Role of radiative transfer in the modeled mesoscale development of summertime arctic stratus

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Abstract. Improvements have been made in the treatment of radiation in Version 5 of the Pennsylvania State University-National Center for Atmospheric Research (NCAR) Mesoscale Model (MM5) to simulate boundary layer stratus observed during the summertime Arctic Stratus Experiment of 1980. Shortwave radiation is treated using a two-stream, delta-Eddington approximation developed for Version 2 of the NCAR Community Climate Model. This code offers many improvements over the original radiative transfer code developed by Dudhia [1989], including a more detailed treatment of surface albedo, solar absorption by ozone, and improved treatment of liquid, ice, and mixed-phase clouds. Longwave radiative calculations are performed with the broadband radiative transfer code currently employed in Version 3 of the European Centre for Medium-Range Weather Forecasts (EC3) model. Improvements offered by EC3 include longwave radiative absorption by ozone and trace gases and the explicit radiative treatment of mixed-phase clouds.

The importance of radiative transfer in the formation of low level Arctic stratus and the evolution of an anticyclone is illustrated with three simulations: (1) baseline simulation using the Dudhia [1989] radiative transfer, (2) improved radiative transfer (i.e., CCM2 shortwave and EC3 longwave), and (3) radiative transfer neglected. The area extent of low clouds is reduced toward observed values when improved radiative transfer is implemented. The temperature, moisture, and cloud water profiles show significant sensitivity to the treatment of radiative transfer as well. Comparisons between observations and model results show that the new radiation package improves the area extent of cloud cover and the quality of the simulated surface radiative fluxes. The importance of radiative cooling in the evolution of a Beaufort Sea anticyclone is demonstrated.

1. Introduction

Arctic climate is complicated by the complex interaction between the inhomogeneous surface and the atmosphere. The effect of clouds on radiative transfer in the Arctic is complicated by the year-round presence of a highly reflective snow and ice surface, low temperatures and water vapor amounts, optically thin clouds, and the presence of multiple, often sub-grid-scale cloud layers. Feedbacks between clouds and large-scale dynamics also contribute to this complexity. Radiative cooling to space by clouds in the Arctic has been shown to be important in the formation of strong anticyclones in winter [Curry, 1987]. Atmospheric dynamics influences the heat and moisture transfer across the air-sea and air-sea ice interface by breaking up the sea ice and exposing relatively warm moist ocean surface. Fluxes of heat and moisture into the atmosphere are also a function of wind speed, and atmospheric stability which depend on large-scale dynamics.

Climate models have shown poor skill in modeling present-day Arctic climate with particular problems associated with the modeled cloud amount. Randall et al. [1985] cite a substantial underestimate of the summertime Arctic Ocean cloud cover by the Colorado State University general circulation model (GCM). Chen et al. [1995] reported discrepancies between five GCMs in the modeled longitudinally and annually averaged cloud cover as large as 60% in the Arctic. Modeled mean wintertime cloud cover varied from 2 to 75%, and summertime cloud cover varied from 5 to 95%. In an intercomparison of GCM results, V.M. Kattsov et al. (unpublished manuscript, 1995) showed modeled Arctic Ocean June surface insolation values to range from 85 to 185 W m\textsuperscript{-2} compared with 310 W m\textsuperscript{-2} observed, indicating a substantial overprediction of cloud amount in the models. These findings are not surprising considering that the parameterization of clouds, radiative transfer, and other processes involved in the life cycle of clouds (e.g., precipitation, boundary layer, and surface processes) have been developed for environments with vastly different thermodynamic characteristics than are found in the Arctic. Lynch et al. [1995] have developed a regional climate model for the western Arctic to examine the sensitivity of modeled regional climate to the parameterizations of physical processes. They found that surface air temperature, an atmospheric parameter often used to describe climate, was quite sensitive to their choice of parameterization of moisture processes in their model.

What is known of low-level Arctic clouds has been ascertained through a limited number of field experiments and modeling studies. Herman and Goody [1976] considered the formation of the summertime Arctic stratus clouds in the relatively warm and moist continental air as it flows over the pack ice and inferred that condensation is induced in an initially unsaturated air mass due to radiative cooling through emission to space and diffusive cooling to the colder surface. Support for this formation mechanism has been provided by the June 1980 Arctic Stratus Experiment (ASE) as discussed by Curry et al. [1988]. Jaya and Ohtake [1971] first described the layered structure of low-level clouds in the Arctic. It appears to be common for summertime Arctic stratus to occur in a number of well-defined layers separated by intervening clear regions which are several hundred meters thick. Herman [1975] and Herman and Goody [1976] cite reports of as many as five distinct, overlapping cloud layers in the lowest 2 km of the atmosphere.
The typical summertime Arctic stratus cloud system consists of a stable fog layer surmounting a well-mixed cloud layer as was observed on June 28, 1980 (see Figure 1), during the Arctic Stratus Experiment (ASE). Results of a radiative-diffusive model led Herman and Goody [1976] to suggest that two cloud layers may result from the absorption of solar radiation by the interior of a surface-based cloud. Using observations from the ASE, Tsay and Jayaweera [1984] hypothesized that the upper cloud layer was an advection fog, while the lower cloud formed due to surface convection. By analyzing the cloud microphysical and turbulence data from the ASE, Curry [1986] and Curry et al. [1988] inferred the following: the top cloud layer is maintained by convection, and vertical condensational growth is induced by cloud-top radiative cooling; surface evaporation plays little if any role in maintaining the cloud; and cloud-top entrainment has little dissipative effect, especially if the air above the cloud is moist as is commonly observed in the Arctic.

The evolution of a two-layer cloud system observed during the ASE was examined by McInnes and Curry [1985a] using a one-dimensional second-order turbulence closure model which includes parameterizations for drizzle and radiation. Their results indicate that the upper cloud layer forms initially by radiative and diffusive cooling of warm, moist air that has been advected into the region. A cloud-topped mixed layer evolves through radiative cooling at cloud top and subsequent mixing within and beneath the cloud; however, this cloud top mixing is insufficiently strong to erode the surface-based inversion layer which, being warmer than the surface and the cloud above, cools radiatively to produce a radiation fog. Modeling results from McInnes and Curry [1995b] suggest that the two layers are most likely to be maintained under conditions of weak, rising motion. McInnes and Curry [1995b] also noted that the modeled downwelling longwave radiative flux at the surface depends strongly on the vertical resolution in the lower atmosphere because of the complexity of the temperature profile.

The radiative properties of Arctic stratus clouds have been determined through observational studies. Using aircraft data obtained during the Arctic Ice Dynamics Joint Experiment (AIDJEX), Herman [1977] determined the shortwave transmittance, reflectance, and absorptance of stratus clouds observed over the Arctic ice pack in summer. Herman determined the broadband albedo of the cloud/surface system to range from 0.60 to 0.75 for several different Arctic stratus clouds. Additional observations of the radiative characteristics of Arctic stratus were obtained during the ASE by Herman and Curry [1984], Curry and Herman [1985a], and Tsay et al. [1989]. Herman and Curry [1984] determined a "reduced" cloud reflectivity and transmissivity for a hypothetical system with zero surface albedo. Their "reduced" cloud reflectivities ranged from 0.20 to 0.68 and transmissivities ranged from 0.25 to 0.80, for optical depths ranging from 2 to 24. Herman [1980] has shown that the emissivity of Arctic stratus clouds is a function of cloud geometrical thickness, indicating that the stratus clouds often did not emit as black bodies. Bulk cloud emissivities determined for summertime Arctic stratus by Curry and Herman [1985a] ranged from 0.4 to 1.0. They noted that the broadband emissivity was not well represented by the window emissivity, due to the relative transparency of the water vapor bands outside the window.

Toward improving the treatment of summertime clouds in global climate models, we employ a mesoscale atmospheric model to simulate a case study of the cloudy boundary layer observed during the ASE. Since the standard treatment of radiation code that is presently available in the mesoscale model is perceived to be deficient under Arctic conditions, a more sophisticated treatment of radiative transfer is incorporated into the mesoscale model. The role of radiative transfer in the evolution of the cloud deck is examined by comparing the results of three mesoscale simulations: one with an improved treatment radiative transfer, one with the original Dudhia [1989] radiative transfer code and one with radiative heating set to zero. It is shown that radiative cooling is vital for the development of low-level clouds over the ice pack and also for the maintenance of the anticyclone. The treatment of radiative transfer in the model determines the vertical structure and evolution of the cloud deck and also affects the synoptic-scale dynamics. Important differences in the modeled net radiation at the surface result from the differing treatments of radiative transfer and feedbacks between radiative transfer and the thermodynamic structure of the cloudy boundary layer.

2. Case Study

The Arctic Stratus Experiment was conducted during June 1980 over the Beaufort Sea to study summertime Arctic stratus clouds. The observing platform was the National Center for Atmospheric Research (NCAR) Electra which was equipped with instrumentation that measured longwave and shortwave fluxes, cloud liquid water content, particle size distributions, and turbulent fluxes along with the air temperature, moisture, and air velocity. The instrumentation is described in some detail by Herman and Curry [1984]. During the experiment, six stratus clouds were sampled. The case chosen for this modeling study is the two-layer cloud system observed on June 28, 1980, which has been extensively analyzed [e.g., Tsay and Jayaweera, 1984; Curry et al., 1988].

The synoptic situation on June 28, 1980, consisted of a mature anticyclone centered over the Beaufort Sea. A deck of stratus clouds in the northwest quadrant of the anticyclone covered an area of about (300 km)² [Curry et al., 1988]. The vertical
thermodynamic structure of the cloudy boundary layer was obtained by several aircraft profiles. The profile depicted in Figure 1 illustrates the vertical structure of this cloud system which consisted of a stable surface fog layer about 200 m deep surmounted by an upper cloud deck with a base around 800 m. The upper cloud had an average depth of 350 m, with liquid water mixing ratios of over 0.38 g kg⁻¹ in the upper portion of the cloud. Between the top of the upper cloud layer and the top of the lower cloud layer atmospheric properties were well mixed. The cloud liquid water mixing ratio within the lower, stable cloud layer is super-adiabatic. Winds were from the southwest at between 5-10 m s⁻¹ through the lowest 1.5 km of the troposphere.

Curry et al. [1988] estimated budgets for the vertically integrated total water content in the boundary layer using the aircraft observations for the June 28 case. They determined that radiative cooling and turbulent flux divergence caused the largest contribution to the equivalent potential temperature budget and moisture budget, respectively. Advection of heat and moisture was determined from the aircraft measurements to be small, although a fairly large residual evident in both the heat and the moisture budgets indicate that potentially significant errors may exist in some of the budget terms. Using aircraft measurements, Curry et al. [1988] estimated a vertical velocity of 0.2 cm s⁻¹; however, they noted that this value was likely not significantly different from zero. This mesoscale estimate is similar to the synoptic-scale estimate of 0.12 cm s⁻¹ determined by the European Centre for Medium Range Weather Forecasts (ECMWF) analysis [Curry and Herman, 1985b], indicating that synoptic-scale forcing was not being overwhelmed by mesoscale dynamics.

By June 30 the anticyclone had weakened somewhat; however, the region was still characterized by weak subsiding motion. The cloud layers in this region were somewhat broken, with predominately stable stratification in the lowest kilometre. On this day a single cloud layer with a base at 500 m and a top of 750 m is observed at 73.3°N, 133.8°W Troy and Jayaweera [1984] believed that this cloud deck, which had microphysics similar to that observed on June 28 in a cloud deck to the northwest, was actually the same cloud. This observation is indicative of the persistence of Arctic stratus clouds.

3. Mesoscale Model

Version 5.1 of the Pennsylvania State University (PSU)/NCAR Mesoscale Model (MM5) is employed to study the Arctic stratus cloud deck described above. For a complete description of the model, see Dudhia [1989, 1993] and Grell et al. [1994]. The diabatic processes, which have been shown by Melness and Curry [1995b] to be important for modeling Arctic stratus, are described in some detail below.

Prognostic equations are used to predict cloud water and precipitation. We employ the computationally efficient bulk microphysics scheme developed by Dudhia [1989] which allows for the treatment of liquid- and ice-phase processes without introducing additional prognostic equations for snow and cloud ice. This is done by assuming that melting/freezing at 273.15 K is instantaneous. The conversion of cloud liquid water to rain is accomplished following Hsieh [1984], while the conversion of cloud ice to snow follows Risedale and Hobbs [1983] and Lin et al. [1983].

In the standard version of the MM5, radiative transfer calculations are treated following Dudhia [1989]. For shortwave radiation, only the downward component is considered following an "integration-differentiation" technique. The downward component of shortwave radiation is computed as a function of attenuation by absorption and scattering by cloud particles [Stephens, 1978], absorption by water vapor [Lactis and Hansen, 1974], and Rayleigh scattering. The upward component of shortwave radiation is determined only at the surface assuming a broadband albedo for determination of the surface energy budget. This scheme must be used with caution over highly reflective surfaces particularly for cloudy situations in which multiple reflections between the surface and the cloud layer occur which significantly affects the net shortwave radiative flux at the surface. For longwave radiation both upward and downward fluxes are evaluated and represent hemispheric integrations of diffuse radiation. Gaseous absorption by water vapor and carbon dioxide is treated using a broadband emissivity method with CO₂ overlap [Stephens, 1984]. Clouds are assumed to be grey bodies, with emissivity determined as an exponential function of cloud water path with separate absorption coefficients for liquid [Stephens, 1978] and ice [Griffith et al., 1980]. The cloud fraction is assumed to be either 0 or 1 in each grid cell, depending on whether or not cloud water is present.

Turbulent exchange within the planetary boundary layer is treated using Blackadar's high resolution planetary boundary layer scheme [Blackadar, 1976, 1979; Zhang and Anthos, 1982]. Surface heat and moisture fluxes are computed from similarity theory for four cases: stable, mechanically forced turbulence, forced convection, and free convection. Prognostic variables in the boundary layer are solved either by first-order closure K theory for the "nocturnal" regimes (i.e., stable, mechanically forced turbulence, forced convection) or by mixing of buoyant plumes originating at the surface with each level in the mixed layer [Blackadar, 1979] for the regime of free convection.

4. Radiative Transfer

Because of perceived problems with the radiation code currently used in the MM5, notably the absence of the upwelling shortwave radiation and the crude treatment of clouds, two alternative radiation codes are examined for incorporation into MM5. We consider the radiation code currently used in Version 2 of NCAR's Community Climate Model (CCM2) [Briegleb, 1992a, b] and that used in Version 3 of the European Centre for Medium-Range Weather Forecasting (EC3) [Morcrette et al., 1986; Morcrette, 1991].

Both the EC3 and the CCM2 shortwave schemes offer several improvements over the Dudhia [1989] shortwave radiation scheme. Solar radiation is partitioned into direct and diffuse radiation. The solar absorption by ozone is explicitly modeled. The radiative properties of the atmosphere and surface are treated as spectrally varying quantities. Delta-Eddington calculations are performed for 18 bands in CCM2, while only two broadband are considered in EC3. The surface albedo is treated using two broadband in both models which is a significant improvement over the single broadband value used in MM5.

More detailed treatment of cloud radiative properties are an added improvement offered by EC3 and CCM2. In EC3 the shortwave radiative properties of liquid clouds are parameterized to be a function of effective radius and liquid water path [Morcrette, 1989]. Ice and mixed-phase clouds are treated by using larger effective radii. In CCM2 the shortwave optical depth of liquid clouds is determined in four bands as a function
of effective radius and liquid water path, while the single-scattering albedo and asymmetry parameter are a function of effective radius following Slingo [1989].

In the longwave the more explicit treatment of clouds and the inclusion of absorption by ozone and trace gases are improvements over the longwave radiative transfer scheme currently in MM5. The EC3 code includes explicit treatment of e and p-type continuum absorption by water vapor, and absorption by ozone, carbon dioxide, and other trace gases (i.e., CH4, N2O, and CFCs). The CCM2 includes the radiative effect of these gases as well and offers greater spectral resolution than the EC3. Both schemes determine a spectrally weighted cloud emissivity as a function of liquid water path and effective radius following Curry and Herman [1985a]. These parameterizations have been incorporated into EC3 as described by Pinto and Curry [1995] and into CCM2 following a similar method.

4.1. Evaluation of the Radiative Transfer Codes

The shortwave and longwave components of the CCM2 and EC3 are compared with the Dudhia [1989] code (hereinafter referred to as MM5) and a narrow band (24 bands in shortwave) model called Streamer (STR). The Streamer model was developed by Key [1995] and has its roots in the STRATS/Disort codes of Tsay et al. [1989, 1990]. The discrete ordinate version of STR is employed for cloudy-sky shortwave calculations. The STR code has the added advantage of allowing the spectral variation in the surface albedo to be a function of surfacetype. Pinto et al. [1996] have extensively tested Streamer for springtime Arctic conditions and have shown that the modeled broadband surface fluxes compare very well with observations obtained for clear-sky episodes observed during the Arctic Leads Experiment (LEADEX). Tsay et al. [1989] computed flux profiles for summertime Arctic stratus clouds. They show that the stratus clouds reduce the downward radiative flux by 130-200 W m⁻², while Arctic haze may decrease the downward radiative flux by an additional 10-15 W m⁻². Tsay et al. [1989] noted that although surface fluxes match the observations well, there is some discrepancy between the model and observed shortwave fluxes at cloud top.

A model intercomparison for clear-sky conditions is conducted to determine the utility of each model for conditions typically observed during the Arctic summer. The temperature and dew point temperature profiles used in this intercomparison are shown in Figure 1. A standard Arctic summer ozone profile is assumed after Ellingson et al. [1991] in EC3, CCM2, and STR. For simplicity of the intercomparison the aerosol optical depth is assumed to be zero. The solar zenith angle is set to 53.8° as determined by Herman and Curry [1984] for 28 June 1980. The surface is assumed to be bare sea ice with a broadband albedo of 0.55 and a temperature of 271.4 K. Each model has a vertical resolution that increases with height from 75 m adjacent to the surface to 4 km at the tropopause.

The specification of surface reflectance varies significantly among the models. In MM5 a single broadband reflectance of 0.55 is used to calculate the upward shortwave flux at the surface. In CCM2 and EC3 broadband reflectances are given for visible (0.28-0.7 μm) and near-infrared (0.7-4.0 μm) bands. From Ebert and Curry [1993] the visible and near-infrared reflectances for bare ice are taken to be 0.178 and 0.345, respectively. The broadband surface reflectance for the entire solar spectrum may be determined by summing the spectral reflectances weighted by the percentage of the total incoming solar flux at the surface in both bands. The spectrally varying surface albedo in STR may be scaled to a desired broadband albedo for model intercomparison.

The shortwave radiative fluxes were computed for reflecting (broadband reflectance of 0.55) and nonreflecting surfaces. Table 1 lists the upwelling and downwelling shortwave fluxes at the surface for each model. For the nonreflecting surface the downward fluxes obtained with STR, EC3, and CCM2 agree to within 7 W m⁻², while that obtained using MM5 is over 125 W m⁻² less. This large negative bias in MM5 has been attributed to an overestimation in the amount of backscattering by gases and aerosols (J. Dudhia, personal communication, 1996). For the reflecting surface the variations between the models in the downward flux at the surface are more dramatic. The differences can be related to the treatment of the spectral reflectance of the surface in each model and the corresponding effect of multiple reflections between the surface and the clear atmosphere. It is seen that the downward flux increases by 30-40 W m⁻² due to multiple reflections in STR and CCM2 and only 6 W m⁻² in EC3. The effect of multiple reflections appears to be underestimated in EC3. It is noted that although the specified spectrally integrated broadband reflectance is 0.55, the clear-sky broadband albedo is larger in the three models with spectrally dependent reflectances. This increase, which is seen to be largest in STR, is due to selective transmission of visible radiation by the clear atmosphere and the selective reflection of visible and near-infrared radiation by the surface.

The net shortwave flux received at the surface, a vital component of the surface energy budget, varies by over 50 W m⁻² among the models. The net surface flux obtained with EC3 and CCM2 compares fairly well with that obtained using STR, while MM5 grossly underestimates the net flux obtained with STR. It is noted that these large differences in the modeled net shortwave flux at the surface would be accentuated through the ice-albedo feedback in a coupled thermodynamic-radiative sea ice model.

The clear-sky downward longwave fluxes at the surface obtained with MM5RAD, EC3, STR, and CCM2 are listed in Table 1. The surface fluxes obtained using EC3 and STR vary by less

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<th>Table 1. Clear-Sky Surface Radiative Flux Model Intercomparison, W m⁻²</th>
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<td>Model</td>
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<td>MM5</td>
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<td>CCM2</td>
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<td>EC3</td>
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<td>STR</td>
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| *Values in parenthesis were obtained for a non-reflecting surface.
than 6 W m\(^{-2}\), while the value obtained with CCM2 is nearly 20 W m\(^{-2}\) less. This "cold" bias of CCM2 for clear sky conditions has been corroborated in discussions with J. Kiehl (personal communication, 1996) and may be a problem caused by under absorption in the far wings of the water vapor bands. The downward longwave flux at the surface in EC3 is greater than STR due to trace gas contributions which are not included in STR. It appears that MM5 may have a slight "warm" bias in the downward longwave flux.

Through comparisons with STR it is inferred that the shortwave radiative parameterization of CCM2 outperforms that of MM5 and EC3. It has been shown that EC3 performs poorly for large surface albedos encountered over the Arctic ice pack due to inadequate treatment of multiple reflections. In addition, the two shortwave bands in EC3 are inadequate for treating the radiative properties of Arctic clouds. Clear-sky longwave fluxes obtained with EC3 are much better than those obtained with CCM2 for the cold and dry conditions typically observed in the Arctic. The longwave results of EC3 are also an improvement over the slight "warm" bias in MM5. The detailed treatment of clouds in the longwave portion of EC3 is an added advantage over MM5. On the basis of these results we have selected the CCM2 shortwave code and the EC3 longwave code for inclusion into MM5 to address the radiation in the Arctic.

4.2 Role of Surface Albedo

The large discrepancies between the shortwave radiative surface fluxes between the models can be partially related to multiple reflections between the surface and the atmosphere. The effect of multiple reflections on the net shortwave radiation at the surface is enhanced by the presence of clouds. To examine this, radiative transfer calculations are performed with STR for a cloudy atmosphere using the atmospheric properties depicted in Figure 1 except only the upper cloud layer is implemented to simplify interpretation of the results. Simulations are conducted for clear-sky broadband surface reflectances ranging from 0 to 0.66. The broadband reflectances of 0.42, 0.59, and 0.66 correspond to surfaces comprised of bare ice with melt ponds, bare ice, and snow, respectively. Selective transmission of visible radiation by the atmosphere and cloud layer alters the spectral distribution of downwelling shortwave radiation reaching the surface thus changing the broadband surface albedo in the model. This effect cannot be accounted for with a single broadband surface reflectance value as done in MM5 and is only partially accounted for in CCM2 and EC3 which use two values to describe the surface reflectance. Computations for clear and cloudy situations using Streamer indicate that clouds increase the surface albedo by about 0.10.

Table 2 gives values of the downwelling, upwelling, and net shortwave flux for four different values of surface albedo. It is seen that both downwelling and upwelling fluxes decrease with decreasing broadband surface reflectance. The net shortwave flux received at the surface increases with decreasing surface albedo; however, this decrease is a somewhat nonlinear function of surface albedo due to multiple reflections. The large amount of downwelling shortwave radiation for the high-albedo case becomes extremely important for heterogeneous surfaces comprised of snow covered ice floes and open water or leads. Absorption of solar radiation by the dark surfaces may be greatly enhanced under these conditions.

Shortwave radiative heating increases within the lower portion of a cloud deck in response to an increase in the surface albedo. It was found that increasing the clear-sky broadband surface reflectance from 0 to 0.66 increased the heating rate within the cloud by up to 0.3 K d\(^{-1}\) with the largest effect near cloud base. Changes in the heating rate profile within the cloud may alter the evolution of the cloud and atmospheric boundary layer characteristics, thus feeding back further onto the surface radiation flux and the thermodynamics of the sea-ice surface.

Further, assuming the cloud microphysics remain unchanged during the transition from bare sea ice to ponded sea ice, the bulk radiative properties of the cloud will change markedly due to the reduced contribution from multiple reflections between the cloud layer and the surface. The bulk radiative properties of the cloud layer were determined from the flux profiles generated by STR and are listed in Table 3. It is seen that a transition in surface type from bare ice to ice with melt ponds results in a decrease in the bulk cloud reflectivity of about 3.7%, while the bulk transmissivity decreases by over 20%. Similar changes in the bulk radiative properties occur during the transition from snow-covered ice to bare ice. Values obtained for the nonreflecting surface are similar to the "reduced" values calculated by Herman and Curry [1984].

5. Role of Radiation in the Evolution of an Arctic Stratus Deck

The three-dimensional hydrostatic version of MM5 is initialized using the ECMWF analysis (available on a 2.5° by 2.5° grid at the standard pressure levels every 12 hours) for 0000 UT on June 28, 1980 and is integrated for 60 hours using a 2-minute time step. prognostic variables are relaxed to the ECMWF analyses at the lateral boundaries at 12-hour intervals. The model domain covers the Beaufort Sea region depicted in Figure 2 with a horizontal grid spacing of 40 km. The vertical coordinate is a terrain-following sigma-coordinate system in which the vertical grid spacing increases with height from approximately 30 m in the lowest 1.5 km to 1000 m at the top of the model (i.e., 100 mbar) resulting in 43 model levels. As a first-order approximation, all water points are assumed to be

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<th>Table 2. Cloudy-Sky Shortwave Radiative Fluxes for Various Surface Albedos Using Streamer</th>
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<tr>
<td>Surface Fluxes, W m(^{-2})</td>
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<tr>
<td>Clear Sky Albedo</td>
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<td>------------------</td>
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<tr>
<td>0.00</td>
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<tr>
<td>0.42</td>
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<td>0.59</td>
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<td>0.66</td>
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Table 3. Bulk Radiative Properties of Upper Cloud Layer Determined Using Streamer

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<tr>
<th>Clear-Sky Albedo</th>
<th>Reflectivity</th>
<th>Transmissivity</th>
<th>Absorptivity</th>
</tr>
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<tbody>
<tr>
<td>0.00</td>
<td>0.741</td>
<td>0.206</td>
<td>0.053</td>
</tr>
<tr>
<td>0.42</td>
<td>0.781</td>
<td>0.328</td>
<td>0.055</td>
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<tr>
<td>0.59</td>
<td>0.812</td>
<td>0.424</td>
<td>0.055</td>
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<tr>
<td>0.66</td>
<td>0.829</td>
<td>0.475</td>
<td>0.056</td>
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Sea ice with a fixed surface temperature of 271.4 K and a broadband surface reflectance of 0.35. The sea ice concentration at each point is assumed to be 100%. Land surface temperatures are allowed to vary following surface energy budget formulation based on the "force-restore" method developed by Blackadar [Zhang and Anthes, 1982].

The computationally efficient cloud microphysical scheme described by Dudhia [1989] is employed since only warm cloud microphysics is necessary for this case. Curry et al. [1995a] have shown that the threshold for autoconversion to occur may be smaller than that specified in the Dudhia [1989] scheme, thus this threshold is set to half its original value. Since conditions were observed to be relatively stable, it is assumed that cloud processes are explicitly resolved at the employed horizontal resolution so that parameterization of convection is unnecessary.

The initial conditions for the model run are specified with the ECMWF analysis for 0000 UT June 28. The initial sea level pressure field and the 850-mbar heights and temperature field are shown in Figure 2. An anticyclone covers much of the eastern Beaufort Sea, setting up a meridional flow at the surface in the central Beaufort between 180 and 150°W. At 850 mbar a relatively strong ridge is evident with its axis oriented north-south along the 140°W meridian. The northeastward bulge in the isotherms at 850 mbar around 180°W provides evidence that warm, moist air originating over the Chukchi Sea is being advected over the ice pack. The initial profiles of temperature, dew point, and u- and v-wind components for the approximate location of the June 28 aircraft observations (77.5°N, 155°W) are plotted in Figure 3. This profile was obtained through in-

![Figure 2](image1)

Figure 2. Initial model (a) sea level pressure and (b) 850-mbar heights (solid) and temperature (dotted) derived from ECMWF analysis valid 28 June at 0000 UT. Asterisk marks location of June 28 aircraft observation.

![Figure 3](image2)

Figure 3. Initial model profiles of (a) temperature and dew point and (b) u- and v-wind components obtained from (ECMWF) analysis at 77.5 N, 155 W valid June 28 at 0000 UT.
terpolation of the ECMWF gridded analysis to the model grid point. The lowest 3 km of the atmosphere are relatively moist and stable with a dew point depression of about 2 K and a temperature lapse rate of about 2.7 K km\(^{-1}\). The winds are rather strong (~10 m s\(^{-1}\)) throughout the lower troposphere, from the south-southwest with significant directional shear that is indicative of warm air advection.

Three simulations are conducted, each using a different treatment of radiation: the original Dudhia scheme (MMSRAD), the new radiation package (NEWRAD), and a null case in which radiative processes are not included (NORAD). For the NORAD run, radiative heating of the atmosphere has been set to zero throughout the entire domain, and the net radiative flux at the surface is not included in the surface energy budget over land points.

The evolution of the synoptic-scale environment is depicted by the sea level pressure field and the 850 mbar height and temperature field obtained for the MMSRAD simulation (Figure 4). The anticyclone initially goes through a period of strengthening over the first 24 hours as the system translates southeastward. The next 36-hour period, ending at 1200 UT June 30, is characterized by anticyclonicity as the system becomes nearly stationary and weakens considerably. The initial ridge at 850 mbar (see Figure 2) strengthens over the first 24 hours as warm air advection continues over much of the Beaufort Sea. Thereafter, heights fall throughout the region as the ridge weakens and the advection of warm, moist air into the region is reduced.

The sea level pressure pattern changes significantly when radiative effects are not considered but only slightly when the improved radiation package is used. These changes demonstrate the importance of radiative processes for the evolution of synoptic-scale features as seen in Figure 5. The elimination of radiative processes results in rapid anticyclonicism of the Beaufort anticyclone with the maximum surface pressure being 5 mbar less than in the MMSRAD after 60 hours of the simulation. The cyclone in the northwest corner of the domain has noticeably strengthened and a trough, which is not evident in the MMSRAD run, has developed across the Beaufort Sea extending southeastward to Banks Island.

Feedbacks between the thermodynamic state of the atmosphere and the radiative processes are noted as well. The presence and vertical distribution of clouds determines the structure...
of the radiative heating profile which determines the temperature profile. Changes in the temperature profile directly affect the longwave flux profiles. The longwave flux at the surface may also be affected by changes in the cloud properties resulting from changes in the temperature profile caused by latent heat release, solar absorption or infrared cooling.

Changes in the treatment of radiative transfer affect the evolution of the cloud field through changes in the heating rate profiles and associated feedbacks between the cloud and the synoptic scale. The cloud water mixing ratio field at 350 m is plotted at 12 and 24 hours for MM5RAD and NewRAD (Figure 6). After forming near 72.5°N, 163.0°W about 6 hours into the simulation the cloud deck expands and translates clockwise around the anticyclone center to the southeast. It is seen that the cloudy area at this level is much greater in the MM5RAD simulation at both times. The cloud water mixing ratios obtained in the NewRAD run are generally less than those obtained in the MM5RAD run, as demonstrated by the area covered by cloud water mixing ratio greater than 0.2 g kg⁻¹. The horizontal structure in cloud water mixing ratio is similar in the two runs, as seen in the location of local maxima. Both models overpredict the horizontal extent of the cloud deck; however, this effect is reduced in the NewRAD run. In the NORAD run (not shown), very little cloud is generated above 300 m which strongly suggests the importance of radiative cooling in the generation of this Arctic stratus cloud deck.

By 60 hours the two-dimensional shape and size of the cloud in both runs are quite similar overall although regional differences are still evident. A tongue of clearing extends into the region where the June 28 observations were made, as seen in the cloud water mixing ratio field generated by the MM5RAD run (Figure 7). The maximum cloud water mixing ratios have translated southeastward over the June 30 observation location with the southern cloud edge extending over the northeastern half of Banks Island.

The maximum liquid water mixing ratio, which originated around 73°N, 165°W, may be used as a tracer to determine the trajectory of air. The cloud water mixing ratio maximum rotates around the anticyclone centered over the southern Beaufort Sea (Figure 6, 7). At 24 hours the maximum cloud water has trans-
lated northeastward to 79°N, 146°W. The parcel then recurves toward the southeast and after 60 hours is located at 74°N, 130°W. This analysis corroborates the conclusion of Tsay and Jayaweera [1984] that the cloud deck observed at 2200 UT June 30 (73.3°N, 133.8°W) was the same deck observed two days earlier at 77.0°N, 154.2°W.

Varying the treatment of radiative transfer also affects the vertical structure of the thermodynamic variables. The time evolution of temperature, dew point and cloud water mixing ratio profiles in the lower atmosphere is depicted for each simulation (Figure 8). The profiles were obtained for the grid point at 77.5°N, 155°W which corresponds with the location of the aircraft measurements obtained on 28 June 1980 at 2200 UT. After 6 hours the cloud has formed in the lowest model level and deepens with time as warm, moist air advected into the region cools through radiation and diffusion. By 12 hours the region has cooled enough through radiation and diffusion to support condensation in a layer from the surface to 600 m in the MM5RAD run and to 550 m in the NEWRAD run. Some of the radiative cooling near the surface can be attributed to the presence of a cloud deck based at 1.5 km for preceding the 6 hours (and still evident in the MM5RAD run at 12 hours) which reduced solar heating of the column below. The cooling of air directly above the surface through turbulent diffusion results in the formation of a shallow (200 m deep) fog layer in the NORAD run, indicating the importance of radiative cooling for the formation and deepening of this Arctic stratus cloud deck.

The stratus cloud deck has reached its mature stage after 18 hours in both MM5RAD and NEWRAD runs, having two distinct maxima in the liquid water mixing ratio and a cloud-top temperature inversion. The vertical structure and depth of the cloud is similar in both radiation runs. The cloud top has

![Figure 8](image_url)  
**Figure 8.** Profiles of temperature, dew point, and liquid water content for the MM5RAD run (left), NEWRAD run (middle), and NORAD run (right) after 12, 18, and 24 hours.
reached nearly 1 km. The two maxima in each run are of the same magnitude (0.3 g kg\(^{-1}\) in the upper layer, 0.38 g kg\(^{-1}\) in the lower layer), while the local minima, located at 850 m, is slightly larger in the NEWRAD run. The cloud-top temperature inversion of 1.6 K in the MM5RAD run is 0.5 K stronger than that obtained in the NEWRAD run. The shallow cloud has deepened to 400 m in the NORAD run as the lowest levels of the atmosphere continue to cool through mixing of cold air adjacent to the surface.

The modeled profiles after 18 and 24 hours for each run may be compared with observations. The depth of the two-layer cloud system of about 0.8-1 km is similar to that observed (Figure 1). The depth of the upper cloud layer is comparable to that observed, while the depth of the lower cloud layer is much too great in both simulations. The modeled cloud water mixing ratios in the surface-based cloud layer are much larger than observed in both simulations, while that in the upper cloud layer is only slightly less than observed. The vertical structure of the cloud at 18 hours of the simulation compares best with the observations. The strength of the temperature inversion is underestimated by both models by about 1 K. The modeled boundary layer is warmer than observed and the modeled depth of the mixed layer extending from the top of the upper cloud layer to the top of the surface-based inversion is smaller than observed. The strength of the modeled surface-based inversion is stronger than that observed.

By 24 hours the upper cloud layer in both model runs has begun to dissipate. The amount of cloud water has decreased in both simulations with a much larger decrease occurring in the NEWRAD run. After 24 hours the cloud top begins to lower (more quickly in the NEWRAD run) and the two-layer cloud system collapses into a single surface-based cloud deck in both simulations. The cloud-top radiative temperature inversion has been eroded away as the temperature profile is warmed from above and is completely gone by 48 hours in both runs (Figure 9). The fog layer which had formed after 6 hours in the NORAD run persists through the entire simulation with its depth varying between 200 and 400 m.

Substantial differences between the two models are evident in the downwelling shortwave flux received at the surface. The diurnal cycle is clearly evident in the downwelling shortwave fluxes, with minima occurring after 12 hours (corresponding with 1200 UT) (Figure 10a). Both shortwave flux time series indicate the presence of clouds during the second 12-hour period. Attenuation of shortwave radiation is much greater in the MM5RAD run with little solar radiation reaching the surface between 6 and 18 hours. The shortwave flux determined in the NEWRAD run is 90-150 W m\(^{-2}\) greater than that in the MM5RAD run. At 22 hours the shortwave flux in the NEWRAD run is 225 W m\(^{-2}\) compared with 75 W m\(^{-2}\) in the MM5RAD run and the observed value of 300 W m\(^{-2}\) [Herman and Curry, 1981].

Figure 9. Profiles of temperature, dew point, and liquid water content for the MM5RAD run after 48 hours.

Figure 10. Time series of downward surface (a) shortwave and (b) longwave radiative fluxes obtained for the NEWRAD and the MM5RAD run.

[Image of profiles and graphs showing cloud water content, temperature, and fluxes over time.]

Variations in the modeled downwelling longwave flux are evident in Figure 10b. The clear-sky longwave fluxes are similar in the two models at around 285 W m\(^{-2}\). The onset of cloud around 4-6 hours results in a large jump in the downwelling longwave flux. Soon after the atmosphere emits as a blackbody in the MM5RAD run with variations in the cloud geometry and...
microphysics having little effect on the downwelling longwave flux. These variations are reflected in the downwelling longwave flux of the NEWRAD run. The observed downward flux at 2200 UT of 327 W m⁻² is slightly underestimated in the NEWRAD run in which the longwave flux varied between 315 and 325 W m⁻² in the preceding 3 hours. The MM5RAD longwave flux which is practically constant after 7 hours is fairly close to the observed flux for this case. The modeled longwave flux is similar to that observed because this cloud layer may be accurately treated as a blackbody. More tenuous clouds in which the cloud emissivity is a function of liquid water path will be handled poorly by the parameterization of longwave radiation given by Duhia [1989]. Evidence for this is seen in the initial stages of cloud development for this case where large discrepancies between the two models are apparent

6. Summary and Conclusions

An intercomparison of radiative transfer codes currently used in GCMs with a narrowband model showed that CCM2 shortwave radiation and EC3 longwave radiation codes performed best in the Arctic. Downwelling longwave fluxes computed using CCM2 were too low for both clear and cloudy sky conditions. The shortwave scheme in EC3 uses only two broadband which is insufficient to accurately parameterize cloud radiative properties using the parameterization of Slingo [1989]. The current shortwave cloud parameterization in EC3 performs adequately; however, the treatment of multiple reflections between the atmosphere and the surface was deemed inadequate. As demonstrated by calculations with Streamer, the downward shortwave radiation is dramatically increased by multiple reflections. The surface albedo is effectively increased by 10% when clouds are present.

A new treatment of radiative transfer has been added to MM5 and has been tested in both off-line and predictive modes for summertime Arctic conditions. The new radiation package is comprised of the CCM2 shortwave radiation scheme and the EC3 longwave code, both of which included more sophisticated treatment of cloud radiative properties than is currently available in MM5.

Three model runs were conducted with different treatments of radiative transfer to simulate Arctic stratus clouds in the vicinity of a summertime anticyclone over the Beaufort Sea. The model is run by using the new radiation code, the Duhia [1989] code, and with radiative effects neglected. The Arctic stratus cloud deck evolves in horizontal extent and vertical structure as the anticyclone goes through several stages of development. When radiation is neglected, the anticyclone weakens much more rapidly than in the runs with full radiation.

The evolution of the cloud deck, both in the vertical and in the horizontal extent is sensitive to the treatment of radiative transfer. Radiative cooling is seen to be vital for the development of the stratus cloud deck observed during the Arctic Stratus Experiment, as very little cloud formed in the NORAD run. Results obtained with the new radiation scheme were improved over the Duhia radiation run. The horizontal extent of cloud after 24 hours is greatly reduced using the more sophisticated treatment of radiation more closely matching the observations. The evolution of vertical structure in the temperature and cloud water profiles was also improved over the Duhia radiation run. In the new radiation run, the upper cloud layer is nearly decoupled from the surface, and the strength of the cloud top temperature inversion is closer to that observed. In addition, the modeled shortwave and longwave fluxes at the surface more closely match those observed near the surface at 2200 UT June 28. Much of the difference between modeled and observed surface fluxes can be attributed to the overprediction of liquid water path, the treatment of multiple reflections and the excessive attenuation of shortwave radiation by clear sky in MM5.

Acknowledgments. Discussions with Jeff Tilley and Yong Zheng with regard to the MM5 modeling system are greatly appreciated. The computer facilities at the University of Colorado and the National Center for Atmospheric Research were used for the numerical simulations and analyses. This research was supported by the Department of Energy Atmospheric Radiation Measurement (ARM) Program and a NASA Global Change Fellowship (Pinto).

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(Received January 23, 1996; revised June 4, 1996; accepted July 1, 1996.)