Atmospheric convective plumes emanating from leads

1. Thermodynamic structure

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Abstract. A one-dimensional model with detailed radiative transfer, a second-order closure scheme for turbulence, and bulk cloud microphysics has been employed to simulate the convective atmosphere above a wide open lead. The development of a thermal internal boundary layer (TIBL) associated with the atmospheric convection above a wide open lead is modeled for several different temperature profiles characteristic of the Arctic winter. Large differences in the predicted top of the TIBL are observed between two simple diagnostic models and the time-dependent model introduced here. The modeled thermodynamic processes (i.e., radiative transfer and condensation/evaporation) occurring within the TIBL above an open lead are shown to account for as much as 20% of the predicted TIBL top height. Radiative heating dominates the heat budget in the lowest model level while sensible and latent heating occur throughout the TIBL. The production of cloud water affects the modeled TIBL top height via enhanced radiative cooling near the top of the TIBL and by latent heat released during condensation and freezing. Under stable conditions the TIBL top height above an open lead increases with time as a decaying exponential. The initial temperature of a surface-based isothermal layer can account for up to 50% of the variations in TIBL top height. The depth of the surface-based inversion layer and the stability of the atmosphere above the inversion are both important controlling factors in TIBL development while the temperature difference across the inversion layer is slightly less important.

1. Introduction

Linear breaks in the arctic ice pack or "leads" tend to form along weak points in the sea ice as a result of stresses incurred on the ice surface by wind and ocean currents. Just above an open lead, extremely large vertical temperature and moisture gradients exist during winter. These gradients lead to large fluxes of heat and moisture into the Arctic boundary layer. Andreas et al. [1979] observed sensible and latent heat fluxes just above open leads exceeding 400 W m$^{-2}$ and 100 W m$^{-2}$, respectively. These values are nearly 2 orders of magnitude greater than those encountered over the central ice pack during winter [Maykut, 1978].

In the central Arctic during winter, leads typically cover at least 1% of the area; however, newly formed leads remain ice free for only a short period of time. Under cold, calm conditions, a 100 m wide lead can freeze over in less than a day [Bauer and Martin, 1983], while convergent sea ice motion can close a lead in a few hours. Typically, the width of an open lead is of the order of 10-100 m, but much larger fractures do occur. Using an airborne near-infrared lidar, Barry et al. [1989] found several leads greater than 1 km wide during a 1300-km flight track over northern Baffin Bay in April. Using infrared satellite imagery and airborne infrared lidar, Schnell et al. [1989] observed several leads in excess of 5 km wide in the central Arctic during January. A summary of lead characteristics is given by Smith et al. [1990].

Until recently, the influence of leads was thought to be confined to the lowest few hundred meters of the atmosphere in their immediate vicinity owing to the lower tropospheric stability ubiquitous during the Arctic winter. It was believed that "steam fog," produced within the moist boundary layer above leads, could not possibly affect regional climate, since the upward motion of buoyant air parcels modified by the warm, moist underlying surface was prevented by a strong capping inversion [Badgley, 1966; Maykut, 1978]. However, using an airborne infrared lidar, Schnell et al. [1989] detected condensate emanating from an open lead that reached heights of up to 4 km. To attain this height there must have been enough moist potential energy available at the surface to overcome the strong capping inversion that was present. Once above the inversion layer the heat, moisture and cloud particles may be transported large horizontal distances downwind. Schnell et al. [1989] observed a lead-induced cloud extending over 250 km downwind of an open lead. The potential for vertical and horizontal development of these lead-induced clouds suggest that the widest leads may impact regional climate.

The thermal internal boundary layer (TIBL) top height downwind of an open lead has been examined by Schnell et al. [1989] and Serreze et al. [1992a]. Using a method developed by Andreas and Murphy [1986] to estimate surface-averaged fluxes from bulk aerodynamic formulas, Schnell et al. [1989] derived the fetch-dependent, temperature and specific humidity of air in the surface layer of the TIBL. The equivalent potential temperature of air in the surface layer is then used to
determine the equilibrium height for convective plumes originating at the surface of an open lead. Schnell et al. [1989] determined that an open lead fetch of 10 km is necessary to warm and moisten the ambient air at the surface sufficiently to produce buoyant plumes capable of reaching 4 km. Using a similar model, Serreze et al. [1992a] found that the derived height of buoyant convection tends to be largest under conditions of low surface wind speed, low surface air temperature, weak low-level temperature inversion, and large lead width. They point out that these conditions are also optimal for the rapid freezing of open leads.

In assessing the climatic importance of leads in the Arctic, the conditions under which leads merely affect the atmosphere in their immediate vicinity must be distinguished from those in which the thermodynamic effects of the leads will influence the atmosphere and surface energy budget over much larger areas. As evidenced by the surprisingly large lead-induced clouds observed by Schnell et al. [1989], convection from wide leads can impact regional climate. The production of lead-induced clouds has the potential to extend the thermodynamic impact of leads over broad regions through their radiative impact [see Pinto and Curry, this issue]. The lead-induced clouds, which have been observed during the polar night, will modulate the infrared radiative flux profile, resulting in enhanced downward infrared radiative flux at the surface and enhanced flux divergence at cloud top. The importance of clouds to the surface radiation balance and tropospheric temperature profiles over the central Arctic ice pack has been described by Curry et al. [1993] and Curry [1983], respectively.

In this paper we utilize a one-dimensional model with detailed radiative transfer, a second-order closure scheme for turbulence, and bulk cloud microphysics to simulate the evolution of a cloudy TIBL above a wide open lead. The model used in this study is considerably more complex than that utilized by Schnell et al. [1989] and Serreze et al. [1992a], allowing investigation of the effects of radiation, turbulent mixing, cloud formation and precipitation on TIBL development. The sensitivity of the TIBL structure and evolution to each of these processes is determined. The model is then run for several temperature profiles characteristic of the midwinter in the Arctic to assess the conditions necessary for the development of a cloudy TIBL which may have a regional impact. In a companion paper [Pinto and Curry, this issue] (herein after referred to as part 2), details of the microphysical and radiative processes that occur within the modeled hydrometeor plume are examined.

2. The Model

To model the convection above an Arctic lead, we adopt the conceptual model of a thermal internal boundary layer. An internal boundary layer in the atmosphere is associated with the horizontal advection of air across a discontinuity in some property of the surface [e.g., Garratt, 1990]. In the case of the Arctic lead the discontinuity arises primarily from changes in surface temperature; the change in surface roughness being less significant when the temperature discontinuity is large. As summarized by Stunder and Sethuraman [1985] and Garratt [1990], numerous studies have addressed the TIBL along coastlines, including analytical solutions, numerical models, laboratory experiments, and field measurements.

The formulation described by Weisman [1976] and Gamo et al. [1983] for the TIBL top height \( H \) assumes that the TIBL is heated uniformly and that the heat flux decreases linearly with height and is given as follows:

\[
H = \left( \frac{2H_s X}{pr_s \gamma U} \right)^{0.5}
\]

where \( H_s \) is the surface heat flux, \( X \) is the fetch, \( \gamma \) is the upwind potential temperature gradient, \( p \) is the density of air, \( c_p \) is the specific heat of air at constant pressure, and \( U \) is the mean wind speed. This formulation was determined by Stunder and Sethuraman [1985] to agree best with observations for conditions of nearly isothermal stratification of the onlflow and is used for comparison with results from our model.

We investigate the TIBL away from the upwind edge of the lead, so that form drag and fetch dependence of the surface fluxes, as discussed by Andreas and Murphy [1986] for Arctic leads, can be neglected. A one-dimensional model is used to simulate the evolution of the potential temperature, water vapor, cloud water (liquid and ice) and snow mixing ratio profiles over a wide, open lead. Implementation of a one-dimensional model requires that the fetch across the lead be great enough so that the inhomogeneous effects associated with the lead edge can be neglected. The fetch across a lead is a function of lead width and wind direction. To examine the evolution of the convection over a lead, the period of integration must be less than the value of \((X/U)\). Assuming a wind speed of 2 m s\(^{-1}\) for a fetch of about 45 km corresponds to a period of 6 hours over which the air is modified by the underlying sea. Since most leads are much longer than they are wide, one could imagine such a fetch if the winds are parallel to the lead.

Prognostic equations for potential temperature \( \theta \), total water vapor mixing ratio \( q_w \) (includes vapor and liquid and ice cloud particles), and snow mixing ratio \( q_s \) are formulated to describe the development of a TIBL over an open lead and are given by

\[
\frac{d\theta}{dt} = \frac{\partial}{\partial z} \left( \langle w' \theta' \rangle + Q^* + P_z \right)
\]

\[
\frac{dq_w}{dt} = \frac{\partial}{\partial z} \left( \langle w' q_w' \rangle - P_z \right)
\]

\[
\frac{dq_s}{dt} = \alpha \frac{\partial(\rho w' q_s)}{\partial z} + P_z
\]

where \( P_z \) and \( P_0 \) are production terms for snow and potential temperature, respectively, \( Q^* \) is the radiative heating rate, \( \alpha \) is the specific volume of air, \( w_0 \) is the terminal velocity of snow, and \( \langle w' q_w' \rangle \) and \( \langle w' \theta' \rangle \) are the turbulent fluxes of total water mixing ratio and potential temperature (angle brackets denotes an ensemble average), respectively. The snow production term \( P_z \) is the summation of all microphysical processes which affect the snow field [Lin et al., 1983]. The production term for potential temperature \( P_0 \) is the latent heating rate due to phase changes within a cloud and during snow production. Snow is removed through gravitational settling (first term in (4)), while turbulent diffusion of snow has been
neglected as a first approximation. $P_3$ and $P_0$ are described in more detail in part 2, along with the partitioning of the total water mixing ratio $q_w$ into the three phases. The turbulent fluxes are parameterized using a second-order closure scheme (level 3) developed by Yasuda and Mellor [1979], in which the mixing length, turbulent kinetic energy, temperature and total water mixing ratio variances, and temperature and total water mixing ratio covariances are predicted variables. Surface fluxes of heat, moisture, and momentum are determined using Moin-Obukhov theory, following Louis [1979].

Equations (2)-(4), including radiative transfer calculations $Q_*$ and cloud microphysics $P_3$ and $P_0$, are solved numerically using a 10-s time step. Profiles of temperature and water vapor mixing ratio serve as input to the model. The lower boundary is specified to be consistent with an open lead where the relatively warm ocean waters, near the freezing point of sea water, are exposed. The model is integrated over a 6-hour period which corresponds with the maximum time an air parcel may be expected to remain over an open lead. The vertical grid spacing of the model is specified by a stretch coordinate system, so that the highest resolution occurs just above the surface. The model domain extends from the surface to 10 km with a total of 80 grid points. The lowest level is at 5 m, above which the grid spacing follows,

$$dz = 2.5 z^{0.5}$$

where $dz$ is the vertical grid spacing which increases with height $z$ above the surface, so that at the top of the inversion (at 1700 m) the grid spacing is about 100 m.

3. Baseline Simulation

For the baseline case we choose idealized profiles of potential temperature and moisture, so we are able to more easily study each physical process (i.e., radiation, turbulence, and cloud microphysics) during the evolution of the TIBL. The potential temperature profile consists of a stable layer (characterized by $\delta \theta/\delta z = 10$ K km$^{-1}$) extending from the surface to about 1700 m and a potential temperature of 241.5 K at 5 m. Glendenning and Burk [1992] used a similar potential temperature profile in their large-eddy simulation of the convection above a narrow lead. The moisture within this layer is specified using a constant relative humidity with respect to liquid water of 85%, corresponding to a water vapor mixing ratio between 0.25 and 0.3 g kg$^{-1}$. Above 1700 m the potential temperature increases with height at 3.8 K km$^{-1}$. The horizontal wind speed and vertical wind shear are assumed to be small to simulate the regime of free convection. The large-scale vertical velocity has been set to zero as well, so that changes in the vertical profile are directly related to local physical processes associated with the lead.

The lower boundary is specified to be consistent with an open lead. A surface temperature of 271.35 K, the approximate freezing point of the ocean water in the Arctic, and a saturated surface, neglecting the solute effect of salt, which corresponds to a water vapor mixing ratio $q_w$ of 3.3 g kg$^{-1}$, are used as bottom boundary conditions. The surface characteristics are assumed to remain constant, corresponding to a lead that remains ice free.

3.1. Evolution of the Boundary Layer

The relatively warm ocean waters exposed in an open lead provide the thermal forcing necessary for convection and the growth of a TIBL. The time-dependent surface sensible and latent heat fluxes obtained using the Louis [1979] surface flux parameterization are shown in Figure 1. The sensible and latent heat fluxes are greatest at the initial time step with values of 258 W m$^{-2}$ and 70 W m$^{-2}$, respectively, decreasing nonlinearly until the end of the simulation. The mean Bowen ratio, which indicates the relative efficiency of warming versus moistening, is 2.8 with a maximum of 3.6 occurring at the initial time step. During the Arctic Ice Dynamics Joint Experiment (AIDJEX) Andreas et al. [1979] estimated surface sensible heat fluxes above a number of open leads obtaining values between 100 W m$^{-2}$ and 550 W m$^{-2}$. Latent heat fluxes were more difficult to estimate due to the lack of valid moisture measurements; however, for the cases in which moisture measurements were available Andreas et al. [1979] obtained Bowen ratios between 2 and 3.5.

Comparisons between our model results and observations is difficult. We assume an infinitely wide open lead so that ice edge effects can be neglected while leads observed by Andreas et al. [1979] were less than 100 m. The modeled fluxes are due primarily to the effect of buoyancy while the observations Andreas et al. include the effects of both vertical wind shear and buoyancy. Finally, the observed surface fluxes vary by as much as 40% depending on the method of estimation. In light of these factors, the modeled surface sensible and latent heat fluxes seem reasonable.

The evolution of modeled potential temperature $\theta$, total water vapor mixing ratio $q_w$, snow mixing ratio $q_b$, and turbulent kinetic energy $TKE$ is shown in Figure 2. The rapid warming and moistening of the lower atmosphere and formation of a TIBL is evident in Figure 2a and 2b. The stability within the TIBL is determined from the potential temperature trace (Figure 2a). It is seen that the lower portion of the TIBL is unstable (i.e., $\delta \theta/\delta z < 0$ K m$^{-1}$) as a result of the strong surface heating while the upper portion of the TIBL is nearly neutral. The top of the TIBL is denoted by the dashed line (Figure 2a) which denotes the transition in stability from near neutral to strongly stable. The total water mixing ratio (Figure 2b), although not well mixed within the TIBL, increases at each level throughout the simulation owing to the continuous flux of moisture from the surface. The vertical extent of moistening is indicated by the dotted 0.33 g kg$^{-1}$ contour. As the TIBL moistens and becomes supersaturated with respect to liquid, condensation occurs. Cloud water is then partitioned...
Figure 2. Time-height plots of modeled (a) potential temperature (b) total water mixing ratio (dotted contour of 0.33 g kg$^{-1}$ denotes top of the TIBL as determined from potential temperature trace), (c) snow mixing ratio, and (d) turbulent kinetic energy TKE (the zero contour is being used figuratively to denote the top of the TIBL) for the baseline simulation.
into liquid and ice producing a mixed-phase cloud as described in part 2. Snow, a total water sink, is produced in the cloud layer and is present throughout the TIBL with maximum snow mixing ratios near the surface that decrease to zero at cloud top (Figure 2c). Snow production, combined with gravitational fallout of snow, provides an important sink for total water.

The TKE field (Figure 2d) indicates that the strongest turbulence is occurring within the lowest 50 m with values in excess of 9 m² s⁻². The top of the TIBL is characterized by a local minimum in TKE (note that although TKE is greater than 0 m² s⁻² at each model level the zero contour has been added for illustrative purposes to indicate the top of the TIBL which corresponds with a local minimum in TKE). It is seen that the TIBL growth rate decreases with time, although steady state is not reached after 6 hours of integration.

The TIBL height H can be determined by using the potential temperature profile or the TKE profile. The top of the TIBL is characterized by an abrupt shift in stability from neutral within the TIBL to strongly stable above the TIBL. This region is also characterized by a minimum in TKE. The hourly TIBL H are determined using both potential temperature and TKE profiles and are plotted in Figure 3. It is seen that the values of H determined using TKE profile are nearly identical to those derived from the potential temperature profiles. In the remainder of this study TKE is used to determine the TIBL height.

Also given in Figure 3 are values of H determined using (1) and the relationship described by Serrize et al. [1992a]. Following Serrize et al. [1992a], the equivalent potential temperature a parcel of air attains as it traverses an open lead ϑₑ is determined. The initial equivalent potential temperature profile ϑₑ(z) is then vertically scanned to determine the height of neutral buoyancy H (where ϑₑ = ϑₑ(z = H)) at each hour of the simulation. This simple model neglects the effects of latent heat release and entrainment; however, it is a useful tool for estimating the TIBL height for large data sets. The TIBL height predicted by the Serrize et al. [1992a] formulation are significantly greater than those found here, while those obtained from (1) are systematically smaller.

### 3.2. Vertical Structure of the Boundary Layer

The modeled heat and moisture flux profiles after 1, 3, and 6 hours are shown in Figure 4. Both fluxes are nearly linear functions of height with minima occurring near the top of the TIBL and maxima in the surface layer (evident in Figure 5). The reversal of the vertical heat flux gradient near the top of the TIBL is indicative of entrainment. Both flux profiles evolve appreciably with time. The sensible heat flux increases with time between 300 m and the base of the "entrainment zone" (i.e., where the heat flux is negative at the top of the TIBL), and decreases with time below 300 m. The moisture flux increases with time throughout the TIBL, except in the surface layer where it remains relatively constant. The vertical gradient in both fluxes decreases with time.

Vertical profiles of the terms in (2) that contribute to the mean profile of potential temperature are shown in the left half of Figure 6 after 1, 3, and 6 hours of integration. The net heating rate at each level depends on turbulent flux divergence of θ or turbulent heating rate, \( H_θ = \partial \left( \omega \theta' \right)/\partial z \), latent heat release, \( P_L \), and radiative heating rate, \( Q^* \). The evolution of the each term contributing to the net heating rate profile is now discussed.

Heating by turbulent flux convergence \( H_θ \) is evident throughout most of the TIBL except at the top where much colder air is being entrained from above. At 1 hour \( H_θ \) exceeds 50 K d⁻¹ throughout most of the TIBL, while at the top of the TIBL, entrainment produces a turbulent cooling rate of -40 K d⁻¹. By 6 hours the turbulent heating rates have been reduced significantly owing to the reduced vertical gradient in the sensible heat flux (Figure 4).

Radiative heating occurs through most of the TIBL except in the upper 200 m. The largest heating rate occurs at 5 m throughout the simulation but decreases with time from 110 K d⁻¹ at 1 hour to about 80 K d⁻¹ after 6 hours. The strong radiative heating at the lowest model level arises from the close proximity to the warm surface. The divergence of infrared radiation due to the presence of cloud water results in strong cooling (maximum of over -20 K d⁻¹ after 6 hours) just below the entrainment zone. The radiative maximum cooling rate near the top of the TIBL increases slightly from 18 at 1 hour to -22 K d⁻¹ by 6 hours.

Latent heat is released throughout most of the TIBL with a maximum of about 20 K d⁻¹ occurring near the top of the TIBL at 6 hours. Note that the increase in maximum \( P_L \) at the top of the TIBL is proportional to the increase in radiative cooling \( Q^* \). Some negative values for \( P_L \) due to the evaporation of cloud liquid water and sublimation of cloud ice and snow are evident in the lower model levels.

The moisture budget (right half of Figure 6) is influenced by the turbulent flux divergence of total water, \( H_{qw} = \partial \left( \omega q_w' \right)/\partial z \), and the production of snow, \( P_s \). The turbulent flux of total water is positive throughout the TIBL at 1 and 3 hours; however, by 6 hours the turbulence term is negative in the lowest 300 m where the TIBL experiences drying. The precipitation flux is negative throughout the TIBL at both 1 hour and 3 hours but becomes positive at the lower levels by 6 hours due to the sublimation of snow (see part 2). The total water mixing ratio is not well mixed through the TIBL (Figure 2b) due to the removal of \( q_w \) by precipitation; however, the entire TIBL is moistening at 1 and 3 hours when \( H_{qw} \) is greater than \( P_s \). By 6 hours the TIBL below 300 m is moistening by sublimation of snow while \( H_{qw} \) moistens the rest of the TIBL.

![Figure 3](image-url)  
**Figure 3.** Time evolution of TIBL height H as determined from the modeled potential temperature profile, the modeled turbulent kinetic energy (TKE) profile, the Serrize et al. [1992a] method, and equation (1) [Weisman, 1976].
4. Impact of Physical Processes on TIBL Evolution

In this section we address the impact of individual thermo-dynamic processes on the evolution of the TIBL above an open lead. In these simulations there is no large-scale vertical velocity and horizontal wind speeds and vertical wind shear are assumed to be small. For the effects of vertical velocity and horizontal advection on lead convection, see Alam and Curry [this issue].

The thermodynamic processes that are examined here include turbulence, radiative transfer, cloud production, and precipitation. The results from five simulations are compared, each simulation consisting of a different combination of physical processes, with all simulations including turbulence. The simulations are denoted as follows: (1) "baseline," as described in section 3; (2) "minimum physics," for which only turbulent transport is considered; (3) "clear sky," for which clouds are not allowed to form but radiative transfer is included (note that the relative humidity is allowed to exceed 100%); (4) "cloud," for which the terms $P_g$ and $Q^*$ are set to zero (but the condensation and evaporation of cloud particles are included); and (5) "no radiation," in which cloud and precipitation processes are turned on but the radiative heating term $Q^*$ is set to zero.

The predicted TIBL height, as determined from the TKE profiles, is plotted as a function of time for each experiment in Figure 7. It is seen that after 3 hours, the baseline $H$ is 10-20% larger than in each of the other simulations. The importance of turbulent transport in the initial stages of TIBL development is evident as $H$ at 1 hour is the same for each simulation. The effects of radiative transfer, condensation, and precipitation become increasingly important after 1 hour of integration.

The TIBL evolves identically for the minimum physics and the clear sky simulations. Although the vertically integrated net heating rate (which can be related to TIBL height) is the same in both cases, it is achieved slightly differently. The inclusion of radiative heating must be balanced by changes in the turbulent flux profiles to produce the same time-dependent $H$ as that achieved during the minimum physics run. The lowest model levels are warmed radiatively, reducing the air-sea temperature difference and the flux of heat into the TIBL so that in the net heating rate profiles in the clear sky TIBL are nearly identical to those achieved in the minimum physics TIBL.

The TIBL height is smallest for the cloud run. In the absence of precipitation, cloud water mixing ratios are an order of magnitude larger than in the baseline simulation (not shown). The increased presence of cloud water, due to the lack of a removal process results in large evaporation rates by 2 hours in the lower model levels which has a stabilizing effect on the TIBL. In addition, the large amount of cloud water present reduces the buoyancy flux owing to the burden of condensed water, thus decreasing the production of TKE and the height $H$ attained by convection.

![Figure 4](image1.png)  
**Figure 4.** Vertical profiles of (a) sensible heat flux and (b) moisture flux after 1, 3, and 6 hours of the baseline simulation.

![Figure 5](image2.png)  
**Figure 5.** The lowest 100 m of (a) sensible heat and (b) moisture flux profiles after 1, 3, and 6 hours of the baseline simulation.
Figure 6. Vertical profiles of terms in the (left) heat and (right) moisture budget equations after 1, 3, and 6 hours for the baseline simulation where $H$ is the sensible heating rate, $Q^*$ is the radiative heating rate, $P$ is the latent heating rate, $P_s$ is the snow production rate, and $H_{qw}$ is the rate of moistening.

The impact of radiative transfer in a cloudy atmosphere is seen by comparing the no-radiation simulation with the baseline case. The TBL $H$ for the no-radiation run is 100 m less than for the baseline case; therefore, radiative transfer through the cloudy TIRL accounts for about 10% of the 6-hour TBL $H$. Since the TBL heights for the no-radiation case are identical to those obtained for the minimum physics run it appears as though cloud and precipitation processes have a
minimal effect on the 6 hour $H$ when feedbacks between radiation and cloud processes (e.g., increased radiative cooling increases the production of cloud water) are not allowed.

Examination of these results shows that the effects of including different thermodynamic processes are not simply additive. The omission of one physical process often resulted in a compensating change in another physical process. Feedbacks between each individual process must also be considered. Nonetheless, it is clear that turbulence is the dominant physical process in the evolution of the TIBL, with radiative transfer, condensation, and precipitation being less important. Neglect of these thermodynamic processes may lead to underestimates in the depth of the TIBL of 10-20%.

5. Sensitivity to Initial Temperature Profile

The influence of the surface-air temperature difference and lower tropospheric stability on the evolution of a TIBL above an open lead is investigated. A determination of the most ideal conditions for rapid TIBL development and maximum TIBL height will be made. Under these conditions we might expect leads to have an impact on a regional scale.

5.1. Air-Sea Temperature Difference

In these experiments the temperature of the surface-based isothermal layer $T_i$ is varied between 233 K and 263 K in 10-K increments to produce air-sea temperature differences between 8 K and 38 K. The initial isothermal layer depth (1700 m) and the relative humidity with respect to liquid water in the Isothermal layer (85%) are the same for each simulation. The surface sensible and latent heat fluxes and Bowen ratio at the initial time step are compared for each model run. The initial surface heat flux increases nonlinearly with decreasing $T_i$ with the sensible heat flux increasing more rapidly than the latent heat flux. Interestingly, the nonlinear dependence between the sensible heat flux and $T_i$ is opposite that found between the latent heat flux and $T_i$ as demonstrated by the sensitivity parameter $\lambda_i = \delta F_i / \delta T_i$, where $\delta T_i$ is the change in isothermal layer temperature (constant at 10 K) and the subscript $j$ indicates sensible or latent heat flux) given in Table 2. The sensible heat flux is the most sensitive to changes in surface-based isothermal layer temperature $\Delta T$, at colder $T_i$, while the latent heat flux is more sensitive to changes in $\Delta T_i$ at warmer $T_i$.

The time evolution of the TIBL height $H$ for varying surface-based isothermal layer temperature $T_i$ is shown in Figure 8. When the TIBL is contained within the surface-based isothermal layer the TIBL height depends on time as a decaying exponential. The TIBL height increases nonlinearly with decreasing $T_i$ (e.g., using results shown in Figure 8 we see that the 6-hour $H$ increases by 360 m for $T_i$ decreasing from 243 K to 233 K and by 260 m for $T_i$ decreasing from 253 K to 243 K). The increase in $H$ has been modulated by interactions between the turbulent fluxes, the radiative fluxes, and cloud and precipitation processes resulting in smaller increments in $H$ for the same 10-K change in $T_i$ when $T_i$ is warmer.

5.2. Lower Tropospheric Stability

To further understand the impact of the initial temperature profile on TIBL height, the influence of lower tropospheric stability is examined. Here we study the effect of varying inversion strength (temperature difference across the inversion $\Delta T$), inversion depth $\Delta z$, and stability above the surface-based inversion layer on TIBL development. The temperature profile characteristics are varied around the median temperature profile for February. Median, wintertime values for inversion depth (about 1200 m) and strength (10 K) as determined by Serreze et al. [1992b] are used to depict conditions typically encountered over the central ice pack. We assume a surface air temperature of 243 K, relative humidity of 85% through the inversion layer, decreasing thereafter, and a moist adiabatic lapse rate (6.7 K km$^{-1}$) immediately above the inversion for an idealized, central Arctic temperature profile. By systematically varying the inversion characteristics and the stability above the inversion, the influence of lower tropospheric stability on TIBL height can be deduced.

The first set of experiments the inversion depth is held constant at 1200 m, while the inversion strength $\Delta T$ is varied from 5 K to 15 K. The time-dependent TIBL height is shown in Figure 9 as a function of inversion strength. As expected, the TIBL deepens most quickly under the weakest inversion. The change in the height of the TIBL after 6 hours shows a linear dependence on $\Delta T$ with a $\pm 5$ K change resulting in a $\pm 130$ m change in $H$ or about $\pm 15$% of the 6-hour $H$ determined under median February conditions.

In the next set of experiments, the inversion strength is held constant at 10 K, while the depth of the inversion $\Delta z$ is varied. Assuming a surface air temperature of 243 K and a

<table>
<thead>
<tr>
<th>$T_i$</th>
<th>Sensible Heat Flux, W m$^{-2}$</th>
<th>Latent Heat Flux, W m$^{-2}$</th>
<th>Bowen Ratio</th>
</tr>
</thead>
<tbody>
<tr>
<td>233</td>
<td>411</td>
<td>87</td>
<td>4.7</td>
</tr>
<tr>
<td>(243)</td>
<td>(258)</td>
<td>(70)</td>
<td>(3.7)</td>
</tr>
<tr>
<td>253</td>
<td>137</td>
<td>50</td>
<td>2.7</td>
</tr>
<tr>
<td>263</td>
<td>44</td>
<td>25</td>
<td>1.8</td>
</tr>
</tbody>
</table>

Values for the baseline case are in parentheses.
Table 2. The Sensitivity of Surface Heat Fluxes to the Temperature of the Surface-Based Isothermal Layer

<table>
<thead>
<tr>
<th>$T_i$ Range, K</th>
<th>$\lambda_{sh}$, W m$^{-2}$ K$^{-1}$</th>
<th>$\lambda_{lw}$, W m$^{-2}$ K$^{-1}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>233 - 243</td>
<td>15.3</td>
<td>1.7</td>
</tr>
<tr>
<td>243 - 253</td>
<td>12.1</td>
<td>2.0</td>
</tr>
<tr>
<td>253 - 263</td>
<td>9.3</td>
<td>2.5</td>
</tr>
</tbody>
</table>

The sensitivity parameter $\lambda$, which is the ratio of the change in surface heat flux ($SH$, sensible heat, $LH$, latent heat) to the change in the initial isothermal layer temperature $T_i$, is computed for three 10-K increments in $T_i$.

moist adiabatic lapse rate above the inversion, the model is run for $\Delta T$ equal to 300 m and 600 m (in addition to the median February run of 1200 m). The importance of inversion depth in determining the TIBL height is evident in Figure 10. It is seen that once the TIBL breaches the inversion layer, it begins to deepen at a much faster rate. For an inversion depth of 600 m (half the median inversion depth) the TIBL depth breaches the inversion after about 4 hours, growing at a rate of 270 m h$^{-1}$. For the shallower inversion of 300 m the TIBL reaches through the inversion after 2 hours, and deepens at a rate of 360 m h$^{-1}$. This accelerated growth can be explained by the reduced amount of energy necessary for destabilizing the more shallow inversion layer and demonstrates the importance of inversion depth in TIBL development.

In the final set of experiments, three different lapse rates of -3.5 K km$^{-1}$, -6.7 K km$^{-1}$ (moist adiabatic), and -10.0 K km$^{-1}$ (dry adiabatic) are used above the surface-based temperature inversion. The surface air temperature is held constant at 243 K, the inversion strength $\Delta T$ is 10 K, and the depth of the inversion $\Delta z$ is 600 m. The results from the three model runs are compared in Figure 11. As previously seen, it takes approximately 4 hours to overcome a 600-m deep temperature inversion of 10 K. The differences in $H$ after 4 hours arise from the varying degrees of stability above the inversion. Once the inversion is breached, the rate of deepening increases with time and displays a strong nonlinear dependence on the stability above the inversion. For the adiabatic lapse rate, $H$ increases at a rate of 580 m h$^{-1}$ after 4 hours, which is more than double that obtained for the moist adiabatic lapse rate (280 m h$^{-1}$). The most stable lapse rate above the inversion layer results in the slowest rate of deepening of 150 m h$^{-1}$. The large variation in $H$ after 6 hours for this set of experiments is similar to that found when inversion depth was varied.

6. Summary and Conclusions

We have presented results from a one-dimensional model with detailed radiative transfer, a second-order closure scheme for turbulence, and bulk cloud microphysics. The model has been used to simulate atmospheric conditions directly above a wide, open lead. We have focused on the evolution of the thermal internal boundary layer (TIBL) associated with the convection above an open lead. Several experiments were undertaken to describe the impact of various thermodynamic processes on TIBL evolution and to illustrate the dependence of TIBL height on temperature and stability.

Figure 8. Time evolution of the modeled TIBL height $H$ for various surface-based isothermal layer temperatures $T_i$.

Modeled surface heat and moisture fluxes are comparable to observations which have been made over leads less than 100 m wide. The maximum modeled sensible heat and moisture fluxes from the lead into the atmosphere occur initially, decreasing with time as the lower atmosphere warms and moistens reducing the air-sea temperature and moisture gradients. Sensible heat and moisture flux profiles above an open lead have been simulated. Unfortunately, the magnitude of the fluxes and the structure of the flux profiles cannot be directly verified since observations of such open leads have not been made. Nonetheless, the simulated profiles exhibit some interesting features. The maximum sensible heat and moisture fluxes occur at the lowest model level, causing a dramatic spike within the surface layer. Both the heat and moisture fluxes decrease linearly with height above the surface layer. Cooling by entrainment is evident at the top of the TIBL.

The TIBL height shows an exponential decay with time for cases of strong stability (i.e., when the TIBL is contained with in the surface-based inversion or isothermal layer). For the range of conditions examined here the TIBL height does not reach steady state after 6 hours. For the baseline case the method discussed by Serreze et al. [1992a] predicts a TIBL height that is nearly double that predicted here, while the equation developed by the Weisman [1976] systematically underpredicts our modeled TIBL height.

The effects of each thermodynamic process are not additive, the omission of one physical process may be compensated by changes in another. In the lowest 20 m of the atmosphere, the radiative heating rates are larger than the sensible heating rates. When radiative transfer is neglected, the lack of radiative heating in the lowest model levels is compensated by an increase in sensible heat flux convergence (i.e., warming). The TIBL height is modulated by cloud and precipitation processes and the complex interactions between clouds, radiative transfer and turbulence. Clouds destabilize the TIBL by radiative cooling at the top of the TIBL and warming due to radiative flux convergence in the lower model layers. Radiative cooling at cloud top is somewhat compensated for by latent heat released during cloud formation and freezing of cloud water. The radiative warming in the lower model levels is compensated for by a reduced turbulent flux convergence. Neglect of radiative transfer, condensation, and precipitation processes results in an underestimate in TIBL depth on the order of 10-20%.

The importance of the air-sea temperature difference and stability within and above the inversion layer has also been
Figure 9. Time evolution of the modeled TIBL height $H$ for various inversion strengths ($\Delta T$), with inversion depth ($\Delta z = 1200$ m) and above inversion stability ($\Delta T/\Delta z = 6.7$ K km$^{-1}$) held constant.

studied. For the range of surface-based isothermal layer temperatures studied here (233 K to 263 K), it was found that the 6-hour TIBL height varied by over 800 m or more than 50%. When the TIBL is contained within the inversion or isothermal layer, the TIBL growth rate decreases with time. Once the top of an inversion is reached the TIBL deepens at a much faster rate. For a shallow, surface-based inversion layer the stability aloft is an important control on TIBL height. The optimal conditions for rapid penetrating development of the TIBL include a weak, shallow, surface-based inversion layer topped by a much less stable layer.

A one-dimensional model has been used to determine the importance of turbulence, cloud formation, long-wave radiative transfer, and precipitation processes on the evolution of a TIBL above a wide open lead. Two dimensional effects such as horizontal inhomogeneity are examined by Alam and Curry [this issue] who use a two-dimensional model that neglects the effects of radiation, condensation, and precipitation. The results from Alam and Curry [this issue] indicate that the TIBL height is enhanced by lead-induced circulations, with rising motion occurring over the lead. Interactions between dynamic and thermodynamic processes in these circulations could be deduced by parameterizing the thermodynamics in the model described by Alam and Curry [this issue] using results from this one-dimensional modeling study. Further, this study has helped to identify which variables (e.g., heat and moisture flux profiles, temperature, moisture, and cloud water profiles and cloud top and surface irradiances) need to be measured above and downwind of open leads to improve parameterizations of air-sea-ice interactions in the vicinity of leads in the Arctic. A better understanding of the turbulent and radiative flux profiles above and downwind of wide open leads through observation and more sophisticated models must be obtained to assess the overall impact of leads on regional climate in the Arctic.

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References


Figure 10. Time evolution of the modeled TIBL height $H$ for varying inversion depths $\Delta z$ with inversion strength ($\Delta T = 10$ K) and above inversion stability ($\Delta T/\Delta z = 6.7$ K km$^{-1}$) held constant.

Figure 11. Time evolution of the modeled TIBL height $H$ for varying degrees of stability above the inversion layer for a constant inversion depth ($\Delta z = 600$ m) and strength ($\Delta T = 10$ K).


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