Atmospheric convective plumes emanating from leads

2. Microphysical and radiative processes

James O. Pinto and Judith A. Curry
Program in Atmospheric and Oceanic Sciences, University of Colorado, Boulder

Abstract. A one-dimensional, second-order turbulence model with bulk cloud microphysics and detailed radiative transfer is used to simulate the evolution of a thermal internal boundary layer (TIBL), which develops above a wide, open lead. A mixed-phase cloud, originally based at the surface, is produced within the TIBL. The cloud initially fills the entire TIBL but is later elevated above the surface with its top coincident with the top of the TIBL. Model-derived cloud ice and cloud liquid water mixing ratios exceed 0.06 g kg\(^{-1}\) directly above the open lead, with a secondary maximum near the top of the TIBL. In addition, precipitating ice particles or snow fills the TIBL with a maximum snow mixing ratio of about 0.05 g kg\(^{-1}\). Radiative flux divergence results in strong cooling at cloud top (which contributes to the local maxima in cloud water mixing ratio at this level) and warming near the surface. The lead-induced cloud increases the downwelling long-wave irradiance received at the surface by up to 70 W m\(^{-2}\) (reducing the surface radiative cooling by over 40%) during the baseline case. This value is quite sensitive to the assumed particle size and cloud particle concentration. The vertical structure and composition of the lead-induced cloud is shown to strongly depend on the rate of snow production and the cloud water partitioning.

1. Introduction

Pinto et al. [this issue] (hereinafter referred to as part 1) discuss the general thermodynamic structure of atmospheric convective plumes emanating from Arctic leads. Clouds or fog which appear to originate from open leads have recently been observed to reach great heights over the central Arctic ice pack during winter [Barry et al., 1989; Schnell et al., 1989; Andreas et al., 1990]. The extraordinary thermodynamic conditions just above an open lead during winter may produce supersaturations with respect to liquid water in excess of 200% [Andreas et al., 1981]. When these conditions are present, deliquescence occurs rapidly forming droplets which continue to grow through diffusion to be 10 \(\mu\)m in diameter in less than 2 s. Ohtake et al. [1982] have observed ice crystals growing at an extremely rapid rate above an open lead as well. The phase of these lead-induced clouds is believed to be predominantly crystalline aloft and liquid near the surface; however, direct measurements of particle phase, size, and concentration within these clouds have not been made.

Estimates of droplet concentrations of lead-induced clouds have been made by Andreas et al. [1990] using the backscatter from airborne infrared lidar. By assuming that the cloud base consisted of water droplets with a mean radius of 5 \(\mu\)m, they retrieved droplet concentrations of 3-6 \(\times\) 10\(^{13}\) m\(^{-3}\). Ice crystals with radii of 40 \(\mu\)m were assumed to be present in the upper portions of the lead-induced cloud to obtain particle concentration of about 1 \(\times\) 10\(^{13}\) m\(^{-3}\). Uncertainties in these estimates due to assumptions of particle phase, size, and shape translate into possible errors in the retrieval of up to a factor of 10.

During the Arctic winter the surface energy budget is dominated by long-wave radiation except in the vicinity of open leads. Clouds significantly alter the downwelling long-wave irradiances received at the surface. Curry et al. [1993] have shown that the long-wave radiative energy budget is significantly affected by low level ice crystals. The contribution by leads to Arctic cloud cover, directly through the production of clouds and indirectly through increasing boundary layer moisture is not well known.

Cloud climatologies for the Arctic region often exclude low-level ice crystal clouds, which may include a significant contribution from lead-induced clouds, when computing wintertime cloud fraction [e.g., Curry and Ebert, 1992]. The importance of obtaining an accurate cloud fraction for the Arctic has been emphasized by Curry et al. [1993] who have shown strong sensitivities of the equilibrium sea ice thickness, a crucial variable in Arctic climate, to low-level cloud cover.

Furthermore, the interrelationship between leads and low-level cloudiness needs to be further explored for possible feedbacks. For example, enhanced downwelling long-wave radiation at the surface due to the presence of lead-induced clouds may slow the freezing of open leads. As leads remain open longer, heat and moisture fluxes continue to warm and moisten the lower levels of the atmosphere, which may maintain or enhance low-level cloudiness and thus enhance downwelling long-wave radiation. Another possible feedback between infrared radiation and turbulence arises from the combination of cloud top cooling and warming near the surface due to long-wave radiative transfer. The radiative heating profile has a destabilizing effect on the cloudy boundary layer above an open lead. Increased stability would enhance turbulent fluxes of heat and moisture into the boundary layer above an open lead and could contribute to the maintenance or enhancement of low-level clouds which would then feed back to radiative transfer through the cloud layer.

Large-scale feedbacks between leads and the atmosphere are possible as well. At the regional climate scale, Ledley...
[1988] describes a lead-temperature feedback whereby increasing the winter minimum lead fraction increases the surface heat fluxes and thus warms the lower atmosphere. The warmer surface air temperatures reduce the coverage and mean thickness of sea ice and increase the lead fraction.

In this paper we utilize the one-dimensional model described in part 1 to investigate the evolution of the cloud microphysical and radiative processes that characterize a convective hydrometeor plume occurring over a wide lead. An analysis is given of the composition, maintenance, and vertical structure of a modeled hydrometeor plume or lead-induced cloud. The sensitivity of the vertical structure and composition of a lead induced cloud to various microphysical parameters is determined. The impact of these lead-induced clouds on the surface radiative energy budget and the sensitivity of the surface radiative energy budget to changes in particle size are also examined.

2. The Model

The one-dimensional model described in part 1 is utilized here. For purposes of reference the prognostic equations for this model are given below

\[ \frac{d\theta}{dt} = \frac{\partial}{\partial z} \left( \rho w' T' \right) + Q^* + P_a \]  
(1)

\[ \frac{dq}{dt} = \frac{\partial}{\partial z} \left( \rho w q' \right) - P_S \]  
(2)

\[ \frac{dq}{dt} = \alpha \frac{\partial (L w q_S)}{\partial z} + P_S \]  
(3)

where \( q_w = q_i + q_i + q_{w} \), \( q_i \), \( q_w \), and \( q_{w} \) are mixing ratios for cloud liquid water, cloud ice, and water vapor, respectively, 0 is the potential temperature, and \( q_S \) is the snow mixing ratio. The first term on the right-hand side of (3) gives the rate of snow removal through gravitational settling where \( \alpha = 1/\rho \), \( \rho \) is the density of air and \( w_S \) is the terminal velocity of snow parameterized after Lin et al. [1983]. Turbulent diffusion of snow has been neglected. In the following sections we describe the production terms for snow \( P_S \) and potential temperature \( P_a \) and the radiative heating term \( Q^* \).

2.1. Cloud Microphysical Model

A saturation adjustment scheme, similar to that described by Curry [1983], has been implemented to diagnose cloud water condensation, evaporation, melting, and freezing (and thus \( q_i, q_w \), and \( q_{w} \)) and the associated latent heat release \( P_a \) at each time step. Condensation occurs as the air becomes supersaturated with respect to liquid water forming cloud liquid water. The cloud liquid water is then partitioned into liquid and ice using a cloud water partitioning function \( f \) which gives the ratio of cloud ice to total cloud water \( (q_{ci} = q_i + q_{ci}) \) as a function of temperature following:

\[ f = 1 - \exp \left[ -0.01(T_c - T) \right] \]  
(4)

where \( T_c \) is the critical temperature above which ice does not nucleate and \( T_c \) is the air temperature. The critical temperature for the baseline case (as described in part 1) is determined from the observations of Curry et al. [1990]. Cloud ice melts instantaneously when the air temperature rises above 273.15 K. Latent heat released during these processes contributes to environmental temperature change through the \( P_a \) term in (1).

In this model condensed water is separated into cloud liquid water, cloud ice, and snow. Snow is a predicted quantity while cloud liquid and cloud ice are diagnosed from the total water mixing ratio. The cloud particles are assumed to be small so that their terminal velocities can be neglected compared with the terminal velocity of snow. In a cloud layer, snow production processes transform vapor, cloud liquid, and cloud ice into snow. These processes are described in more detail below. Snow is removed from the atmosphere by gravitational fallout (the second term in (3)).

The determination of \( P_S \) can be made either by using a detailed cloud microphysical model that computes the spectral evolution of cloud particles [e.g., Berry and Reinhardt, 1974; Lee, 1990] or by using a bulk microphysical model, whereby only the hydrometeor mixing ratios are used [e.g., Cotton, 1972; Orville and Kopp, 1977; Hsue et al., 1980; Lin et al., 1983; Rutledge and Hobbs, 1983, Lee, 1992]. The relative simplicity of the bulk microphysical model is an obvious advantage; however, previous bulk microphysical models have been derived using time-dependent cumulus-convection models under conditions that are significantly different from those characteristic of the environment in which lead-induced hydrometeor plumes occur.

For this study we have adopted the framework of the Lin et al. [1983] bulk microphysics parameterization, with some modification to make it more suitable for the Arctic environment. The equation for the production of snow \( P_S \) is divided into two temperature regimes

\[ P_S = P_{S_{\text{Sat}}} + P_{S_{\text{Cur}}} + P_{S_{\text{Acc}}}, \]  
(5)

\[ P_{S_{\text{Sat}}} = \delta (1 - \delta) + P_{\text{deg}} \delta \]  
(6)

\[ P_{S_{\text{Cur}}} = T > 0 \degree C \]  
(7)

where \( P_{S_{\text{Sat}}} \) is ice particle aggregation, \( P_{S_{\text{Cur}}} \) is the accretion of cloud ice by snow, \( P_{S_{\text{Acc}}} \) is the accretion of cloud water by snow, and \( P_{S_{\text{Deg}}} \) is the Bergeron process in which cloud water and cloud ice, respectively, are converted to snow. The conversion of vapor to snow or deposition is given by \( P_{\text{deg}} \) and \( P_{\text{Dep}} \) is the sublimation of snow. Sublimation of snow is not allowed within a cloud layer. That is, if cloud ice or cloud liquid is present, then only deposition is allowed to occur (\( \delta = 1 \)), otherwise snow is allowed to sublimate (\( \delta = 0 \)). The use of \( \delta \) insures that cloud water (liquid and ice) will be removed through evaporation or sublimation before snow. Finally, \( P_{\text{Sat}} \) is the rate at which snow particles melt when the temperature is above freezing. The formulation of each process is given by Lin et al. [1983]. Processes involving the production of rain and graupel have been neglected for this study. For the base-line case all temperature-independent collection efficiencies were assumed to be 100%. The snow production rates, \( P_{S_{\text{Sat}}} \) and \( P_{S_{\text{Cur}}} \), are set to occur at 20% of the values used by Lin et al. [1983]. The reason for this reduction
is discussed in section 4.2. Latent heat released during the production/depletion of snow is summed over all processes and contributes to the predicted potential temperature through $P_v$.

2.2. Radiative Transfer

For these simulations we examine only wintertime conditions and thus consider only the long-wave radiative fluxes. The radiation model used to compute infrared fluxes was developed by Morcrette et al. [1986] and is further updated by Morcrette [1991] for use in the European Centre for Medium-Range Weather Forecasts model. This model has six spectral intervals in the infrared and computes the long-wave radiative transmission functions for each of the radiatively important atmospheric gases and aerosols. Gaseous CO$_2$, CH$_4$, chlorofluorocarbons (CFCs), and NO$_2$ are assumed to be uniformly mixed throughout the troposphere. An O$_3$ profile for the sub-Arctic region is specified after McClatchey et al. [1972], and a background marine aerosol profile is obtained from Taras et al. [1984]. The concentrations of each radiatively active gas, except water vapor, are constant with time.

Clouds are determined to exist if condensate is present as determined through the prognostic equation for total water (2) and the saturation adjustment scheme (4). The influence of clouds on the radiative transfer through the column depends on cloud fractional coverage, geometric depth of the cloud layer, and the amount, phase, and size of cloud water particles. The geometric depth of the cloud layer, the amount of cloud water, and the phase of the cloud particles are explicitly modeled here, while the cloud particle size spectra are externally specified in terms of effective radii. The cloud fraction is assumed to be 100%. The bulk long-wave absorption coefficients for liquid and ice cloud particles are parameterized following Curry and Herman [1985] and Ehret and Curry [1992], respectively, to be functions of the particle effective radius. The bulk absorption coefficients are weighted according to the amount of each cloud constituent present.

3. Baseline Case

The same initial conditions employed for the baseline case discussed in part 1 apply here. The critical temperature ($T_c$ in (4)) for ice to nucleate is specified to be 260 K, so that only ice will nucleate at temperatures below 240 K, while mixed-phase clouds are allowed at temperatures between 240 K and 260 K, with equal amounts of cloud ice and cloud liquid being formed at 252 K (see (4)). The rate of snow production through Bergeron processes is parameterized to be 20% of the rate defined by Lin et al. [1983]. The snow size distribution is determined by an intercept parameter given by Lin et al. [1983]. We assume snow crystals are unrimed side planes. The size distribution and particle shape are used to determine a bulk fall speed after Locatelli and Hobbs [1974]. For radiative calculations the effective radii for cloud liquid and cloud ice particles are somewhat arbitrarily assumed to be 7 µm and 20 µm, respectively. The sensitivity of this choice to long-wave radiative transfer is discussed in section 4.4.

Evolution of the cloud water (ice and liquid) profile is described by the time-height plot given in Figure 1a. Rapid moistening of the lower troposphere above an open lead gives rise to highly supersaturated conditions and condensation within the first few time steps. After the first few time steps a maximum cloud water (ice and liquid) mixing ratio of 0.24 g kg$^{-1}$ is reached just above the surface. During the first 2 hours of the simulation cloud water is present throughout the modeled mixed layer. The depth of the condensate layer increases to nearly 800 m by 2 hours, filling the entire thermal internal boundary layer (TIBL), (see part 1), with a maximum cloud water mixing ratio exceeding 0.06 g kg$^{-1}$ near the top of the TIBL. After 0.5 hours the cloud water mixing ratio begins to decrease at the lower model levels and has been entirely depleted from the surface to 600 m by 6 hours. The reduction in cloud water mixing ratio near the surface arises from the conversion of cloud water to snow and a reduction in the moisture flux at low levels (see Figure 6 in part 1). Snow is present throughout the TIBL with the largest values occurring just above the surface (Figure 1b). Combining the snow field and the total cloud water field gives the total hydrometeor mixing ratio (Figure 1c). This field decreases with height until 1.5 hours, at which time a layer of enhanced hydrometeor mixing ratio is evident near the top of the hydrometeor plume. The artificial steps in cloud water mixing ratio at the top of the cloud layer arise from the model grid spacing which increases with height (i.e., the steps become more pronounced with height). Ideally, we would like to eliminate this numerical result by reducing the grid spacing; however, this is not computationally practical. The general structure of

Figure 1. Time-height plots of modeled (a) total cloud water mixing ratio ($q_i + q_l$) (b) snow mixing ratio $q_s$ and (c) hydrometeor mixing ratio ($q_i + q_l + q_s$) for the baseline simulation.
modeled lead-induced cloud is similar to the large hydrometeor plume observed by Schnell et al. [1989] emanating from a wide lead north of Ellesmere Island.

3.1. Plume Composition

The cloud ice, cloud liquid, and snow mixing ratio profiles give a more detailed view of the vertical structure of the modeled lead-induced cloud. Profiles of cloud liquid water, cloud ice and snow are shown in Figure 2 at 1, 3, and 6 hours. It is seen that the cloud ice mixing ratio exceeds the cloud liquid water mixing ratio at all levels throughout the model run. This reflects the choice of critical temperature $T_c$ in determination of the cloud water partitioning function $f$ in (4). After 1 hour, the cloud liquid water mixing ratio exceeds 0.05 g kg$^{-1}$ just above the surface which is about half of the total cloud water present at this level. By 3 hours an elevated cloud layer is evident with maximum liquid water mixing ratios of about 0.01 g kg$^{-1}$. The fraction of total cloud water that is liquid ranges from 20% near the base of the cloud to less than 10% near cloud top. By 6 hours, over 95% of the elevated cloud layer is crystalline. The maximum cloud water mixing ratio evident near cloud top increases from 0.08 g kg$^{-1}$ at 1 hour to 0.11 g kg$^{-1}$ by 6 hours. The lower levels of the lead-induced cloud are drying out with a single, elevated cloud layer present after 3 hours and a continually rising cloud base.

Evolution of the snow field is also evident. Snow is produced within the cloud layer and increases with distance below cloud top. At 1 hour, a maximum snow mixing ratio of just over 0.04 g kg$^{-1}$ exists near the surface. By 6 hours it is evident that the snow mixing ratio increases to a maximum value just below cloud base, then decreases toward the surface as a result of sublimation in a layer which is subsaturated with respect to ice (as will be shown later); however, the total, vertically integrated mass of snow increases with time. Assuming a snow particle density of 0.1 g cm$^{-3}$, snow accumulates at the surface to a depth of 0.65 cm after 6 hours of model integration.

In situ measurements of cloud particle phase, size, and concentration within a hydrometeor plume originating from an open lead are lacking; however, indirect measurements of these microphysical properties have been made. Andreas et al. [1981] made measurements of water droplet diameter and concentration directly above the open lead. They found a modal droplet diameter of 10 μm and particle concentrations ranging between 50 and 220 cm$^{-3}$. Using the modeled liquid water mixing ratio of 0.065 g kg$^{-1}$, which occurs near the surface at 1 hour (Figure 2a) and assuming a mean liquid particle size of 10 μm we obtain particle concentration of 25 cm$^{-3}$.

Lidar-derived ice crystal concentrations were determined by Andreas et al. [1990] above a lead north of Ellesmere Island. The derived concentrations rarely exceeded 0.04 cm$^{-3}$ for assumed ice spheres of 40 μm in radius. It is noted that these concentrations are accurate only to within an order of magnitude and are quite dependent on assumed particle size. The modeled total ice mixing ratio (cloud ice and snow) is between 0.05 and 0.15 g kg$^{-1}$ throughout the TIBL (Figure 1c). Using an equivalent-sphere radius of 40 μm for the cloud ice and snow particles and a ice density of 0.85 g cm$^{-3}$, we obtain ice particle concentrations between 0.3 cm$^{-3}$ and 1.0 cm$^{-3}$, more than an order of magnitude greater than that observed by Andreas et al. [1990]. However, given the uncertainty in the lidar-derived concentrations due to assumptions in particle shape and size these particle concentrations do not seem unreasonable.

3.2. Snow Production Processes

The snow field evolves through the contribution of each snow production process described in section 2.1. A detailed look at the time dependent snow production rates has been done at 20 m and 700 m (Figure 3). At 20 m each snow production term in (5), with the exception of aggregation, contributes to the evolving snow field (Figure 3a). The dominant snow-producing processes at 20 m convert cloud ice to snow through the ice Bergeron process and accretion ($P_{3h}$ and $P_{3ac}$, respectively). The snow production processes which convert cloud ice to snow are at least an order of magnitude greater than those that convert cloud liquid to snow. This difference in snow production rates is due, in part, to the abundance of cloud ice. The most important process at 20 m, with a peak snow production rate of $1.5 \times 10^{-7}$ s$^{-1}$, is the ice-Bergeron process which transfers cloud ice to snow through the depositional growth of cloud ice particles. As the layer warms, the amount of cloud liquid water present increases so that the production of snow via accretion of cloud water $P_{3ac}$ eventually exceeds that produced by the accretion of cloud ice $P_{3ac}$. Each snow production process, except deposition, ceases by 4 hours, as the layer becomes devoid of cloud water but remains supersaturated with respect to ice. After 4.5 hours this level is subsaturated with respect to ice allowing sublimation $P_{3sub}$.
Several trends in the radiative transfer within a lead-induced cloud are evident. The maximum cooling rate is collocated with the maximum total cloud water mixing ratio near the top of the cloud. The increase in the maximum cooling rate at cloud top is proportional to the increase in the total cloud water mixing ratio. The radiative heating rate at the lowest model level decreases with time as the lower atmosphere warms. The majority of the TIBL warms by 1-2 K d$^{-1}$ with a growing heating spike near cloud base owing to radiative flux convergence.

The radiative energy budget at the surface is affected as well. The ocean waters within an open lead have an emissivity of 0.92 at 271.4 K which produces an upward flux of $-280$ W m$^{-2}$. The downwelling flux of long-wave radiation at the surface increases by nearly 70 W m$^{-2}$ (from 120 to 190 W m$^{-2}$) by the end of the simulation due to the warming and moistening of the temperature profile above the lead and the "insulating" effect of the lead-induced cloud (solid curve in Figure 5). Thus, the magnitude of the radiation deficit at the surface decreases from -160 to -90 W m$^{-2}$ reducing the cooling of the surface by 44%.

The radiative impact of a cloud is determined by the vertically integrated amounts of cloud ice and liquid (ice water path IWP and liquid water path LWP, respectively) present. In the Arctic, cloud water mixing ratios and integrated cloud water amounts are often quite small due to the limited supply of

3.3 Radiative Effects

Studies have shown that low-level ice crystals play an important role in modifying the temperature profile and surface energy balance in the Arctic [Curry, 1987; Curry et al., 1990; Curry and Ebert, 1992]. Curry et al. [1990] have shown that a uniform cooling rate of about -2.5 K d$^{-1}$ throughout the lower troposphere can be associated with "clear air" ice crystal precipitation. Figure 4 depicts the radiative heating rate and total cloud water mixing ratio (cloud ice and cloud liquid) profiles after 1, 3, and 6 hours of model integration. At cloud top the radiative cooling rate exceeds 20 K d$^{-1}$ by 3 hours, with the magnitude of cooling increasing with time. Radiative flux convergence at the lowest model level results in heating rates in excess of 100 K d$^{-1}$ at 1 hour decreasing to 80 K d$^{-1}$ by the end of the simulation.
moisture over the ice pack. Ebert and Curry [1992] have shown that due to small cloud IWP typical of low-level ice crystal clouds in the Arctic the emissivity of these clouds is much less than unity and varies nonlinearly with changes in particle size and concentration. Curry and Herman [1985] have shown the same result for summertime Arctic stratus clouds which are composed entirely of liquid water droplets. The modeled ice and liquid water paths obtained during the baseline run were 50 g m\(^{-2}\) and 10 g m\(^{-2}\), respectively. Using Figure 5 of Ebert and Curry [1992] and Figure 6 of Curry and Herman [1985] it is seen that the assumed particle size and modeled water paths correspond to emissivities much less than unity and lie in a nonlinear, rapidly changing region of the curves. Because of the strong dependence of emissivity on cloud water path in lead-induced clouds it is crucial to accurately model the microphysical properties of these clouds. Even more important is the direct observation of microphysical properties of lead-induced clouds for improved parameterizations and model verification.

4. Sensitivity Studies

The sensitivity of the vertical structure and composition of a lead-induced cloud to our choice of microphysical parameters is determined. The critical temperature is varied to account for variability in the ice nucleating process. The efficiency of the Bergeron processes is also varied as this process appears to be the most important snow producing mechanism in mixed-phase clouds. The sensitivity of surface radiative fluxes to variations in the specified particle effective radii is examined as well.

4.1. Critical Temperature

Curry et al. [1990] have noted the large natural variability in the dependence of cloud particle phase on temperature in the Arctic. To test the robustness of the baseline case results to the choice of critical temperature two additional simulations have been performed using critical temperatures of 255 K and 265 K. The higher (lower) critical temperature may indicate an increased (decreased) presence of ice-forming nuclei. The profiles of cloud liquid, cloud ice, and snow after 2 hours of each simulation are shown in Figure 6. The cloud ice mixing ratio increases throughout the TLBL as the critical temperature is increased from 255 K to 265 K while the liquid water mixing ratio is reduced. Since the reduction in liquid water is greater than the increase in cloud ice the net effect of increasing the critical temperature is to decrease the total cloud water present. This decrease in total cloud water arises from the more efficient conversion of cloud ice (rather than cloud liquid water) to snow.

It is interesting to note that there is little change in the maximum cloud ice mixing ratio between \( T_c \) of 260 K and 265 K; however, the vertically integrated amount of cloud ice changes noticeably (Figure 6). This is shown in Table 1 where the 6-hour mean IWP and LWP are given for each critical temperature. The IWP changes by about 10% for a 5-K change in critical temperature. It is also seen that the mean LWP is more sensitive to changes in critical temperature than the mean IWP. The total water path shows sensitivity to a reduction in critical temperature. These changes in cloud water partitioning and cloud water amount will have an appreciable effect on the radiative properties of the lead-induced cloud.

Although the snow field appears fairly robust as \( T_c \) is varied, the snow production rates are somewhat sensitive to changes in \( T_c \). The 6-hour mean snow production rate at two levels is given in Table 1 for critical temperatures of 255 K, 260 K, and 265 K. The mean snow production rate at 20 m is more sensitive to changes in \( T_c \) than the mean snow production rate at 700 m. At 20 m, the production of snow is slower when \( T_c \) is smaller. The change in snow production rate with varying critical temperature is somewhat reduced at higher levels where the air temperature is much less than the critical temperature.

![Figure 6. Comparison of vertical profiles of (top) cloud liquid water, (middle) cloud ice, and (bottom) snow mixing ratio after 2 hours for the following critical temperatures: 255 K (dotted curve), 260 K (solid curve), and 265 K (dashed curve).]
Table 1. Microphysical Properties of the Hydrometeor Plume for Various Critical Temperatures.

<table>
<thead>
<tr>
<th>Tc (K)</th>
<th>P3 (10^4 s^-1)</th>
<th>IWP (g m^-2)</th>
<th>LWP (g m^-2)</th>
</tr>
</thead>
<tbody>
<tr>
<td>265</td>
<td>4.5</td>
<td>40</td>
<td>47</td>
</tr>
<tr>
<td>260</td>
<td>3.4</td>
<td>3.9</td>
<td>43</td>
</tr>
<tr>
<td>255</td>
<td>2.6</td>
<td>4.8</td>
<td>38</td>
</tr>
</tbody>
</table>

Parameters have been averaged over 6 hours of model integration. Tc is the critical temperature where P3 is the net, 6-hour mean snow production rate (given at 2 heights). IWP and LWP are the 6 hour mean ice and liquid water paths, respectively.

Further analysis shows that when the critical temperature is 265 K the dominant snow production processes are accretion of cloud ice onto snow P_{Scal} and the ice-Bergeron process P_{3n} due to the increased availability of cloud ice. When a critical temperature of 265 K is used the liquid-Bergeron P_{Swb} process and accretion of cloud liquid water by snow P_{Snew} become more important than the conversion from cloud ice to snow as the amount of cloud liquid water available increases.

4.2. The Bergeron Processes

The ice-Bergeron process P_{3n} describes the conversion of cloud ice to snow through the deposition of vapor onto cloud ice embryos, while the liquid-Bergeron process P_{Swb} is described as the conversion of cloud liquid water to snow through riming and deposition. The two processes are interrelated, as both depend on a temperature-dependent time constant and the cloud ice mixing ratio. Lin et al. [1983] parameterize the two conversion rates as

\[ P_{3n} = B \frac{q_i}{\Delta t} \]  

\[ P_{Swb} = B \frac{q_i \Delta t}{M_{50} \Delta t} \left( a_1 M_{50}^{a_1} + \pi E_{iw} q_i r_{50}^2 U_{50} \right) \]

where B, the efficiency factor, has been added for this study, \( E_{iw} \) is the collection efficiency of cloud ice for cloud liquid, \( r_{50}, M_{50}, \) and \( U_{50} \) are the radius, mass, and terminal velocity of an ice particle with a 50 \( \mu m \) radius, \( a_1 \) and \( a_2 \) are temperature-dependent parameters \[ Koenig, 1971 \]. \( \Delta t \) is the model time step and \( \Delta t \) is a temperature-dependent time constant defined by

\[ \Delta t = \frac{1}{a_1(1-a_2)} \left[M^{(1-a_1)}_{50} - M^{(1-a_2)}_{50} \right] \]

where \( M_{50} \) is the mass of a 40-\( \mu m \) ice crystal. The time constant is the time required for a cloud particle to grow from 40 \( \mu m \) to 50 \( \mu m \) by vapor deposition, with 50 \( \mu m \) being the delimiter between cloud ice and snow. Because of the importance of this snow production mechanism (see Figure 3) and because it had been tuned to fit the observations of Itie et al. [1980], a series of sensitivity studies are done to determine the optimal formulation for this process in the environment encountered above an Arctic lead.

Figure 7 shows vertical profiles of cloud liquid, cloud ice, and snow for three ice-Bergeron process efficiency factors (B = 0.01, 0.20, and 1.0). It is seen from Figure 7 that when B is set to 1.0 the total cloud water mixing ratios are the smallest and several discrete cloud layers form, rather than a continuous cloud as observed \[ e.g., Andreas et al., 1990 \]. In addition, several unrealistic discontinuities are apparent in the snow mixing ratio profile as a result of the multiple cloud layers. When the efficiency factor B is set to 0.20 the modeled profiles of cloud water (ice and liquid) and snow are continuous and much more realistic than when B is 1.0. When B is set to 0.01, the cloud water to snow conversion rates are almost negligible. The cloud water mixing ratio becomes quite large during this simulation which has a noticeable impact on the radiative transfer through the lead-induced cloud (not shown).

The time-averaged, liquid water path LWP and ice water path IWP and the time-averaged net snow production rate P3 are given in Table 2 for the three Bergeron efficiencies. The effect of varying B on the amount of condensate is evident in the mean LWP and IWP. Note that changing B by 2 orders of magnitude (from 1.0 to 0.01) results in an increase in total cloud water amount of by a factor of 18. The total amount of cloud ice increases by a larger percentage (nearly 2000%).

Figure 7. Comparison of vertical profiles of (top) cloud water, (middle) cloud ice, and (bottom) snow mixing ratio after 2 hours for the following Bergeron efficiencies: 0.01 (dotted curve), 0.20 (solid curve), and 1.0 (dashed curve).
Table 2. Microphysical Properties of the Hydrometeor Plume for Various Bergeron Process Efficiencies

<table>
<thead>
<tr>
<th>$B_r$</th>
<th>$P_S$</th>
<th>IWP, $\times 10^8$ s$^{-1}$</th>
<th>LWP, g m$^{-2}$</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>20 m</td>
<td>700 m</td>
</tr>
<tr>
<td>1.0</td>
<td>6.9</td>
<td>0</td>
<td>10</td>
</tr>
<tr>
<td>0.20</td>
<td>3.4</td>
<td>3.9</td>
<td>43</td>
</tr>
<tr>
<td>0.01</td>
<td>2.7</td>
<td>2.0</td>
<td>208</td>
</tr>
</tbody>
</table>

Parameters have been averaged over 6 hours of model integration where $B$ is the Bergeron efficiency parameter and all other abbreviations are the same as in Table 1.

than the amount of cloud liquid (350%). Unlike the effect of a varying critical temperature, changes in $B$ result in significant changes in the mean snow production rate at all levels. In addition, changes in the Bergeron efficiency result in nonlinear changes in the time-averaged net snow production rate $P_S$. These nonlinear changes vary with height as well, demonstrating the complicated nature of this system.

4.3. Atmospheric Temperature

The sensitivity of the amount and phase of cloud water in the lead-induced cloud to air temperature is examined. The surface-based isothermal layer temperature is varied between 233 K and 253 K with the initial relative humidity within this layer constant with height at 85%. The critical temperature in (4) is 260 K and the Bergeron efficiency parameter is set to 0.2 for each simulation. A time series of the vertically-integrated amounts of cloud liquid water and cloud ice for each isothermal layer temperature is given in Figure 8. It is seen that the relative proportion of cloud ice increases with decreasing atmospheric temperature with almost all ice present for the coldest case. During the warmest case a greater proportion of the cloud is liquid.

The time evolution of the total cloud water path (LWP + IWP) varies between the three experiments. During the first hour of integration, the total cloud water path is largest for the coldest atmospheric temperature. This arises from the enhanced surface flux of moisture associated with the large air-sea temperature and moisture differences (see part 1). However, by 6 hours the total cloud water path is greatest for the warmest atmospheric temperature due to the increased presence of cloud liquid water which is converted to snow more slowly than cloud ice.

4.4. Radiative Properties

The sensitivity of the downwelling long-wave radiative flux at the surface to cloud particle size is examined here. Considerable uncertainty exists in the cloud particles size spectra (and thus the effective radius $r_e$) within a lead-induced cloud. Recall that for the baseline case the effective radii for cloud ice and cloud liquid were specified as 7 μm and 20 μm, respectively. In this sensitivity study the effective radii of both cloud liquid and cloud ice particles used in the baseline case are reduced by half 0.5$r_e$ and doubled 2$r_e$ to illustrate the importance of obtaining particle size spectra for a lead-induced cloud. Figure 5 shows the time evolution of the downwelling long-wave radiative flux at the surface for the baseline case, and particle sizes of 2$r_e$ and 0.5$r_e$. The initial (before 0.1 hour) jump in downwelling long-wave radiation, corresponding with the formation of cloud, decreases from 40 W m$^{-2}$ to 15 W m$^{-2}$ with increasing particle size. On average, the downwelling long-wave flux is 20 W m$^{-2}$ greater than the baseline case when the particle size is reduced by half and 16.5 W m$^{-2}$ less than that for the baseline case when the particle is doubled. Evidence for feedbacks between the lead-induced cloud and the surface sensible and latent heat fluxes are evident but small. The profiles of cloud water show slight changes as well (not shown).

The radiative effects of snow are invariably neglected in modeling studies. In the baseline case of this study, snow was not allowed to affect the radiative transfer through the lead-induced cloud. To deduce the consequences of this assumption snow is allowed to affect the long-wave radiative flux profiles. The radiative effect of snow is parameterized using a method analogous to that for cloud ice, with the effective radius for snow specified to be 100 μm. The result of including the radiative effects of snow was to increase the downwelling long-wave flux at the surface by 8 W m$^{-2}$ or about 5% throughout the simulation.

![Figure 8](image-url)  
**Figure 8.** Time evolution of (a) ice water path and (b) liquid water path for the following isothermal layer temperatures: 233 K (dotted curve), 243 K (solid curve), and 253 K (dashed curve).
5. Summary and Conclusions

A one-dimensional model with a second-order closure scheme for turbulence, bulk cloud microphysics, and a sophisticated radiative transfer scheme has been used to investigate the microphysical and radiative processes occurring in a lead-induced cloud. In addition to the turbulent fluxes of sensible heat and moisture into the lower troposphere, cloud-producing leads significantly affect the long-wave radiative heating rate profile, with strong cooling at the top of the lead-induced cloud and reduced long-wave cooling by the surface.

The model has improved our understanding of the microphysical and radiative processes occurring in a lead-induced cloud. A deep, mixed-phase cloud has been modeled for idealized Arctic conditions over the center of a wide open lead. Condensate formed preferentially at the top of the TIBL due to enhanced cooling in this region via radiative flux divergence (described in part I). Condensate near the surface was mixed phase, while condensate at higher levels was predominantly crystalline. As the lower part of the boundary layer warmed, cloud liquid and cloud ice water mixing ratios declined but snow originating from the elevated cloud layer continued to reach the surface resulting in an accumulation of 0.65 cm by 6 hours. Characteristics of the modeled cloud layer showed reasonable agreement with the limited observations of lead-induced clouds in terms of particle phase, cloud particle concentration, and vertical structure.

The ice-Bergeron process and the accretion of cloud ice on snow are the most important snow-producing mechanisms throughout most of the lead-induced cloud, except near the surface where the availability of liquid condensate increases the importance of the cloud liquid-to-snow conversion processes (accretion of cloud water onto snow and the liquid-Bergeron process). The modeled cloud layer and snow field were sensitive to the efficiency of the Bergeron processes. When the Bergeron efficiency factor B is increased, cloud particles are more quickly depleted so that only thin layers of cloud can be maintained. When B is reduced, the lead-induced cloud fills the entire TIBL and cloud particle concentrations become unrealistically large. Variations in the partitioning of cloud water into liquid and ice and/or atmospheric temperature significantly changed the composition of the cloud and associated snow field as well.

Radiative cooling rates were at a maximum near cloud top, with maximum values around -20 K d⁻¹. The downwelling flux at the surface increased by up to 70 W m⁻² because of the lead-induced cloud layer in the baseline case. This value was quite sensitive to the assumed particle size. When the radiative effect of snow was included the downwelling long-wave radiation at the surface increased by 8 W m⁻². Such increases in downwelling long-wave radiation due to lead-induced cloudiness may have a profound impact on thermodynamic sea-ice processes (e.g., slow the freezing of a lead). Infrared heating due to the convergence of long wave radiation was modeled through much of the TIBL with heating rates as high as 110 K d⁻¹ in the lowest model level.

In spite of the many uncertainties in model parameterizations, a reasonable, mixed-phase cloudy TIBL has been produced. The model exhibits sensitivity to several microphysical parameters including the critical temperature, the rate of snow production due to the Bergeron processes, and the assumed particle size. In situ measurements of the microphysical properties of lead-induced clouds are needed, particularly with regard to particle size, phase, and concentration. These measurements would improve our understanding of the microphysical processes occurring within lead-induced clouds and would aid in determining the impact these clouds have on the surface energy budget. Our ability to accurately model the radiative, sensible and latent heat fluxes at the surface of leads and the profile of these fluxes through the cloudy TIBL generated by some Arctic leads is a necessary step in determining the coupling between the atmosphere, ocean, and sea ice in the Arctic and ultimately the climate of this region.

Acknowledgments. This research was supported by ONR grant N0014-91-J-1387 and the DOE ARM program.

References


---

I A Curry and J O Pinto, Program in Atmospheric and Oceanic Sciences, University of Colorado, Campus Box 311, Boulder, CO 80309-0311.

(Received January 18, 1994; revised August 12, 1994; accepted August 20, 1994.)