Modeling clouds and radiation for the November 1997 period of SHEBA using a column climate model

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Abstract. A column version of the Arctic regional climate system model (ARCSYM) has been developed for testing general circulation model parameterizations in the Arctic. The ARCSYM column model has been employed for a 23 day period in November to simulate conditions over a multiyear ice floe that has been the site of intensive observations as part of the Surface Heat Budget of the Arctic (SHEBA) project. The large-scale tendencies of temperature, moisture, and wind are specified with values obtained from a special column data set obtained from the European Centre for Medium-Range Weather Forecasting. Comparisons between the ARCSYM column simulations and SHEBA data reveal that modeled temperature profiles are too cold aloft and generally too warm in the boundary layer. The occurrence of low clouds is severely underpredicted while the high cloud fraction is over predicted. The modeled longwave radiative cooling at the surface is 1.5–3 times as large as that observed. Much of this bias is related to problems with the treatment of clear-sky radiative transfer and in the simulated cloud optical properties. At the same time, the magnitude of modeled downward sensible heat flux at the surface is much too large. This has been related, in part, to the method for scaling temperature at the lowest modeled level to its surface air value under conditions of strong static stability. The importance of properly treating longwave radiative transfer under extremely cold, clear-sky conditions is evident in the sensitivity studies. The best simulation of cloud properties was achieved by assuming liquid cloud processes and properties at temperatures above 255 K. This temperature is significantly colder than that used in many climate models. The occurrence of supercooled clouds in the simulation dramatically reduced longwave cooling at the surface due to increases in the optical depth and fractional coverage of clouds. Results from a coupled sea ice-atmosphere simulation reveal that improvements in the atmospheric parameterizations are enhanced when the system is coupled.

1. Introduction

In most general circulation models (GCMs), the atmospheric temperature response to a perturbation in the radiative energy budget is enhanced in the polar regions [e.g., Manabe et al., 1991]. The Atmospheric Model Intercomparison Project (AMIP) [Gates, 1992] has summarized results from approximately 30 GCMs that have completed a 10-year (1970-1988) simulation using observed monthly averaged values of sea surface temperature and sea ice extent. The AMIP results have been analyzed for the Arctic by Tao et al. [1995], Chen et al. [1995], Bormwch et al. [1994], Walsh and Khattsov et al. (Voeikov Main Geophysical Observatory, St. Petersburg, Russia, unpublished manuscript, 1995) (hereinafter referred to as unpublished manuscript). Tao et al. [1995] examined Arctic surface air temperatures from the AMIP simulations and found a range among the models in the zonally and seasonally averaged surface air temperature of 8°C in summer and 17°C in winter. Chen et al. [1995] reported that discrepancies in the longitudinally and annually averaged modeled cloud cover are as large as 60% with larger variability seen in seasonal comparisons. Khattsov et al. (unpublished manuscript, 1995) showed that the modeled mean June surface insolation over the Arctic Ocean ranged from 85 to 185 W m^-2 in the AMIP runs. These values are significantly less than the observed mean insolation for June of 310 W m^-2 obtained from Russian drifting stations [Serreze et al., 1997].

These discrepancies arise potentially for many reasons [e.g., Randall et al., 1998], but are in part associated with unique conditions occurring in the polar regions: strong static stability and complex vertical structure of the lower troposphere, very cold temperatures and low moisture contents, lack of sunlight for half the year, and an extremely inhogeneous underlying surface. These conditions strongly influence the physical processes occurring in the region. For example, several unusual types of boundary layer clouds occur with regularity in the Arctic [e.g., Curry et al., 1996]. Clear-sky ice crystal precipitation is now thought to be nearly ubiquitous in the Arctic throughout the cold half of the year. Stratiform clouds over the melting ice pack often occur in multiple layers [e.g., Curry et al., 1988]. Low clouds may also form in the convective plume above and downwind of wide, open...
leads in winter when the surface-air temperature contrast is large as observed by Schnell et al. [1989]. Mixed-phase clouds occur in the Arctic boundary layer with some regularity [Curry et al., 1996], and appear to be able to persist for extended periods of time as described by Pinto [1998].

The paucity of in situ measurements over the Arctic Ocean is a major obstacle to evaluating and improving satellite retrievals and climate model parameterizations of polar physical processes. To address this lack of data, the Surface Heat Budget of the Arctic (SHEBA) field program, the Atmospheric Radiation Measurement (ARM) program, and the FIRE Arctic Clouds Experiment are being conducted simultaneously over the Arctic Ocean [Randall et al., 1998]. The collective goal of these projects is to study the surface energy budget and sea ice mass balance of a typical multiyear ice floc in the central Arctic over an annual cycle. It is anticipated that observations from the SHEBA ice camp can be processed with the aid of satellite data and in situ aircraft data to be representative of a GCM grid cell. A primary goal of this unprecedented observational data set is to improve the simulation of Arctic physical processes in GCMs. To achieve this goal, the SHEBA data will be used to develop, test, and improve parameterizations of processes occurring in the Arctic.

An important tool in the development and evaluation of parameterizations for use in GCMs is the single-column model (SCM) [Randall et al., 1996]. Single-column models have been used to improve understanding of GCM parameterizations and to provide guidance for GCM parameterization development. Because of their efficiency, SCMs allow for the systematic evaluation of model parameterizations. Parameterizations tested in a SCM can be transferred directly into a three-dimensional GCM. The fact that the SCM is isolated from large-scale dynamics is an advantage because it allows parameterizations to be tested in a controlled local environment without propagation of errors from other grid cells. At the same time, SCMs can only be used to test parameterizations which do not involve dynamical feedbacks. Another disadvantage is the amount and type of data required to perform SCM simulations. These include time-varying profiles of mean vertical motion and horizontal advective tendencies, which are very difficult to observe directly. Instead, these quantities must be obtained from large-scale numerical weather prediction models.

Single-column models may be derived from existing GCMs or developed a prior using various parameterizations to describe the physical processes in a 1-D thermodynamic model. Numerous SCM studies have been performed with the column version of the Colorado State University GCM [e.g., Randall et al., 1991; Hu and Randall, 1994]. A single-column version of the Community Climate Model, Version 3 (CCM3) has been developed by Hack et al. [1997] and has been used to study cloud radiative forcing in the tropical western Pacific [Petch and Kiehl, 1997]. Petch and Dudhia [1998] have used output from a mesoscale model to force the single-column CCM3 (SCCM) to study the effect of horizontal advection of hydrometeors on climate simulations. Jacobellis and Somerville [1991] developed a SCM with an ocean mixed-layer to study interactions between the physical processes that affect the onset of the Indian summer monsoon. This SCM has also been employed by Randall et al. [1996] to study the impact of various cloud parameterizations on the predicted cloud amount and the net shortwave radiation at the surface for the ARM Southern Great Plains site.

Single-column modeling efforts in the polar regions have recently been undertaken. Holland et al. [1997] performed simulations using a coupled ice-ocean SCM to determine the relative strength of various thermodynamic sea ice-albedo feedback pro-

cesses. Bitt et al. [1996] use the SCM described by Moritz et al. [1992] to study low-frequency variability of sea ice thickness in the central Arctic. Abegg et al. [1997] used a 1-D version of the high-resolution regional climatic model HIRHAM to determine how the treatment of boundary layer physics affects the simulation of mean wintertime profiles in the Arctic. A. H. Lynch et al., (Surface energy balance on the Arctic tundra: Measurements and models, submitted to International Journal of Climatology, 1998) have recently employed a SCM to study the interaction between the atmosphere and the Arctic tundra.

The goal of this paper is to provide a basis for setting climate model improvement priorities and to identify the changes that will lead to first-order improvements in GCM simulation of the wintertime Arctic using a column climate model and observations from SHEBA. The focus of this study will be on the parameterization of cloud processes, boundary layer processes, and radiative transfer in the central Arctic during winter.

2. Observations and Model-generated Data

2.1. Observations

The SHEBA experiment is described in the SHEBA operations plan (http://sheba.apl.washington.edu/sheba/ShebaOpPlan.html). Observations from the SHEBA field program commenced in October 1997 as the Canadian ice breaker, the Des Groseilliers, was moored to a large multiyear ice floe in the Beaufort Sea. Data collected during the first data intercomparison period (DIP), November 1–23, 1997, are used in this study. On November 1, 1997 the ship was located at 75.8°N, 145.0°W. Over the next 23 days the ship drifted with the ice floe about 80 km toward the northwest, reaching 76.2°N, 147.6°W by November 23, 1997.

Various observational instruments were operated at a meteorological site about 250 m from the ship through a cooperative effort between NOAA/Environmental Technology Laboratory (NOAA/ETL), the Army Cold Regions Research and Engineering Laboratory (CRREL), and the Naval Postgraduate School (NPS). This equipment provided measurements of the atmospheric surface layer, surface and subsurface properties, and snowfall. Atmospheric surface layer measurements are obtained from a 20-m meteorological tower mounted with five levels of sonic anemometers, collocated mean temperature and humidity sensors, and one Ophir fast humidity probe. These instruments provided five levels of 1-minute mean temperatures, humidity and wind data, five levels of direct measurements of momentum and sensible heat fluxes, and one level of direct turbulent latent heat flux. For this study, the tower data has been averaged to hourly values and interpolated to 2.5 m and 10 m. The latent heat flux is determined from the surface, 2.5-m, and 10-m data using the bulk method with transfer coefficients taken from [Andreas, 1987]. The flux data has been quality controlled to eliminate data contaminated by riming. Surface radiative properties at this site were provided by pairs of Fpplpy pyranometers and pygrometers mounted approximately two meters above the surface 30 m from the base of the tower. They provided 1-min values of incoming and outgoing solar and infrared radiation and the radiative snow surface temperatures. The latter were determined following the method of Fairall et al. [1998] using the pygrometer measurements and assuming a snow emissivity of 0.99. The snow-surface temperature is assumed to be representative of the multiyear snow-covered ice floe. Though some freezing and riming of the radiometers did affect the data, the radiometers at the tower site were cleaned frequently and the measurements appear to be of generally good
and the DABUL lidar are used for determining cloud boundaries, cloud fraction, and cloud microphysical properties. Rawinsondes were launched at least twice per day to obtain profiles of temperature, moisture, and wind up to 50 hPa. These are summarized by Moritz et al. [1999]. Preliminary analysis of the cloud radar data for the November DIP have been presented by T. Uttal (NOAA/Environment Technology Laboratory). Returns from the cloud radar indicated that clouds were present about 82% of the time during November 1–23. Layer cloud fractions may also be determined; however, clouds and precipitation cannot be easily distinguished so that precipitation is classified as cloud. This will have the greatest impact on the low cloud fraction, which will likely be significantly greater than that obtained via a surface observer. The radar data indicated that high clouds (between 5–10 km) were present 34% of the time while low clouds and/or precipitation (between 0–2 km) were observed 73% of the time. Many of the observed clouds were either based at the surface or precipitating and some were layered in the vertical. The DABUL indicated that mixed-phase or liquid clouds were often present at low levels at temperatures far below freezing [Intrieri et al., 1999].

2.2. Model-generated Data

Daily forecasts from the operational European Centre for Medium-Range Weather Forecasting (ECMWF) model are used to produce the ECMWF single-column SHFRA (ECSS) data set. This ECSS data set provides 1) initialization data, 2) model comparison data, and 3) the large-scale 3-D wind field, and the horizontal and vertical advection fields required by the column model. The ECSS data describes the state of the atmosphere and surface in a column near the SHEBA site. The location of the ECSS column changes as the ship drifts with the ice floe. The ECSS data are from 76.1°N, 140°W for November 1–5 and from 75.5°N, 144°W for November 6–25. The ECSS data are given at 31 sigma levels and the surface. Data for a given day are obtained from the 12- to 36-hour ECMWF model forecast to minimize the effect of model spin-up on the large-scale tendencies. The SHEBA rawinsonde data have been assimilated into the ECMWF operational analyses prior to the model forecast.

The ECSS data are organized into four main categories: basic state, flux, tendency, and surface variables. The basic state variables include profiles of wind, temperature, specific humidity, cloud liquid and ice water content, cloud fraction, and vertical velocity at each hour. The flux variables, which are given at the 31 sigma levels and the surface, include hourly values of the sensible heat flux, moisture flux, net longwave flux, and net shortwave flux. The surface variables include the air temperature and specific humidity at 2 m, the $u$ and $v$ wind components at 10 m, the surface skin temperature, and the downwad shortwave and longwave fluxes at the surface. Total and parameterized tendencies of temperature, specific humidity, and $u$ and $v$ wind components are available for each sigma level. The total tendency includes both large-scale advective tendencies and the tendencies resulting from parameterized processes (e.g., radiative transfer). The basic-state and surface variables are used to initialize the ARCSYM column model as described in the next section. All four categories of ECSS data can be used as model comparison data, though this is only done sparingly in this study. Of most importance, the large-scale horizontal and vertical advective tendency required by the column model is obtained by subtracting the parameterized ECSS tendency from the total ECSS tendency.

The parameterized tendencies obtained with the ARCSYM column model are added to the large-scale ECSS tendencies to

![Figure 1](image.png)
Figure 2. Time series of advective tendencies of (a) temperature and (b) water vapor mixing ratio at 975 hPa. The ECSS advective tendencies are given as hourly (thin solid lines) and daily averages (dashed connected by thick solid line). The ECSS and ARCSYM physical tendencies (thick solid line and thick dashed line, respectively) are given as daily averages. The ECSS large-scale vertical velocity at 975 hPa is given in (c).

determine the total tendencies during the simulation. The hourly ECSS tendencies are assumed to vary linearly between hours. Using the ECSS wind data, the horizontal winds are specified at 12-hour intervals and the vertical velocities are specified at 1 hour intervals. The three-dimensional wind field is assumed to vary linearly with time between the intervals.

Figure 2 shows the ECSS large-scale advective tendencies and the physical tendencies of temperature and specific humidity from both ARCSYM and the ECSS data for the entire DIP. The tendencies in Figures 2a and 2b are shown at 975 hPa, roughly corresponding to 290 m, which is a height within the low-level arctic inversion and generally above any surface mixed layer. The ECSS advective tendencies show a mixture of low and high frequency variability. The 3-D temperature advection tendency at 975 hPa varies between $-26$ and $+40$ K d$^{-1}$. Temporal variations in temperature advection at 875 hPa (not shown), which is
generally near the top of the low-level inversion, are 50% less than at 975 hPa. Variations in the 3-D specific humidity advection tendency are ±10 g kg\(^{-1}\) d\(^{-1}\) between 975 hPa (see Figure 2b) and 700 hPa with the magnitude of the variations decreasing with height above 700 hPa. The advective tendencies are often coherent in the vertical for depths of over 200 hPa (not shown). The ECSS vertical velocities (Figure 2c) are about ±1 cm s\(^{-1}\), at 975 hPa. Larger vertical velocities of up to 10 cm s\(^{-1}\) occur aloft near 200 hPa, with vertical coherence over depths similar to those found in the advective tendencies. The ECSS advective tendencies are inferred to be reasonable for the period November 1-23 based on comparisons between ECSS and observed mean temperature and moisture profiles. The ECSS physical tendencies are similar in magnitude and often of a sign opposite the ECSS advective tendencies. This is evident when comparing daily average physical and advective tendencies from the ECSS data set (Figures 2a,b). In the ECSS data set the total tendency at the SHEBA gridpoint is alternately dominated by periods of large ECSS advective tendency and large ECSS physical tendencies. At times, the physical tendencies for the ARCSYM single-column model deviate from those of the ECMWF model. The physical temperature tendency of the ARCSYM model at 975 hPa is opposed in sign to the ECMWF tendency on November 4 and a 3-day period beginning November 18. For the physical moisture tendency at 975 hPa (Figure 2h), the ARCSYM and ECMWF tendencies are well correlated but there are periods when discrepancies of up to a factor of two exist (e.g., November 10–13). Implications of combining the ECMWF advective tendencies and the ARCSYM physical tendencies to obtain the total tendency in the ARCSYM column model are discussed in section 7.

### 3. Model Description

The column model employed here was developed from version 3.0 of the Arctic regional climate system model (ARCSYM) [Lynch et al., 1995]. The most recent version of ARCSYM is described in detail by Lynch et al. [this issue]. The ARCSYM is based on version 2 of the NCAR regional climate model (RegCM2), which is described in detail by Giorgi et al. [1993a, b]. It is a hydrostatic model with a terrain-following sigma coordinate system (31 vertical levels). The column represents a grid box of (90 km)\(^2\). The relevant physical parameterizations in ARCSYM are described in some detail below.

The treatment of cloud processes is separated into parameterizations of stratiform and convective clouds. The cloud water amount in stratiform clouds is predicted using a bulk microphysics scheme [Hsie et al., 1984]. Cloud water and precipitation are treated with two prognostic equations. A temperature threshold is used to determine whether the clouds and precipitation are liquid or ice. In the baseline case, cold cloud microphysical processes are implemented at temperatures below 273.15 K, as is currently implemented in the 3-D version of ARCSYM. Convective cloud processes are determined using the scheme developed by Grell [1993].

A cloud fraction parameterization is necessary for radiative transfer. For stratiform clouds a cloud fraction of 0 or 1 is assumed depending on whether or not the cloud water mixing ratio exceeds some threshold. This assumption is made because the column represents an area much smaller than the scale of stratiform clouds. In this study a cloud water mixing ratio greater than 10\(^{-4}\) g kg\(^{-1}\) corresponds with a cloud fraction of 1. The cloud fraction obtained from the convective cloud scheme is allowed to vary between 0 and 1.

Longwave radiative transfer is determined with the CCM2 radiation scheme [Riegle et al., 1992]. This scheme is a modified Maxmuus random-band model that treats gaseous absorption by H\(_2\)O and CO\(_2\) and most of the trace gases. Clouds are treated as gray bodies, with the absorption coefficient being a function of cloud particle phase. The cloud emissivity is given by

\[
\epsilon = 1 - \exp(-\alpha WP)
\]

where \( WP \) is the liquid or ice water path, and \( \alpha \) is the absorption coefficient, where \( \alpha = 0.0735 \text{ m}^2 \text{ g}^{-1} \) for ice and 0.1 m\(^2\) g\(^{-1}\) for liquid. Random overlap has been assumed to deal with multiple cloud layers in the vertical. The shortwave parameterizations are not described since the Sun is at or below the horizon during the period of interest.

The rapid radiative transfer model (RRTM) developed by Mlawer et al. [1997] is tested during the sensitivity analysis. This model uses the correlated \( k \) method for 16 bands to treat the radiatively active gases (H\(_2\)O, CH\(_4\), CO\(_2\), CO, N\(_2\)O, O\(_3\), and O\(_2\)) over wavelengths from 3.3 to 1000 \( \mu \)m. This longwave scheme has been shown to perform well under extremely cold, clear-sky Arctic conditions [Pinto et al., 1997].

Various options for parameterizing the optical properties of clouds have recently been implemented into RRTM. The parameterization of the longwave radiative properties of clouds ranges from the simple CCM2 gray body treatment to spectrally dependent treatments in which the cloud radiative properties are a function of particle phase and size [Hu and Stamnes, 1993, Ebert and Curry, 1992]. For the detailed treatments the effective radii characterizing liquid and ice clouds have been set at 7 and 40 \( \mu \)m, respectively. The effect of multiple cloud layers is determined assuming random overlap.

The turbulent exchange of heat and moisture between the surface and the atmosphere is obtained with bulk aerodynamic formulae. The surface-air temperature and specific humidity in these formulae are determined by scaling the quantities at the lowest sigma level, which fluctuates between 27 and 30 m, to their surface air values. The turbulent exchange coefficients are determined to be a function of the near-surface stability following the method of Louis [1979]. Fluxes over open water and sea ice are area weighted to obtain the grid box average flux to the atmosphere. Atmospheric boundary layer processes are parameterized following Holtslag et al. [1990].

Two treatments of the surface are employed. In one set of simulations the surface temperature is specified with observations obtained at the SHEBA ice station. In another set of simulations the surface is coupled with the atmosphere using the sea ice thermodynamics of Parkinson and Washington [1979]. In all simulations, the surface is initialized as a mixture of open water and sea ice assuming a climatological sea ice concentration of 95% an ice thickness of 2.0 m, and a snow depth of 0.2 m based on observations. The ocean surface temperature is specified to be the freezing point of seawater, 271.2 K. In the runs with specified surface temperatures sea ice thickness, ice concentration and snow depth are constant. The time series of surface temperature used in the specified runs is shown in Figure 1. In the coupled runs, sea ice concentration, snow water equivalent, and snow depth are a function of the sea ice thermodynamics. Sea ice dynamics cannot be calculated with a SCM and sea ice divergence was not yet available from observations; therefore, the sea ice concentration and thickness is a function of the sea ice thermodynamics only.

The Parkinson and Washington [1979] scheme accounts for three surface types: snow-covered sea ice, bare ice, and open water. The temperature at the top of the snow layer is determined
4.1. Components of the Surface Energy Budget

Turbulent fluxes of heat and moisture in the surface layer affect the properties of the boundary layer. Time series of the observed and modeled sensible and latent heat flux over sea ice are shown in Figure 4, and mean values are given in Table 1. Negative values indicate a turbulent flux toward the surface. It is seen that the magnitude of the sensible heat flux in the column model is much greater than observed. The mean value of the modeled sensible heat flux for the period (-25.3 W m\(^{-2}\)) differs from the observed mean value (-1.0 W m\(^{-2}\)) by more than an order of magnitude. This large bias results from an excessively warm boundary layer. The situation is exacerbated when standard interpolation procedures are used to determine the 2-m air temperature from the temperature at the lowest model sigma level (near 28 m).

Discrepancies between modeled and observed latent heat flux are also evident (Figure 4b). The observed latent heat flux has been estimated using the bulk aerodynamic formula for surface fluxes assuming that the surface is saturated with respect to ice and using the observed 2.5-m specific humidity and 10-m wind speed. A negative bias in the modeled latent heat flux is evident in the time series and averaged over the simulation (see Table 1). The magnitudes of the modeled and observed latent heat fluxes are significantly different.

4. Baseline Simulation

The baseline simulation is performed using ARCSYM with specified ECSS advective tendencies, specified surface conditions, CCM2 radiative transfer [Briggs, 1992], bulk cloud microphysics of Hsie et al. [1984], Grell convection and the atmospheric boundary layer scheme of Holtslag et al. [1990]. Cloud microphysics and radiation are dealt with assuming that only ice exists at temperatures below 273.15 K. The simulation is essentially a 23-day forecast which is nudged toward ECSS mean fields via the advective terms. The results of the simulation are compared with observations obtained with rawinsondes, the 20 m tower, the cloud radar and DARTII.

Figure 3. Temperature and relative humidity profiles observed on October 31 at 2315 UTC (dots) and those obtained from the ECSS data set for November 1 at 0000 UTC (dashed lines).

Figure 4. Time series of modeled (solid line) and observed (dots) (a) sensible heat flux and (b) latent heat flux at the surface. The observed sensible heat flux was obtained via eddy correlation, while the latent heat flux was determined from the bulk aerodynamic equation using the observed 10-m wind speed and 2-m water vapor mixing ratio and assuming an exchange coefficient of 0.0012. Negative values represent fluxes toward the surface.
Table 1. Observed and Modeled Parameters Averaged Over the November 1–23 Period.

<table>
<thead>
<tr>
<th>Run</th>
<th>Temperature, K</th>
<th>Mixing Ratio, g kg$^{-1}$</th>
<th>Cloud Fraction</th>
<th>Water Path, cm</th>
<th>Accum. Precip.*, cm</th>
<th>Downward Longwave, W m$^{-2}$</th>
<th>Net Longwave, W m$^{-2}$</th>
<th>Sensible Heat, W m$^{-2}$</th>
<th>Latent Heat, W m$^{-2}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Observed</td>
<td>255.0</td>
<td>0.710</td>
<td>0.82</td>
<td>N.A.</td>
<td>0.28 (0.45)†</td>
<td>217.2</td>
<td>-15.7</td>
<td>-1.0</td>
<td>0.48‡</td>
</tr>
<tr>
<td>Baseline</td>
<td>259.0</td>
<td>0.947</td>
<td>0.64</td>
<td>14.1</td>
<td>0.48</td>
<td>171.5</td>
<td>-64.1</td>
<td>-25.3</td>
<td>-2.3</td>
</tr>
<tr>
<td>LRBL</td>
<td>259.5</td>
<td>0.942</td>
<td>0.53</td>
<td>10.1</td>
<td>0.53</td>
<td>166.8</td>
<td>-69.1</td>
<td>-21.4</td>
<td>-2.3</td>
</tr>
<tr>
<td>M250</td>
<td>258.6</td>
<td>0.894</td>
<td>0.73</td>
<td>32.0</td>
<td>0.38</td>
<td>185.7</td>
<td>-50.1</td>
<td>-23.6</td>
<td>-1.5</td>
</tr>
<tr>
<td>M255</td>
<td>258.8</td>
<td>0.905</td>
<td>0.69</td>
<td>19.4</td>
<td>0.42</td>
<td>178.6</td>
<td>-57.3</td>
<td>-24.2</td>
<td>-1.7</td>
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<td>0.946</td>
<td>0.66</td>
<td>14.7</td>
<td>0.48</td>
<td>174.8</td>
<td>-61.6</td>
<td>-25.3</td>
<td>-2.3</td>
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<td>RTMC2C</td>
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<td>0.945</td>
<td>0.65</td>
<td>15.3</td>
<td>0.49</td>
<td>192.8</td>
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<td>-25.4</td>
<td>-2.3</td>
</tr>
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<td>RTMSC</td>
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<td>0.948</td>
<td>0.64</td>
<td>13.5</td>
<td>0.50</td>
<td>179.2</td>
<td>-56.6</td>
<td>-25.3</td>
<td>-2.3</td>
</tr>
<tr>
<td>Coupled</td>
<td>259.3</td>
<td>0.971</td>
<td>0.69</td>
<td>21.3</td>
<td>0.53</td>
<td>196.8</td>
<td>35.3</td>
<td>21.2</td>
<td>-0.03</td>
</tr>
<tr>
<td>IVRBL</td>
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<td>0.936</td>
<td>0.71</td>
<td>16.1</td>
<td>0.56</td>
<td>197.0</td>
<td>-33.6</td>
<td>-16.4</td>
<td>-1.3</td>
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<tr>
<td>ECSS</td>
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<td>0.757</td>
<td>0.75</td>
<td>26.7</td>
<td>1.26</td>
<td>212.0</td>
<td>-31.8</td>
<td>-6.3</td>
<td>0.61</td>
</tr>
</tbody>
</table>

*liquid water equivalent accumulated snowfall  
†estimated from bulk aerodynamic formula  
‡obtained from Neph sensor and in parentheses optical raingage (assuming snow density of 0.1 g cm$^{-3}$).

The surface energy components obtained over the sea ice are given. Temperature and mixing ratio are at 28 m with observations from rawinsonde data. Fluxes are obtained at the surface with flux observation from 20-m tower. Abbreviations are LRBL, low resolution boundary layer; M250 and M255, cold cloud microphysics at temperatures less than 250 K and 235 K, respectively; CCM2L, cloud radiative properties assuming CCM2 liquid bulk absorption coefficient; RTMC2C and RTMSC, rapid radiative transfer model using CCM2 cloud radiative properties and using wavelength-dependent properties, respectively; IVRBL, increased vertical resolution in boundary layer; and ECSS, from the ECMWF single-column SHEBA data set.
fluxes are both small (Table 1), though the signs of their means differ and their variation appears to be poorly correlated (Fig. 4b). The variations in the modeled latent heat flux are much larger than observed. This is particularly evident during the 4-day period beginning November 9. However, their generally small magnitudes indicate that errors in this term are of secondary importance.

The mean modeled fluxes of sensible and latent heat over open leads into the atmosphere are 105 and 50 W m\(^{-2}\), respectively. These values combined with those given in Table 1, and assuming a sea ice concentration of 95\%, give area-weighted sensible and latent heat fluxes of -18.8 and +1.3 W m\(^{-2}\), respectively. Thus the sea ice-ocean surface is modeled as a heat sink and a weak source of moisture for the atmospheric boundary layer.

The longwave radiative flux is a vital component of the surface energy budget during the Arctic winter. The downward \( F' \) net longwave \( F_n \) flux at the surface are shown in Figures 5a and 5b, respectively. The temporal variability in \( F' \) results from variations in cloud and temperature profiles (Figure 5a). A large negative bias in \( F' \) is evident throughout the simulation. This leads to overestimates of the surface cooling (Figure 5b). Recall that \( F' \) is specified using the observed surface temperature in the baseline run. The observed \( F_n \) is positive during several brief periods of time corresponding with the presence of clouds [Persson et al., 1999] while the modeled \( F_n \) is greater than zero only once. The baseline case overestimates the mean longwave cooling of the surface by over 40 W m\(^{-2}\) compared to observations (Table 1). Only a small fraction of this discrepancy may be explained by contamination due to riming of the pyrgeometers.

The longwave cooling bias is evidently caused by errors in the calculation of the clear-sky downward longwave flux and errors in the modeled cloud properties. The modeled clear sky \( F' \) is about 20 W m\(^{-2}\) less than observed, as estimated from Figure 5a. Because cloud optical properties were not measured directly during the November DIP we use \( F' \) greater than 240 W m\(^{-2}\) as a proxy for optically thick clouds. The observed \( F' \) exceeds this threshold about 31% of the time. Persson et al. [1999] have related these periods of time to the occurrence of optically thick liquid water clouds in the boundary layer. For comparison, a similar method is used to analyze the model results. Here a threshold of 220 W m\(^{-2}\) is used to diagnose the presence of optically thick clouds. This lower threshold is used to compensate for the 20 W m\(^{-2}\) clear-sky bias in the CCM2 radiation. It is found that optically thick clouds are present in the simulation only 12% of the time, indicating that the model underestimates the occurrence of optically thick clouds in addition to having a clear-sky bias.

4.2. Boundary layer structure

Time-height plots of the modeled temperature and water vapor mixing ratio are shown in Figure 6 for the lowest 4 km of the atmosphere. The temperature field reveals several warming and cooling episodes, the former have been related to cloudy periods and increased longwave radiation at the surface [Overland and Guest, 1991; Persson et al., 1999]. Warmest temperatures occur between 0.5-1.3 km, with the warmest period being November 10-12. Similar features are seen in the water vapor field. Both fields are characterized by an inversion of fluctuating strength and depth. The strength of the inversion may be quantified according to the total increase in temperature or moisture across the inversion layer. The modeled mean temperature inversion strength for the entire simulation period is 2 K, with a maximum of about 10 K. A water vapor inversion is not evident in the modeled mean water vapor profile, but water vapor inversions in the instantaneous model profiles were as large as 0.6 g kg\(^{-1}\). The modeled boundary layer stability varies widely from strongly stable surface-based inversion layers to elevated well-mixed layers with and capping inversions.

The modeled evolution of boundary layer temperature structure is evaluated using 12 hourly rawinsonde data. Data from rawinsondes within ±5% of the model level were averaged to obtain an observation for a given model height. It is evident in Figure 7 that the modeled 28-m temperature is consistently warmer than observed with a mean bias of about 4 K (see Table 1). The modeled boundary layer is warmer than observed, but the warm bias thickens with height as seen at 780 m. A much hotter winter.
between modeled and observed surface turbulent fluxes are caused by errors in the implementation of the bulk aerodynamic formula when strong surface-based inversions are present. The surface heat flux is proportional to the difference between the surface temperature, $T_s$, and the surface air temperature, $T_a$ (typically the air temperature 2 m above the surface). In the model, the surface sensible heat flux is found using $H_s = \rho c_p C H / (T_s - T_{\sigma 0})$ where $T_{\sigma 0}$ is the temperature at the lowest model sigma level and $\chi$ is a correction factor that adjusts $T_{\sigma 0}$ to that at 2 m assuming an adiabatic lapse rate. This correction factor is typically equal to about 1.01. The validity of this approach is tested by comparing the observed surface-air temperature difference to that used by the model.

The approximate modeled surface-air temperature difference, $T_s - T_{\sigma 0}$, varies between -15 and +4 K, with the difference exceeding -7 K 43% of the time and the positive values occurring about

Figure 6. Modeled (a) temperature and (b) water vapor mixing ratio as a function of time and height for the baseline case.

Figure 7. Time series of modeled 3-hourly air temperature (solid line) from baseline case and observed air temperature (diamonds) obtained from rawinsonde data at several heights. Temperatures from rawinsonde data have been vertically averaged around the model height indicated.
5% of the time. In the observations, $T_z - T_{IM}$ is always negative, but the magnitude of the difference is never greater than 7 K (Figure 1). This error in the determination of the surface–air temperature difference in the model is a major cause of error in the modeled turbulent heat fluxes at the surface which contributes to differences between modeled and observed boundary layer structure. Other errors arise from using a $C_H$ and $U$ that have not been scaled to 10 m height. Further, the stability dependent exchange coefficient, $C_H$, is not well-known under strongly stable conditions.

4.3. Clouds and precipitation

The simulated cloud water and precipitation mixing ratios are shown in Figure 8. In contrast to the lidar observations, the modeled clouds are composed entirely of ice since the modeled temperatures within the cloudy regions are below the cold cloud threshold of 273.15 K. The modeled cloud ice water mixing ratios are typically less than 0.05 g kg$^{-1}$ and snow mixing ratios are less than 0.01 g kg$^{-1}$. A total cloud fraction may be diagnosed from the vertical integral of the predicted cloud and precipitation water amount or total water path. The snow water path has been included in the determination of cloud fraction to be consistent with the radar observations. Cloud is assumed to be present if the cloud optical depth (from the total water path) exceeds the background aerosol optical depth (i.e., clouds having a significant impact on the downward radiative flux at the surface). A background aerosol optical depth for the central Arctic of 0.25 is assumed based on aerosol profiles given by Key [1995]. Using the CCM2 absorption coefficients for ice and liquid indicates that the cloud water path must exceed 2.3–3.3 g m$^{-2}$ before the radiation field is affected significantly beyond that expected from aerosols. Since all clouds are crystalline in this simulation we use the threshold of 3.3 g m$^{-2}$ throughout the column. Using this criteria for the presence of cloud a mean cloud fraction of 64% is obtained from the modeled cloud water and precipitation fields shown in Figure 8. Much of the underestimated cloud fraction can be linked with a severe underestimate in the model low cloud fraction (28% compared with observed value of 73%). The modeled cloud fraction above 5 km (45%) is somewhat greater than that observed (34%). The model does produce several multiple layer cloud events which are evident in Figure 8 and precipitation often fills the column.

Modeled precipitation is predominantly crystalline. The snow often sublimes as it falls into the relatively dry lower atmosphere; however, there are four snow events at the surface. Some snow does melt as it falls through a warm layer near the surface on November 11 (see Figure 6 for the location of warm layer). The four simulated precipitation events are also illustrated in the time series of liquid water equivalent precipitation rate (Figure 9). The modeled precipitation rate and accumulation may be compared with data collected with the optical rain gauge and the Nipher shielded snow gauge system. The optical rain gauge data have been scaled by a factor of 0.1 g m$^{-2}$ (assumed density of snow) to obtain liquid water equivalent values. The two observation methods agree fairly well after optical rain gauge data obtained under high winds are removed. Both measuring systems indicate that four significant events took place during the November 1–23 period. The timing of each event is well simulated by the model; however, the total accumulated amount of precipitation is almost 50% greater than observed with the Nipher system. Part of this discrepancy may be due to uncertainties in the measurements as the shielded Nipher system tends to underestimate snowfall by a factor of 2 in the Arctic (D. Moritz, personal communication, 1998).
Figure 9. Time series of liquid water equivalent accumulated precipitation from Nipher shielded snow gauge (dashed line), optical rain gauge (dash-dotted line), and baseline run (dashed line) and precipitation rates from optical rain gauge (dots) the model simulation (solid line). The observations were obtained by an optical rain gauge. Modeled precipitation rates are 3-hourly instantaneous values, while observed rates are hourly mean values.

4.4. Thermodynamic structure

The modeled temperature and water vapor mixing ratio profiles are compared with rawinsonde data to assess the simulated thermodynamic structure of the atmosphere. Comparisons between the modeled profiles and the observations are made for each day at 1145 and 2345 UTC, the nominal rawinsonde launch times. The ability of the model to simulate vertical variations in temperature and moisture is assessed. Observed and modeled mean profiles for the entire period are also compared.

Averaged over the length of the simulation the modeled temperatures are too cold aloft and too warm near the surface and the model water vapor mixing ratios are too small throughout the profile (not shown). An example of these biases is seen on Julian day 316, temperatures below the inversion layer are 2–3 K too warm and temperatures aloft are up to 10 K too cold (Figure 10c). These biases effectively smooth out the inversion layer resulting in an underestimated inversion strength. Hence, the bulk stability of the atmosphere is too small in the model which results in excessive mixing between the free atmosphere and the boundary layer. The inability of the model to reproduce the proper thermodynamic structure of the atmosphere is exacerbated by potential positive feedbacks between weak bulk stability of the atmosphere and vertical mixing. The mean water vapor mixing ratio is most poorly simulated between the surface and 2 km where a water vapor inversion is observed. At the peak of the observed water vapor inversion, the mean modeled water vapor is 40% less than observed. These features are extremely important in the formation of low clouds.

The coarse initial conditions obtained from the ECSS data set do not have the pronounced vertical variations observed in the temperature and moisture fields (see Figure 3). The observed surface-based temperature inversion is missed entirely with no hint of the inversion in the simulated temperature field after 24 hours (Figure 10a). Large vertical fluctuations in the observed initial water vapor profile below 850 hPa are smoothed considerably in the model initial conditions. Despite the poor initial conditions the model is able to produce some of the observed structure in the vertical profiles.

The modeled temperature profiles often exhibit large-scale structure similar to that observed, but smaller scale variations are often missed. The model has increased difficulty in simulating more complicated temperature profiles (e.g., Figure 10b). Errors in this profile may be related to errors in the simulated cloud field including their occurrence, vertical distribution, and radiative properties. The model is able to properly develop a cloud top inversion layer if the modeled cloud layer is accurately depicted (e.g., Figure 10c). The general shape of the temperature profile formed under strong clear-sky radiative cooling conditions is well simulated (e.g., Figure 10d), however, the same vertical distribution of biases is evident with the profile being 2–5 K too cold aloft and up to 11 K too warm in the boundary layer.

The modeled water vapor mixing ratio profile lacks the small scale structure that is observed; however, the model does seem to capture the larger-scale features. For example at 1200 UTC on Julian day 305, the observed standard deviation in the surface to 900 hPa water vapor mixing ratio of 0.19 g kg⁻¹ is much larger than the 0.06 g kg⁻¹ obtained by the model (Figure 10e). This smoothing of the vertical variations in the modeled water vapor profiles holds throughout the simulation and leads to an underprediction of saturated conditions in the modeled boundary layer. On Julian day 316, the modeled vertical structure in the water vapor mixing ratio is similar to that observed, with a strong inversion based at 920 hPa (Figure 10g).

5. Sensitivity Studies

Several simulations have been conducted to determine the utility of cloud, radiation, and boundary layer parameterizations in the Arctic. Each simulation is identical to the baseline case except for the change made to the parameterization of interest. Mean quantities obtained for each simulation are given in Table 1. Mean values obtained from the ECSS data are also given in Table 1 for comparison. The mean ECSS temperature and moisture profiles compare very well with the observations (not shown), this includes the boundary layer as evidenced by the mean values at 28 m. The ECSS cloud fraction is also less than observed while the mean downward longwave flux at the surface from ECSS data set is within 2.5% of the observations. However, the magnitude of net longwave cooling in the ECSS is much greater than observed indicating a problem with the modeled surface temperature (Table
Figure 10. Profiles of modeled temperature and water vapor mixing ratio from baseline case (thin lines) and from rawinsonde observations (bold lines) at two times for November 1, 6, 12, and 16 (Julian days 305, 310, 316, and 320, respectively). The earlier times (typically around 1145 UTC) are indicated by solid lines and the later times (around 2345 UTC) are indicated by dashed lines. Horizontal dashed lines denote modeled cloud locations.

1) The ECSS sensible and latent heat fluxes are also much better simulated than in the baseline ARCSYM run. Possible reasons for such discrepancies are briefly discussed in section 7.

5.1. Cloud parameterizations

To determine the importance of convection in the simulation, the model is run with the parameterization for convection turned off. The results for this run and the baseline case which uses the Grell scheme [Grell, 1993] are identical, indicating that convection does not occur during the simulation.

In a second set of cloud sensitivity runs the temperature at which the cold cloud microphysical processes are initiated is reduced. Two simulations are performed using temperature thresholds for cold cloud microphysics, $T_c$, of 250 and 255 K (hereafter referred to as M250 and M255). The cloud radiative properties are the same as in the baseline case (i.e., the bulk cloud absorption coefficient for ice is used) so that differences between these simulations and the baseline run are directly related to the treatment of cloud microphysics. Observations indicate that many low clouds in the Arctic contain ice at temperatures above 250 K; however, an abundance of supercooled water is often present as well [Curry et al., 1997; Pinto, 1998]. The reduction in $T_c$ increases the occurrence of supercooled liquid water in clouds. The increased presence of supercooled water allows for liquid water paths and cloud lifetimes to increase due to the relative in-
efficiency of precipitation production processes in liquid clouds. This is similar to the findings of Beesley and Morice [1998]. The cloud water path in the M250 run is 2.2 times greater than that in the baseline case and the cloud fraction increased by 0.09. At the same time, the total amount of accumulated liquid water precipitation in the M250 run has been reduced by 20%. The presence of liquid clouds reduces precipitation, thus prolonging their existence and allowing them to have a greater influence on the surface energy budget through modulation of the radiative fluxes.

The increased presence of optically thick clouds results in a \(14.7 \, \text{W m}^{-2}\) increase in the downward longwave flux at the surface in M250. Because the cloud radiative properties have been held constant, this increase is related to increases in the water path and cloud fraction. These effects are slightly reduced in the M255 simulation.

Changes in the treatment of the cloud microphysical properties are far-reaching. This is demonstrated in the mean fields given in Table 1 for M250. The low-level water vapor mixing ratio in run M250 is 5.5% lower than that in the baseline case. This can be related to an enhanced depletion of water vapor in liquid cloud processes. This reduction in boundary layer moisture results in a decrease in the downward flux of latent heat.

5.2. Radiation

Three simulations are performed to study the model sensitivity to the treatment of cloud optical properties and clear-sky longwave radiative transfer. The three simulations are performed using (1) CCM2 longwave scheme assuming liquid cloud optical properties (CCM2L), (2) RTM with baseline CCM2 cloud optical properties (RTMC2C), and (3) RTM with spectrally dependent cloud properties (RTMSC).

The CCM2L run is used to determine the sensitivity of the downward longwave flux at the surface to the treatment of cloud optical properties in the CCM2 scheme. Comparison between the CCM2L run and the baseline case (in which clouds were predominantly crystalline) reveals that the radiatively, errors in the assumed cloud phase accounts for an uncertainty in the downward longwave flux at the surface of only 3.3 W m\(^{-2}\). This because the liquid and ice absorption coefficients used in the CCM2 scheme are similar. Therefore simulations using CCM2 cloud radiative properties will be fairly insensitive to changes in cloud phase. This insensitivity of the model to phase changes may be important when considering climate change scenarios where potential feedbacks between temperature and cloud phase must be properly simulated.

The effect of using a completely different parameterization of longwave radiation is determined by comparing the baseline simulation with the RTM2C run. The mean downward longwave flux at the surface in the RTM2C run is 20 W m\(^{-2}\) greater than that in the baseline run (Table 1). This dramatic improvement is due to the improved treatment of clear-sky radiative transfer.

The RTMSC run is compared with the RTM2C run to determine the model's sensitivity to the treatment of cloud optical properties. On average, the downward longwave flux at the surface in the RTMSC run is much less than in the RTM2C run (Table 1). The time series of downward longwave radiation at the surface for the two runs reveals that the two runs differ substantially only when clouds are present (Figure 11). The instantaneous downward longwave flux in the RTM2C run is as much as 30 W m\(^{-2}\) greater than in the RTMSC run (see Figure 11) due to differences between the bulk and spectral treatment of the absorption coefficient for cloud ice (since temperatures are below 273.15 K).

This comparison indicates that the spectral treatment of cloud radiative properties results in a much lower bulk emissivity for ice clouds than is used in the RTM2C run. The downward longwave fluxes in the RTM2C run are closer to the observations than that obtained with the RTMSC run because of compensating errors in the RTM2C run, namely, assuming the wrong cloud phase and an erroneously large absorption coefficient for ice. Further sensitivity studies with RTM indicate that the surface radiative fluxes in simulations using the spectral treatment of cloud radiative properties are much more sensitive to the treatment of cloud phase than those using the gray absorption coefficients.

5.3. Boundary layer

The importance of the treatment of the boundary layer is illustrated by comparing results from two different boundary layer schemes. The two schemes used in this study are described by Giorgi and Bates [1989] (LRBL) and Holtslag et al. [1990]. Both boundary layer schemes use eddy diffusivity to relate mean profiles to vertical flux through a layer. The Holtslag et al. [1990] scheme employs a counter-gradient term to allow for non-local mixing and is employed only within the diagnostically determined depth of the boundary layer. Results of the LRBL run are significantly different from those obtained in the baseline case which employed the boundary layer scheme of Holtslag et al. [1990] (Table 1). The magnitude of the sensible heat flux in the LRBL run is less than in the baseline case. This reduction leads to a warming of the lowest model level, which warmed by 0.5 K over the baseline case. At the same time, the mean latent heat flux remained unchanged so that the low-level water vapor mixing ratio did not change significantly. The net effect is a lowered relative humidity in the boundary layer which affects the simulation of clouds and precipitation. The LRBL cloud fraction and cloud water amount were reduced over the baseline run, while the LRBL accumulated precipitation increased over the baseline run.

The reduced cloud fraction and cloud water amounts lowered the downward longwave flux at the surface by nearly 5 W m\(^{-2}\).

The importance of vertical resolution near the surface on the surface fluxes is tested by adding eight levels in the lowest 400 m of the model. The model setup is identical to the coupled run discussed below except for an increased number of levels.
below 400 m. In this sensitivity study, two layers are added below the lowest sigma level in the previous simulations. The lowest three sigma level are now at 6, 12, and 28 m. Despite a reduction in the downward sensible heat flux at the surface in this run, the magnitude of this flux is still much too great (Table 1). The modeled mean 28-m air temperature is improved significantly over that obtained in the other simulations while other mean variables are nudged only slightly in the direction of the observations (e.g., cloud fraction) compared with the coupled run described below.

6. Coupled Simulation

A coupled simulation is performed using the optimal combination of parameters determined from the previously described sensitivity studies. The sensitivity studies indicated that the temperature for cold cloud microphysics should be reduced to 255 K and that the longwave radiative transfer should be handled using a version of RRTM. The lower threshold for cold cloud microphysics allows for supercooled clouds to exist. The version of RRTM with a spectral treatment of clouds, which performed well when supercooled liquid water clouds were allowed to form (results not shown) is implemented. The surface temperature is determined from the surface energy budget following Parkinson and Washington [1979].

The simulated 3-hourly surface temperature is compared with the hourly mean observed surface temperature (Figure 12). It is seen that the low-frequency variability is captured reasonably well; however, a period of excessive warming is evident in the simulation on November 9. During this period the modeled surface temperature is as much as 8 K warmer than observed. The coldest period is captured by the model as well; however, the surface temperatures during this time average about 3 K too warm. The model has less skill at simulating the higher-frequency variations. The modeled surface temperature averages about 1.4 K warmer than the observed surface temperature.

The evolution of the surface energy budget for the coupled simulation is shown in Figure 13. The sensible heat and conductive fluxes are generally anticorrelated with the net longwave radiative flux. This indicates that the model sets up a strong downward turbulent flux of heat toward the surface to offset strong radiative cooling when skies are clear. A much different scenario is operating under cloudy skies.

Using the observed snow surface temperature and ice-snow interface temperature, a conductive heat flux may be estimated from the observations. A mean conductive heat flux of 17.1 W m$^{-2}$ is estimated assuming a thermal conductivity of snow of 0.31 W m$^{-1}$ K$^{-1}$ and a constant snow depth of 0.2 m (same as in the model). The mean modeled conductive heat flux at 8.6 W m$^{-2}$ is much less than observed. This discrepancy is caused by an underestimate of the sea ice-snow surface temperature difference in the model. The modeled conductive flux was positive throughout the simulation, indicating that the sea ice surface tended to warm the snow layer from below. This is similar to that seen in the observations.

The modeled mean net surface heat flux is negative (−8.8 W m$^{-2}$), indicating a general surface cooling in the model during the DIP (Figure 13b). The magnitude of the modeled surface cooling is generally less than that observed. Its mean is about 65% of the observed mean net surface energy flux of −8.8 W m$^{-2}$. However, periods of positive net surface heat flux occur in the simulation, especially during the period November 7-15 (Figure 13b). Periods of positive net heat flux are also seen in the observations at this time and towards the end of the DIP. In the model, the periods of positive net heat flux are generally periods of

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**Figure 12.** Time series of 3-hourly surface temperature from the coupled simulation (solid line) and hourly-mean surface temperature determined from pygerometer measurements from the 20-m tower (dots).

**Figure 13.** Time series of (a) the surface energy budget terms obtained from the coupled simulation including the net longwave radiative flux (solid line), the sensible heat flux (dashed line, positive values are toward the surface), and the conductive heat flux (dash-dotted line), and (b) the modeled (solid line) and observed (dots) net surface energy budget flux. The latent heat flux, which is much smaller than the plotted terms, has been omitted for clarity.
near-zero net longwave radiation and positive conductive flux and sensible heat flux. November 10, (and arguably November 19 and 22) is an exception, when the positive net heat flux coincided with a negative net longwave radiation, a positive conductive flux and a very large positive sensible heat flux. The modeled net surface heat flux on this day was much larger (even of opposite sign) than the observed net surface heat flux. The periods of near-zero net longwave radiation correspond to peaks in downward longwave radiation and warming events (not shown, but see Persson et al. [1999]). The negative periods correspond with cooling at the surface and are generally associated with modeled clear-sky periods. The modeled period of warmer surface temperatures from November 10-13 (Figure 12) occurs during the long period of positive net surface heat flux (Figure 13b). The mean modeled sensible and latent heat fluxes in the coupled simulation are significantly improved over the baseline case (Table 1); however, the magnitude of the mean sensible heat flux is still much too large, as the above discussion suggests.

The cloud water mixing ratio field for the coupled run is shown in Figure 14. The cloud microphysical properties differ significantly from those obtained in the baseline run. The similarities with the M255 run suggest that this is primarily due to lowering the temperature at which liquid cloud processes are allowed to occur. It is seen that lowering this temperature allows the cloud water mixing ratio in the lower clouds to increase. The vertical transition from liquid to ice clouds is clearly evident in Figure 14, where ice clouds are the lighter gray colors aloft. The height of the liquid to ice transition varies appreciably with time as the troposphere warms and cools. The total cloud fraction is closer to that observed as seen in Table 1 due to an increase in the low cloud fraction (cloud between 0-7 km) to 0.38. The mean total water path is also significantly greater than in the baseline case (Table 1). The effect of these increases in low cloud fraction and cloud water path (i.e., optical thickness) along with the effect of an improved treatment of clear sky radiative transfer is evident in the mean downward longwave flux at the surface which is 25 W m$^{-2}$ greater than in the baseline case.

7. Discussion

The baseline simulation using the ARCSYM single column model and the array of sensitivity tests have revealed some of the model improvement priorities necessary for accurate simulations of the wintertime Arctic atmosphere. The baseline simulation shows that modeled temperatures below the Arctic inversion are too warm and that the temperatures above the inversion are too cold. The ARCSYM baseline simulation produced too few clouds, especially at levels below 2 km. The downward longwave radiation is too small, partly because of the lack of optically thick clouds in the model and partly because of problems with the clear sky radiation scheme. The effect of the small downward radiation is a net surface radiation that is too small. The sensible heat flux responds in the warm lower troposphere by greatly increasing the downward heat flux over that observed. Some difficulties also exist in the latent heat flux, but the magnitudes of this term are so small that errors in the other terms dominate. Despite the lack of clouds, the modeled precipitation was greater than that observed.

The sensitivity tests showed that lowering the temperature threshold for cold cloud microphysics from 273.15 K to 255 K or 230 K produced noticeable improvements in the model simulation. This change significantly improved the downward longwave radiation because allowing abundant supercooled water increased the liquid water path and the longevity of the clouds (i.e., cloud fraction). This also decreased the precipitation and the low level water vapor mixing ratio. Another test showed that simulations using CCM2 cloud radiative properties will be fairly insensitive to changes in cloud phase. The implementation of an improved parameterization of longwave radiation significantly improved the downward longwave flux at the surface. Surprisingly, the inclusion of a spectral treatment of cloud radiative properties into same
advanced scheme resulted in a less accurate simulation. However, when this spectral treatment of clouds was combined with a lower temperature threshold for cold cloud physics and a surface energy budget scheme in the coupled simulation, the best results were obtained. Some minor improvements on these coupled results were obtained by adding eight levels to the lowest 400 m of the model. Although this simulation did nudge several of the parameters in Table 1 closer to their observed values, the improvements were only significant for the 28 m air temperature and the surface sensible heat flux. The low-level temperature was still too warm, the net longwave radiation was still too low, and the turbulent sensible heat flux transported too much energy to the surface. Hence, some fundamental problems still exist that were not addressed in this study.

The sensitivity test in which additional model levels were added may have hinted at one possible problem. Though run IVBHL focused on improving the turbulent heat transport from the surface to the base of the Arctic inversion, the remaining errors showing temperatures too warm below the inversion top and too cold above suggest that the heat exchange in the ARCSYM model between the free troposphere and the surface based Arctic inversion below is too large. Unless carefully considered, spurious vertical heat exchange may occur in a model across a sharp temperature maximum or minimum due to vertical numerical diffusion included to control model noise. The impact of such numerical diffusion will be greater if the vertical resolution of the feature is poor. Hence, for the Arctic, improving the vertical resolution near the top of the inversion may provide a larger impact than improving the resolution near the inversion base. Similarly, improving the vertical resolution at the levels of the radiatively important clouds will allow the representation of thinner clouds, which may still have a large optical depth if they consist of supercooled liquid water. The radiatively important clouds are also likely to be principally located near the top of the Arctic inversion.

Though the sensitivity tests performed on the ARCSYM model have highlighted some physical processes important for improving simulations of the Arctic atmosphere, the results from the ARCSYM model have generally been significantly worse than those from the ECMWF model as given by the ECSS data (Table 1). However, several aspects of the experiment design may have influenced the results of the ARCSYM. Because the momentum field is entirely based on the ECMWF model, while the mass field (temperature and specific humidity) is based on the combined tendencies from the ECMWF three-dimensional model and the column ARCSYM, the airflow used in ARCSYM is not affected by the temperature field it develops. Hence, vertical velocities used for producing the cloud field in the ARCSYM are only those obtained from the ECMWF model and do not reflect the thermal structure developed by physical processes in the ARCSYM model. This is especially true in this case since no convection was initiated. The vertical motions are only produced by synoptic or mesoscale circulations, which are dependent on the stability in the ECMWF model, rather than convection, which would be represented by the ARCSYM physics at the SHEBA grid point. Therefore, the vertical positions and timing of the clouds are only partially determined by the ARCSYM model. Also, the ECMWF physical tendencies have an indirect but probably significant impact on the ARCSYM model results because the integrated effect of the ECMWF physical tendencies along the trajectories leading to the SHEBA grid point undoubtedly have a large impact on the ECSS adveective tendencies, which are often larger in magnitude than the ARCSYM physical tendencies at the SHEBA grid point (Figure 2). In addition, the adveective tendencies will not reflect any vertical variations the ARCSYM physics might produce unless the ECMWF physics had also produced them. As seen in Figure 2, the ARCSYM and ECMWF physics respond differently to the same forcing at times. Therefore, running the column model in this way only provides a first test of the ARCSYM physics. A design in which the ARCSYM physics could act more independently and have a greater feedback on the kinematic field would provide a better test. It is not clear whether these complex interactions between various components of the ECMWF model and the ARCSYM acted linearly, implying that the ARCSYM results would have been worse if the ARCSYM model had been more autonomous, or if highly nonlinear interactions degraded the effects of the ARCSYM physics because of inconsistencies between the various schemes in the two models.

8. Summary and Conclusions

A column version of ARCSYM has been developed and tested under Arctic winter conditions. The model is initialized and forced with a special ECMWF data set that includes large-scale three-dimensional advection tendencies of temperature and moisture. The model's ability to simulate Arctic winter conditions has been evaluated using preliminary observations from a 20 m tower and a 35-GHz cloud radar at SHEBA. Results of the coupled simulation indicated the utility of the model in simulating wintertime Arctic conditions. The sensitivity runs and coupled simulations revealed several areas that are in need of further research. Of particular importance is the model's tendency toward underestimating the bulk stability of the atmosphere and overestimating the downward sensible heat flux at the surface. These results suggest improvements are necessary in the surface flux parameterization, vertical resolution near the inversion top, and in treatment of vertical numerical diffusion. Surface turbulent flux parameterizations need to account for strong stability which effects both the turbulent exchange coefficients and the scaling of temperature at the lowest modeled level to the surface air temperature.

Despite shortfalls in using ECSS data to force the ARCSYM column model discussed in section 7, the results of improving the treatment of longwave radiative transfer and supercooled water clouds in Arctic climate simulations have been demonstrated. It has been shown that because of the extreme cold and low absolute humidity in the wintertime Arctic, the longwave scheme must be able to handle the spectral emission of radiation in the water vapor rotation bands at the far end of the infrared spectrum. The treatment of cloud phase has also been shown to be very important in Arctic simulations with transition temperatures of between 250 and 255 K resulting in the best simulation of cloud water path (as deduced indirectly from downward longwave flux) and cloud fraction. The use of such cold transition temperatures is supported by the observations of Curry et al. [1997] who reported the presence of liquid in boundary layer clouds at temperatures as cold as 255 K.

In the coupled sea-ice-atmosphere simulation the net surface energy budget flux had a mean bias of +3 W m$^{-2}$. Such an error, averaged over the cold half of the year, would result in a substantial reduction in the mean sea ice thickness [Schramm et al., 1997]. Improvements in the treatment of the stable boundary layer, radiative transfer, and cloud microphysics resulting from analysis of data collected during the Beaufort and Arctic Storrs Experiment (BASE) and SHEBA and process modeling studies will result in first-order improvements in climate simulations of the Arctic.
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