for the current climate, it is difficult to assess the net cloud forcing in a different climate regime, such as may result from doubled atmospheric CO\textsubscript{2} levels. As the issue of global warming and its impact on society is a relevant problem, assessments of current climate change have been conducted using numerous general circulation models (GCMs), like CCM3 ((Kiehl et al., 1998)), the GISS climate model ((Somerville et al., 1974); (Hansen et al., 1974)) and the GFDL climate model ((Manabe and Wetherald, 1975); (Wetherald and Manabe, 1988)), with varying results. In an intercomparison of various models Cess et al. (1990) found that though most models show a reduction in the net cooling effect of clouds, classified as a positive feedback, some move in the other direction. These differences in model simulations are
a consequence of differences in the internal physical and dynamical processes embedded in the models, which in turn reflect our current knowledge of clouds.

The radiative impact of a cloud depends on many properties of the cloud, including spatial extent, phase, temperature, microphysical structure, and resulting optical properties, as well as the surrounding atmospheric conditions. Quantifying contributions of changes in each of these parameters to radiative feedback processes is a difficult problem. Therefore, assessments of the radiative impacts of cloud type (Hartmann et al., 1992; Poetzsch-Heffter et al., 1995); cloud amount (Ohring and Clapp, 1980; Somerville and Remer, 1974), moisture and temperature (Stephens and Webster, 1981) and microphysical properties (Zuidema and Hartmann, 1995)) are continuously being conducted in an effort to identify and quantify cloud feedback processes. Simultaneously, efforts to improve the internal physics of GCMs to match observations result in on-going cloud parameterization developments (Slingo, 1989; Stephens, 1978b; Stephens et al., 1984; Platt, 1997; Fu and Liou, 1993). While implementation into GCMs of new parameterizations based on observational cloud and radiation studies has resulted in a marked reduction in disagreements between models (Cess and Coauthors, 1996), there still remain a lot of uncertainties in GCM model physics which need to be resolved.

Of the different cloud types low-level water clouds, including stratocumulus and stratus, cover 34% of the ocean and 18% of the land at any given time (Considine et al., 1997). These clouds normally have a high albedo when compared to land and ocean surfaces with temperatures that are comparable to them. Consequently, low stratiform clouds provide about 60% of the annually averaged net cloud radiative forcing (Hartmann et al., 1992). We focus our study on this cloud type given its radiative importance
to climate. Two quantities that are used to quantify the radiative impact of these clouds are the cloud particle effective radius and the cloud liquid water path ((Stephens, 1978a); (Hu and Stamnes, 1993)). Numerous studies ((Charlson et al., 1987); (Han et al., 1994); (Ramaswamy and Chen, 1993); (Chen and Ramaswamy, 1996); (2001)) have tried to assess the impact of individual variations in these parameters on radiative forcing and climate feedback. Most studies of this type are satellite-based and assessment of simultaneous variations in the two parameters and the capacity of one to offset changes in the other within natural limits is difficult to gauge with data having coarse spatial resolution.

Over the years numerous in-situ measurements of stratus microphysics have been made using various probes fitted to research aircraft ((Albrecht et al., 1995); (Duda et al., 1991); (Martin et al., 1994); (Noonkester, 1984)). While the purpose of most measurements was to study the growth and distribution of water droplets in warm clouds, the radiative impacts of the clouds were also observed in many cases. Even though in-situ aircraft measurements are complementary to satellite data, the expense of such studies results in small temporal samples that preclude climatological assessments.

Surface-based remote sensors, unlike airborne instruments, are capable of providing continuous data at a smaller cost. Using this capability of remote sensors, the Department of Energy (DOE) Atmospheric Radiation Measurements (ARM) Program set up a network of such instruments to gather long term datasets that would ultimately be useful in solving problems related to clouds and climate. The biggest and most intensively instrumented measurement site in the network is the ARM Clouds and Radiation Testbed (CART) site in the Southern Great Plains (SGP) of Oklahoma. With a variety
of measurements being available from this site for multiple years climatological studies on clouds and radiation can now be performed.

As we are interested in studying the radiative properties of boundary layer clouds at a climatological level, we must build a multi-year database of the parameters used to describe the radiative properties of warm boundary layer clouds. To quantify the radiative properties of boundary layer cloud we require information regarding a minimum of 3 parameters: liquid water path (amount of water in a column containing the cloud), effective radius (the ratio of the third to the second moment of the cloud drop radii) and distribution width (the standard deviation of the cloud drop size distribution). While the clouds are detectable using radars and lidars, or a combination of both, inferring the size distributions of cloud droplets is a complex task. Possible methods involve the use of radar reflectivities (Kato et al., 2001) and surface radiation measurements (Min and Harrison, 1996) with the column liquid water path retrieved from a microwave radiometer (Liljegren et al., 2001) as a constraint.

Ackerman et al. (1999) developed a paradigm for assessing the accuracy of parameterizations and retrievals of cloud microphysics from a radiative transfer perspective. The paradigm involves calculating surface solar fluxes using cloud microphysics from different sources and comparing the results with actual radiation measurements. We first adapt the paradigm to assess retrievals of boundary layer cloud liquid water paths and cloud drop effective radii. We subsequently assume the cloud liquid water path retrievals to be accurate and evaluate the particle effective radius retrievals against a climatological baseline value appropriate for stratus at the ARM SGP CART site. We also assess the sensitivity of surface flux to each of the three parameters that we used to describe the
radiative properties of boundary layer clouds. As we use a large dataset, the results can be treated as robust and representative of the particular cloud type.

Twomey (1977), Twomey et al. (1984) and Twomey (1991) postulated that changes in the microphysical properties of stratiform cloud due to an increase in anthropogenic aerosols and cloud condensation nuclei (CCN) will result in increased cloud albedo, thereby ameliorating the warming effects of greenhouse gases. This indirect aerosol effect is still uncertain and studies like Brenguier et al. (2000) attempt to gather evidence regarding this effect. While Platnick and Twomey (1994) argue that marine stratus clouds are more susceptible, Charlock and Sellers (1980) have surmized that continental stratus would also be prone to this indirect aerosol effect. Our assessment of the sensitivity of surface flux to the three parameters that we used to describe the radiative properties of warm boundary layer clouds also provides an insight into the significance of the indirect aerosol effect.

We do not explicitly consider horizontal inhomogeneities in our cloud descriptions even though it can be important to boundary layer cloud radiative properties (Cahalan et al., 1994a). By ignoring horizontal inhomogeneities, we have implicitly assumed that horizontal photon transport, while leading to errors in our instantaneous flux calculations, does not lead to significant biases over averages of many temporally consecutive flux calculations (Cahalan et al., 1994b). Therefore, radiative transfer calculations in a cloud column, or at a specific time using retrieved cloud liquid water path, can be treated as independent of surrounding columns, or times, with meaningful comparisons to observations being made only over averages of the results. This independent-column
approximation is tested for applicability to time series of cloud liquid water path, for both broken and overcast boundary-layer clouds.

The availability of long term data at high temporal resolution also makes it possible to analyze and develop statistically meaningful relationships between the observed parameters. As radiative transfer calculations are time consuming, a parameterization of radiation in the presence of boundary layer cloud is a useful tool for computational purposes. Furthermore, simple, but robust, relationships between surface flux and cloud parameters may be used as a baseline for evaluating potential new boundary layer cloud radiative parameterizations. Additionally, some types of research require quick estimates of surface radiation in the presence of clouds. For all of these reasons, as well as a succinct way of summarizing our results, we develop a simple parameterization of boundary layer cloud radiative properties as a function of solar zenith angle and cloud liquid water path using the observations in our study and we evaluate it against explicit radiative transfer calculations.

The values of geophysical parameters obtained from remote sensing data are influenced by calibration errors, instrument noise and retrieval errors. Del Genio and Wolf (2000) point out that there is an apparent positive bias in the SGP ARM site cloud liquid water path retrievals using traditional statistical retrievals based on monthly-mean retrieval coefficients. The liquid water paths used in our study are from a new retrieval algorithm that attempts to eliminate such errors. Nonetheless, these errors may still be present, so we investigate an alternative method for retrieving cloud liquid water path that uses infrared radiances. This alternative retrieval is applicable for low liquid water path clouds where the errors in the microwave radiometer cloud liquid water
path retrievals can be particularly severe. Hence, the infrared radiance and microwave radiometer retrievals of cloud liquid water path are potentially complementary and we investigate in the appendix the performance of the two retrievals on low liquid water path clouds.
Chapter 2

Methodology

The main focus of the current study is development of a climatology of the radiative properties of boundary layer clouds and the use of these radiative properties to assess retrievals of boundary layer cloud microphysics. Both aspects of this study depend upon accurate radiative transfer calculations through low-altitude liquid water clouds, which in turn depend upon accurate model treatments of absorption and scattering and accurate measurements of the atmospheric state. In the first part of this chapter we describe the radiative transfer model and its required input parameters, while in the second part we discuss the measurements of the atmospheric state, their availability, and their limitations.

2.1 The Radiative Transfer Model

The radiative transfer model that we used in this study is based on the $\delta-2$ stream numerical algorithm presented by Toon et al. (1989) and Kato et al. (1999a), which bears the acronym RAPRAD (Rapid Radiative Transfer) to signify its speed of computation. The model has 32 spectral intervals ranging from 0.24 $\mu$m to 4.6 $\mu$m in the shortwave and near-infrared, using absorption coefficients based on $k$ distributions and a correlated-$k$ approximation as explained in detail by Kato et al. (1999b). (See Table 2.1 for a detailed listing of the wavelength range for each of the 32 spectral bands.) The top
of atmosphere spectral solar irradiance in the RAPRAD model is based on the solar irradiance of MODTRAN3 (Berk et al., 1989). The RAPRAD model incorporates ozone, oxygen, carbon dioxide and water vapor absorption, as well as water vapor continuum absorption. The molecular scattering optical depth is computed using the Rayleigh optical depth calculation as shown in Hansen and Travis (1974).

The atmospheric layers in the RAPRAD model are arbitrary. We set the top of atmosphere at 70 km and divided the atmosphere below 16 km into 250 m thick layers. Above 16 km model layer thicknesses increased with altitude. Clouds can be represented by additional higher resolution layers, if necessary, in order to handle specified in-cloud variation of properties. Clouds within a layer are then modeled as homogeneous, with cloud properties allowed to vary from one layer to the next. In the model simulations we used a surface albedo of 0.2 that was invariant with wavelength; this value is typical of the surface albedo at the DOE ARM SGP CART site in the mid-visible.

Clear-sky RAPRAD irradiance calculations require vertical profiles of pressure, temperature, water vapor, ozone, and aerosol particles. When a cloud is inserted into the RAPRAD model, cloud optical depth is a required input, together with the asymmetry parameter and single scattering albedo of the cloud particles for each model wavelength interval. For RAPRAD model inputs based on DOE ARM SGP CART site measurements, calculated surface fluxes are generally within 15-20 W m$^{-2}$ of surface flux measurements (Kato et al., 1997). The RAPRAD model is a one-dimensional model and can treat only plane-parallel, homogenous clouds.
Table 2.1. Wavelengths used for radiative transfer calculations and corresponding complex refractive indices of water.

<table>
<thead>
<tr>
<th>Sl. No.</th>
<th>Wavelength midpoint $\mu m$</th>
<th>Wavelength range $\mu m$</th>
<th>Refractive Index</th>
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<td>$0.13534104e+01$</td>
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<tr>
<td>3</td>
<td>$2.951274e-01$</td>
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<td>$0.13499745e+01$</td>
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<tr>
<td>4</td>
<td>$3.173065e-01$</td>
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<td>$0.13469232e+01$</td>
</tr>
<tr>
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<td>$3.451361e-01$</td>
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<tr>
<td>6</td>
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<tr>
<td>7</td>
<td>$4.297729e-01$</td>
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</tr>
<tr>
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<tr>
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<tr>
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</tr>
<tr>
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</tr>
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<tr>
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</tr>
<tr>
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<td>$1.045744e+00 - 1.194188e+00$</td>
<td>$0.13270000e+01$</td>
</tr>
<tr>
<td>24</td>
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</tr>
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<td>$0.12901646e+01$</td>
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<td>$3.001893e+00 - 3.635417e+00$</td>
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<tr>
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<td>$3.635417e+00 - 3.991003e+00$</td>
<td>$0.13614935e+01$</td>
</tr>
<tr>
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<td>$3.991003e+00 - 4.605654e+00$</td>
<td>$0.13340000e+01$</td>
</tr>
</tbody>
</table>
2.2 Measurements of Atmospheric State at the SGP CART Site

As outlined in the previous section, the RAPRAD model requires an atmospheric thermodynamic profile, aerosol optical depth, and cloud location and composition information as inputs. The DOE ARM program provides long-term, continuous measurements of these quantities at its SGP CART site central facility situated in Lamont, Oklahoma (36.605°N, 97.485°W). In Table 2.2 we match the required inputs of the RAPRAD model with the instruments at the SGP CART site that are providing measurements of them.

We make use of data from 1997 through 2000 in our four-year study of the microphysical and radiative properties of boundary layer clouds at the SGP CART site. The radar-derived and surface radiation-derived cloud drop effective radii, however, are available only for a one-year period extending from 1997 through 1998. We now describe the measurements and retrievals in more detail.

2.2.1 Atmospheric Thermodynamic Profiles

An accurate vertical profile of the atmospheric thermodynamic state, especially of the vertical distribution of water vapor, is important for a reliable computation of broadband irradiance at the surface. At the SGP CART site multiple sets of measurements are used to construct atmospheric profiles with the highest possible temporal resolution (Gerald G. Mace, personal communication). Radiosondes are launched from the site approximately five times a day and they are used to scale vertically integrated measurements of water vapor retrieved either from a microwave radiometer (Liljegren, 1994) or
Table 2.2. Summary of atmospheric variables, their source of measurement and the resolution at which the data are available. The acronyms are listed in the next table.

(†)Rapid Update Cycle (RUC) Weather Forecast Model from National Oceanic and Atmospheric Administration (NOAA) Forecast System Laboratory

(§) Atmospheric Remote Sensing of Clouds

<table>
<thead>
<tr>
<th>Variables</th>
<th>Source</th>
<th>Time Resolution</th>
</tr>
</thead>
<tbody>
<tr>
<td>cloud boundaries</td>
<td>MMCR, MPL, BLC (ARSCL§ data)</td>
<td>10 seconds</td>
</tr>
<tr>
<td>liquid water path</td>
<td>MWR</td>
<td>20 seconds</td>
</tr>
<tr>
<td>water vapor column</td>
<td>MWR</td>
<td>20 seconds</td>
</tr>
<tr>
<td>temperature profile</td>
<td>BBSS, AERI RUC†</td>
<td>1 minute</td>
</tr>
<tr>
<td>pressure profile</td>
<td>MWR , SMOS</td>
<td></td>
</tr>
<tr>
<td>mixing ratio profile</td>
<td></td>
<td></td>
</tr>
<tr>
<td>aerosol optical depth</td>
<td>MFRSR, TOMS</td>
<td>30 minutes</td>
</tr>
<tr>
<td>effective radius</td>
<td>MMCR, MWR</td>
<td>20 seconds</td>
</tr>
<tr>
<td>(Kato et al., 2001)</td>
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<td></td>
</tr>
<tr>
<td>effective radius</td>
<td>MFRSR MWR</td>
<td>5 minutes</td>
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<tr>
<td>(Min and Harrison, 1996)</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
Table 2.3. List of instruments or suite of instruments at the SGP ARM CART site used for our study.

<table>
<thead>
<tr>
<th>Instrument/Suite</th>
<th>General Type</th>
<th>Acronym</th>
</tr>
</thead>
<tbody>
<tr>
<td>Millimeter Wave Cloud Radar</td>
<td>radar</td>
<td>MMCR</td>
</tr>
<tr>
<td>Micropulse Lidar</td>
<td>lidar</td>
<td>MPL</td>
</tr>
<tr>
<td>Belfort Ceilometer</td>
<td>ceilometer</td>
<td>BLC</td>
</tr>
<tr>
<td>Microwave Radiometer</td>
<td>radiometer</td>
<td>MWR</td>
</tr>
<tr>
<td>Balloon Borne Sounding System</td>
<td>radiosonde</td>
<td>BBSS</td>
</tr>
<tr>
<td>Atmospheric Emitted Infrared Interferometer</td>
<td>thermal interferometer</td>
<td>AERI</td>
</tr>
<tr>
<td>Surface Meteorological Observation System</td>
<td>thermometer barometer hygrometer anemometer rain gauge snow depth sensor</td>
<td>SMOS</td>
</tr>
<tr>
<td>Solar Infrared Radiation System</td>
<td>pyrheliometer shaded pyranometer unshaded pyranometer shaded pyrgeometer</td>
<td>SIRS</td>
</tr>
<tr>
<td>Multi-filter Rotating Shadowband Radiometer</td>
<td>radiometer</td>
<td>MFRSR</td>
</tr>
</tbody>
</table>
a combination of interferometer and GOES satellite radiance measurements across the thermal infrared (Schmit et al., 2002). For those periods when ground-based observations are not available thermodynamic state profiles from numerical weather prediction models forced by observations are used. Surface meteorological stations provide high temporal resolution measurements of surface pressure, temperature, and moisture.

2.2.2 Aerosol Optical Depth

The Multi-Filter Rotating Shadowband Radiometer (MFRSR) (Harrison et al., 1994) provides continuous measurements of direct normal solar irradiance across six wavelength intervals, five of which can be used to retrieve the corresponding atmospheric extinction optical depth (Michalsky et al., 2001). Aerosol optical depth for each wavelength interval is estimated by subtracting Rayleigh scattering and ozone absorption optical depths from the total extinction optical depth. In the approach of Michalsky et al. (2001) Rayleigh optical depth is estimated after Hansen and Travis (1974), i.e,

\[
\tau = 0.008569\lambda(1 + 0.0113\lambda + 0.00013\lambda)(p/p_\circ),
\]

where wavelength \(\lambda\) is in microns and the pressure correction is the ratio of surface pressure \(p\) to sea-level pressure \(p_\circ\). The ozone absorption coefficient tables of Shettle and Anderson (1995) for visible and infrared wavelengths, along with column measurements of ozone, are used to estimate the ozone absorption optical depth for the relevant MFRSR wavelengths. The Total Ozone Mapping Spectrometer (TOMS) aboard the Earth Probe satellite (McPeters et al., 1998) is one global source of ozone column data. The aerosol
optical depths retrieved at the five wavelengths are subsequently used to retrieve the parameters of the Angstrom relationship (Angstrom, 1929)

\[ \tau(\lambda) = \beta \lambda^{-\sigma}, \]  

(2.2)

where \( \beta \) and \( \sigma \) are constants determined by a linear least squares fit in natural logarithm coordinates.

The asymmetry parameter and single scattering albedo of aerosol particles are computed using the Toon and Ackerman (1981) Mie code, which requires aerosol particles to be treated as spheres of known size with known refractive indices as a function of wavelength. At the SGP CART site the major aerosol constituent is mineral dust (Kato et al., 1997). The refractive indices of mineral dust are shown in Table 2.4), which we assume to have a mean radius of 0.58 \( \mu \text{m} \) with a standard deviation of 1.35.

### 2.2.3 Cloud Boundaries and Height Assignments

Identifying the vertical distribution of cloud hydrometeors is accomplished from multiple active remote sensors consisting of a ceilometer, a micropulse lidar and a millimeter-wavelength cloud radar. The ceilometer and lidar are more sensitive to large concentrations of small particles, as opposed to small concentrations of large particles, compared to the radar, thereby providing better estimates of cloud base height. The radar beam is not attenuated as much as the ceilometer and lidar beams when propagating through cloud. Consequently, cloud top height estimates from radar are generally more dependable in those cases when the lidar beam suffers severe attenuation.
Table 2.4. Wavelengths used for radiative transfer calculations and corresponding complex refractive indices of mineral dust which typically is the main constituent of the aerosols at the SGP site.

<table>
<thead>
<tr>
<th>Sl. No.</th>
<th>Wavelength midpoint $\mu m$</th>
<th>Wavelength range $\mu m$</th>
<th>Refractive Index</th>
</tr>
</thead>
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<td></td>
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<td>2.724815e−01 − 2.834140e−01</td>
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<tr>
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<td>3.277722e−01 − 3.625000e−01</td>
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<td>3.625000e−01 − 4.075000e−01</td>
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</tr>
<tr>
<td>16</td>
<td>6.943129e−01</td>
<td>6.841772e−01 − 7.044486e−01</td>
<td>0.1530e+01 0.31233380e−07</td>
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<tr>
<td>17</td>
<td>7.235312e−01</td>
<td>7.044486e−01 − 7.426139e−01</td>
<td>0.1530e+01 0.75018065e−07</td>
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<tr>
<td>18</td>
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<td>7.426139e−01 − 7.914788e−01</td>
<td>0.1530e+01 0.15587445e−06</td>
</tr>
<tr>
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<td>8.444581e−01 − 8.889693e−01</td>
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<td>3.635417e+00 − 3.991003e+00</td>
<td>0.1267e+01 0.33975327e−02</td>
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<tr>
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<td>3.991003e+00 − 4.605654e+00</td>
<td>0.1260e+01 0.84386226e−02</td>
</tr>
</tbody>
</table>
Millimeter-wave cloud radar sensitivity is often not sufficient to detect small cloud particles at high altitudes and the most accurate retrievals of cloud top height occur for optically thin clouds through which the micropulse lidar beam can penetrate.

Clothiaux et al. (2000) provide a detailed description of a method for combining the active remote sensor data that yields an estimate of the vertical distribution of cloud particles. A cloud, binary mask is generated from these data such that each altitude bin containing a significant hydrometeor detection is labelled with a 1 and all other altitude bins are labelled with a 0. With the cloud mask in hand identification of cloud layers and their boundary heights is straightforward.

<table>
<thead>
<tr>
<th>Number</th>
<th>Type of cloud</th>
<th>Temperature (K) based classification</th>
</tr>
</thead>
<tbody>
<tr>
<td>(1)</td>
<td>ice clouds (cirrus)</td>
<td>cloud base $\leq 233$</td>
</tr>
<tr>
<td>(2)</td>
<td>warm clouds (stratus clouds)</td>
<td>cloud base $\geq 273$ &amp; cloud top $\geq 253$</td>
</tr>
<tr>
<td>(3)</td>
<td>mixed phase clouds</td>
<td>neither (1) nor (2)</td>
</tr>
</tbody>
</table>

Our cloud classification methodology uses the cloud layer boundary height estimates and atmospheric temperature profiles. We implement simple classification criteria based only on temperature at the boundaries of the cloud layers (Table 2.5).
2.2.4 Cloud Liquid Water Path

Cloud liquid water path is the amount of condensed water in a vertical column of the atmosphere. When we consider the effects of boundary layer clouds on solar radiation, changes in cloud liquid water path are the dominant cause of variations in the flux reaching the surface. Therefore, accurate cloud liquid water path measurements are essential for arriving at reasonable model estimates of observed surface solar irradiance. Within the DOE ARM program, cloud liquid water paths are retrieved using a dual channel microwave radiometer (Liljegren, 1994).

The original site-specific cloud liquid water path retrievals (Westwater, 1993) used site-dependent, monthly-mean atmospheric radiating temperatures and water absorption coefficients at the two radiometer wavelengths of 23.8 GHz and 31.4 GHz. Deviations from the mean radiating temperature and water absorption coefficients, as well as temperature variations in the cloud liquid water, caused significant errors in this retrieval technique. Liljegren et al. (2001) developed a new statistical retrieval that attempts to estimate the atmospheric state at each retrieval time step in order to reduce retrieval error. A complicating factor of this new retrieval is that it requires surface meteorology, radiosonde, and millimeter-wave cloud radar measurements as inputs. Nonetheless, cloud liquid water path retrievals from this new approach are used in the current study.
2.3 Quantifying the Microphysical and Radiative Properties of Warm Clouds

Underlying all of the important microphysical and radiative properties of warm clouds is the warm cloud drop size distribution \( n(r, z) \), where \( n(r, z) \) is the density of cloud drops as a function of drop radius \( r \) and vertical height \( z \) in the atmosphere. Early research demonstrated that warm cloud drop size distributions are expected to be unimodal with a potentially long tail at larger sizes (Pruppacher and Klett, 1997). For this reason cloud drop size distributions used in remote sensing retrievals and recorded by in situ probes are generally parameterized using either a gamma or lognormal distribution, two distributions with long tails at larger particle sizes (Flatau et al., 1989). Following Miles et al. (2000), we use a lognormal distribution in this study. Below, we present the form of the lognormal distribution that we use and characterize several important moments of the drop size distribution in terms of lognormal distribution parameters.

2.3.1 The Lognormal Distribution

The form of the lognormal distribution that we use to parameterize \( n(r, z) \) is

\[
n_{\log}(r, z) = \frac{N_t(z)}{\sqrt{2\pi}\sigma_{\log}(z)} r \exp \left\{ \frac{-[\ln(r/r_{n,\log}(z))]}{2\sigma_{\log}^2(z)} \right\},
\]

(2.3)

where \( N_t(z) \) is the total number of cloud particles per unit volume, \( r_{n,\log}(z) \) is the median radius of the distribution, and \( \sigma_{\log}(z) \) is the logarithmic width of the distribution (i.e., the logarithm of the standard deviation of the drop size distribution). Note that the lognormal distribution has three free parameters: \( N_t(z), r_{n,\log}(z) \) and \( \sigma_{\log}(z) \). Since the
cloud liquid water path and the average size of the cloud drops throughout the extent of the cloud layer are of primary importance to cloud radiative effects, we neglect in this study the dependence of the three parameters on height $z$.

### 2.3.2 Cloud Liquid Water Path

Mathematically, the cloud liquid water path of a boundary layer cloud, i.e., the amount of condensed water in a vertical column through the cloud, can be expressed as

$$\text{LWP} = \frac{4}{3} \pi \int_{z_b}^{z_t} \int_{0}^{\infty} n(r, z) r^3 \rho_w dr dz,$$  \hspace{1cm} (2.4)

where $z_b$ and $z_t$ are the bottom and top heights of the cloud layer boundaries, respectively, and $\rho_w$ is the density of water. Substituting Equation 2.3 into Equation 2.4, we obtain an expression of the cloud liquid water content in terms of the parameters of the lognormal distribution:

$$\text{LWC}_{\text{log}} = \rho_w N_t \left(\frac{4\pi}{3}\right) r_{n,\text{log}}^3 \exp\left(\frac{9}{2} \sigma_{\text{log}}^2\right),$$  \hspace{1cm} (2.5)

where we assumed that the parameters are independent of height.

### 2.3.3 Cloud Particle Effective Radius

Early studies by Hansen and Travis (1974), Stephens et al. (1978) and Hu and Stamnes (1993) demonstrated that the radiative effects of warm clouds are well-characterized by the cloud liquid water path and the cloud particle effective radius. The effective radius $r_e$ is defined as the ratio of the third to second moments of the particle size distribution
and is given mathematically by

\[ r_e(z) = \frac{\int_0^\infty r^3 n(r, z) dr}{\int_0^\infty r^2 n(r, z) dr}. \]  (2.6)

In terms of the three parameters of the lognormal distribution the expression for the height-independent effective radius becomes

\[ r_{e,\log} = r_{n,\log} \exp\left(\frac{9}{2} \sigma_{\log}^2\right), \]  (2.7)

while the mean radius is given by

\[ r_{m,\log} = r_{n,\log} \exp\left(\frac{1}{2} \sigma_{\log}^2\right). \]  (2.8)

2.3.4 Cloud Optical Depth

The optical depth of a cloud at a particular wavelength is defined as

\[ \tau(\lambda) = \int_{z_t}^{z_b} \int_0^\infty n(r, z) \alpha_{\text{ext}}(r, \lambda) dr dz, \]  (2.9)

where \( \alpha_{\text{ext}}(r, \lambda) \) is the extinction cross section of the cloud drops and is a function of drop radius \( r \) and wavelength \( \lambda \). If the clouds are treated as vertically homogeneous, the expression for the optical depth becomes

\[ \tau(\lambda) = \sigma_{\text{ext}}(\lambda) N_t \Delta z, \]  (2.10)
where $\sigma_{\text{ext}}(\lambda)$ is the extinction cross section $\alpha_{\text{ext}}(r, \lambda)$ integrated across the drop size distribution. $N_t$ is the number of cloud drops per unit volume, and $\Delta z$ is the thickness of the cloud.

In terms of the extinction efficiency

$$q_{\text{ext}}(r, \lambda) = \frac{\alpha_{\text{ext}}(r, \lambda)}{\pi r^2}$$ (2.11)

and the cloud liquid water path LWP, the cloud optical depth becomes

$$\tau(\lambda) = \frac{3Q_{\text{ext}}(\lambda)\text{LWP}}{4\rho_w r_e}$$ (2.12)

where $Q_{\text{ext}}(\lambda)$ is the extinction efficiency $q_{\text{ext}}(r, \lambda)$ integrated over the drop size distribution. Since $Q_{\text{ext}}(\lambda)$ asymptotes to 2 as particle sizes become much larger than the wavelength of the incident radiation and shortwave radiation generally contains wavelengths much smaller than the radii of cloud drops, we set $Q_{\text{ext}}(\lambda)$ equal to 2 in Equation 2.12 to obtain

$$\tau = \frac{3 \text{LWP}}{2\rho_w r_e}$$ (2.13)

### 2.4 Retrieving the Microphysical Properties of Clouds

Based on numerous in situ observations of warm clouds available in the literature, Miles et al. (2000) reported climatological values of 5.4 $\mu$m and 0.38 for the effective radius and logarithmic width, respectively, of continental boundary layer clouds. We can, and will, use these values, along with measured cloud liquid water path, in our modeling
studies of surface broadband downwelling irradiance. Comparing computed and observed
downwelling surface fluxes, we will be able to adjust the effective radius to reduce any
bias that occurs in the comparisons, thereby providing a test of the appropriateness
of the Miles et al. (2000) value for boundary layer clouds over the SGP CART site.
We can subsequently use these climatological estimates of cloud drop effective radius to
assess the performance of boundary layer cloud microphysical retrievals that operate on
the data streams available from the ARM SGP CART site. We now describe the two
retrievals that we will investigate in the current study.

2.4.1 Retrieval Based on Radar and Microwave Radiometer Measurements

Kato et al. (2001) use radar reflectivity and Doppler velocity from the millimeter-
wavelength cloud radar, as well as cloud liquid water path estimated from a microwave
radiometer, to infer the size distribution of boundary layer cloud particles. As radar
reflectivity is proportional to the sixth moment of the drop size distribution, while liquid
water content is related to the third moment, the two can be related to each other as
shown by Frisch et al. (1998). Consequently, radar reflectivity and cloud liquid water path
can be used to retrieve cloud liquid water content LWC for each radar sample volume.
Kato et al. (2001) modify this approach so that the liquid water content distribution is
weighted by a power of $r_{n,\log}$. Variations in Doppler velocity are used as a measure of in-
cloud variations in vertical velocity to estimate $r_{n,\log}$ from the statistical model developed
by Considine and Curry (1996). The cloud drop effective radius $r_e$ is then calculated
using a relation between radar reflectivity, $r_{n,\log}$ and LWC. Note that the value of
effective radius so derived is independent of any measurement of surface radiation.
2.4.2 Retrieval Based on Narrowband Transmittance and Microwave Radiometer Measurements

In addition to direct normal measurements, the MFRSR makes measurements of narrowband total and diffuse hemispheric downwelling surface irradiance at the six wavelengths. Min and Harrison (1996) use these measurements to estimate stratus cloud optical depth. In the method developed by Min and Harrison (1996) they use a Langley regression of the direct normal irradiance to estimate the top of atmosphere instrument response for the 415 nm band. This calibration is then applied to the total hemispheric downwelling surface irradiance at this wavelength, leading to estimates of atmospheric transmittance for cloudy conditions.

Min and Harrison (1996) use the 415 nm passband (nominal midpoint of an approximately 10 nm band) of the MFRSR in preference to other bands as the shorter wavelength results in a greater Rayleigh scattering contribution and greater accuracy. For example, the surface albedo, which they estimate from ratios of direct to diffuse irradiances, together with assumptions about the aerosol present, on a clear-sky day, is low at this wavelength, minimizing errors in estimates of the surface albedo. Furthermore, this band is minimally affected by the ozone Chappuis-band and the asymmetry parameter and single scattering albedo, which are obtained from the technique of Hu and Stamnes (1993), are less sensitive to changes in effective radius at this wavelength. They use measurements from the 862 nm band of the MFRSR in conjunction with the 415 nm band to ensure detection of clouds.
Min and Harrison (1996) consider periods of boundary layer clouds when the direct beams are totally attenuated in both bands as potential cloud cases for the retrieval. These conditions ensure greater accuracy in determination of optical depth as the retrieval treats all transmitted radiation as diffuse. The retrieval assumes that cloud optical depth is constant over either a five or thirty-minute interval and estimates cloud optical depth by minimizing the squares of the errors in the one-minute transmittances under varying solar zenith angles. Finally, they compute the effective radius using Equation 2.13 and measurements of liquid water path provided by the microwave radiometer.

Note that the Min and Harrison (1996) technique retrieves a cloud optical depth and subsequently computes a cloud drop effective radius that is constrained to provide the correct optical depth given an estimate of the liquid water path. Therefore, we expect modeled values of broadband downwelling surface irradiance based on the Min and Harrison (1996) cloud drop effective radius and cloud liquid water paths retrieved from microwave radiometer measurements to produce close agreement with measurements.

2.5 Comparing Modeled and Measured Surface Solar Irradiance

To assess the reasonableness of retrieved cloud liquid water path and cloud drop effective radius we use them to compute downwelling broadband surface irradiances and we compare these irradiances with observations. A pyrheliometer provides observations of direct normal downwelling shortwave irradiance, while unshaded and shaded pyranometers measure the total and diffuse downwelling shortwave irradiances, respectively. In order to remove unknown and constant biases in the observations (e.g., (Kato et al.,
26

(1997)), as well as solar zenith angle effects in the comparisons, we opt to compare normalized cloud forcings instead of irradiances directly.

### 2.5.1 Normalized Cloud Forcing

We define normalized cloud forcing as

\[
NCF_c = \frac{CF(F_{c,\text{cld}} - F_{c,\text{clr}})}{F_{c,\text{clr}}},
\]

(2.14)

where \(F_{c,\text{cld}}\) is the computed downwelling surface shortwave irradiance during a cloudy period and \(F_{c,\text{clr}}\) is what the computed downwelling surface shortwave irradiance would be if the sky were actually clear or free of clouds. The cloud fraction \(CF\) weights the computed normalized cloud forcing in order to account for broken cloud sky conditions. In the case of observations the normalized cloud forcing is simply

\[
NCF_o = \frac{F_{o,\text{cld}} - F_{o,\text{clr}}}{F_{o,\text{clr}}},
\]

(2.15)

where we have replaced the “c” with an “o” and removed the explicit cloud fraction \(CF\) since the observed irradiance \(F_{o,\text{cld}}\) implicitly accounts for all sky conditions. As cloudy sky flux is generally less than clear sky flux, apart from exceptional broken cloud sky cover cases, the normalized cloud forcing generally lies between -1 and 0. Observed cases with values larger than 0 are omitted from the data set as modelling such cases using a plane-parallel radiative transfer model is impossible.

Computing modelled normalized cloud forcing is straightforward, as producing estimates of \(F_{c,\text{clr}}\) simply requires a radiative transfer calculation without any model
clouds in it. However, computing observed normalized cloud forcing is more difficult, as one needs an estimate of the clear-sky irradiance $F_{o,clr}$ that is based on observations during the same cloudy sky period. Long (2000) developed one method for deriving the clear-sky irradiance $F_{o,clr}$ using an empirical fit of the form

$$Y = a\mu^b,$$  \hspace{1cm} (2.16)

where $Y$ is the clear-sky total downwelling shortwave irradiance, $\mu$ the cosine of the solar zenith angle and $a$ and $b$ are regression coefficients. The regression coefficients in Equation 2.16 are obtained from a robust least squares estimation procedure that iteratively removes outliers by calculating absolute deviations about the median. In his approach Long (2000) applies the empirical fit to the total downwelling shortwave irradiances of all the clear-sky periods of each day, arriving at estimates of $a$ and $b$ for the day. Days with no clear-sky periods fail to yield estimates of $a$ and $b$ for the day; hence, Long (2000) interpolates the $a$ and $b$ coefficients derived on surrounding days to each cloudy-sky day.

### 2.5.2 Skill Score Assessment of Modelled Normalized Cloud Forcing

To assess how well modelled normalized cloud forcings match observed normalized cloud forcings we make use of the skill score paradigm developed by weather forecasters that is outlined in appendix A. In our implementation of the skill score paradigm modelled normalized cloud forcings based on a climatological effective radius are designated
as the “control.” Modelled normalized cloud forcings obtained from the two sets of retrieved cloud drop effective radii are labelled as the “forecast.” Computing the mean square errors MSE (Equation A.1) between the control and observed data and each forecast data set and observed data, we can compute the skill scores (Equation A.5) of the forecast data sets relative to the control. This paradigm allows us not only to evaluate the retrieved effective radii relative to the climatological estimate, but it also enables us to evaluate the performance of the two different retrieval techniques relative to each other.

2.5.3 Applicability of the Independent Column Approximation for Radiative Transfer

As we use a plane-parallel, horizontally homogeneous radiative transfer model, we expect more accurate modelled normalized cloud forcing values during overcast cases when the effects of horizontal photon transport are not significant. To estimate cloud fractional sky cover we use a technique developed by Long et al. (1999). Long et al. (1999) estimate cloud fraction using a relationship between cloud cover fraction and the difference between the measured and estimated clear-sky downwelling shortwave diffuse irradiances.
Chapter 3

A Climatology of Warm Boundary Layer Clouds

Using the four years of measurements described in the previous section, we now characterize the properties of warm boundary layer clouds observed during the period. The properties that we consider include cloud base height, cloud liquid water path, cloud drop effective radius, cloud visible optical depth, and normalized shortwave cloud forcing. While we considered producing cloud top height statistics for the observed warm boundary layer clouds, we decided that such statistics would be unreliable because of insect contamination of the radar returns, especially during the warm seasons. While the cloud drop effective radius in the Min and Harrison (1996) approach is a derived quantity, we nonetheless include statistics of it for comparison purposes with the Kato et al. (2001) retrieval results. Since we are interested in the radiative properties of warm boundary layer clouds, we only include in our study the properties of single layer clouds close to the surface with no cloud overhead. Cloud base height and cloud liquid water path statistics are produced for both day and night, while all of the quantities considered in the study are analyzed for daytime periods.

We searched for trends in the statistics both across years and between the different seasons. The four-year data set is sub-divided into four seasons for each year using groups of three months. The subdivisions are December-January-February (DJF), March-April-May (MAM), June-July-August (JJA) and September-October-November
Table 3.1. The annual distribution of liquid water path, cloud base height and cloud forcing from all cases of single layer warm clouds for the period 1997-2000. Note that the actual daytime LWP points (197145) are reduced to 166286 points for cloud forcing results due to intermittent missing flux measurements.

<table>
<thead>
<tr>
<th>Year</th>
<th>Time of day</th>
<th>No. of Points</th>
<th>Liquid Water Path (mm)</th>
<th>Cloud Base (m)</th>
<th>Cloud Forcing</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td>Mean</td>
<td>Median</td>
<td>Mean</td>
</tr>
<tr>
<td>1997</td>
<td>day</td>
<td>47112/42189</td>
<td>0.121</td>
<td>0.050</td>
<td>1084</td>
</tr>
<tr>
<td></td>
<td>night</td>
<td>45900</td>
<td>0.156</td>
<td>0.103</td>
<td>1003</td>
</tr>
<tr>
<td></td>
<td>all</td>
<td>93012</td>
<td>0.139</td>
<td>0.070</td>
<td>1044</td>
</tr>
<tr>
<td>1998</td>
<td>day</td>
<td>54655/43634</td>
<td>0.128</td>
<td>0.054</td>
<td>943</td>
</tr>
<tr>
<td></td>
<td>night</td>
<td>47918</td>
<td>0.147</td>
<td>0.091</td>
<td>790</td>
</tr>
<tr>
<td></td>
<td>all</td>
<td>102573</td>
<td>0.137</td>
<td>0.073</td>
<td>872</td>
</tr>
<tr>
<td>1999</td>
<td>day</td>
<td>53036/45275</td>
<td>0.116</td>
<td>0.037</td>
<td>1076</td>
</tr>
<tr>
<td></td>
<td>night</td>
<td>45387</td>
<td>0.167</td>
<td>0.081</td>
<td>979</td>
</tr>
<tr>
<td></td>
<td>all</td>
<td>98423</td>
<td>0.140</td>
<td>0.053</td>
<td>1032</td>
</tr>
<tr>
<td>2000</td>
<td>day</td>
<td>42342/35188</td>
<td>0.140</td>
<td>0.050</td>
<td>915</td>
</tr>
<tr>
<td></td>
<td>night</td>
<td>31410</td>
<td>0.169</td>
<td>0.091</td>
<td>783</td>
</tr>
<tr>
<td></td>
<td>all</td>
<td>73752</td>
<td>0.152</td>
<td>0.073</td>
<td>859</td>
</tr>
<tr>
<td>All data</td>
<td>day</td>
<td>197145/166286</td>
<td>0.126</td>
<td>0.047</td>
<td>1007</td>
</tr>
<tr>
<td></td>
<td>night</td>
<td>170615</td>
<td>0.159</td>
<td>0.091</td>
<td>896</td>
</tr>
<tr>
<td></td>
<td>all</td>
<td>367760</td>
<td>0.141</td>
<td>0.067</td>
<td>955</td>
</tr>
</tbody>
</table>
Summary statistics for the four-year and annual observation periods are presented in Table 3.1 with seasonal statistics presented in Table 3.2. As the distributions of both cloud base height and liquid water path do not follow a normal distribution we use a non-parametric test, the two-sample Wilcoxon rank-sum test (Goon et al., 1993), at a 5% level of significance, to quantititively verify whether certain interannual or interseasonal differences are significant. We now consider these results in more detail.

3.1 Cloud Base Height

3.1.1 Overall Statistics

Probability of occurrence histograms of cloud base height, for clouds with bases below 3000 m, for the four-year data set are presented in Figure 3.1. Both the daytime and nighttime distributions have peaks in the lowest kilometer of the atmosphere with relatively constant, low-percentage cloud base height occurrences above about 2000 m. The nighttime distribution has a weak secondary peak at approximately 1500 m which is not obvious in the daytime distribution. We demonstrate in the next section that this bimodality is explained by seasonal differences in cloud base height occurrence. The nighttime cloud base heights are observed to be consistently smaller than daytime values (Table 3.1). Results from the Wilcoxon rank-sum test established that the nighttime and daytime cloud base height distributions are different.

3.1.2 Interannual Trends

As all the cloud base height distributions are positively skewed the median height is a better measure of central tendency than the mean height. From Table 3.1 we find
Table 3.2. The seasonal distribution of liquid water path, cloud base height and cloud forcing from all cases of boundary layer clouds for the period 1997-2000.

(†) DJF = December January February
(‡) MAM = March April May
(¶) JJA = June July August
(§) SON = September October November

<table>
<thead>
<tr>
<th>Season</th>
<th>Time of day</th>
<th>No. of Points</th>
<th>Liquid Water Path (mm)</th>
<th>Cloud Base (m)</th>
<th>Cloud Forcing</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td>Mean</td>
<td>Median</td>
<td>Mean</td>
</tr>
<tr>
<td>DJF†</td>
<td>day</td>
<td>30870</td>
<td>0.110</td>
<td>0.072</td>
<td>587</td>
</tr>
<tr>
<td></td>
<td>night</td>
<td>41508</td>
<td>0.140</td>
<td>0.102</td>
<td>601</td>
</tr>
<tr>
<td></td>
<td>all</td>
<td>72378</td>
<td>0.127</td>
<td>0.091</td>
<td>595</td>
</tr>
<tr>
<td>MAM‡</td>
<td>day</td>
<td>71988</td>
<td>0.129</td>
<td>0.051</td>
<td>995</td>
</tr>
<tr>
<td></td>
<td>night</td>
<td>55195</td>
<td>0.178</td>
<td>0.106</td>
<td>911</td>
</tr>
<tr>
<td></td>
<td>all</td>
<td>127183</td>
<td>0.151</td>
<td>0.075</td>
<td>959</td>
</tr>
<tr>
<td>JJA¶</td>
<td>day</td>
<td>50169</td>
<td>0.085</td>
<td>0.021</td>
<td>1336</td>
</tr>
<tr>
<td></td>
<td>night</td>
<td>19539</td>
<td>0.126</td>
<td>0.036</td>
<td>1250</td>
</tr>
<tr>
<td></td>
<td>all</td>
<td>69708</td>
<td>0.097</td>
<td>0.024</td>
<td>1312</td>
</tr>
<tr>
<td>SON§</td>
<td>day</td>
<td>44118</td>
<td>0.178</td>
<td>0.068</td>
<td>943</td>
</tr>
<tr>
<td></td>
<td>night</td>
<td>54373</td>
<td>0.165</td>
<td>0.090</td>
<td>980</td>
</tr>
<tr>
<td></td>
<td>all</td>
<td>98491</td>
<td>0.171</td>
<td>0.081</td>
<td>964</td>
</tr>
</tbody>
</table>
Fig. 3.1. Histograms of daytime, nighttime, and all cloud base heights. The left column consists of the consolidated 4 year dataset while the data is divided into years in the right column.
that 1997 and 1999 have the highest median cloud base heights while 2000 has a median height which is at least 100 m below the other years. Comparing 1997, 1998 and 1999 with 2000 using a Wilcoxon rank-sum test our observation regarding the difference is found to be significant. We now check if there is a seasonal influence on this interannual difference.

### 3.1.3 Seasonal Trends

The cloud base height distributions for the different seasons are shown in Figure 3.2. These histograms illustrate that the cloud base heights for the DJF season are generally lower than for the other seasons, while the convective JJA season has clouds pretty much evenly distributed from the surface up to 3000 m. The weak bimodality seen in the nighttime cloud base height statistics is a result of the JJA distribution, which has an average higher than all other seasons. The cloud base heights do not vary from year to year for DJF. In the MAM season of 2000 we find a preponderance of clouds below 500 m, driving the mean and median to be lower than for all other years and influencing the interannual trends. This observation is confirmed using a Wilcoxon rank-sum test. For JJA the cloud bases do not display any significant interseasonal difference. In SON, though, 1999 has significantly higher bases than other years, which is confirmed using the Wilcoxon rank-sum test.
Fig. 3.2. Histograms of seasonal cloud base height.
3.2 Cloud Liquid Water Path

3.2.1 Overall Statistics

Distributions of retrieved liquid water paths for daytime, nighttime and both periods together are illustrated in Figure 3.3. In this figure liquid water paths are truncated at 1 mm; the higher liquid water path samples represent 6% of the total data set. Close inspection of radar reflectivity for those periods with liquid water paths greater than 1 mm generally indicates precipitation at the surface, in which case the microwave radiometer window is most likely wet and the retrieved value of cloud water is biased high. Consequently, we neglect these values of cloud liquid water path both in our figures and analysis.

The mean cloud liquid water path of the four-year data set is 0.14 mm with a standard deviation of 0.18 mm, while the median value is 0.067 mm. Daytime median liquid water paths are seen to be lower than the nighttime values by around 50%. This observation is confirmed as significant using the Wilcoxon rank-sum test. If we look at the histograms in Figure 3.3, the distributions are highly skewed with a modal value in the lowest bin from 0.00-0.02 mm of water. As the accuracy of two-channel microwave radiometer retrievals of cloud liquid water path is on the order of 0.01-0.02 mm, the most common value of cloud liquid water path across the four-year data set is on the order of the instrument accuracy.
Fig. 3.3. Histograms of daytime nighttime and all liquid water path. The left column consists of the consolidated 4 year dataset while the data is divided into years in the right column.
3.2.2 Interannual Trends

Annual distributions of liquid water path for single layer warm clouds for daytime, nighttime and all periods are shown in Figure 3.3. The clouds for 2000 contain more water on average than the clouds for the other years. Applying the Wilcoxon rank-sum test we find this observation to be significant. In the following section we investigate whether this high liquid water for 2000 is a result of particular seasons. The nighttime liquid water path is also seen to be consistently higher than daytime for all four years.

3.2.3 Seasonal Trends

The seasonal probability of occurrence of liquid water path for all boundary layer cloud cases are shown in Figure 3.4. As this figure demonstrates, the $JJA$ season daytime clouds contain less liquid water on average (Table 3.2). A comparison of the daytime and nighttime averages shows that the diurnal liquid water path variation is higher in spring and summer ($MAM$ and $JJA$) than fall and winter ($SON$ and $DJF$). Following up from the previous section, only $MAM$ of 2000 has higher liquid water path than other years and therefore is the main cause behind the higher than average annual value for 2000.

3.3 Cloud Drop Effective Radius

3.3.1 Overall Statistics

In Sections 2.4.1 and 2.4.2 we outlined two separate retrieval techniques for particle effective radius that use liquid water path retrievals from a microwave radiometer.
Fig. 3.4. Histograms of seasonal liquid water path.
Stratus Results from January 1997 to January 1998

Effective Radii (1 minute averaged) from Radar Based Retrievals

Number of data points: 5539; Bin Size: 0.5 µm
Median = 6.3000 µm; Mean = 6.3404 µm; st. dev. = 2.3277 µm

(a) Radar-derived radii

No. of occurrences

0 2 4 6 8 10 12 14 16 18 20
0
100
200
300
400
500
600
700

Stratus Results from January 1997 to January 1998

Effective Radii (1 minute averaged) from Radiometer Based Retrievals

Number of data points: 5539; Bin Size: 0.5 µm
Median = 7.1800 µm; Mean = 7.3869 µm; st. dev. = 2.3646 µm

(b) Transmission-derived radii

Effective Radius (µm)

No. of occurrences

0 2 4 6 8 10 12 14 16 18 20
0
100
200
300
400
500
600
700

Fig. 3.5. Histograms showing effective radius distributions for radar and transmission-derived effective radii. The dataset has 5539 one-minute averaged points. Bin size is 0.05 µm.
Table 3.3. Distribution of effective radii for one-minute averaged stratus cases. The number of points is 5539.

<table>
<thead>
<tr>
<th>Source of Radii</th>
<th>Minimum radius μm</th>
<th>Maximum radius μm</th>
<th>Mean radius μm</th>
<th>Median radius μm</th>
<th>Standard Deviation μm</th>
</tr>
</thead>
<tbody>
<tr>
<td>Radar Derived</td>
<td>1.41</td>
<td>19.69</td>
<td>6.34</td>
<td>6.30</td>
<td>2.33</td>
</tr>
<tr>
<td>Narrowband Transmittance Derived</td>
<td>2.92</td>
<td>14.82</td>
<td>7.39</td>
<td>7.18</td>
<td>2.37</td>
</tr>
</tbody>
</table>

and either radar reflectivity or narrowband solar transmission measurements. We label the results from the Kato et al. (2001) retrieval as radar-derived, while the Min and Harrison (1996) results are labelled as transmission-derived. We currently have over a year of retrieved effective radii from the two techniques, covering the period from January 1997 to January 1998 (Table 3.3; Figure 3.5). To compare the effective radii from the two retrievals at one-minute intervals, we averaged the 20 s resolution radar-derived values to one minute and we interpolated the five minute transmission-derived values to one minute. As the table and figures indicate, while the two sets of retrieved effective radii have similar, symmetric distributions, there is a clear difference between their mean values with the radar-derived values being smaller.
Fig. 3.6. Histogram showing optical depth distribution for transmission-derived radii. The dataset is one minute averaged with 5539 points and the histogram bin size is 2.5.
3.4 Cloud Visible Optical Depth

3.4.1 Overall Statistics

The transmission-derived cloud optical depths (Figure 3.6) have a mean of 19.0, a standard deviation of 17.3 and a median of 22.7. The distribution is slightly skewed and the frequency of occurrence at optical depth of 100 represents all optical depth values greater than 97.5, with the highest value being 167. As the cloud liquid water paths have a median value of 0.054 mm during the daytime, it is evident that a small amount of liquid water path has a huge influence on the cloud optical depth.

3.5 Normalized Shortwave Cloud Forcing

3.5.1 Overall Statistics

Normalized cloud forcing (Section 2.5.1) for the four-year period is plotted in Figure 3.7 for all the daytime points for which both a positive retrieved value of liquid water path and a surface flux measurement exist. As the figure illustrates, most normalized cloud forcing values lie between -0.6 and -0.9 with a secondary peak close to 0.0. The secondary peak is a result of the microwave radiometer and radar detecting a cloud overhead but with the direct beam impinging on the shortwave pyranometer with almost no attenuation due to breaks in cloud cover. The mean and median of the cloud forcing are -0.64 and -0.69 (Table 3.1). These results indicate that approximately 60% of the flux which would have reached the surface on a clear-sky day is either absorbed or reflected back to space in the presence of a warm boundary layer cloud with a typical liquid water path of around 0.05 mm.
Fig. 3.7. Histograms of consolidated, annual and seasonal cloud forcing.
3.5.2 Interannual Trends

The annual forcing histogram shown in Figure 3.7 shows that the bin counts for 2000 have slightly higher values in the region between -0.8 and -0.9. This is also reflected in Table 3.1, which shows 2000 to have a greater median forcing than the other years. This is consistent with the slightly higher average cloud liquid water path for 2000 that we showed earlier. The small differences in interannual values suggest that on a statistical level there are no significant annual variations.

3.5.3 Seasonal Trends

Normalized cloud forcing probability of occurrence histograms from one season to the next are illustrated in Figure 3.7. The seasonal cloud forcing (Table 3.2) follows the trends in daytime liquid water path with $JJA$ having the least negative cloud forcing and lowest liquid water.

3.6 Summary of the Four-Year Climatology

Overall, the liquid water path distribution is highly positively skewed when the data are subdivided into years or seasons. The medians are therefore a better measure for comparison of liquid water paths as the means are sensitive to high values. In terms of annual variations the year 2000 has a slightly higher average mainly due to higher liquid water paths in the $MAM$ season. Nighttime liquid water paths are consistently higher than daytime values irrespective of annual or seasonal comparisons. Comparing seasons, the day-night difference is higher in spring and summer than fall and winter.
Clouds that occur in the summer months of June, July and August always contain the least amount of water.

Cloud base heights had modes below one kilometer irrespective of the time of year. A weak bimodality at 1500 m in the cloud base height distribution is attributable to the summer months, which had consistently higher averages than other seasons. The cloud bases were generally the lowest in winter (DJF) months. In 2000 the average cloud base was lower than for other years primarily because the MAM season had a much lower average than for other years. We have already observed that the same MAM 2000 season also had a higher than average liquid water path. When we consider the cloud forcing statistics, we find that the results follow the liquid water path trends. In particular, there were no significant interannual variations in cloud forcing. We conclude that any particular year of data will be representative of any other year.

The effective radii distributions showed marked differences, depending on the instrument used in the retrieval, with the radar-based retrievals having a smaller average size than transmission-based radii. The standard deviations, though, are similar. Later, we use flux computations from a radiative transfer model and the retrieved effective radii in comparison with observations to check the accuracy of the retrievals.
Chapter 4

Radiative Properties of Warm Boundary Layer Clouds

While we presented statistics of cloud base height, cloud liquid water path, cloud drop effective radius, cloud visible optical depth and normalized shortwave cloud forcing in the observational climatology of warm boundary layer clouds, we did not attempt to use the observed physical properties of the clouds to model their radiative properties. Min and Harrison (1996) use visible wavelength downwelling irradiance measurements at the surface over five-minute intervals to retrieve cloud optical depth. Combining cloud optical depth and microwave radiometer retrievals of cloud liquid water path, they subsequently retrieve cloud drop effective radius. So, modeled normalized shortwave cloud forcing using Min and Harrison (1996) derived cloud drop effective radii and microwave radiometer retrieved cloud liquid water paths ought to produce model cloud forcing values in close agreement with observed values. The only major differences between such modeled and observed cloud forcing values should arise from interpolations due to differences in the temporal resolutions of the Min and Harrison (1996) retrieval and the shortwave broadband measurements, as well as the use of narrowband retrievals to broadband calculations.

The Min and Harrison (1996) retrieved cloud drop effective radii provide a means for assessing the quality of other estimates of cloud particle size in terms of correctly reproducing the downwelling shortwave irradiance at the surface. In the current analysis
we use a climatological estimate of cloud drop effective radius to evaluate both the Min and Harrison (1996) results and the cloud drop effective radii retrieved by Kato et al. (2001). We use the period from January 1997 to January 1998 for this study because both the Min and Harrison (1996) and Kato et al. (2001) retrievals are available for this year. As we illustrated in the observational climatology, the cloud statistics for this period are similar to other years. Hence we assume that our results for this year will be fairly representative of what we would obtain for other years at the ARM SGP site.

4.1 Using Radiation to Assess Stratus Cloud Microphysical Retrievals

To compare the effects of varying effective radii on downwelling shortwave irradiance at the surface we applied the RAPRAD radiative transfer model (Section 2.1) to all cases of single layer, overcast, warm stratus clouds from January 1997 to January 1998. We identified the cases by visual inspection and obtained their cloud layers and boundaries according to the procedure detailed in Section 2.2.3. Each of the selected case study periods has over a 90% instantaneous cloud cover as determined by the procedure in Section 2.5.3. Since the downwelling shortwave surface irradiance measurements had ten-minute temporal resolution, we reduced the 10-s ARSCL data (Table 2.2) statistics to one-minute resolution and we averaged the 20-s microwave radiometer retrievals of cloud liquid water path to 1-min resolution. We defined a single layer cloud over a 60-s interval as those clouds with at least 5 of the 6 ARSCL-based cloud profiles as having only a single cloud layer. If at least one hour of data for a day passed these filters, we included the data for the particular day into our analysis. The final stratus data set
Fig. 4.1. Histogram of one-minute averaged LWP data. All 5539 have stratus during the daytime. The width of each interval in the histogram is 0.02 mm. The mean LWP is 0.116 mm, the median is 0.096 mm and the standard deviation is 0.105 mm.
that we arrived at consisted of 19 days and over 80 hours of data for a total of 5539 one-minute samples.

Using only single layer, overcast cloud conditions was important to this aspect of the study since we wanted to isolate the impact of cloud particle size while minimizing the impact of three-dimensional radiation transport effects. The cloud liquid water paths of the samples in the final pool of data had mean, mode and median values of 0.116 mm, 0.05 mm and 0.096 mm (Figure 4.1; Table 4.2), illustrating that clouds occurring during overcast conditions have, on average, larger liquid water paths than clouds associated with broken cloud cover. The statistics of the liquid water paths for each day are illustrated in Table 4.1. In the radiative transfer calculations we combined these liquid water paths with either the transmission-derived (Section 2.4.2), radar-derived (Section 2.4.1) or climatological (Section 2.4) effective radii to produce three sets of normalized cloud forcings (Section 2.5.1) for the 19 days. We subsequently evaluated the normalized cloud forcings using retrieved radii as the “forecast” and the climatological effective radius as the “control” in the skill score methodology of Section 2.5.2.

4.2 Comparison of Modeled and Observed Normalized Cloud Forcing

We chose the value of the climatological effective radius so as to substantially reduce the bias between the modeled and observed normalized cloud forcing. We started with a 5.4 µm climatological effective radius after Miles et al. (2000). Through a trial and error iterative procedure, we adjusted this value until the bias between the modeled and observed normalized cloud forcing was reduced to approximately 2% of the mean square error in the forcing differences (Table 4.5). The value for the climatological effective
radius that emerged was 7.5 \( \mu \text{m} \). For comparison purposes note that the transmission-derived effective radius has a mean of 7.39 \( \mu \text{m} \), while the radar-derived effective radius has a mean of 6.34 \( \mu \text{m} \).

Independent sources like in-situ aircraft measurements (Roger Marchand, personal communication) from the Atmospheric Radiation Measurements Enhanced Shortwave Experiment II (ARESE II), as well as results from comparisons by Dong et al. (2000), corroborate a value of 7.5 \( \mu \text{m} \) for the climatological effective radius. The lowest value of transmission-derived effective radius was around 3 \( \mu \text{m} \), (Figure 3.5) while continental stratus in the collection of Miles et al. (2000) had a significant number of cases with even lower effective radii. As the Miles et al. (2000) dataset contains a global distribution of case studies, some cases may not be characteristic of the clouds comprising our study. In fact, ignoring the reclassification of certain continental cases as marine by Miles et al. (2000), as well as cases with radii below 4 \( \mu \text{m} \), the mean effective radius of the data in Miles et al. (2000) becomes 7.4 \( \mu \text{m} \).
Table 4.1. Cloud forcing and LWP means, as well as errors in cloud forcing when compared to observations. The data have one-minute resolution. Only days with a minimum of one hour of single layer stratus are included in the study.

(¶) fixed effective radius of 7.5 µm.
(§) Radar-derived effective radius (Kato et al., 2001).
(†) Transmission-derived effective radius (Min and Harrison, 1996).
(‡) Errors are calculated with respect to observations.

<table>
<thead>
<tr>
<th>Serial Number</th>
<th>Date</th>
<th>No. of Points</th>
<th>Mean LWP</th>
<th>Mean Forcing Observations</th>
<th>Mean Forcing Radar†</th>
<th>Mean Forcing Radiometer‡</th>
<th>Fixed‡</th>
<th>Mean Square Error‡</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>January 1, 1997</td>
<td>429</td>
<td>0.140</td>
<td>-0.78</td>
<td>-0.78</td>
<td>-0.77</td>
<td>-0.76</td>
<td>0.0017</td>
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<tr>
<td>2</td>
<td>March 4, 1997</td>
<td>274</td>
<td>0.071</td>
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<td>-0.58</td>
<td>-0.56</td>
<td>-0.56</td>
<td>0.0179</td>
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<td>March 17, 1997</td>
<td>330</td>
<td>0.099</td>
<td>-0.70</td>
<td>-0.57</td>
<td>-0.70</td>
<td>-0.62</td>
<td>0.0303</td>
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<td>4</td>
<td>March 19, 1997</td>
<td>291</td>
<td>0.108</td>
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<td>-0.68</td>
<td>-0.63</td>
<td>-0.63</td>
<td>0.0101</td>
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<td>5</td>
<td>May 26, 1997</td>
<td>222</td>
<td>0.060</td>
<td>-0.64</td>
<td>-0.37</td>
<td>-0.58</td>
<td>-0.45</td>
<td>0.1063</td>
</tr>
<tr>
<td>6</td>
<td>June 4, 1997</td>
<td>347</td>
<td>0.187</td>
<td>-0.64</td>
<td>-0.61</td>
<td>-0.60</td>
<td>-0.63</td>
<td>0.0316</td>
</tr>
<tr>
<td>7</td>
<td>August 8, 1997</td>
<td>229</td>
<td>0.107</td>
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<td>-0.60</td>
<td>-0.58</td>
<td>-0.58</td>
<td>0.0237</td>
</tr>
<tr>
<td>8</td>
<td>August 19, 1997</td>
<td>377</td>
<td>0.083</td>
<td>-0.48</td>
<td>-0.60</td>
<td>-0.56</td>
<td>-0.55</td>
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<tr>
<td>9</td>
<td>September 5, 1997</td>
<td>240</td>
<td>0.049</td>
<td>-0.42</td>
<td>-0.54</td>
<td>-0.46</td>
<td>-0.47</td>
<td>0.0499</td>
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<tr>
<td>10</td>
<td>September 25, 1997</td>
<td>149</td>
<td>0.139</td>
<td>-0.73</td>
<td>-0.80</td>
<td>-0.75</td>
<td>-0.74</td>
<td>0.0102</td>
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<td>11</td>
<td>October 10, 1997</td>
<td>319</td>
<td>0.165</td>
<td>-0.64</td>
<td>-0.82</td>
<td>-0.68</td>
<td>-0.76</td>
<td>0.0464</td>
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<td>12</td>
<td>October 19, 1997</td>
<td>389</td>
<td>0.082</td>
<td>-0.74</td>
<td>-0.60</td>
<td>-0.74</td>
<td>-0.64</td>
<td>0.0230</td>
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<td>October 20, 1997</td>
<td>128</td>
<td>0.270</td>
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<td>-0.86</td>
<td>-0.87</td>
<td>0.0016</td>
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<td>14</td>
<td>November 6, 1997</td>
<td>276</td>
<td>0.038</td>
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<td>-0.68</td>
<td>-0.59</td>
<td>-0.48</td>
<td>0.0191</td>
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<td>November 7, 1997</td>
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<td>0.051</td>
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<td>-0.75</td>
<td>-0.63</td>
<td>-0.54</td>
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<td>16</td>
<td>December 3, 1997</td>
<td>447</td>
<td>0.148</td>
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<td>-0.88</td>
<td>-0.81</td>
<td>-0.80</td>
<td>0.0057</td>
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<tr>
<td>17</td>
<td>December 12, 1997</td>
<td>233</td>
<td>0.248</td>
<td>-0.73</td>
<td>-0.82</td>
<td>-0.72</td>
<td>-0.73</td>
<td>0.0193</td>
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<tr>
<td>18</td>
<td>December 22, 1997</td>
<td>308</td>
<td>0.079</td>
<td>-0.68</td>
<td>-0.82</td>
<td>-0.66</td>
<td>-0.66</td>
<td>0.0231</td>
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<tr>
<td>19</td>
<td>January 5, 1998</td>
<td>275</td>
<td>0.148</td>
<td>-0.76</td>
<td>-0.81</td>
<td>-0.75</td>
<td>-0.78</td>
<td>0.0068</td>
</tr>
</tbody>
</table>
Table 4.2. Simple statistics of liquid water path for one-minute averaged stratus cloud LWP’s below 0.5 mm. The number of points is 5539.

<table>
<thead>
<tr>
<th>Dataset Source</th>
<th>dual-channel MWR</th>
</tr>
</thead>
<tbody>
<tr>
<td>Number of points</td>
<td>5539</td>
</tr>
<tr>
<td>mean</td>
<td>0.1160 mm</td>
</tr>
<tr>
<td>median</td>
<td>0.0960 mm</td>
</tr>
<tr>
<td>standard deviation</td>
<td>0.1054 mm</td>
</tr>
</tbody>
</table>

Histograms of normalized shortwave cloud forcing for the observations and the model computations using retrieved cloud liquid water path and the three sets of cloud drop effective radii are illustrated in Figure 4.2. The means, medians and standard deviations for the points in these histograms are listed in Table 4.3. Scatter plots of the observed versus modeled normalized cloud forcing are presented in Figure 4.3, with the one-to-one lines represented in blue, least square fits that are constrained to have zero normalized cloud forcing in magenta and unconstrained least square fits in black. The corresponding line parameters and correlation coefficients are listed in Table 4.4. Applying the skill score methodology to these observations using the climatological effective radius results as the control produced the results listed in Table 4.5.

Inspecting the figures and tables, we find that the climatological effective radius out-performs the radar-derived effective radii in explaining the observed normalized cloud forcing and it almost performs as well as the transmission-derived effective radii. For example, the observed normalized cloud forcing (Figure 4.2a) is clearly bimodal and the transmission-derived results (Figure 4.2d) capture this bimodality. While the
Table 4.3. Distribution of cloud forcing for one-minute averaged stratus data. The number of points is 5539.

<table>
<thead>
<tr>
<th>Source of forcing</th>
<th>Mean forcing</th>
<th>Median forcing</th>
<th>Standard deviation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Climatological (7.5 µm)</td>
<td>-0.64</td>
<td>-0.69</td>
<td>0.20</td>
</tr>
<tr>
<td>Radar Derived</td>
<td>-0.69</td>
<td>-0.77</td>
<td>0.21</td>
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<td>Narrowband Derived</td>
<td>-0.67</td>
<td>-0.70</td>
<td>0.17</td>
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<tr>
<td>Observations</td>
<td>-0.66</td>
<td>-0.68</td>
<td>0.17</td>
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</table>

Table 4.4. Linear regression coefficients for least square fit between the three different cloud forcing calculations and observations. The slope of a constrained fit with the intercept forced to 0 is also shown.

† All comparisons are with respect to observations

<table>
<thead>
<tr>
<th>No.</th>
<th>Dataset compared†</th>
<th>Correlation Coefficient</th>
<th>Unconstrained coefficients</th>
<th>Constrained coefficient</th>
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</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td>intercept $a_0$   slope $a_1$</td>
<td>slope $a_1$</td>
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<tr>
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<td>Transmission-derived</td>
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<td>-0.1138</td>
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<td>3</td>
<td>Climatological</td>
<td>0.79</td>
<td>-0.0235</td>
<td>0.9374</td>
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Fig. 4.2. Histograms showing cloud forcing distributions. The climatological radius is 7.5 µm.
Fig. 4.3. Scatter plots of observed cloud forcing versus modeled cloud forcing using (a) a fixed effective radius of 7.5 µm, (b) radar-derived radii and (c) narrowband transmission-derived radii. Two regression lines, one constrained to 0 forcing, are also drawn.
Table 4.5. The mean square error in a comparison of observed and modeled cloud forcing is split into its components. Subscript $c$ (control) represents fixed effective radius calculations, $o$ observations and $fk$ and $fm$ (forecast) represent radar-derived and transmission-derived effective radii, respectively. The number of one-minute averaged datapoints is 5539.

<table>
<thead>
<tr>
<th>Datasets compared</th>
<th>Square of bias</th>
<th>Random Error</th>
<th>Variance Error</th>
<th>Total Error</th>
</tr>
</thead>
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<td>$\text{MSE}_{c,o}$</td>
<td>0.000324</td>
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<tr>
<td>$\text{MSE}_{fk,o}$</td>
<td>0.001016</td>
<td>0.023248</td>
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<td>0.025361</td>
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<tr>
<td>$\text{MSE}_{fm,o}$</td>
<td>0.000011</td>
<td>0.010548</td>
<td>0.000012</td>
<td>0.010571</td>
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</table>

climatological effective radius forcings (Figure 4.2b) do not capture the bimodality as well the transmission-derived results, they are clearly better than those produced by the radar-derived effective radii (Figure 4.2c). Considering the scatter plots, we find that the radar-derived normalized cloud forcing values are the most biased relative to the observations. Furthermore, the primary improvement of the transmission-derived forcings over those based on the climatological effective radius is a reduction in the random error relative to the observations. Relative to the forcings based on the climatological effective radius, the skill scores for the radar-derived and transmission-derived effective radii forcings are -0.71 and 0.30, respectively. As a positive skill score means improvement relative to the control and a negative value indicates degradation relative to the control, we conclude that in shortwave radiation studies the climatological effective radius is preferable to the radar-derived effective radii of Kato et al. (2001) while the transmission-derived
effective radii of Min and Harrison (1996) are preferable to the climatological effective radius.

For each of the 19 case study periods we present time series of normalized cloud forcing in Figure 4.4 and summary statistics in Table 4.1. Inspection of the figure and table confirms some of our earlier conclusions. The transmission-derived effective radii produce normalized cloud forcings in best agreement with observations, while the forcings based on the climatological effective radius out-perform the forcings based on the radar-derived effective radii. An observation that emerges from inspection of these individual case study days is that the normalized cloud forcings based on the climatological effective radius are consistently biased relative to the observations in one direction for extended intervals of time within a case study period. These biases are most likely the result of variations in stratus cloud drop effective radii within extended intervals of a case study period and from one case study period to the next. While the effects of inhomogeneities in cloud structure were not explicitly accounted for in either the transmission-derived effective radius retrievals or the normalized cloud forcing computations, we attempted to reduce their influence by selecting only overcast cases.

4.3 Stratus Cloud Radiative Properties: Are Drop Sizes Important?

The overall superior performance of the transmission-derived effective radii is not surprizing as these radii are chosen to match the observed transmissions in a visible solar band and we expected them to produce the most accurate downwelling shortwave irradiances at the surface. What we did find surprizing was that the climatological effective radius was almost as skilled as the transmission-derived effective radii in modeling the
Fig. 4.4. Time series of cloud forcing for individual days. The date is in the YYYYMM-DD format. Linear features in the figures (eg. Figure k) are a result of interpolating missing liquid water path values.
downwelling shortwave irradiances at the surface. This result suggests that for understanding the importance of warm clouds to the surface radiation budget knowing the liquid water paths for these clouds is of paramount importance and the cloud particle size is only of secondary importance. These results suggest that measuring the influence of changes in drop size on solar transmission and cloud albedo, as suggested by Twomey et al. (1984) and Twomey (1991), will be difficult, as liquid water path uncertainties may mask the effect.

4.3.1 The Range of Stratus Cloud Liquid Water Paths and Effective Radii

Cloud liquid water path and cloud drop effective radius are two important geophysical parameters for quantifying the microphysical and radiative properties of a stratus cloud. These two parameters have natural limits dictated by nature, above and below which their values seldom occur. Knowledge of the range of values for these two parameters is quite important to theoretical sensitivity studies based on them because sensitivity studies based on parameter values that never occur in nature are of little value. In this section we assess the sensitivity of normalized cloud forcing to changes in cloud liquid water path and cloud drop effective radius. As a prelude to the analysis, we first consider the range of cloud liquid water paths and cloud drop effective radii that we expect to find in nature.

In an analysis of continental stratus microphysics based on aircraft in situ measurements reported in the scientific literature Miles et al. (2000) arrived at a mean cloud drop effective radius of 5.4 $\mu$m with a standard deviation in the measurements of 2.0 $\mu$m. In the current analysis we arrived at a value of 7.5 $\mu$m for the climatological cloud drop
effective radius, while the mean and standard deviation generated by the transmission-derived effective radius retrieval are 7.39 µm and 2.37 µm, respectively. While Miles et al. (2000) provide mean liquid water content, we base our study on the liquid water path. The liquid water paths in our dataset have a mean of 0.116 mm, a standard deviation of 0.105 mm and a median of 0.096 mm. For the radiative transfer calculations in the current study we set the cloud thickness to 500 m, which leads to a mean liquid water content of 0.2 g/m³ for a liquid water path of 0.1 mm. This value of the liquid water content is similar to the mean liquid water content of 0.18 g/m³ presented by Miles et al. (2000). Using a lognormal distribution to characterize stratus cloud drop sizes, Miles et al. (2000) arrived at a mean lognormal distribution width of 0.38 with a standard deviation of 0.14.

4.3.2 Sensitivity of Optical Depth to Changes in LWP and \( R_e \)

Using Equation 2.12, together with the approximation that the extinction coefficient \( Q_{\text{ext}} \) of cloud drops is 2 at solar wavelengths, leads to the relation

\[
\frac{\delta \tau}{\tau} = \frac{\delta \text{LWP}}{\text{LWP}} - \frac{\delta r_e}{r_e}.
\] (4.1)

(Note that the approximation \( Q_{\text{ext}} \approx 2 \), while reasonable for visible wavelengths, will lead to errors in the near infrared region and at longer wavelengths as can be seen from Figure 4.5.) Even though Equation 4.1 shows that the liquid water path and effective radius are equally important in causing changes in the optical depth, and thereby solar transmission and downwelling shortwave irradiance at the surface, we do not find strong
evidence of this effect in our observations. The explanation for this finding lies in the fact that the downwelling shortwave irradiance at the surface is most sensitive to changes in optical depth for small values of the optical depth. Given the typical values of cloud drop effective radii that occur in stratus at the ARM SGP site, the values of liquid water path that most often occur in stratus over the ARM SGP site are sufficiently small and sufficiently varied to produce optical depths that fall across the range where the downwelling irradiance is most sensitive to changes in them. Changes in the magnitude of the cloud drop effective radius that occur naturally in continental clouds, however, are not sufficient to produce such large variations in the downwelling shortwave irradiance at the surface.

Consider a stratus cloud with a liquid water path of 0.1 mm that contains drops with a collective effective radius of 8 \( \mu \text{m} \). These values are typical of what we would expect at the ARM SGP site. Using Equation 4.1, we find that a 0.02 mm change in liquid water path, which is much smaller than the standard deviation in Table 4.2, would need an opposing change of 2 \( \mu \text{m} \) in the effective radius, which equals the standard deviation in Table 3.3, to keep the optical depth invariant. That is, for small, unremarkable changes in typical stratus cloud liquid water path values, the effective radius would have to make a relatively large, remarkable change to keep the optical depth invariant.

Another way of looking at this issue would be to consider equal percentage changes to typical values of cloud liquid water path and cloud drop effective radius that are within the natural range of variability of these two geophysical parameters and then compare the corresponding effect on optical depth. To this end we fix the range of variability to one standard deviation on either side of the corresponding mean value. This leads to
liquid water paths that range from 0.0-0.2 mm and cloud drop effective radii that range from 5.5-9.5 µm. Breaking these two ranges into 20 equal intervals and computing the change in optical depth across each of the two parameter ranges independently, while holding the other parameter fixed at its mean value, we find that the rate of change of optical depth with effective radius varies from 0.2 to 0.8 across the corresponding interval (Figure 4.6) while the rate of change of optical depth with cloud liquid water path is approximately 3.0 across the corresponding interval (Figure 4.7). The optical depth is at least three times as sensitive to variations in liquid water path as effective radius for the values we would expect at the ARM SGP site.

Fig. 4.5. The extinction coefficient of cloud droplets for different effective radii plotted at wavelengths used in our radiative transfer model (Table 2.1).
Fig. 4.6. The rate of change in optical depth as a function of effective radii is. The optical depth was calculated using an extinction coefficient of 2, which is good for visible wavelengths. The rate of change of optical depth $\tau$ is per 0.02 $\mu$m change in effective radius. The liquid water path is 0.1 mm.

Fig. 4.7. The rate of change in optical depth as a function of liquid water path. The optical depth was calculated using an extinction coefficient of 2, which is good for visible wavelengths. The rate of change of optical depth $\tau$ is per 0.01 mm change in liquid water. The effective radius used for the calculations is 7.5 $\mu$m.
Fig. 4.8. Normalized cloud forcing calculated using our radiative transfer model is plotted as function of liquid path and effective radius. The solar zenith angle used for the calculations is 60°. The arrows on the axes demarcate the observed ranges.

Fig. 4.9. Normalized cloud forcing calculated using our radiative transfer model is plotted as function of liquid water path for different fixed effective radii represented by the lines. The solar zenith angle used for the calculations is 60°.
Fig. 4.10. Normalized cloud forcing calculated using our radiative transfer model is plotted as function of effective radii for different fixed liquid water paths represented by the lines. The solar zenith angle used for the calculations is 60°.

Fig. 4.11. Normalized cloud forcing calculated using our radiative transfer model is plotted as function of liquid water paths for different distribution widths and a fixed effective radius represented by the lines. The solar zenith angle used for the calculations is 60° and the effective radius was 7.5 µm.
Fig. 4.12. The rate of change in normalized cloud forcing with (a) effective radius and (b) liquid water path. The rate of change of cloud forcing in (a) is per 0.02 µm change in effective radius for a fixed liquid water path of 0.1 mm. The rate of change of cloud forcing in (b) is per 0.01 mm change in liquid water path for a fixed effective radius of 7.5 µm. The solar zenith angle is 60° for both (a) and (b).
Fig. 4.13. The fluxes at the surface for different effective radii between 8 and 9 µm are plotted for the 32 different bands of the radiative transfer model. The solar zenith angle is 60° and the liquid water path is 0.1 mm.

Fig. 4.14. The difference in the extinction coefficients, for effective radii between 8 and 9 µm with 8 µm as the base, for the 32 wavelengths of the radiative transfer model.
4.3.3 Sensitivity of Cloud Forcing to Changes in LWP and $R_e$

While the previous section demonstrated that the optical depth is at least three times more sensitive to liquid water path as it is to effective radius, the sensitivity might be different if we were to compare normalized cloud forcing, as multiple scattering effects would then be included as well. To test this possibility we used RAPRAD (Section 2.1) to compute the downwelling irradiance at the surface for different stratus cloud liquid water paths and cloud drop size distributions confined to a 500 m cloud layer located from 0.5 km to 1.0 km above ground level. The cloud drop size distributions were modeled with a lognormal distribution with a width of 0.38. We obtained the atmospheric thermodynamic state from the U.S. standard atmospheric profile and set the solar zenith angle to $60^\circ$. We computed the scattering and absorption coefficients, as well as the single-scattering albedo and asymmetry parameter, using Mie theory with the refractive indices provided in Table 2.1.

The normalized cloud forcings that resulted from our calculations, as a function of effective radius and liquid water path, are shown in Figure 4.8. As the figure illustrates, the change in forcing is a non-linear function of liquid water path with a larger magnitude slope, and hence sensitivity, at the lower values. In Figure 4.9 we emphasize this nonlinearity by plotting the dependence of normalized cloud forcing versus liquid water path for a number of different cloud drop effective radii. As it turns out, at the ARM SGP site most of the naturally occurring liquid water paths lie in the high-sensitivity region (Figure 4.1). The variation of forcing with effective radius is much more linear for typical values of cloud liquid water path (Figure 4.10) and hence much less sensitive
to changes in cloud drop effective radius across the whole range of effective radii that we expect at the ARM SGP site. Note that the results presented here are not dependent on the width of the lognormal distribution that we use to represent the cloud drop size distribution (Figure 4.11).

To quantify the sensitivity of normalized cloud forcing to changes in cloud liquid water path and cloud drop effective radius we computed the changes in normalized cloud forcing both as a function of cloud drop effective radius with the liquid water path held fixed at 0.1 mm (Figure 4.12a) and as a function of cloud liquid water path with the cloud drop effective radius held fixed at 7.5 $\mu$m (Figure 4.12b). Note that the scatter in Figure 4.12a about the regression line is a result of the non-linear variation of the extinction coefficient with effective radius (Figure 4.14) that leads to abrupt variations with wavelength of downwelling irradiance at the surface (Figure 4.13; Table 2.1). The results in Figure 4.12a illustrates how the rate of change of normalized cloud forcing varies from 0.001 to 0.006 (per 0.2 $\mu$m change in drop size) over the cloud drop effective radius range of 4-15 $\mu$m. This is much smaller than the 0.01-0.30 rate of change of normalized cloud forcing (per 0.02 mm change in liquid water path) over the range of typical values of cloud liquid water path (Figure 4.12b). Overall, the normalized cloud forcing, which includes the effects of multiple scattering, is approximately six times more sensitive to changes in the liquid water path as it is to changes in the effective radius for typical values of these geophysical parameters for stratus clouds over the ARM SGP site. We conclude that liquid water path is the dominant parameter in correctly ascertaining continental stratus cloud solar transmission and the cloud drop effective radius is of secondary importance.
Chapter 5

A Parameterization for Stratus Cloud Normalized Cloud Forcing

In the climatology that we developed using ARM SGP data from 1997 to 2000 (Chapter 3) we found little interannual variation in cloud liquid water paths for boundary layer cloud cases, although some seasonal variations did exist. Using a climatological approach (Chapter 4), we found that the downwelling irradiances at the surface were primarily dependent on cloud liquid water path, with a much weaker dependence on cloud drop effective radius. These results motivated us to develop a parameterization of normalized cloud forcing that is based, in part, on cloud liquid water path and no explicit dependence on cloud drop effective radius. In this chapter we present the parameterization that we developed using the four-year dataset described in Chapter 3 and we subsequently compare its performance with the results from the transmission-derived normalized cloud forcing calculations in Chapter 4.

5.1 Normalized Cloud Forcing and Cloud Liquid Water Path

To illustrate the dependence of normalized cloud forcing on cloud liquid water path we plot the observed and modeled normalized cloud forcings as a function of cloud liquid water path in Figure 5.1. The multiple values of observed normalized cloud forcing for a fixed value of cloud liquid water path (Figure 5.1a) are a result of varying solar zenith angle, cloud drop effective radius and cloud structure inhomogeneities that impact
Fig. 5.1. Plot showing cloud forcing as a function of liquid water path. Cloud forcings are either (a) observed or calculated using (b) climatological effective radius, (c) radar based effective radii or (d) narrowband transmission based effective radii. The dataset is 1-minute averaged with 5539 points.
the measured downwelling irradiances at the surface. The scatter in the computed radar-derived (Figure 5.1c) and transmission-derived (Figure 5.1d) normalized cloud forcing versus liquid water path graphs result only from varying solar zenith angle and cloud drop effective radius, while the scatter in the computed normalized cloud forcing based on the climatological effective radius (Figure 5.1b) results only from a varying solar zenith angle, as all other parameters are fixed. The resemblance of Figure 5.1b to Figure 5.1a is what motivated us to develop an expression for the dependence of normalized cloud forcing on cloud liquid water path together with the solar zenith angle.

5.2 Development of a Model for Normalized Cloud Forcing

To model the variation of normalized cloud forcing with cloud liquid water path and solar zenith angle we assumed an exponential relation of the form

$$NCF(\mu, \text{LWP}) = e^{[a_0(\mu) + a_1(\mu)\text{LWP}]} - 1,$$

(5.1)

where $\mu$ is the cosine of the solar zenith angle, LWP is the cloud liquid water path and $a_0$ and $a_1$ are regression coefficients that depend on solar zenith angle. We transformed Equation 5.1 by taking the logarithm of both sides after moving the constant factor to the left hand side, thereby obtaining

$$\log[NCF(\mu, \text{LWP}) + 1] = a_0(\mu) + a_1(\mu)\text{LWP},$$

(5.2)

which is the model that we used for the parameterization.
Fig. 5.2. Least square fit to cloud forcing as a function of liquid water path for eight solar zenith angle bins. The black dashed line is from a single exponential fit (Equation 5.1) to the whole dataset, while the solid line is from a combination of three fits (Equations 5.6, 5.7 and 5.8) for different liquid water path ranges.
Table 5.1. Comparison of regression coefficients for the three different equations needed to cover the liquid water path regime as well as the coefficients for the exponential fit covering the whole liquid water path domain.

<table>
<thead>
<tr>
<th>Solar Zenith Angle of Points</th>
<th>LWP range (mm)</th>
<th>$a_{10}$</th>
<th>$a_{21}$</th>
<th>$a_{22}$</th>
<th>$a_{23}$</th>
<th>$a_{24}$</th>
<th>$a_{30}$</th>
<th>$a_{31}$</th>
<th>$a_0$</th>
<th>$a_1$</th>
</tr>
</thead>
<tbody>
<tr>
<td>13 - 29</td>
<td>20402</td>
<td>77.34</td>
<td>-0.7327</td>
<td>0.0325</td>
<td>0.4132</td>
<td>12.98</td>
<td>-0.9247</td>
<td>-1.6962</td>
<td>-0.5460</td>
<td>-3.1933</td>
</tr>
<tr>
<td>29 - 39</td>
<td>21099</td>
<td>70.53</td>
<td>-0.7245</td>
<td>0.3128</td>
<td>0.4252</td>
<td>13.53</td>
<td>-1.0094</td>
<td>-2.1101</td>
<td>-0.5543</td>
<td>-3.5510</td>
</tr>
<tr>
<td>39 - 48</td>
<td>20803</td>
<td>71.80</td>
<td>-0.8847</td>
<td>-0.2308</td>
<td>0.5861</td>
<td>10.93</td>
<td>-1.0958</td>
<td>-2.2535</td>
<td>-0.5969</td>
<td>-3.7816</td>
</tr>
<tr>
<td>48 - 55</td>
<td>22782</td>
<td>80.79</td>
<td>-0.7404</td>
<td>0.3939</td>
<td>0.4437</td>
<td>18.85</td>
<td>-1.1209</td>
<td>-2.4690</td>
<td>-0.6591</td>
<td>-3.9373</td>
</tr>
<tr>
<td>55 - 62</td>
<td>24875</td>
<td>86.33</td>
<td>-0.6943</td>
<td>0.6619</td>
<td>0.3780</td>
<td>23.46</td>
<td>-1.1831</td>
<td>-2.2942</td>
<td>-0.7186</td>
<td>-3.9218</td>
</tr>
<tr>
<td>62 - 68</td>
<td>20807</td>
<td>100.41</td>
<td>-0.7064</td>
<td>0.6024</td>
<td>0.3747</td>
<td>26.13</td>
<td>-1.1408</td>
<td>-2.8456</td>
<td>-0.7641</td>
<td>-4.2538</td>
</tr>
<tr>
<td>68 - 74</td>
<td>15871</td>
<td>137.62</td>
<td>-0.7327</td>
<td>0.6301</td>
<td>0.3813</td>
<td>27.44</td>
<td>-1.2389</td>
<td>-3.1125</td>
<td>-0.8056</td>
<td>-4.8537</td>
</tr>
<tr>
<td>74 - 80</td>
<td>12231</td>
<td>152.07</td>
<td>-0.8046</td>
<td>0.3654</td>
<td>0.3836</td>
<td>23.25</td>
<td>-1.4481</td>
<td>-2.8851</td>
<td>-0.9258</td>
<td>-5.1057</td>
</tr>
</tbody>
</table>
5.3 Fitting the Model of Normalized Cloud Forcing to Observations

To fit the model of normalized cloud forcing to the observations we first divided the data into 8 bins equally spaced across cosines of the solar zenith angle ranging from $\cos(0^\circ)$ to $\cos(80^\circ)$. We did not use observations with solar zenith angles greater than $80^\circ$ so as not to have to deal with the numerous problems that one encounters at these large angles, including the break-down of assumptions used in plane-parallel radiative transfer calculations and inherent problems in measuring the downwelling irradiance at these large solar zenith angles. We then fit the model to the observations using a linear least square regression for each solar zenith angle bin. We made the fitting procedure “robust” by incorporating a filter on outliers (Huber, 1981) so that large negative values of normalized cloud forcing did not dominate the fit at small liquid water path values. A set of regression coefficients $(a_0(\mu), a_1(\mu))$ emerged from the fit for each solar zenith angle bin (Table 5.1). We present the resulting regression line for each solar zenith angle bin in Figure 5.2 as the black dotted line. Note that the regression fit follows the binned median cloud forcing values for higher liquid water paths, while for liquid water paths of 0 the regression yields non-zero cloud forcing values for every zenith angle bin.

To finish the model we developed analytical expressions for the coefficients $(a_0(\mu), a_1(\mu))$ generated by the regression fits (Figure 5.3). Fitting a first order, linear polynomial to the $a_0(\mu)$ coefficients and a second order, quadratic polynomial fit to the $a_1(\mu)$ coefficients, we obtained

$$a_0(\mu) = -1.0 + 0.53\mu$$ (5.3)
Fig. 5.3. Plot of regression coefficients (a) \( a_0 \) and (b) \( a_1 \) (Equation 5.1) for different zenith angles. Regression coefficient \( a_0 \) is fit linearly while \( a_1 \) is fit quadratically as a function of the cosine of the solar zenith angle.
and

\[ a_1(\mu) = -6.04 + 4.65\mu - 1.84\mu^2. \] (5.4)

Substituting these equations into Equation 5.1 leads to

\[ CF(\mu, \text{LWP}) = e^{[-1.0+0.53\mu+(-6.04+4.65\mu-1.84\mu^2)\text{LWP}]} - 1 \] (5.5)

as the model for normalized cloud forcing as a function of \( \mu \) and LWP.

### 5.4 Evaluation of Parameterized Normalized Cloud Forcing

To evaluate the parameterization of normalized cloud forcing derived in the previous section we compare its predictions with the normalized cloud forcing results of Chapter 4. Overall the performance of the parameterization is similar to that of the climatological effective radius for all 19 case study periods from 1997 to 1998. The parameterization has a skill score of 0.19 against the climatological effective radius normalized cloud forcings as the control, indicating better performance against the observations. Plotting the parameterized versus observed normalized cloud forcing (Figure 5.4a), we found a bias in the parameterization at low (i.e. approaching 0) normalized cloud forcing values. These samples occurred when there was a relatively high liquid water path cloud in the field-of-view of the microwave radiometer and little cloud between the sun and the pyranometer. Periods when there was little to no cloud in the field-of-view of the microwave radiometer and high liquid water path cloud between the sun and the pyranometer are not represented in Figure 5.4a because the parameterization approaches a non-zero forcing of -0.4 as the liquid water path goes to 0.
Results (Parameterization) from January 1997 to January 1998

Scatter Plot for Cloud Forcing calculated using 1 minute averaged inputs

Number of data points 5539

Mean difference: 0.0180; Mean square difference: 0.0148

Regression Coefficients

(a) exponential fit
\[ a_0 = -0.3217, \ a_1 = 0.5335; \text{Constrained} \ a_0: 0.9900 \]

(b) 3-equation fit
\[ a_0 = -0.2058, \ a_1 = 0.7116; \text{Constrained} \ a_0: 1.0037 \]

---

Fig. 5.4. Scatter plot of (a) exponential (Equation 5.1) and (b) 3-equation (Equations 5.6, 5.7 and 5.8) parameterized cloud forcings against observations.
Analyzing the components of the skill score, we found that the mean square error for the parameterization is primarily a result of bias, while its random error is lower than for the forcings based on the climatological effective radius. The correlation between the parameterization and observations is 76% and increases to 83% if we exclude observations with normalized cloud forcings lying between 0.0 and -0.4. (Note that excluding observations with normalized cloud forcings between 0.0 and -0.4 led to a loss of 8% of the total cases.) Recomputing the parameterization skill score against the climatological effective radius results as control with these low cloud forcing points removed, we found that it increased to 0.55.

In Chapter 4 we showed that the transmission-based cloud drop effective radius retrieval performed best against the observations. Considering the whole dataset with the parameterization as control, the skill score for the transmission-derived normalized cloud forcings is 0.19. However, when we eliminate the observed normalized cloud forcing values between 0.0 and -0.4, the skill score for the transmission-derived results drops to −0.22, implying a better performance by the parameterization. In this case the correlation between the observations and transmission-derived results is 0.82, which is similar to the parameterization. Hence, we conclude that the parameterization provides a reasonable approximation to the actual observations, especially for normalized cloud forcing values between -0.4 and -1.0.

### 5.5 A More Physical Model for Normalized Cloud Forcing

One drawback to the parameterization developed in the previous section is that the normalized cloud forcing does not go to 0 as the cloud liquid water path goes to 0.
Fig. 5.5. Time series of parameterized cloud forcing forcing for individual days using Equations 5.6, 5.7 and 5.8. The date is in the YYYYMMDD format.
Furthermore, the shape of the fit does not quite match the data for any solar zenith angle bin, always overestimating the magnitude of the normalized cloud forcing at small and large values of cloud liquid water path while always underestimating it at intermediate values of cloud liquid water path. To improve the overall fit of the parameterization to the data we divided the cloud liquid water paths into three regions and we used combinations of functions of the form \(\exp(a \text{ LWP})\) and \(2/(2 + a \text{ LWP})\) (Bohren, 1987) to fit the data in each region. The three equations that emerged from this process are:

For LWP < 0.015

\[
NCF(\mu, \text{LWP}) = \frac{2}{(2 + a_{10}(\mu) \text{LWP})} - 1, \tag{5.6}
\]

for 0.015 \(\leq\) LWP < 0.15

\[
NCF(\mu, \text{LWP}) = a_{21}(\mu)e^{(a_{22}(\mu) \text{LWP})} + a_{23}(\mu)e^{(-a_{24}(\mu) \text{LWP})}, \tag{5.7}
\]

and for 0.15 \(\leq\) LWP

\[
NCF(\mu, \text{LWP}) = e^{(a_{30}(\mu) + a_{31}(\mu) \text{LWP})} - 1. \tag{5.8}
\]

For each of the eight solar zenith angle bins we solve Equations 5.6 and 5.8 for the regression coefficients using a linear least square fit, while for Equation 5.7 we use the non-linear Newton-Gauss fitting procedure (Table 5.1). As before, we fit polynomial functions to the coefficients obtained by linear least square fits (Figure 5.6), obtaining

\[
a_{10}(\mu) = 238 - 429\mu + 275\mu^2, \tag{5.9}
\]
The coefficients produced by the Newton-Gauss fitting procedure were not smooth, so we used the actual regression coefficients for each solar zenith angle range in Table 5.1 to estimate the normalized cloud forcing for liquid water paths in this intermediate range. The final parameterization is also illustrated in Figure 5.2 by the solid colored lines with the colors representing the different liquid water path regimes represented by Equations 5.6, 5.7 and 5.8.

Assessing the quality of this second parameterization, we find that it slightly improves performance compared to Equation 5.1 with the skill score and correlation against the climatological effective radius improving to 0.26 and 79% respectively. Time series of the observed and model estimated normalized cloud forcing for all 19 case study periods from 1997 to 1998 (Figure 5.5) shows that the performance of the parameterization is similar to that of the climatological effective radius. In terms of estimating the normalized cloud forcing the second parameterization does not produce significantly biased normalized cloud forcings at small values of cloud liquid water path (Figure 5.4b), unlike the first one. Moreover, the variation of liquid water path with cloud forcing (Figure 5.7) closely resembles the climatology (Figure 5.1b). The transmission-based retrieval results are marginally better than the three-equation parameterization with a skill score of 0.06 when compared with this parameterization. Nonetheless, both parameterizations should be useful as baselines for testing sophisticated stratus cloud radiative transfer and
Fig. 5.6. Plot of regression coefficients (a) $a_{10}$, (b) $a_{30}$ and (c) $a_{31}$ for different zenith angles. Regression coefficients $a_{30}$ are fit linearly while $a_{10}$ and $a_{31}$ are fit quadratically as a function of the cosine of the solar zenith angle.
Fig. 5.7. Scatter plot of normalized cloud forcing against liquid water path using the 3-equation parameterization (Equations 5.6, 5.7 and 5.8).
cloud parameterizations for performance improvements. Researchers looking for quick estimates of surface solar radiation in the presence of low clouds might also find these parameterizations to be of interest.
Chapter 6

The Independent-Column Approximation

All of the radiative transfer calculations of downwelling shortwave irradiance in this study implicitly assumed the independent-column approximation (ICA). The independent-column approximation states that first computing the fluxes and heating rates column-by-column across a domain and then averaging the results leads to accurate estimates of the domain-averaged fluxes and heating rates (Cahalan et al., 1994a). Results from a number of studies since the Cahalan et al. (1994a) study have confirmed that ICA methods produce reasonably accurate domain-averaged quantities. For example, Marshak et al. (1998) demonstrated that for overcast cases ICA calculations lead to the same results derived from calculations using a Monte Carlo method.

To date ICA radiative transfer has been applied to three-dimensional spatial distributions of cloud at a fixed time but not to estimating the time-averaged fluxes and heating rates within a single column using time series of cloud properties from the column. However, if time series of remotely sensed atmospheric variables within a column represent the spatial inhomogeneities across many columns, temporal averaging can be treated as synonymous with spatial averaging. We now assess if the temporal averaging that we applied to the cloud liquid water paths before inserting them into the RAPRAD
model computations leads to differences from the corresponding time-averaged observations that are consistent with ICA theory when applied to spatial distributions of cloud.

To understand why different cloud liquid water path averaging schemes might impact calculations of shortwave irradiance consider cloud reflectance and transmittance defined as

$$R = \frac{F^\uparrow}{F^\downarrow}$$

and

$$T = \frac{F^\downarrow_b}{F^\downarrow_t},$$

respectively, where $F$ is shortwave irradiance, $t$ and $b$ represent locations at the top and bottom of the cloud, and $\uparrow$ and $\downarrow$ represent upwelling and downwelling shortwave irradiances. As it turns out, $R$ is a convex-upward function of cloud liquid water path, while $T$ is a concave-upward function of cloud liquid water path. From Jensen’s inequality (Royden, 1968) if $f$ is a convex function ($f'' \leq 0$) on an interval $I$, $x_1, x_2, \ldots, x_n$ lie within $I$ and $0 \leq t_1, t_2, \ldots, t_n \leq 1$, then

$$f\left(\sum_{i=1}^{n} t_i x_i\right) \geq \sum_{i=1}^{n} t_i f(x_i)$$

(6.3)

when

$$\sum_{i=1}^{n} t_i = 1.$$

The inequality is reversed in Equation 6.3 when the function is concave ($f'' \geq 0$). So, in theory, the averaging schemes that we apply to the cloud liquid water paths before
computing the shortwave irradiances may introduce biases into our comparisons with observations.

6.1 Radiative Transfer Input Averaging Schemes

In the normalized cloud forcing studies of Chapter 4 the observed forcings were at one-minute resolution while we averaged the 20-second liquid water path observations to one-minute resolution before applying the RAPRAD model to them. The question now becomes whether or not we significantly biased our radiative transfer model results by first averaging the liquid water paths. We now consider three different approaches to averaging cloud liquid water paths before using them in RAPRAD model calculations and we evaluate their performance in terms of reproducing the observed normalized cloud forcing.

6.1.1 Liquid Water Path Averaging Scheme (LWPAS)

This scheme involves averaging all available cloud liquid water paths over the selected time interval prior to computing fluxes using the radiative transfer model. If there are \( N_l \) measurements over an averaging interval, we compute the average liquid water path in this scheme as

\[
\langle LWP \rangle = \frac{1}{N_l} \sum_{i=1}^{N_l} LWP_i.
\] (6.4)
Note that the cloud liquid water path is evenly distributed over the averaging domain. Measurements lying within the time period with no cloud liquid water path, representative of clear-sky gaps in the cloud, are included in the average with a value of zero. In this scheme the shortwave irradiances are computed with the average cloud liquid water path given by Equation 6.4. This scheme is the one that we used in the calculations of Chapter 4.

6.1.2 Plane-Parallel Homogeneous Averaging Scheme (PPHAS)

The difference between the LWPAS and PPHAS schemes is that the PPHAS scheme first partitions measurements across the averaging interval into clear and cloudy-sky samples and then averages only the cloud liquid water paths in the cloudy-sky samples. If there are $N_{l,cld}$ positive liquid water path measurements among the $N_l$ measurements across the averaging interval, the average liquid water path is defined as

$$\langle LWP \rangle_{cld} = \frac{1}{N_{l,cld}} \sum_{i=1}^{N_{l,cld}} LWP_i. \quad (6.5)$$

The corresponding cloud fraction is defined as

$$CF = \frac{N_{l,cld}}{N_l}. \quad (6.6)$$

and the shortwave irradiances are computed as the cloud fraction weighted average

$$F_{pph} = CF \times F_{cld} + (1 - CF) \times F_{clr} \quad (6.7)$$
where \( F_{\text{cld}} \) and \( F_{\text{clr}} \) represent the modeled shortwave irradiances for a cloud having a liquid water path given by Equation 6.5 and clear sky, respectively. For cloud fractions \( CF \) of unity, i.e., overcast, the LWPAS and PPHAS averaging schemes are equivalent.

### 6.1.3 Independent-Column Approximation Averaging Scheme (ICAAS)

As we mentioned above, the ICAAS is an equal weighted average of individual irradiances computed with the highest resolution cloud liquid water path measurements that are available, which in this case are at 20-second resolution. The shortwave irradiances in this scheme are computed as

\[
F_{\text{ica}} = \frac{1}{N_1} \sum_{i=1}^{N_1} F(\text{LWP}_i),
\]

(6.8)

where \( F(\text{LWP}_i) \) is the shortwave irradiance computed with the high resolution liquid water path \( \text{LWP}_i \). For clear-sky periods \( \text{LWP}_i \) has a value of zero.

### 6.2 Comparisons of Model Calculations to Observations

The liquid water path data from the ARM SGP site have a resolution of 20 seconds. For comparisons against the observations we first computed the shortwave irradiances for each 20-second interval during our selected periods of stratus using the 20-second liquid water path data together with an effective radius of 7.5 \( \mu \text{m} \) for the cloud droplets. This procedure is tantamount to the ICAAS with 20-second resolution data. We subsequently calculated the shortwave irradiances using the LWPAS and PPHAS methods for one-minute, ten-minute and sixty-minute intervals. We averaged the
Table 6.1. Comparison of reflectance, transmittance and normalized cloud forcing for stratus cases using LWPAS, PPHAS and ICAAS for a sixty-minute averaging period.

<table>
<thead>
<tr>
<th>Cloud Fraction</th>
<th>Average Reflectance</th>
<th>Average Transmittance</th>
<th>Cloud Forcing</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>&lt; 0.95</td>
<td>≥ 0.95</td>
<td>&gt; 0</td>
</tr>
<tr>
<td>No. of points</td>
<td>52</td>
<td>84</td>
<td>136</td>
</tr>
<tr>
<td>LWPAS</td>
<td>0.412</td>
<td>0.569</td>
<td>0.509</td>
</tr>
<tr>
<td>PPHAS</td>
<td>0.378</td>
<td>0.568</td>
<td>0.495</td>
</tr>
<tr>
<td>ICAAS</td>
<td>0.363</td>
<td>0.560</td>
<td>0.485</td>
</tr>
</tbody>
</table>

Table 6.2. Comparison of reflectance, transmittance and normalized cloud forcing for stratus cases using LWPAS, PPHAS and ICAAS for a ten-minute averaging period.

<table>
<thead>
<tr>
<th>Cloud Fraction</th>
<th>Average Reflectance</th>
<th>Average Transmittance</th>
<th>Cloud Forcing</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>&lt; 0.95</td>
<td>≥ 0.95</td>
<td>&gt; 0</td>
</tr>
<tr>
<td>No. of points</td>
<td>143</td>
<td>595</td>
<td>738</td>
</tr>
<tr>
<td>LWPAS</td>
<td>0.346</td>
<td>0.549</td>
<td>0.510</td>
</tr>
<tr>
<td>PPHAS</td>
<td>0.320</td>
<td>0.549</td>
<td>0.504</td>
</tr>
<tr>
<td>ICAAS</td>
<td>0.318</td>
<td>0.544</td>
<td>0.500</td>
</tr>
</tbody>
</table>
Table 6.3. Comparison of reflectance, transmittance and normalized cloud forcing for stratus cases using LWPAS, PPHAS and ICAAS for a one-minute averaging period.

<table>
<thead>
<tr>
<th></th>
<th>Average Reflectance</th>
<th>Average Transmittance</th>
<th>Cloud Forcing</th>
</tr>
</thead>
<tbody>
<tr>
<td>Cloud Fraction</td>
<td>&lt; 0.95</td>
<td>≥ 0.95</td>
<td>&gt; 0</td>
</tr>
<tr>
<td>Cloud Fraction</td>
<td>&lt; 0.95</td>
<td>≥ 0.95</td>
<td>&gt; 0</td>
</tr>
<tr>
<td>No. of points</td>
<td>176</td>
<td>6202</td>
<td>6378</td>
</tr>
<tr>
<td>LWPAS</td>
<td>0.316</td>
<td>0.535</td>
<td>0.528</td>
</tr>
<tr>
<td>LWPAS</td>
<td>0.549</td>
<td>0.268</td>
<td>0.276</td>
</tr>
<tr>
<td>LWPAS</td>
<td>-0.22</td>
<td>-0.63</td>
<td>-0.61</td>
</tr>
<tr>
<td>PPHAS</td>
<td>0.285</td>
<td>0.535</td>
<td>0.528</td>
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<tr>
<td>PPHAS</td>
<td>0.587</td>
<td>0.268</td>
<td>0.277</td>
</tr>
<tr>
<td>PPHAS</td>
<td>-0.16</td>
<td>-0.63</td>
<td>-0.61</td>
</tr>
<tr>
<td>ICAAS</td>
<td>0.274</td>
<td>0.532</td>
<td>0.525</td>
</tr>
<tr>
<td>ICAAS</td>
<td>0.601</td>
<td>0.272</td>
<td>0.281</td>
</tr>
<tr>
<td>ICAAS</td>
<td>-0.14</td>
<td>-0.62</td>
<td>-0.61</td>
</tr>
</tbody>
</table>

One-minute observations and 20-second ICAAS modeled irradiances to these degraded resolutions and then compared the resulting reflectances, transmittances and normalized cloud forcings to their LWPAS and PPHAS equivalents. From Jensen’s equality and our definitions we expect

\[ R_{\text{icaas}} \leq R_{\text{pphas}} \leq R_{\text{lwpas}} \]  

(6.9)

and

\[ T_{\text{icaas}} \geq T_{\text{pphas}} \geq T_{\text{lwpas}}, \]  

(6.10)

where the subscripts represent the cloud liquid water path averaging scheme that was applied to the data before the RAPRAD model computations. We also expect the observations to fall near \( R_{\text{icaas}} \) and \( T_{\text{icaas}} \), but with small differences with \( R_{\text{lwpas}} \) and \( T_{\text{lwpas}} \) for the one-minute resolution comparisons. Since normalized cloud forcing is
analogous to transmittance, we also expect the normalized cloud forcing to follow the
trends in Equation 6.10.

We begin with the sixty-minute comparisons of R, T and NCF for overcast, broken
cloud cover, and all cases combined (Figure 6.1; Table 6.1). For broken cloud cover the
reflectance from the LWPAS is 0.412, which is around 9 % and 14 % higher than from the
PPHAS and ICAAS methods, respectively. Changes in the reflectance for overcast cases
are extremely small, with almost no difference between LWPAS and PPHAS and just
over a 1 % change between these two schemes and ICAAS. Considering transmittance
(Figure 6.1) for broken cloud cover the ICAAS results are higher than the PPHAS by ap-
proximately 10 %, while the PPHAS results are, in turn, higher than the LWPAS results
by about 5 % . As for reflectance, there is almost no difference in the transmittances
generated by the three averaging schemes for overcast conditions.

Comparisons of the averaging schemes using the ten-minute data are summarized
in Table 6.2 and Figure 6.3 while results for the one-minute comparisons are presented in
Table 6.3 and Figure 6.4. Inspecting Tables 6.2 and Table 6.3, we find that the trends for
these two averaging periods are similar to those for the sixty-minute averaging period,
but the changes are much smaller. For overcast conditions and one-minute averaged data
all three averaging schemes yield almost identical results. This conclusion is reinforced by
the histograms that are presented in Figure 6.4 as the histograms for the three different
averaging schemes are nearly identical.

Considered together, these comparison studies produced results that are consis-
tent with Jensen’s inequality and earlier ICA studies on inhomogeneous cloud fields.
Fig. 6.1. Reflectance (panels a, b and c) and transmittance (panels d, e and f) of stratus cases with different cloud fractions for a time resolution of 60 minutes using LWPAS, PPHAS and ICAAS.
Fig. 6.2. Normalized cloud forcing of stratus cases with (a) cloud fraction above 0.95, (b) below 0.95 and (c) all cases for a time resolution of 60 minutes using LWPAS, PPHAS and ICAAS.
Fig. 6.3. (a) Reflectance, (b) transmittance and (c) cloud forcing of stratus cases for all overcast cases with time resolution of ten minutes using LWPAS, PPHAS and ICAAS.
Fig. 6.4. Reflectance (a, b), transmittance (c, d) and cloud forcing (e, f) for 1 minute overcast stratus cases using LWPAS, PPHAS and ICAAS.
For broken cloud fields prior averaging of the cloud liquid water path before the radiative transfer calculations led to biases relative to the ICA averaging scheme and these biases generally increased with the averaging interval. For overcast cloud conditions, while similar trends were evident in the data, they were so small for the one-minute and ten-minute results as to be negligible. Most importantly, for overcast conditions the LWP and ICA averaging schemes yield almost identical results and we conclude that our normalized cloud forcing calculations in Chapter 4 are not significantly biased by averaging the cloud liquid water paths to one-minute resolution.
Chapter 7

Summary and Conclusions

The importance of warm boundary layer clouds to the radiation budget of the Earth led us to build a climatology of their properties, including cloud base heights, liquid water paths and normalized cloud forcings, and to assess their impact on solar radiation. With continuous data from the ARM SGP site we built a dataset consisting of over 3000 hours of data at 20-second resolution over the period of four years from 1997 through 2000. As it turns out, this dataset is equally divided between daytime and nighttime boundary layer cloud cases. Cloud drop effective radii and visible cloud optical depth data from radar-derived and narrowband transmission based retrievals were available for over 80 hours of daytime, overcast boundary layer clouds at 1-minute resolution between January 1997 and January 1998. We statistically analyzed the cloud base heights and liquid water paths, two significant parameters of stratus clouds, for diurnal, interannual and seasonal variations, while we limited our analysis of the normalized cloud forcing data to interannual and seasonal variations.

We found the cloud liquid water paths in the dataset to be positively skewed with a mean of 0.14 mm, a standard deviation of 0.18 mm and a median of 0.067 mm. Daytime liquid water paths were generally lower than nighttime values by approximately 0.035 mm, with the differences being larger in spring and summer than fall and winter. Warm boundary layer clouds that occurred during the summer months, however, had
the smallest liquid water paths. There was little variation in the cloud liquid water paths from one year to the next, except for the March, April and May season of 2000, which had the largest liquid water paths and lowest cloud bases in the dataset.

We found the cloud base heights to lie mostly in the lowest kilometer of the atmosphere, except for the summer months when the cloud base heights were more evenly distributed across the lowest two kilometers of the atmosphere. The differences in the summer month cloud base heights from the rest of the year led to a slight bimodality in the overall cloud base height distribution. The winter months of December, January and February consistently had the lowest cloud base heights from one year to the next.

The normalized cloud forcing statistics were correlated with the liquid water path statistics. Consequently, the normalized cloud forcing did not exhibit any interannual trends and the occurrence of small magnitude forcings was at a maximum during the summer months. The magnitudes of the cloud droplet effective radii varied depending on the instruments used for their retrieval. The narrowband transmission-based retrieval produced a mean effective radius of 7.39 µm with a standard deviation of 2.37 µm, while the radar-based retrieval produced a mean and standard deviation of 6.34 µm and 2.33 µm, respectively. The standard deviations of the cloud drop effective radii for both retrievals matched a climatological standard deviation of around 2 µm as reported by Miles et al. (2000). Overall, the climatology suggested that the warm boundary layer cloud properties were stable from one year to the next, so that results based on any one year would most likely be representative of the entire four-year period.

The primary purpose of our study was to assess the sensitivity of downwelling shortwave irradiance at the surface to variations in cloud droplet effective radii and
cloud liquid water paths of warm boundary layer clouds. To this end we identified 19 daytime, overcast boundary layer cloud cases consisting of over 80 hours of 1-minute resolution data. Using overcast cloud cases in this part of the study was important to eliminating three-dimensional radiative transfer effects that might contaminate the model computations based on changing drop sizes and cloud liquid water paths. The mean and median liquid water paths for the 19 case study periods were 0.116 mm and 0.096 mm, which are slightly higher than the values for the 4-year dataset. Furthermore, for these overcast cloud cases the modal value increased from 0.0-0.02 mm up to approximately 0.04-0.06 mm, suggesting that the lowest liquid water path clouds are associated with broken cloud in the boundary layer.

Using a climatological cloud drop effective radius of 7.5 µm, as well as cloud drop effective radii produced by the Kato et al. (2001) and Min and Harrison (1996) retrieval algorithms, we assessed the impact of these 19 overcast cloud cases on the downwelling shortwave irradiance at the surface. We found that the normalized cloud forcing at the surface varies nonlinearly with cloud liquid water path and is most sensitive to changes in liquid water path for values from 0.00-0.10 mm, which is the range of values that most often occurs at the ARM SGP site. The narrowband transmission-based effective radii (Min and Harrison, 1996) provided the best estimates of surface irradiance, but the climatological effective radius of 7.5 µm was nearly as skilled in estimating the downwelling shortwave irradiance at the surface. Taken together, these results provide evidence that the indirect aerosol effect proposed by Twomey (1984) will be a phenomenon hard to detect, being overwhelmed by small variations in cloud liquid water paths.
Investigating the apparent insensitivity of downwelling shortwave irradiance at the surface to variations in effective radius, we found that stratus cloud optical depth is 3 times more sensitive and downwelling shortwave irradiance at the surface is 6 times more sensitive to liquid water path changes than to cloud particle size changes of the same percentage magnitude. On a side note, we did not find any bias in the downwelling shortwave irradiances at the surface using the transmission-derived radii as RAPRAD model inputs. As the transmission-derived effective radii were retrieved using a single wavelength at a non-absorbing wavelength and our model calculations were for broadband shortwave irradiance, we did not find any evidence for anomalous absorption ((Pilewskie and Valero, 1995); (Valero et al., 1997)) that is purported to occur somewhere across the shortwave but not at the narrowband wavelength that is used in the Min and Harrison (1996) retrieval.

Del Genio and Wolf (2000) showed that a bias in microwave radiometer cloud liquid water path retrievals yielded positive values even in the absence of clouds. We attempted to remove this bias by implementing a new microwave radiometer retrieval of cloud liquid water path that is more robust. Nonetheless, the uncertainty in the low cloud liquid water path values motivated us to further develop an alternative infrared-window cloud emission based retrieval of cloud liquid water path. This alternative retrieval was only applicable for clouds with emissivities less than 1, implying cloud liquid water paths lower than approximately 0.05 mm. While this new approach is complementary to existing microwave radiometer retrievals of cloud liquid water path, comparisons of low cloud liquid water paths retrieved by the two different approaches were inconclusive.
Overall, we were unable to demonstrate conclusively the accuracy of either retrieval at low cloud liquid water path values.

All of our radiative transfer calculations in this study implicitly assumed that the independent column approximation to radiative transfer theory was applicable to our high temporal resolution time series of boundary layer cloud properties over the ARM SGP site. Furthermore, we assumed that averaging overcast cloud properties to 1-minute resolution did not introduce any bias into the radiative transfer calculations relative to the observations at this resolution. We demonstrated that these assumptions were valid for overcast continental boundary layer cloud conditions, but began to fail for broken cloud conditions. Since our radiative transfer studies were based on 1-minute data during periods of overcast boundary layer cloud conditions, we concluded that our averaging scheme did not introduce any significant biases into the study.

We developed a parameterization for normalized cloud forcing using both a simple exponential model and a combination of power law and exponential models that depended only on the solar zenith angle and cloud liquid water path. These parameterizations were found to perform nearly as well as the detailed radiative transfer model calculations with narrowband transmission-based effective radii as inputs. We proposed these parameterizations as baselines for testing sophisticated schemes for estimating downwelling shortwave irradiances, as well as a means for providing quick estimates of downwelling shortwave irradiance for researchers who need results without running complicated radiative transfer models.
Appendix A

Mean Square Error and Skill Score

The mean square error of a forecast is defined as

\[ \text{MSE}_{f,o} = \frac{1}{N} \sum_{i=1}^{N} (F_i - O_i)^2 \]  

(A.1)

with \( N \) being the number of observations and \( F \) and \( O \) denoting the forecast and observation data, respectively. The correlation coefficient between the forecast and observations using the same notation can be written as

\[
R_{f,o} = \frac{\text{Cov}_{f,o}}{\sqrt{\frac{1}{N} \sum_{i=1}^{N} (F_i - \bar{F})^2} \sqrt{\frac{1}{N} \sum_{i=1}^{N} (O_i - \bar{O})^2}}.
\]

(A.2)

where \( \bar{F} \) and \( \bar{O} \) are the means of the forecast and observed data, respectively. As seen from equation A.2, the correlation coefficient can also be represented as

\[
R_{f,o} = \text{Cov}_{f,o}(\sigma_f \sigma_o)^{-1}.
\]

(A.3)
where $Cov_{f,o}$ is the covariance of the forecast and observations and $\sigma_f$ and $\sigma_o$ the standard deviations of the forecast and observations, respectively. Expansion of equation A.1 and use of equation A.3 gives us

$$MSE_{f,o} = (\bar{F} - \bar{O})^2 + (1 - R_{f,o})(\sigma_f^2 + \sigma_o^2) + R_{f,o}(\sigma_f - \sigma_o),$$

where the three terms are useful in determining the sources of error.

Having defined the mean square error in terms of the forecast and observed variables, we need to determine whether our forecast has any skill given that we have a value for $MSE_{f,o}$. This comparison is done with a control, which is generally a climatological value of some sort. The skill score is then defined as

$$SS = \frac{(MSE_{f,o} - MSE_{c,o})}{(0 - MSE_{c,o})} = 1 - \frac{MSE_{f,o}}{MSE_{c,o}},$$

where $MSE_{c,o}$ is the mean square error of the control when compared to the observations. As can be seen from equation A.5, a positive value reflects skill in the forecast. Negative values imply that our forecast is worse than the climatological value, implying the need for improvement.
Appendix B

Infrared Radiance Approach to Cloud Liquid Water Path Retrieval

In Chapter 3 we mentioned that the uncertainty in cloud liquid water paths derived from ARM dual-channel microwave radiometers was around 0.02 mm. As a large number of retrieved cloud liquid water path values at the ARM SGP site fall around 0.02 mm, alternative methods for retrieving small amounts of cloud liquid water path need to be considered. One alternative approach is to use the emissivity of clouds in the infrared window region to deduce the cloud liquid water path (Paltridge, 1974). This technique is applicable only when the emissivity of the cloud is less than unity, which corresponds to liquid water paths less than about 0.05 mm. This chapter presents a development of the infrared radiance retrieval technique and presents some results from it. As there is currently no independent means of assessing the quality of these new retrievals, we present comparisons of them with liquid water paths derived from the microwave radiometers and show where there are consistencies and inconsistencies between the two retrievals.

B.1 Theory of LWP Retrieval Using Infrared Radiances

The emissivity of a cloud layer at an absorbing wavelength is defined as

\[ \epsilon_{\text{cld}} = 1 - e^{-\tau_a}, \]  

(B.1)
where $\tau_a$ is the absorption optical depth of the cloud layer. The absorption optical depth is defined as

$$\tau_a = \int_0^\infty \sigma_a(r)n(r)dr\Delta z \quad \text{(B.2)}$$

where $\sigma_a(r)$ is the size-dependent absorption cross section of the cloud drops at the wavelength of interest, $n(r)$ is the cloud drop size distribution and $\Delta z$ is the thickness of the cloud. During cloudy conditions, the downwelling infrared radiance at the surface has contributions from the cloud, the atmosphere above and below the cloud, and surface emitted radiation that is reflected by the cloud back to the surface, which we currently neglect.

With the assumptions above an expression for the downwelling infrared radiance at the surface is

$$I_m = t_b (1 - \epsilon_{cld}) I_{cld,t} + t_b \epsilon_{cld} I_{cld,bb} + I_{cld,b} \quad \text{(B.3)}$$

where $I_{cld,t}$ is the downwelling radiance at cloud top due to the atmosphere above the cloud, $I_{cld,b}$ is the downwelling radiance at the surface as a result of atmospheric emission between the surface and cloud base, $I_{cld,bb}$ is the blackbody radiance at mid-cloud temperature, and $t_b$ is the atmospheric transmissivity between the surface and cloud base. As we are dealing with thin clouds with small vertical extents, the clear-sky radiance can be approximated as

$$I_{\text{clear}} = t_b I_{cld,t} + I_{cld,b} \quad \text{(B.4)}$$
Using Equations B.3 and B.4, the cloud emissivity can alternately be represented as

\[ \epsilon_{\text{cld}} = \frac{I_m - I_{\text{clear}}}{t_b (I_{\text{cld, bb}} - I_{\text{cld, t}})} . \]  

(B.5)

For a lognormal cloud droplet size distribution, the number concentration of cloud droplets \( N \) becomes

\[ N = \frac{\text{LWC}}{\rho_l \left( \frac{4}{3} \pi r_e^3 \exp \left( -3 \sigma_{\log}^2 \right) \right)} . \]  

(B.6)

where \( \text{LWC} \) is the cloud liquid water content, \( r_e \) is the effective radius of the cloud drop distribution, \( \sigma_{\log} \) is the width of the cloud drop distribution and \( \rho_l \) is the density of water. Moreover, the absorption optical depth associated with the lognormal distribution of cloud drops becomes

\[ \tau_a = N \left[ \int_0^\infty \frac{\sigma_a(r)}{\sqrt{2\pi}\sigma_{\log}r} \exp \left\{ \frac{-[\ln(r/r_n, \log)]^2}{2\sigma_{\log}^2} \right\} dr \right] \Delta z . \]  

(B.7)

Representing the quantity in brackets by \( \sigma_a \), we can write the liquid water path, which is the product of cloud liquid water content and cloud thickness, as

\[ \text{LWP} = \frac{\ln(1/e_{\text{cld}} \rho_l (\frac{4}{3}) \pi r_e^3 \exp(-3\sigma_{\log}^2))}{\sigma_a } . \]  

(B.8)

Therefore, given estimates of \( r_e, \sigma_{\log}, \epsilon_{\text{cld}} \) and \( \sigma_a \), one can estimate \( \text{LWP} \).

Using refractive indices from Downing and Williams (1975) and the Toon and Ackerman (1981) Mie code, we computed \( \sigma_a \) for a range of cloud drop effective radii (Figure B.1). Since we kept the liquid water path fixed in the calculations for Figure B.1,
Fig. B.1. Variation of absorption coefficient $\sigma_a$ and number of drops with effective radius for a liquid water content of 0.125 g/m$^3$ at a wavelength of 9.5 $\mu$m. A lognormal distribution with a width of 0.38 is used in the calculations.

Fig. B.2. Variation of absorption optical depth $\tau_a$ with effective radius for a liquid water content of 0.125 g/m$^3$, cloud depth of 200 m (LWP = 0.025 mm) at a wavelength of 9.5 $\mu$m. A lognormal distribution with a width of 0.38 is used in the calculations.
Fig. B.3. Variation of emissivity $\epsilon_{\text{cll}}$ with effective radius for a liquid water content of 0.125 g/m$^3$, cloud depth of 200 m (LWP = 0.025 mm) at a wavelength of 9.5 $\mu$m. A lognormal distribution with a width of 0.38 is used in the calculations.

Fig. B.4. Variation of absorption optical depth $\tau_a$ with effective radius for various fixed LWP at a wavelength of 9.5 $\mu$m. A lognormal distribution with a width of 0.38 is used in the calculations.
Fig. B.5. Variation of emissivity $\epsilon_{\text{cld}}$ with LWP for various fixed effective radii at a wavelength of 9.5 $\mu$m. A lognormal distribution with a width of 0.38 is used in the calculations. The green band is representative of the values derived in chapter 4.
as the cloud drop effective radius increased, which led to increases in $\sigma_a$, the drop concentration $N$ decreased. As a result, the absorption optical depth, which depends on the product of $\sigma_a$ and $N$, first increased with increasing cloud drop effective radius (up to values of about 3 $\mu$m) and decreased thereafter (Figure B.2). Since cloud emissivity depends upon absorption optical depth, for fixed cloud liquid water path the cloud emissivity will first increase and then decrease with increasing cloud drop effective radius (Figure B.3).

With an estimate of the emissivity the liquid water path LWP follows directly from Equation B.8. As Figure B.4 illustrates, at small values of liquid water path the retrieval is not that sensitive to errors in the cloud drop effective radius. This point is further demonstrated in Figure B.5.

### B.2 Infrared Radiance Data at the ARM SGP Site

The best type of infrared radiance data for retrieving cloud liquid water paths would have both high accuracy and high temporal resolution. Unfortunately, these two aspects of infrared radiance data do not go hand-in-hand; rather, as the temporal resolution of the radiance increases, the accuracy of the radiance decreases. The ARM SGP site infrared thermometer data have a relatively high 20-second temporal resolution, but these data are reported as brightness temperatures for a relatively wide spectral band from 9.6 to 11.5 $\mu$m with an accuracy of 0.5 K. The conversion of the measured brightness temperature to an average radiance across the band is not straightforward and produces errors of unknown magnitude. Furthermore, the infrared thermometer cannot measure accurately temperatures below about 223 K, which leads to missing
values in clear, dry atmospheres. The Atmospheric Emitted Radiance Interferometer
(AERI) measures radiance at a high spectral resolution of 1 cm\(^{-1}\) with an accuracy of
around 1 mW/(m\(^2\)str) in our region of interest. However, it produces only one 2-minute
resolution radiance value every 10 minutes. To overcome the limitations of the two
instruments we used the AERI spectral radiances to calibrate the infrared thermometer
data at half-hourly intervals. Once the high temporal resolution thermometer data were
calibrated, we used these data in the retrieval.

Table B.1. List of days and the number of points in each day for which the retrieval is
performed.

<table>
<thead>
<tr>
<th>No.</th>
<th>Date</th>
<th>No. of points</th>
</tr>
</thead>
<tbody>
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<td>1</td>
<td>March 19 1997</td>
<td>692</td>
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B.3 Implementation of LWP Retrieval Using Infrared Radiances

In application of the retrieval to data we assumed that \( r_e, \sigma \log \) and \( \sigma_a \) were known and equal to 7.5 \( \mu m \), 0.38 and 66.39 \( \mu m^2 \), respectively. We then used the measured radiance to estimate \( \epsilon_{cld} \). However, estimation of the cloud emissivity in Equation B.5 required an estimate of the clear-sky radiance \( I_{\text{clear}} \), as well as estimates of the above cloud radiance \( I_{\text{cld},t} \), mid-cloud blackbody radiance \( I_{\text{cld},bb} \) and below cloud transmittance \( t_b \). To estimate the clear-sky radiance we ran the MODTRAN 4 radiative transfer code (Berk et al., 1989) to produce radiances at 1 cm\(^{-1}\) resolution from 870 cm\(^{-1}\) to 1041 cm\(^{-1}\), the wavelength range of the infrared thermometer. The inputs to the MODTRAN 4 radiative transfer code were atmospheric profiles of temperature, pressure and water vapor mixing ratio (Chapter 3), where the water vapor mixing ratios had been scaled to produce integrated amounts of vapor in agreement with the microwave radiometer retrievals. We subsequently averaged the MODTRAN 4 radiances to produce an estimate of the radiance that the infrared thermometer would measure if the skies were cloud free. We also used the clear-sky MODTRAN 4 radiative transfer code simulations to estimate the below cloud transmittance \( t_b \) and the above cloud radiance \( I_{\text{cld},t} \). The mid-cloud blackbody radiance \( I_{\text{cld},bb} \) is calculated from the Planck function using the radar and lidar observations of the cloud boundaries to determine the temperature.

Note that accurate water vapor profiles are a prerequisite for accurate radiance calculations in the atmospheric infrared window. A dry bias sometimes occurred in the microwave radiometer retrieved water vapor which translated into an inability to
Fig. B.6. Liquid water path from microwave radiometer and infrared radiance retrievals as a function of cloud emissivity at 20-s resolution. The dates are in YYYYMMDD format.
match measured and MODTRAN 4 computed clear-sky radiances. We used these clear-sky offsets to adjust the modeled radiances that occurred during neighboring clear and cloudy periods.

Using Equation B.8 to solve for LWP becomes problematic when the cloud emissivity $\epsilon_{cld}$ approaches unity. Although we applied the retrievals up to emissivities of 0.98, cloud emissivities above about 0.8 lead to relatively large uncertainties in retrieved cloud liquid water path. Also, only liquid water paths below 0.06 mm from the microwave radiometer are considered and the retrieval procedure in the previous section applied to a few days on which thin single layer stratus is present. These days are part of the days considered in Chapter 4 and the subset is listed in Table B.1. As Figure B.6 shows, the infrared radiance retrieval of liquid water path is comparable to the liquid water path from the microwave radiometer. While May 26, 1997, and September 05, 1997, show distinct biases, between the two sets of liquid water paths, but in opposite directions, the rest of the days show uniform scatter of varying degrees with March 19, 1997, having the smallest disagreement. As we cannot independently verify whether there is any improvement in liquid water path when using the infrared technique we can only say that this is a viable technique usable in the presence of low liquid water path clouds. As the infrared thermometer is less sensitive than the AERI, availability of high resolution AERI data in the infrared window region would reduce uncertainties in the retrieval.
References


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Vita

Manajit Sengupta was born in Calcutta, India in 1966. He got his B.S. with Honors in Physics from Presidency College, Calcutta in 1989. He then worked in the Revenue Services of the Indian Government before he quit job to come to the US in 1996 to pursue a Ph.D. in Meteorology. His field of specialization is clouds and atmospheric radiation. His current interests are in remote sensing of the properties of stratus clouds, their impact on solar radiation and their representation in climate models.