The Impact of Arctic Cloud Water and Ice on Cloud Radiative Forcing during the
Arctic Summer Cloud-Ocean Study in August 2008

by

Elizabeth A. Maroon

Submitted to the Department of Earth, Atmospheric and Planetary Sciences
in Partial Fulfillment of the Requirements for the Degree of
Bachelor of Science in Earth, Atmospheric and Planetary Sciences
at the Massachusetts Institute of Technology

May 7, 2010 [June 2010]

Copyright 2010 Elizabeth A. Maroon. All rights reserved.

The author hereby grants to MIT permission to reproduce and to
distribute publicly paper and electronic copies of this thesis document in whole or in part
in any medium now known or hereafter created.

Signature redacted
Department of Earth, Atmospheric and Planetary Sciences
May 7, 2010

Kerry Emanuel
Thesis Supervisor

Signature redacted
Samuel Bowring
Chair, Committee on Undergraduate Program
The Impact of Arctic Cloud Water and Ice on Cloud Radiative Forcing during the Arctic Summer Cloud-Ocean Study in August 2008

Elizabeth Maroon

May 7, 2010
I would like to thank my mentors Dr. Amy Solomon, Dr. Matthew Shupe and Dr. Ola Persson at the NOAA Earth Systems Research Lab in Boulder, Colorado for all their guidance. I greatly appreciate all the time they put into helping me. I would also like to thank my senior thesis advisor Professor Kerry Emanuel for his help with this project and in writing this paper. Thank you also to Jane Conner for her support and ever-useful writing advice.

I also appreciate all the computing help from Linda Meinke and Scott Blomquist while trying to set-up various radiative transfer models on the 16th floor computers. Thanks also to my Boulder officemate Cassie Wheeler for sharing her space (and much research advice) with me during Summer 2009 and January 2010, as well as to the many PAOC graduate students (Brian Tang, Neil Zimmerman, Allison Wing, Morgan O’Neill, Michael Byrne...) who answered my many questions, helped me with compiling issues, and occasionally lent me a place to work on their couch.

I would also like to thank the NOAA’s Office of Education, who sponsored my Hollings Scholarship during Summer 2009. Much of the research in this paper was conducted thanks to their generous support.
# Contents

1 Background 6
   1.1 Importance of Studying the Arctic Climate 6
   1.2 Radiative Transfer in the Arctic Atmosphere 7
   1.3 Cloud Microphysics and Aerosol Effects 8
   1.4 Cloud Radiative Effects in the Arctic 9

2 Observation Sources and Radiative Models 12
   2.1 Arctic Summer Ocean-Cloud Study 12
   2.2 The Discrete Ordinate Method and Rapid Radiative Transfer Model 13
   2.3 Case Studies and Sensitivity Studies 14

3 Results 17
   3.1 Summary of Results 17
   3.2 Case Study Results 17
   3.3 Validation against Surface Radiation Observations 20
   3.4 Cloud Forcing as a function of Liquid Water Path 23
   3.5 Liquid and Multi-cloud Water Sensitivity Studies 25
   3.6 Liquid Water Path Partitioning between Multiple Clouds 34
   3.7 Drop Size Studies 35
   3.8 Minimization of LW and SW surface radiation with Liquid Effective Radius 39

4 Discussion of Results 41
   4.1 Notable features during Case Studies 41
   4.2 Effectiveness of RRTM at capturing the day’s trends 43
   4.3 Importance of Cloud forcing versus Liquid Water Path 44
   4.4 Radiative interactions between upper and lower clouds 44
   4.5 Relations of liquid drop size studies to indirect effects 45
   4.6 Discussion of minimization results 46

5 Concluding Remarks 47
### List of Figures

<table>
<thead>
<tr>
<th>Figure</th>
<th>Description</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>ASCOS Trajectory</td>
<td>13</td>
</tr>
<tr>
<td>2</td>
<td>August 25, 2008 Cloud Forcing Case Study</td>
<td>18</td>
</tr>
<tr>
<td>3</td>
<td>August 29, 2008 Cloud Forcing Case Study</td>
<td>19</td>
</tr>
<tr>
<td>4</td>
<td>August 25 Longwave Validation</td>
<td>21</td>
</tr>
<tr>
<td>5</td>
<td>August 25 Shortwave Validation</td>
<td>22</td>
</tr>
<tr>
<td>6</td>
<td>August 29 Longwave Validation</td>
<td>23</td>
</tr>
<tr>
<td>7</td>
<td>August 29 Shortwave Validation</td>
<td>24</td>
</tr>
<tr>
<td>8</td>
<td>Cloud forcing as a function of liquid water path</td>
<td>25</td>
</tr>
<tr>
<td>9</td>
<td>Cloud forcing as a function of liquid water path (individual days)</td>
<td>26</td>
</tr>
<tr>
<td>10</td>
<td>Longwave heating rate cloud forcing with increasing upper cloud liquid water content</td>
<td>28</td>
</tr>
<tr>
<td>11</td>
<td>Shortwave heating rate cloud forcing with increasing upper cloud liquid water content</td>
<td>29</td>
</tr>
<tr>
<td>12</td>
<td>Longwave heating rate cloud forcing with varying liquid water content in lower cloud</td>
<td>30</td>
</tr>
<tr>
<td>13</td>
<td>Shortwave heating rate cloud forcing with varying liquid water content in the lower of two clouds</td>
<td>31</td>
</tr>
<tr>
<td>14</td>
<td>Longwave heating rate cloud forcing as ice content of the upper cloud is increased</td>
<td>32</td>
</tr>
<tr>
<td>15</td>
<td>Shortwave heating rate cloud forcing as ice content of the upper cloud is being phased in</td>
<td>33</td>
</tr>
<tr>
<td>16</td>
<td>Partitioning of liquid water path between two clouds</td>
<td>34</td>
</tr>
<tr>
<td>17</td>
<td>Liquid and ice water content profiles as liquid water content is partitioned between two clouds</td>
<td>35</td>
</tr>
<tr>
<td>18</td>
<td>Longwave net flux differences (cloudy-clear) with varying liquid drop radius</td>
<td>36</td>
</tr>
<tr>
<td>19</td>
<td>Sensitivity of LW net flux differences to changes in drop size</td>
<td>37</td>
</tr>
<tr>
<td>20</td>
<td>Shortwave net flux difference (cloudy-clear) with varying liquid drop radius</td>
<td>38</td>
</tr>
<tr>
<td>21</td>
<td>Sensitivity of SW net flux difference to changes in drop size</td>
<td>39</td>
</tr>
<tr>
<td>22</td>
<td>Comparison of LW downwelling surface radiation before and after minimization</td>
<td>40</td>
</tr>
<tr>
<td>23</td>
<td>Comparison of Shortwave downwelling surface radiation before and after minimization</td>
<td>41</td>
</tr>
<tr>
<td>24</td>
<td>Comparison between original liquid drop radii and minimized RRTM drop size</td>
<td>42</td>
</tr>
</tbody>
</table>
Abstract

The Arctic atmosphere is especially sensitive to changes in climate forcing; however, Arctic processes and feedbacks are not understood well enough to accurately predict how the Arctic environment might change under anthropogenic forcing. Further study of the basic atmospheric processes is needed, especially due to uncertainties in modeling cloud feedbacks. August and September are the months when the Arctic sea surfaces begin to freeze; clouds play an important role in determining when this process begins. In this study, the radiative properties of Arctic stratocumulus are studied by comparing measurements for two days in August 2008 during the Arctic Surface Cloud Ocean Study (ASCOS) with simulations using the Rapid Radiative Transfer Model (RRTM). Cloud radiative forcing for both days is examined, and the modeled radiative fluxes were found to compare well to observations. Sensitivity studies are conducted on single and multi-level stratocumulus clouds to study their radiative interactions with each other. Cloud-top cooling in upper clouds is found to radiatively turn off cloud-top cooling in clouds below it. The RRTM and the surface radiative observations are used together to constrain estimates of liquid droplet radius; constraining these radii shows the sensitivity of shortwave cloud radiative forcing and the insensitivity of long wave cloud forcing to changes in drop size.
1 Background

1.1 Importance of Studying the Arctic Climate

The impacts of global climate change are being seen acutely in the Arctic region (Solomon et al., 2007). The region north of 65°N has seen a temperature change twice that of the global average from 1965-2005 (Lemke et al., 2007). The paleoclimate record shows that cryosphere regions change greatly during periods of significant climate variation on time scales from long ice ages to shorter periods like the Little Ice Age and the Younger Dryas (Lemke et al., 2007). In fact, there was a sensitive response of high-latitudes to the last major period of a sustained warm temperature equilibrium, the mid-Pliocene (3.3-3.0 Mya). CO₂ was estimated to be 360 – 400 ppm, and sea levels were 15 – 25 m higher than current levels. Paleoclimate proxies also show that while the high latitudes were significantly warmer than present, tropical sea and surface temperatures were not significantly different. Although there is uncertainty in the paleoclimate reconstructions, global circulation models have shown the same sensitivity in the high latitudes relative to the tropics Jansen et al. (2007).

Studying the paleoclimate can provide a context for our current changing climate; already the present period of anthropogenically-forced heating is showing profound changes in the Arctic. Today, decreasing sea ice is related to changes in the surface air and ocean temperature, which varies due to natural variability, as well as with anthropogenic forcing. Serreze et al. (2007) argue that observations and models both point to an Arctic that will be free of sea-ice as the global atmosphere and ocean warm, and that the remaining questions relate to the time frame of these changes and its impacts on the system. In 2007, due to an anomalous (but not unprecedented) atmospheric circulation pattern, record sea ice melt created a Northwest Passage for the first time since records began. Based on observations from NASA's A-train satellite platform Kay et al. (2008) showed that anomalous cloud reductions during 2007 were associated with increases in the downwelling radiation. This increase is linked to greater sea ice melt, which can accelerate when the ocean becomes ice-free due to ice-albedo feedback. Kay et al. (2008) suggest that the effect of clouds and downwelling radiation may play an increasing role in determining sea ice extent. It is clear that more study of the mechanics of the Arctic system, from clouds to sea ice, is necessary to understand the coming changes to this environment and to the rest of the planet.
1.2 Radiative Transfer in the Arctic Atmosphere

The transfer of heat in the Arctic atmosphere affects the surface and the sea ice. Radiation from an incoming source (in our case, solar insolation) is either absorbed or reflected. A blackbody in thermal equilibrium at a specific temperature will produce energy fluxes \( F \) according to the Stefan-Boltzmann Law,

\[
F = \sigma T^4, \tag{1}
\]

where \( T \) is temperature and \( \sigma \) is the Stefan-Boltzmann constant \( (5.67 \times 10^{-8} \text{W/m}^2) \). All incoming radiation of a particular wavelength is absorbed and then emitted. Known as Kirchoff's Law, this can be summarized as

\[
A_\lambda = \epsilon_\lambda = 1, \tag{2}
\]

for all wavelengths, where \( A_\lambda \) is the absorptivity, a dimensionless quantity that describes the fraction of incident radiation that is absorbed, and \( \epsilon_\lambda \) is the emissivity, the fraction of radiation that is emitted. Grey bodies exhibit imperfect absorption and emission: \( A_\lambda = \epsilon_\lambda < 1 \) (Liou, 1988).

By conservation of energy through a medium that scatters and absorbs, it is clear that the fractions of radiation that have been reflected/back-scattered \( (R_\lambda) \), absorbed, and transmitted \( (t_\lambda) \) through this medium should sum to one for each wavelength (Liou, 1988):

\[
t_\lambda + A_\lambda + R_\lambda = 1. \tag{3}
\]

When radiation travels through a medium of thickness \( ds \), matter will interact with it, and change its original intensity from \( I_\lambda \) to \( I_\lambda + dI_\lambda \). This change in intensity is described by:

\[
dI_\lambda = -k_\lambda \rho I_\lambda ds, \tag{4}
\]

where \( k_\lambda \) is the mass extinction cross section for the wavelength \( \lambda \) and \( \rho \) is the density of the medium. However, the intensity may be strengthened by emission in the medium or by multiple scattering. This increase can be described by \( dI_\lambda = j_\lambda \rho ds \), where \( j_\lambda \) is the source function coefficient. The full transfer equation is then

\[
dI_\lambda = -k_\lambda \rho I_\lambda ds + j_\lambda \rho ds. \tag{5}
\]

The source function \( J_\lambda \) is defined as \( j_\lambda / k_\lambda \). This allows us to write the general
equation of radiative transfer as

\[ \frac{dI_\lambda}{k_\lambda} \rho ds = -I_\lambda + J_\lambda. \]  

(6)

In a plane-parallel atmosphere this becomes

\[ \cos(\theta) \frac{dI(z, \theta, \phi)}{k \rho dz} = -I(z, \theta, \phi) + J(z, \theta, \phi), \]  

(7)

which can be solved for the upward and downward intensities in a finite atmosphere bounded on two sides. For a more in-depth treatment of atmospheric radiative theory, see Liou (1988).

For the purposes of this project, it is useful to think of radiation as being either shortwave (SW) or longwave (LW). SW radiation in the atmosphere consists of solar insolation. Downwelling SW comes from the sun, while upwelling SW is reflected from a surface. Given the high latitude of the Arctic, the solar zenith angle of the sun has a great effect on how much solar insolation the surface receives. The higher albedo of snow and ice create a greater upwelling SW component. In contrast, downwelling and upwelling LW radiation are radiated by anything with mass: the Earth, clouds, aerosols, atmospheric gases, particles, etc. Increased downwelling LW (the greenhouse effect) is an example of this effect.

1.3 Cloud Microphysics and Aerosol Effects

The atmosphere in the Arctic is particular pristine, with few relatively few aerosol particles (compared to more industrialized regions) to serve as cloud condensation nuclei (CCN) (Leck et al., 2008). CCN are hygroscopic aerosols that are used as centers for condensation once the air has become saturated (Rogers and Yau, 1996). The number of aerosols and CCN available affects the nature of clouds and the size of droplets or ice crystals. Changes in the aerosol concentration in the Arctic will likely have an effect on the climate of the Arctic through cloud processes. Aerosols may directly have a cooling effect on a global scale (as with major volcanic eruptions), that helps to offset global warming (Solomon et al., 2007). However, aerosols may have an indirect effect that causes either a warming or a cooling; more study is needed on aerosol effects.

Although aerosols in the atmosphere can be from natural sources, the population of anthropogenic aerosols has increased since the dawn of the industrial age (Forster et al., 2007). The increasing anthropogenic population of aerosols will have an effect on the atmosphere through many direct and indirect effects. Direct effects of aerosols include
scattering, absorption and emission of radiation. It is the indirect effects of aerosols that play a more important role, however. The first indirect effect (the Twomey effect) occurs when aerosols act as CCN or ice nuclei (IN). Ice nuclei are rare and consist of a very small subset of aerosols (Rogers and Yau, 1996). Increasing the number of CCN creates more, smaller cloud droplets with an overall greater surface area. Due to this greater surface area, more solar radiation will be reflected, decreasing the top of atmosphere (TOA) and surface forcing negatively from -0.5 to -1.9 W/m², as reviewed by Lohmann and Feichter (2005). The second indirect effect has an impact on the lifetime of the clouds: with more CCN cloud droplets are smaller, which decreases the precipitation efficiency, and increases the lifetime of the cloud (Albrecht, 1989). This effect, as shown by the literature, also has a negative influence on TOA and surface net radiation.

A glaciation indirect effect, where an increase in IN increases precipitation, is also potentially important in the Arctic. Although it is the first indirect effect that is cited most often, all of the indirect effects need to be examined further to understand the overall effect that increasing the CCN/IN in the atmosphere may have on cloud formation (Lohmann and Feichter, 2005).

1.4 Cloud Radiative Effects in the Arctic

The role of clouds in the Arctic is especially important for understanding the radiation budget, as well as ice-albedo and cloud-radiative feedback mechanisms. The ice-albedo feedback occurs when ice reflects back SW insolation; with more SW reflected, the surface cools, allowing more ice to form. This new ice, in turn, increases the albedo, which causes more solar radiation to be reflected to space, causing a positive feedback. Cloud radiative forcing (CF) is a measure of how much clouds heat or cool the surface (or any other layer in the atmosphere) relative to the clear sky. Surface cloud forcing, as defined by Ramanathan et al. (1989), is

\[ CF = F(allsky) - F(clearsky), \]  

where \( F \) is the net LW or SW flux at the surface. A positive CF indicates that clouds are heating the surface, while a negative CF indicates a cooling of the surface. In general, clouds in the atmosphere, including the Arctic, warm the surface by emitting LW radiation (greenhouse effect) and cool the surface by shading the incoming SW radiation (cloud shading effect). The dry atmosphere of the Arctic emits less clear-sky radiation than at lower latitudes, and LW emitted by clouds becomes all the more
important. The Arctic atmosphere has strong temperature inversions (Curry et al., 1993), which can allow for clouds to emit LW at cooler temperatures than the rest of the lower troposphere.

As seen in Shupe and Intrieri (2004), radiative flux equations can be reduced to first order to better understand how CF changes in the Arctic. To simplify matters, the atmosphere is divided into three sections: below-cloud \((b)\), cloud \((c)\), and above-cloud \((a)\), as done in Shupe and Intrieri (2004). Total CF consists of the sum of the LW and SW CF. As cloud forcing consists of the differences in net fluxes, it is these quantities that need to be compared to first order. The upwelling LW fluxes for both clear sky (no clouds) and all-sky (includes clouds) are approximately equivalent when considering the instantaneous impact of clouds on radiation, and the difference is ignored here. The LW net flux then only consists of downwelling (which is considered positive – \(NetFlux = Upwelling - Downwelling\)) fluxes based on the transmittances and effective temperature in the two respective layers. As described by the Stefan-Boltzmann Law:

\[
F_{LW}^{down}(clearsky) = t_{bl} \sigma T_a^4 + \sigma T_b^4
\]

\[
F_{LW}^{down}(allsky) = t_{bl}(1 - \epsilon) \sigma T_a^4 + t_{bl}\epsilon \sigma T_c^4 + \sigma T_b^4 + R
\]

\((9)\)
\((10)\)

\(T\) is temperature, \(t_{bl}\) is LW broadband transmittance for the \(b\) layer, \(\epsilon\) is emissivity, and \(R\) is reflectance. For the purpose of this first-order explanation, we assume that the cloud fraction of the all-sky case is 1. Reflectance in the infrared (LW) is very small when compared with the LW emitted; thus, \(R\) can be neglected. As seen earlier, \(t + \epsilon = 1\) for a specific frequency, and here \(t_b = 1 - \epsilon_b\). \(t\) can be related to the microphysical processes of the clouds as \(t = exp(-\tau)\) where \(\tau\) is the optical depth. To first order,

\[
CF_{LW} \approx t_{bl}\epsilon \sigma \left( T_c^4 - T_a^4 \right).
\]

\((11)\)

For the SW component of atmospheric radiative fluxes, all downwelling radiation will be from solar insolation, while upwelling radiation at the surface will be due to the

---

1A note on sign conventions, net fluxes and heating rates: As mentioned, a positive CF indicates a net warming (a convergence of radiation due to the cloud) while a negative CF shows net cooling (divergence). Net fluxes are defined with upwards being positive. Example: under a cloud, downwelling LW will be greater than without a cloud present above (greenhouse effect); assuming that all fluxes are taken instantaneously, upwelling flux from the surface will be the same in both case the cloudy and clear cases. Net flux in both cases is positive because the amount of radiation from the surface is greater than that which is absorbed and re-emitted downward by the cloud. With the cloud present, the net flux will have a smaller magnitude, because of a greater downwelling flux. There is a positive cloud forcing under the cloud because the magnitude of radiation entering into this region is greater, however the net flux cloud forcing will be negative. It is easier to use the differences in cloudy and clear heating rates to show region of positive or negative cloud forcing. Heating rates are calculated using the convergence or divergence of fluxes into a layer.

---
albedo of the surface; upwelling radiation above the surface depends on the albedo of
the surface and any other intervening reflectors in the atmosphere. This tells us that
the net flux for clear and all sky approximations will be

\[ F_{SW}^{(clearsky)} = t_{bsw} S \mu (1 - \alpha), \]

and

\[ F_{SW}^{(allsky)} = t_{bsw} S \mu (1 - \alpha) t_{csw}, \]

where \( t_{bsw} \) is the SW broadband transmittance for the b layer, \( t_{csw} \) is the SW broadband
transmittance for the c layer, \( S \) is the solar constant, \( \mu \) is the solar zenith angle \( \cos(\theta) \),
and \( \alpha \) is the surface albedo. The SW CF then becomes

\[ CF_{SW} = t_{bsw} S \mu (1 - \alpha) (t_{csw} - 1), \]

and the total radiative impact the cloud has on the surface is

\[ CF = CF_{SW} + CF_{LW} \]

\[ CF \approx t_{bsw} e \sigma \left( T_c^4 - T_a^4 \right) + t_{bsw} S \mu (1 - \alpha) (t_{csw} - 1). \]

From this first-order equation we can see the main components of how clouds affect
the surface: i.e., microphysical processes (through the transmittances and emissivity of
the cloud), the temperature of the cloud, the temperature above the cloud (which can
be determined by the height of the cloud), the zenith angle, and the surface albedo. By
varying these parameters we can modify how the clouds affect the surface radiation,
and study its impacts.

Shupe and Intrieri (2004) analyzed observations taken from the Surface Heat Budget
of the Arctic (SHEBA) field campaign and a radiative transfer model to compute
clear sky conditions. In the field portion of SHEBA observations were collected for a
full year from October 1997-1998 to study ocean-ice-atmosphere processes and cloud-
radiative effects (Uttal et al.). It was a major field campaign in the Arctic with a
sophisticated suite of atmospheric and oceanic sensors. Shupe and Intrieri (2004) found
that increasing Arctic clouds produced a net surface warming effect for the majority
of the year, except during a short period during August when the SW cloud shading
outweighed the LW warming. SW CF (in the form of cloud shading) increases with the
low solar zenith angles and low albedo, which occurs in the summer: for this reason
this period corresponds with the time that the Arctic sea ice begins to refreeze. Other
important results included that 95% of all clouds that had a significant LW contribution
were at an altitude of 4.3 km or lower in the atmosphere, and that the LW CF increases with the temperature of the cloud. Given the Arctic's frequent temperature inversion, this result becomes all the more important. Upon examining cloud properties, Shupe and Intrieri found that LW CF increased with the liquid water path (LWP) of the cloud up to 30 $W/m^2$; after that point, any additional LWP did not change the LW CF. Conversely, with additional LWP, the SW CF continued to become more negative (although with less gains) and did not have a threshold level of LWP as did the LW CF.

2 Observation Sources and Radiative Models

2.1 Arctic Summer Ocean-Cloud Study

Natural variability in the polar regions, as well as the response in this region to anthropogenic warming is not well understood; full sets of observations in the Arctic help to shed light on the processes of this region. SHEBA’s goal was to collect complete observations of Arctic cloud-ocean-ice processes for a full year Uttal et al.. Similarly during August 2008, an international group of scientists conducted the Arctic Summer Ocean-Cloud Study (ASCOS) aboard the Swedish icebreaker Oden (see Figure 1). From August 12 to September 2, the Oden was frozen into the ice floe in the central Arctic Ocean. The multi-faceted nature of the subject at hand, the Arctic system, required an interdisciplinary approach. Meteorologists, atmospheric chemists, oceanographers, marine biologists and other scientists observed and studied the Arctic region for the purpose of better understanding what processes connect the ocean, atmosphere, and sea-ice. One specific goal was to study the mechanisms that govern cloud processes in the Arctic. To understand how the climate in the Arctic will change, and to model it more accurately, clouds in this region must be understood thoroughly (Leck et al., 2008).

Radiosondes were released every six hours. They typically reached heights of around 20km (lower-middle stratosphere). Occasionally, the radiosondes would reach a maximum height and then fall back down again. Once a maximum height was attained, data after this point (if the radiosonde descended again) were ignored. The maximum height used for interpolation of the case studies was 23km.

Vertical cloud locations, ice water content (IWC) and ice crystal size were derived from data collected by a cloud doppler radar (Leck et al., 2008). Retrieval methods used for ASCOS were tested during SHEBA and compared against SHEBA aircraft retrievals (Shupe et al., 2005). Shupe et al. (2005) showed that the uncertainty in the
method for these retrievals for ice water content for single-phase ice clouds at SHEBA was 73% and for ice crystal radius was 40%. They did not expect mixed-phase clouds to be significantly different. Shupe et al. (2006) compared the difference between mixed and single-phase ice clouds and find that mixed-phase microphysical properties are all larger than those in their single-phase counterparts. Cloud type was determined from ceilometer, microwave radiometer, radar and radiosonde data as in Shupe (2007). Microwave radiometer data was used to derive cloud liquid water content (LWC). Liquid water profiles were derived given ceilometer cloud-base height, an assumed adiabatic profile, radiosonde profiles and then constrained with the microwave radiometer observations as described in Schofield et al. (2007) and Shupe et al. (2005). When comparing liquid water content retrievals against aircraft observations at SHEBA, retrievals had a root square deviation of 44 - 49% as calculated by Shupe et al. (2005). Water droplet effective radius is assumed to be 8 microns.

2.2 The Discrete Ordinate Method and Rapid Radiative Transfer Model

The Rapid Radiative Transfer Model (RRTM), developed by Atmospheric and Environmental Research (AER), Inc., is a radiative transfer model, based on the discrete ordinate method for radiative transfer (DISORT) by Chandrasekhar (1960). Stamnes et al. (1988) developed the numerical method for solving plane-parallel radiative transfer (see Equation 7) in a vertically inhomogeneous layered media that is used here.

Their model calculates the transfer of radiation in a plane-parallel medium that reflects, emits and absorbs radiation differently at each frequency. It was the first discrete ordinate method that dependably and efficiently computed the eigenvalues
and eigenvectors of the solution and inverted the matrix necessary for the constants of
integration in the problem.

AER's RRTM calculates longwave and shortwave fluxes as well as cooling rates (Mlawer et al., 1997) (Clough et al., 2005). The RRTM divides frequencies into a
finite number of bands: 16 bands in the LW (10 – 3250 cm⁻¹) and 14 bands in the SW
(820 – 50000 cm⁻¹). Bands were chosen given absorptions in major atmospheric gases.
The RRTM contains parameterizations for both clear-sky and liquid and ice clouds. It
has been verified against other AER models (such as the line-by-line radiative transfer
model, LBLRTM) as well as measurements. Given its accuracy and rapidity, the LW
RRTM has been used operationally in the ECMWF and NCAR GFS models. Also,
the SW has been used as a reference in the development of other SW radiative transfer
models. Gases included as absorbers in the LW RRTM are water vapor, carbon dioxide,
ozone, nitrous oxide, methane, oxygen, nitrogen, and common halocarbons. Likewise,
the gaseous sources of extinction included by the SW RRTM are similar: water vapor,
carbon dioxide, ozone, methane, oxygen, and aerosols.

2.3 Case Studies and Sensitivity Studies

To facilitate the incorporation of observations into the RRTM, a Matlab wrapper script
was created that takes in the raw observations, interpolates them to a size and format
that the RRTM can understand, calls the RRTM, and saves the results for each time
step or individual sensitivity study. The benefit of having such a script is that it can
easily loop through many instances and automatically format observations, allowing
for RRTM runs for long time periods, or with fine sensitivity increments.

August 25 and August 29, 2008 were chosen for case studies as they both had
instances of multiple-layers of stratified clouds that could have significant effects on
the evolution of each cloud and interesting effects on the surface.

Radiosondes were typically released every 6 hours. For August 25 and 29, pressure,
temperature, and relative humidity fields were linearly interpolated to each minute
during the entire day from 0 UTC to 0 UTC using the radiosondes from 18 UTC the
previous day, all radiosondes from the day in question, and the radiosonde from 6
UTC the following day. (Other interpolation schemes, besides linear, were considered,
hence the need for data points outside each day.) Cloud liquid and ice effective radius
and content already existed in minute resolution. All fields, cloud and radiosonde
alike, were interpolated in height from 510 layers to 200 equally space layers 115m
apart, starting at 95m above the surface. This interpolation was necessary as the SW
RRTM could handle a maximum of 200 layers (while the LW RRTM could handle
much larger). Drop and crystal radii were interpolated using a weighted average for the layer in question, while liquid and ice content, as well as radiosonde data were linearly interpolated.

The RRTM was run for every minute in each day; it was called for both cloudy and clear-sky with both the SW and LW RRTMs - four different calculations for each minute of simulated day. On average, approximately 6 minutes of day were calculated per one minute of run time, and calculating the radiative profiles for an entire day (1440 minutes) usually took approximately 4 hours on the NOAA server Scarecrow that is used by the Polar Processes Group.

Model output was radiative fluxes between each layer. LW and SW upward, downward, and net fluxes (W/m²) were calculated in each of the 200 layers, as well as the heating rate (K/day) for each layer. At ASCOS, radiation observations at the surface provided a useful check for the results of the RRTM. Surface observations of upward LW and SW fluxes were calculated from their respective downwelling observations with a set albedo of 0.75. Due to this dependence, we can use either the downwelling or the upwelling LW and SW observations, and not both. Model surface results were compared against these surface results to verify the legitimacy of the RRTM results.

After completing the case studies, multi-cloud sensitivity studies were conducted using portions of the August 29 case. Using averaged atmospheric profiles, 15z-16z was chosen for the upper stratocumulus cloud and 19z-20z for the lower stratocumulus cloud. LWP and IWP (ice water path) were varied through the clouds from 0 – 200% through the lower (upper) cloud while holding the upper (lower) cloud at its original LWP. The entire LWP for both clouds was phased between the two clouds to determine the effects of having moisture at different places between the clouds. As the LWC was determined in each layer, given an adiabatic profile and the original LWP, changing the LWP in each cloud used the same adiabatic profile and multiplied it by a constant factor for the entire cloud. RRTM results from these runs were analyzed in the portion of the atmosphere containing the clouds to determine what changes occurred from altering the LWP or IWP.

Uncertainty in the observed drop size has the potential to significantly change the radiative effects of the clouds and the effect on the surface. To simulate the impact of aerosols on clouds, the drop sizes in the clouds were varied for reasonable cloud drop sizes. The RRTM could not handle drop sizes below 2.5 microns, and above 20 microns was deemed too large for a reasonable Arctic cloud; as a result, a single-cloud profile was studied for cloud droplet sizes from 3-20 microns (in 1 micron increments) at various LWPs from 0 – 200% of the original LWP. The chosen profile was a single-layer cloud during August 29 from 5:30-6:00 UTC. All other quantities (ice effective radius,
Table 1: Parameters used for all RRTM runs.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Dates</td>
<td>August 25/29</td>
</tr>
<tr>
<td>Julian Dates</td>
<td>238/242</td>
</tr>
<tr>
<td>Average Latitude</td>
<td>87.4°/87.2°</td>
</tr>
<tr>
<td>Average Longitude</td>
<td>−7.4°/−9.4°</td>
</tr>
<tr>
<td>Surface Altitude</td>
<td>0 km</td>
</tr>
<tr>
<td>First Layer Height</td>
<td>95 m</td>
</tr>
<tr>
<td>Number of Layers</td>
<td>200</td>
</tr>
<tr>
<td>Top of Atmosphere Altitude</td>
<td>100 km</td>
</tr>
<tr>
<td>Layer thickness (excluding TOA layer)</td>
<td>115.1 m</td>
</tr>
<tr>
<td>CO2 Concentration</td>
<td>387 ppm</td>
</tr>
<tr>
<td>Number of Angles for Radiance Computation</td>
<td>3</td>
</tr>
<tr>
<td>LW Surface Emissivity</td>
<td>0.985</td>
</tr>
<tr>
<td>LW Albedo</td>
<td>0.015</td>
</tr>
<tr>
<td>SW Surface Emissivity</td>
<td>0.25</td>
</tr>
<tr>
<td>SW Albedo</td>
<td>0.75</td>
</tr>
<tr>
<td>Radius of Earth</td>
<td>6356.9 km</td>
</tr>
<tr>
<td>Scattering</td>
<td>DISORT</td>
</tr>
<tr>
<td>Number of Streams</td>
<td>4 (default)</td>
</tr>
<tr>
<td>Compensating for instrumental cosine response</td>
<td>No</td>
</tr>
<tr>
<td>SW Downwelling flux type</td>
<td>'True’ Diffuse Downwelling</td>
</tr>
</tbody>
</table>

ice content, etc.) were averaged during this period. Later, a minimization algorithm was written to minimize

\[
\left( LW_{down}^{model} - LW_{down}^{obs} \right)^2 + \left( SW_{down}^{model} - SW_{down}^{obs} \right)^2 ,
\]

while varying the liquid drop size from an effective radius of 3μm to 20μm; these two radii were used as they are the limits of reasonable cloud drop sizes. Only the downwelling components were used, as the upwelling components of both the LW and SW surface radiation depend on them. As varying the drop size changes the surface radiation in both the LW and SW, both were included in the minimization scheme. If any meaningful drop size change or patterns arise, then these could potentially be used as a retrieval for drop size from the radiation observations.
3 Results

3.1 Summary of Results

Studying the effect of clouds while varying their parameters suggests much about the clouds and their radiative interactions. However, seeing the actual evolution first is instructive in showing if the radiative model is capturing the processes involved. The greenhouse effect and cloud shading effects are clearly seen, as well as the radiative impact of liquid versus ice water in the clouds and on the surface. Varying liquid and ice water content as well as liquid drop size shows the sensitivity of the cloud to each parameter.

3.2 Case Study Results

As seen in the third and fourth panels of Figures 2 and 3, a low stratocumulus deck is persistent through both days. In the first two panels of Figures 2 and 3, black outlines encircle regions where the cloud fraction was observed to be 1. Observations were defined for cloud fraction to be either 0 or 1; regardless, the RRTM will not accept partial cloud fractions (rrt).

On August 25 (Figure 2), we see that the initial stratocumulus deck is at a higher level (2-3 km) than it was through most of the day when the deck spanned the region from 100 m to 700 m. Later in the day, the deck’s height increases. From 10 UTC to 14 UTC, a higher level cloud consisting mostly of ice evolves. From 14 UTC to 15 UTC, observations show that there is no liquid water content present; this is an error in the observations. Likely, the microwave radiometer had rimed over and had not been cleaned off due to difficulty reaching the ice camp instruments. The clouds are mixed-phase, but lower clouds generally contain more liquid water, while the higher clouds have a higher ice content.

For August 29 (Figure 3), we see a stratocumulus deck that lasts through the entire day with a higher layer of stratocumulus that lasted from 14 UTC to 16 UTC. Throughout this day there are higher (4-11 km) clouds comprised mostly of ice.

The upper panels in both figures show the LW and SW heating rate forcing for the two days examined. Heating rate forcing is defined with the same convention as cloud forcing. Layers of negative forcing are defined as having a net outward flux of radiation - a cooling. Positive forcing in a layer indicates a convergence of radiation: a warming.

With both of these days, we can immediately see a few processes worth noting. In regions where the liquid water content retrievals have concentrated liquid water, there is stronger LW and SW forcing. This is especially noticeable in the upper level clouds.
Figure 2: August 25, 2008 Time by Height Case Study. Panels show the RRTM-calculated LW (top panel) and SW (second panel) cloud forcing results through each layer in the atmosphere up to 7 km for the entire day; outlines of portions where cloud fraction is greater than 0 are shown. The ice water content (third panel) and the liquid water content (bottom panel) through the same time period are included. Note that regions with more liquid water show greater SW and LW cloud forcings. Greenhouse effects are visible under clouds in the LW cloud forcing and cloud shading effects are seen in the SW cloud forcing. The lack of liquid water from 14-15 UTC is due to an instrument error.
Figure 3: August 29, 2008 Time by Height Case Study. Panels show the RRTM-calculated LW (top panel) and SW (second panel) cloud forcing results through each layer in the atmosphere up to 7 km for the entire day; outlines of portions where cloud fraction is greater than 0 are shown. The ice water content (third panel) and the liquid water content (bottom panel) through the same time period are included. Note that regions with more liquid water show greater SW and LW cloud forcings. Greenhouse effects and cloud shading effects are both visible. At 14-17 UTC, note how the second stratocumulus cloud takes over the function of the lower stratocumulus deck.
While they are mostly ice, they do have some radiative impact on the atmosphere underneath them, but only in the areas where liquid water also exists. During the 14-15 UTC gap in water observations on August 25, we see that the lower layer clouds no longer produce CF in either the LW or SW, despite ice water content not being effected.

The top of the stratocumulus deck shows the most LW and SW cloud forcing for those cloud type. In the LW forcing panels we see that there is a net cooling at the top of the clouds. Underneath clouds, we generally see a slight LW warming, the greenhouse effect that translates to the surface when considering the low cloud deck. There is a net SW warming at the top of the stratocumulus clouds. Underneath clouds in the SW, we see a cooling; this is the cloud shading effect.

It is useful to note that when examining instances where there are multiple stratocumulus layers (at 0-3 UTC on August 25 and at 14-16 UTC on August 29), it is only the top of the upper cloud that shows any LW cooling or SW warming. The top of the lower layer shows no forcing relative to its base or the base of the upper cloud.

### 3.3 Validation against Surface Radiation Observations

Figures 4-7 compare the model’s surface fluxes with the surface fluxes observed at the ice camp. Both the upward and downward fluxes are presented for the SW and LW verifications, although there are some dependencies between them. The upward SW fluxes is dependent on the downward SW flux and albedo in both the model and observations (0.75); neither is considered to change here. The upward LW flux is dependent on the downward LW flux, the surface temperature and the emissivity. The surface temperature is dependent on the LW downward flux and emissivity is held fixed.

For August 25, (Figures 4 and 5) we see that the model LW fluxes (upward and downward) are both slightly less (by 2-5 W/m² in the LW upward and 5-10 W/m² in the LW downward) than the observed LW fluxes; however, the general trends are captured effectively. The only area of concern was around 14-15 UTC when the liquid water observations were faulty; the model acted as if there was no liquid in the clouds. This same error due to bad liquid water content input is visible in the SW upward and downward validation plots (Figure 5). At the times where the instruments were reading no liquid water, we can see the SW upward and downward surface fluxes acting as if there were no clouds present, showing the envelope of the possible maximum SW radiation (given clear skies). Again, SW model surface results followed those of the observed surface fluxes well and captured the trends of the day effectively. At some
Figure 4: August 25 LW Validation. LW upwelling surface flux (upper panel) and downwelling surface flux (lower panel) observations (blue) are compared against RRTM surface results (green). LW upward fluxes are dependent upon the downward component. Overall, LW flux model results were 5-10 W/m² lower than observed, but overall trends were captured well. At points where liquid water observations were highly suspect (see Figure 2) model results deviate strongly from observations.
Figure 5: August 25 SW Validation. SW upwelling surface flux (upper panel) and downwelling surface flux (lower panel) observations (blue) are compared against RRTM surface results (green). SW upward fluxes are also dependent upon the downward component (by surface albedo) and were calculated as 0.75 in both the model and the surface observations. Overall trends were captured nicely by the model, except at times where the liquid water observations (included in the model) are suspect.
Figure 6: August 29 LW Validation. LW upwelling surface flux (upper panel) and downwelling surface flux (lower panel) observations (blue) are compared against RRTM surface results (green). LW upward fluxes are dependent upon the downward component. LW flux model results were lower than observed.

August 29’s LW and SW model and observation comparisons show similar results (Figures 6 and 7). LW upward and downward surface fluxes were often underestimated (especially in the downwelling). SW upward and downward surface flux trends are captured generally by the model, but both upward and downward model results are 10-20 W/m² greater than the observations from 0-7 UTC. The magnitude of the trends is also not captured as-well, although the existence of the fluctuations in the flux are captured.

3.4 Cloud Forcing as a function of Liquid Water Path

Cloud forcing was plotted against the liquid water path to see if the same features are present as in Shupe and Intrieri (2004). Cloud forcing was plotted with the days combined (Figure 8) and distinguished (Figure 9). Matlab fitting scripts were used
Figure 7: August 29 SW Validation. SW upwelling surface flux (upper panel) and downwelling surface flux (lower panel) observations (blue) are compared against RRTM surface results (green). SW upward fluxes are dependent upon the downward component (through surface albedo). Trends are captured well but the model results are significantly higher than the observations early in the day.
3.5 Liquid and Multi-cloud Water Sensitivity Studies

The interactions between the layers of stratocumulus clouds was studied as seen on August 29 from 15-16 UTC (see Figure 3). However, the liquid water content between these two clouds has been retrieved from the liquid water path as being either in one layer or the other. It is possible that the liquid water was divided between these two clouds, and that the retrieval for liquid water content is not capturing this detail. To create a liquid water profile for the two stratocumulus clouds, the adiabatic liquid water profiles from the lower cloud and upper cloud were averaged over a short amount of
Figure 9: Liquid water path sensitivity for both days individually. LW Cloud forcing levels off after a liquid water path of 30g/m² and does not increase. SW Cloud forcing as a function of LWP continues to decrease, with diminishing returns. No significant difference between the CFs on the two days is noticeable.
time (20 minutes) and an average liquid water path for this time period was spread over this profile. As the levels for the lower and upper cloud are distinct, the amount of liquid water in the two clouds and how it was partitioned between them was explored to see the radiative impact on the surface and each other.

The first study (Figures 10 and 11) left the lower cloud with its original liquid water path, while varying the liquid water path of the upper cloud from 0 – 200% of the original value. With 0% of the upper cloud present, the lower cloud has LW cooling at its top as visible by the negative heating rate forcing in Figure 10. With 10% of the upper cloud’s LWP present the lower cloud no longer has LW cooling at its top and very little heating rate forcing throughout. The upper cloud with 10% of its liquid water has cloud top cooling in the LW instead. As the amount of water in the upper cloud is increased, the magnitude of this negative heating rate forcing at the top of the cloud increases from $-30K/day$ (at 10% of the original liquid water path) to $-90K/day$ (at 200% of the original liquid water path). In the lower cloud we see a small LW heating rate. Overall, we see that the cloud has a small positive radiative forcing (warming) of the surface, but there is no noticeable change as the upper cloud’s liquid water path is varied.

The SW results for this study shows similar trends (see Figure 11). With only 0% of the upper cloud’s liquid water present, the lower cloud shows a positive SW heating rate (warming). As the upper cloud’s liquid water is varied from 10 – 200% this warming changes to a mostly constant heating rate through the lower cloud. The upper cloud gains the SW heating that the lower cloud lost as more liquid water is added to it. The atmosphere above the clouds sees a little to no SW heating rate cloud forcing. The surface sees a negative cloud forcing, and varies slightly, increasing its negative forcing on the surface.

The overall effect of varying the amount of liquid water in the upper cloud is stronger in the LW than in the SW. The strongest heating rate forcing we see in the LW is $-90K/day$ while the maximum forcing in the SW is $2K/day$.

The same time on August 29th was used to study the effects of varying the lower cloud liquid water path. The upper cloud liquid water path was held fixed as the lower cloud liquid water path was varied from 0 – 200% (see Figures 13 and 12). With no liquid water in the cloud, we see that there is a constant small positive forcing (of $2K/day$) through the lower cloud. As the cloud’s liquid water is increased, we see a warming with a maximum in the middle of the cloud that approaches $10K/day$. At 10% of the original liquid water content there is a peak of LW heating rate forcing at 0.7km that is larger than for larger values of liquid water content. When varying the liquid water path in the lower cloud, there is no perceptible effect on the LW cloud
Figure 10: LW heating rate forcing as upper cloud liquid water content is increased. Upper cloud is varied from 0% (red) to 200% (blue) of the original (observed) LWP in increments of 10%. LWP of lower cloud is held constant. Note lower cloud-top cooling when the upper cloud has no liquid content; with 10% of the upper cloud’s liquid water present, cloud-top cooling shifts to the upper cloud.
Figure 11: SW heating rate forcing as the upper cloud liquid water content being increased. Upper cloud is varied from 0% (red) to 200% (blue) of the original (observed) LWP in increments of 10%. LWP of lower cloud is held constant. Lower cloud shifts from a positive warming to a cooling as the upper cloud's LWP increases.
Figure 12: LW Heating Rate of clouds as liquid water content is increased in the lower cloud. Lower cloud is varied from 0% (red) to 200% (blue) of its original (observed) LWP in increments of 10%. LWP of upper cloud is held constant. The upper cloud is not affected by the lower cloud.
forcing of the upper cloud or the surface.

In the LW cloud forcing, the lower cloud does change with its increased LWP; it gains two layers of warming; perhaps these are due to the bottom of the cloud feeling the effects of surface upward LW radiation, and the top of the cloud absorbing radiation emitted downward by the upper cloud.

We see similar effects when examining the SW heating rate forcing after varying the liquid water in the lower cloud. There is no effect on the forcing of the upper cloud and a very small effect on the surface. The lower cloud’s forcing is negative (with a maximum magnitude of 0.5K/day) and becomes less negative (approaching a neutral forcing) with greater liquid water in the cloud; however, it never becomes positive.

The effects of varying ice content in the upper cloud were also examined. When varying the ice water content in the upper cloud from 0 – 200% of its original values, we see no change on the heating rate forcing of either cloud, the surface, or above the clouds. See Figures 14 and 15.
Figure 14: LW Heating rate forcing as the ice content of the upper cloud is increased from 0 – 200% of its original IWP. There is no noticeable effect on either cloud. Ice does not matter much to the radiative heating rates.
Figure 15: SW Heating rate forcing as the ice content of the upper cloud is increased from 0 – 200% of its original IWP. There is no noticeable effect on either cloud. Ice does not matter much to the radiative heating rates.
Figure 16: Partitioning of LWP between upper and lower clouds. Red shows 0% LWP in lower cloud and 100% in upper cloud, while blue shows 100% in the lower cloud. 10% LWP in the upper cloud is enough to turn off LW lower cloud-top cooling.

### 3.6 Liquid Water Path Partitioning between Multiple Clouds

The liquid water path in this experiment instead is partitioned between the two clouds, keeping the total LWP constant. For this study, the liquid water path was instead chosen from one instant in time (between the two stratocumulus levels on August 29 at 14 UTC), with the liquid water content profiles being chosen from the instants before and after the transition. The original hope was that partitioning the liquid water between the two clouds might alter the RRTM’s radiation results enough that a certain partition of water between the two clouds might better match the surface observations at particular times.

Although this did not prove effective, partitioning between the two clouds shows the same sorts of results as phasing in the upper cloud did. In the LW (see Figures 16 and 17) when there is a significant amount of liquid water present in the upper cloud (> 10%), the LW heating effect at the top of the upper cloud dominates over any other effects that were present when the lower cloud had all the liquid water. For this time step \((t = 884 \text{ minutes into the day or at 14:43 UTC})\), the liquid water path was 129 g/m\(^2\). For reference, a profile of the ice water profile at this time has also been included (see Figure 17).
Figure 17: Liquid and ice water content as liquid water is partitioned between the upper and lower clouds. Red shows 0% LWP in lower cloud and 100% in upper cloud, while blue shows 100% in the lower cloud.

### 3.7 Drop Size Studies

As there is uncertainty in both the liquid water content and the liquid effective radius, cloud forcing will vary with changes in both quantities. Liquid water path for a single cloud was varied (usually in increments of 10% of the original liquid water path), as was the cloud’s drop size. The profile chosen for this study was a single cloud observed during August 29 from 5:30-6:00 UTC (see Figure 3). The liquid water profile, ice content, ice crystal and other non-varying quantities were averaged through this time period. Drop sizes were varied from 3 – 20μm. The difference in net fluxes is related to the cloud forcing. At the surface, if this quantity is increasing upward, it implies that flux is diverging from the surface. Diverging surface fluxes imply a negative cloud forcing for this period in the LW. The gradient of the net flux difference does not change significantly (the divergence – and cloud forcing – does not change significantly) and becomes less negative (weaker) with increasing drop size (see 18). With increasing LWP, the SW net flux forcing profiles become stronger, more positive. The spread between drop size increments changes with different LWP path. The intermediate LWPs (see the 50% and 100% on Figure 18) show the greatest change in cloud forcing between small and large drop sizes, especially at the surface.

Here, the sensitivity of cloud forcing to drop size is defined as $dF/dr \approx \Delta NetFlux/\Delta r$. As the drop size increases, we see that there is a greater sensitivity to drop sizes with
Figure 18: Difference in LW net flux profiles (cloudy-clear) are examined over selected liquid water paths as the liquid effective radius is varied from 3μm (red) to 20μm (green) by increments of 1μm on August 29 at 5:30-6:00 UTC. Overall, LW net flux differences increase with increasing LWP path, but becomes less negative (weaker) with increasing drop size. There is greater variation in the cloud forcing with each successive μm for intermediate percentages of liquid water path (see 50% and 100%).
smaller drop sizes (red) than larger drop sizes. When examining 50% and 100% of the original LWP, the sensitivity is greater than with either of the extremum LWP, 10% and 20% of the original values (see Figure 19).

In the SW, the change in net flux (cloudy-clear) is positive and becomes less positive (weaker) with increasing drop size (see Figure 20). Under the cloud this net flux difference is increasing upward from the surface, implying that there is diverging flux from the surface, and a negative surface cloud forcing. Overall, with increasing LWP, the change in net fluxes becomes stronger; its gradient under the cloud also increases, implying that this negative cloud forcing is becoming stronger as well. However, the magnitude of the change in cloud forcing from 3\(\mu\)m to 20\(\mu\)m increases with increasing liquid water content with the largest cloud forcing profile being for 200% of the original LWP and 3\(\mu\)m drop radius. SW net flux differences has the greatest sensitivity to smaller drop sizes and to larger LWPs (see Figure 21).
Figure 20: SW net flux difference (cloudy-clear) height profiles are examined over selected liquid water paths as the liquid effective radius is varied from 3μm (red) to 20μm (green) on August 29 at 5:30-6:00 UTC. The difference in cloudy and clear SW net fluxes becomes less positive (weaker) with increasing drop size. As liquid water path increases in the cloud, the total change from 3μm to 20μm becomes greater (i.e., there is a greater change in cloud forcing from red to green droplet sizes for 200% than for 10%).
Figure 21: Sensitivity \((dF/dr)\) of SW net flux difference (cloudy-clear) to changes in drop size. The difference in SW cloudy and clear net fluxes is more sensitive (greater magnitude) to smaller drop sizes (red) than larger drop sizes (yellow), and for the greatest LWP percentage (200%).

3.8 Minimization of LW and SW surface radiation with Liquid Effective Radius

When using the difference between the RRTM’s output with the original liquid water and ice water observations and the observed surface radiation observations, the minimized radius was kept to a reasonable range of liquid radii (3 - 20\(\mu\)m). When examining how well the RRTM’s LW downwelling surface radiation changed to match the observed radiation (see Figure 22), we see that the optimized RRTM (green) generally follows the first RRTM run (red). It sometimes even increases away from the observed radiation (blue).

The changes in SW downwelling surface radiation after this minimization are much more dramatic (see Figure 23). When changing the drop size to find one that better fits the observed radiation (blue), this drop-size optimized curve (green) follows the observed surface radiation almost exactly. The minimization was able to find the drop size that better matched observed surface SW radiation, but had little to no effect on matching the LW radiation.

In general for the first part of the day (see Figure 24), the retrieved drop sizes
Figure 22: Comparison of LW downwelling surface radiation from observations (blue), from the RRTM with original liquid drop radii (red), and from the RRTM with drop sizes being optimized (green) to minimize the difference between the first two through the entirety of August 29. The minimization does not greatly change the RRTM model surface LW fluxes in favor of fluxes more like those observed.
Figure 23: Comparison of SW downwelling surface radiation from observations (blue), from the RRTM with original liquid drop radii (red), and from the RRTM with drop sizes being optimized (green) to minimize the difference between the first two through the entirety of August 29. The minimized RRTM followed the observed surface SW flux very well: SW cloud forcing is sensitive to changes in drop size.

(red) matched the minimized drop sizes (green) within a few $\mu m$. After 7 UTC, the difference between these two (bottom panel of Figure 24) have increased significantly and fluctuate from the consistently sized retrieved radii. The liquid water path for the day is included above for comparison.

4 Discussion of Results

4.1 Notable features during Case Studies

Only areas with liquid water show a significant cloud forcing; ice water does not seem to play a significant role. When qualitatively examining the case studies (Figures 2 and 3), the radiative properties of liquid versus ice water are fairly clear. In portions of the cloud where the retrieved liquid water is concentrated, we can see a correspondingly strong signal in the LW and SW cloud forcing. Where there is a greater amount of ice water in the cloud, the same strong radiative effects are not present. For example, on August 25, from 11-13 UTC there is a cloud above the stratocumulus deck starting at 5 km consisting of mostly ice and some water. The retrieval for the liquid and ice
Figure 24: Comparison between original liquid drop radii and minimized RRTM drop size. A time series of August 29th's liquid water path (panel 1) is included above the time series of original and minimized drop sizes (panel 2). The third panel shows the time series of differences between the original radius and its minimized counterpart. In the first seven hours of the day the difference in the two radii is small. During the rest of the day, the minimized liquid radius fluctuates much more.
content concentrates the liquid water in a few regions while it spreads the ice content smoothly for the entire cloud.

Later when sensitivity to ice water path is examined, ice water content is seen to not be as radiatively effective as liquid water. There is almost no difference between the 0% and 200% ice water cases, holding liquid water fixed. Further study should be conducted with all liquid water removed to see if differences in ice water path would then have a more noticeable effect on cloud forcing. This, however, would be unrealistic for summertime Arctic stratocumulus clouds, which are mixed-phase.

On both days, stratocumulus deck persists through the entire day. Under these clouds, LW greenhouse and SW cloud shading effects are clear. The effect on the surface, however, is not as strong as immediately under the cloud: surface CF varies with the height cloud. The higher ice clouds (at or above an altitude of 5km) show CF under them; this CF does not connect down to the stratocumulus deck and affect it significantly. Unlike the the two layers of stratocumulus clouds which affect each other through cloud-top cooling, these high clouds do not strongly affect the clouds underneath them, likely due to their height and their mostly icy composition (rather than liquid-water).

### 4.2 Effectiveness of RRTM at capturing the day’s trends

Liquid drop and ice crystal effective radii and other cloud properties were observed remotely using passive instruments. Qualitatively, the RRTM surface downward and upward LW and SW fluxes match their observed counterparts well, given the estimated cloud parameters. Both days show LW upward and downward fluxes a few W/m$^2$ lower than observed values; the predicted variability however agrees quite well with observations. It would appear that using the LW RRTM to determine surface fluxes is reliable. Errors in this estimate likely stem from the observations themselves. Varying cloud droplet size (and liquid water content) could point to consistent biases in the observations.

The SW RRTM fluxes also correspond well to the observed surface fluxes. The general trends and variability are captured; however, the magnitude of variability is not always simulated. Between 8 UTC and 10 UTC on August 25, the SW upward flux is more than 20W/m$^2$ different from the observed SW upward flux. This difference is too great to be explained by a simple error; either some process is not being covered or some initial parameter is incorrect. On August 29, the downward and upward SW fluxes are considerably higher than the observations (> 20W/m$^2$) from 0 to 7 UTC. Perhaps the transmissivity (or some other parameter) needs to be adjusted during this period.
only. Later in the day, the differences between observed and modeled are less, and the RRTM's fluxes is following the general surface trend of the observations. Occasionally, these later day SW fluxes miss the entire height of an hour-long fluctuation, but still indicate a fluctuation at that point. The modeled surface fluxes are not negatively or positively biased about the observations. Distinguishing whether the errors are due to drop or ice crystal sizes or liquid/ice content is difficult, especially given potential model biases.

On August 25th, there are times when the liquid water content was not collected due to an instrument malfunction; the SW fluxes spike unreasonably to the top of a SW envelope. The surface fluxes would follow this envelope if there were no clouds present. Clouds have a shading effect (reflecting back sunlight) and decrease the amount of SW radiation relative to this envelope. This envelope of maximum SW flux will vary with the sun’s zenith angle.

4.3 Importance of Cloud forcing versus Liquid Water Path

Cloud forcing varies differently with LWP for LW and SW. As seen in Figures 8 and 9, surface cloud forcing is significant in both the LW (up to 110W/m² for these days) and SW (up to -50W/m²). The amount of liquid water available in Arctic clouds can vary dramatically and with this comes a variation in their radiative effects.

After the cloud has reached approximately 30g/m², there is no further increase in the LW cloud forcing. Adding more water to the cloud will not have any greater effect on the surface. Extra water does not allow for any greater emission of LW radiation. However, with increasing liquid water, the SW cloud forcing continues to slowly decrease (becoming more negative). While SW cloud forcing does decrease with increased liquid water path, it does so with diminishing returns. In a moister atmosphere, such differences in the SW cloud forcing could be important. August and September are also the months when the Arctic sea ice begins to re-freeze. The moisture content of the atmosphere has an effect on how much SW cloud shading occurs and affects how soon refreezing begins.

Shupe and Intrieri found similar results using year-round surface observations from SHEBA. The only difference was the magnitude of the LW asymptote; it is 50W/m², in contrast to the 100W/m² seen in for these two days in August 2008.

4.4 Radiative interactions between upper and lower clouds

At times when there was a higher stratocumulus deck, the upper cloud’s top showed a radiative cooling in the LW. The top of the lower cloud (if present) does not show this
LW cooling. In the sensitivity studies with two layers of stratocumulus clouds, we see that if a little liquid water (as little as 10% the original LWP) is present, the upper cloud will show LW cooling. This LW radiative cooling in the cloud’s top points toward the mechanism that makes these stratocumulus clouds remaining in place through hours and days.

When considering only radiative effects, cooling at the top of the cloud occurs because radiation is equally emitted both down and up, but above the cloud, there is no mass to absorb and re-emit this radiation back down. (Empty atmosphere does not absorb and re-emit as much radiation as do clouds with liquid water.) The net flux divergence cools the cloud parcels at the top. These parcels become negatively buoyant and start sinking relative to the cloud below them. The warm air underneath these cool cloud tops is pushed upwards to the cloud top in a kind of upside-down convection. As cloud-top cooling is a radiative process and will occur as long as there is a cloud, it can continue for hours and entire days, as seen on both the 25th and 29th.

In the sensitivity study, the upper cloud shuts off the lower cloud’s cloud-top cooling (given the layers of upper cloud that are above it). This effectively shuts off the mechanism that was sustaining the lower cloud. The lower cloud begins to dissipate while the upper cloud top cools and begins to experience convection.

For this example on the 29th at 14-15z, the upper cloud did not last as long as lower stratocumulus deck that reformed. Why did the upper cloud not continue for the rest of the day? It is likely that dynamical processes (not used in this 1-D column model) are also involved that perturbed the state to create the upper cloud.

4.5 Relations of liquid drop size studies to indirect effects

Neither the observations of liquid drop size nor the liquid water path is completely certain, and the difference in net fluxes between cloudy and clear skies (and with it cloud forcing) will vary with slight changes in both of these quantities. Figures 18 and 19 show how this magnitude increases with decreasing drop size (given fixed LWP) and increases with LWP. For a given LWP, a smaller drop size implies that there are more drops (a greater concentration of drops and thus, more CCN) with a greater surface area with which to absorb radiation. As seen in Figure 8, LW surface cloud forcing increases with increasing LWP, and Figure 18 shows that the greatest magnitude of cloud forcing (for the surface and entire cloud) is seen with the greatest LWP. When examining the changes in CF with LWP, the amount that cloud forcing increases by with growing LWP is decreasing (with diminishing returns): a property already seen for LW cloud forcing versus LWP.
In other profiles, the LW surface cloud forcing was positive where we were measuring in K/day. Here we are measuring in W/m² and see a negative cloud forcing at the surface; since cloud forcing is defined here as the difference in cloudy and clear net fluxes and a positive net flux is defined as upward; by having a negative cloud forcing, the cloud is actually causing a greater negative net flux than without the cloud, and the surface is receiving a net flux due to the cloud.

Cloud liquid drop sizes are also uncertain in observations and do cause a change in the cloud forcing profiles with the greatest sensitivity being seen in the cloud itself and for LWPs that are close to the original (100%) LWP of the cloud. Cloud forcing, related to the change in net flux between cloudy and clear skies, is more sensitive to changes in liquid drop size for changes of intermediate LWPs, rather than with the largest LWPs. Drop size becomes less important at small and large LWPs.

The greatest SW negative net flux differences, and negative cloud forcings, occurs with small drop sizes and large LWPs, which is expected given the first indirect effect. The cloud forcing decreases with increasing drop size. The first indirect effect states that larger drops (smaller aerosol/CCN concentration) with a smaller total surface area will reflect less radiation and will have less forcing on the surface than will the smaller drops. Smaller drop sizes have greater reflectivity, increasing the upward flux, decreasing the downward flux, and increasing the net flux for the cloud (relative to clear skies). This greater net flux (relative to clear skies) is a negative cloud forcing (cloud shading) as more SW is being reflected. Also, greater LWP provides more surface area with which to reflect, hence there is a stronger upward net flux difference (Figure 20) and more divergence of SW radiation (net cooling) from the layers below relative to a clear sky. Figure 21 shows that clouds are the most negatively sensitive for small drop sizes (red). Increasing the drop size makes the cloud forcing less sensitive to drop size.

4.6 Discussion of minimization results

Varying the liquid effective radius shows what different (realistic) drop sizes could do to the observed LW and SW fluxes. Given that most drop sizes were estimates, if successful, retrievals of liquid drop size could be created given the minimization script used. The majority of the retrievals of the liquid and ice effective radii on August 29 were reasonable estimates (within 3 – 20μm); however, there were a few portions of the day when the liquid radius (that minimized the differences between observed and modeled downward fluxes) was larger than 20μm. Although the script did evaluate past 20μm, it would be interested to see just how high these values peak.

The RRTM drop-size minimized SW downward flux almost exactly matched the
observed SW surface flux, but there was very little change if any to the LW downward surface fluxes. The new drop-size minimized RRTM LW fluxes hardly changed from the RRTM LW fluxes with the original liquid drop sizes. For the day’s LWPs, SW fluxes are very sensitive to changes in drop sizes, and a change in drop size will significantly affect how much the cloud shades the surface. However, LW fluxes are not particularly sensitive to changes in drop size for these values of LWP. LW CF does not change much for LWPs greater than 30 g/m², nor for changes in drop size. The hope was that both LW and SW surface fluxes would match the surface observations after finding an optimal drop size; the RRTM could then have constrained drop sizes based on radiative observations. Discussions in January 2010 in Boulder indicated that the RRTM might have a positive bias in its LW surface fluxes and that other single-column radiative transfer models (such as the Single-Column Atmospheric Model used in the Community Atmospheric Model) might provide an alternative.

5 Concluding Remarks

Studying ASCOS cloud observations from August 25 and 29, 2008 with a radiative model reveals many properties of the stratocumulus clouds present. The RRTM is able to adequately simulate the radiative fluxes for both days and captures much of the variability of the observed surface radiative fluxes. Qualitatively, expected LW and SW effects (such as the greenhouse effect) are visible and provide a starting point from which to study individual portions of the day for sensitivity to various microphysical properties. Varying the liquid water path between two layers of stratocumulus clouds shows the importance of liquid water in the upper cloud: a small amount of liquid water in an upper cloud will create a divergence of LW radiation, creating a cooling at the top of this cloud. As there is no longer LW divergence in the lower cloud, the lower cloud no longer sustains convection, while the upper cloud does. Single-cloud sensitivity studies also show the importance of liquid drop size in the clouds; different cloud drop sizes have an effect on the surface and indicate the radiative impact of the indirect effects of aerosols. There is also a potentially useful tool in the single-column radiative model in retrieving a more accurate liquid drop size from surface radiation observations; although the RRTM did not constrain drop size with the LW surface observations, perhaps other models could.

Through the course of this project, various microphysical properties have been examined radiatively during two Arctic summer days; however, this is still much work that can be done for this project. There is still a wealth of ASCOS observations
available for analysis; and even with these two days, more case studies and individual profiles can be examined for different situations. Another radiative transfer model (single-column or otherwise) can be employed to remove the RRTM's possible LW bias and potentially better retrieve drop size using surface radiation observation. Dynamics of the system have been disregarded in these single-column studies, but microphysical properties are certainly affected by the dynamics and a study of any correlations between dynamical and cloud conditions could prove enlightening. More accurate Arctic observations, especially those involving clouds, are necessary to improve the accuracy of climate modeling at high latitudes.
References


