Evolution of new ice and turbulent fluxes over freezing winter leads

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Abstract. A model is presented that describes the coevolution of surface turbulent fluxes and new ice growth during the freezing of leads. The model includes a sophisticated parameterization of the surface sensible and latent heat fluxes. The new ice growth model distinguishes between the congelation and frazil regimes. During frazil growth, heat loss from the open water part of the lead results in formation of new ice which is advected to the downwind edge. With time, the ice covers the upwind lead edge, and the lead is gradually covered with ice. Over the ice-covered portions, the turbulent heat loss results in ice consolidation, and thereafter ice growth occurs. The turbulent heat flux depends on the surface characteristics which vary across the lead surface during frazil growth. Therefore in the frazil regime, ice concentration, ice thickness, surface temperature, and the surface turbulent flux vary across the lead surface. Even after consolidation, frazil ice has a different surface roughness length from congelation ice for the same ice thickness up to an ice thickness of 10 cm. We have used this model to determine the evolution of surface turbulent heat fluxes under various atmospheric conditions and for different lead widths. In the frazil regime, there is a considerable fetch dependence of the surface characteristics, as the ice is advected to the downwind edge and slowly covers the entire lead. This fetch dependence is greatest for the higher wind speeds and larger lead widths. There is significantly higher ice production under conditions when frazil formation occurs because the ice transport to the downwind edge leaves the surface of the lead open, allowing the warmer sea surface to exchange heat with the atmosphere. The rapid growth rates result in large salt release to the ocean, with implications for ocean dynamics. We have done a sensitivity study to investigate the effect of oceanic heat flux at the underside of the ice, which results from the salt rejection upon ice formation in freezing leads, on the evolution of new ice and turbulent fluxes.

1. Introduction

Differential motion of the atmospheric winds and oceanic currents, internal ice stresses, and topographical effects can cause the sea ice to fracture and expose the ocean, resulting in leads. Vertical temperature gradients over winter leads can reach 30°-40°C over a distance of only a few meters, causing large losses of sensible and latent heat from the ocean [e.g., Andreas et al., 1979]. Although open and newly frozen leads compose only a small fraction of the central Arctic surface area, they are a major mechanism by which heat and moisture are exchanged between the ocean and atmosphere. The significant heat loss from the surface of arctic leads, particularly in winter, results in freezing at the lead surface and the production of new ice. Most previous attempts at modeling lead effects have assumed that the leads are entirely open water [Alam and Curry, 1995, 1997; Gtrending, 1995; Andreas and Murphy, 1986] or assumed that if there is freezing, it occurs evenly over the entire lead [Maykut, 1986]. However, leads have been observed to freeze over starting from the downwind side as the frazil ice crystals are advected downwind to pile up against the lead edge [Martin, 1981]. Although this freeze over reduces the heat fluxes substantially, significant heat fluxes still occur over newly frozen leads. Therefore understanding the effects of the freezing process on turbulent surface fluxes is crucial to understanding the entire surface heat budget of the Arctic in winter. Winter rates of ice growth, turbulent heat exchange with the atmosphere, and salt rejection to the ocean can be up to 2 orders of magnitude larger over a refreezing lead than over perennial sea ice. If an ice pack were composed of 90% perennial ice and 1% refreezing leads, winter ice production in the small area covered by leads would roughly be equal to that in the area covered by the perennial ice [Maykut, 1982]. Although decreasing rapidly with increasing ice thickness, ice growth rates can still be an order of magnitude larger for 30-cm-thick ice than for 3-m-thick ice [Maykut, 1986]. As a lead freezes over, the exchanges of heat, water vapor, and salt between the ice and the adjacent atmosphere and ocean gradually diminish. Once sea ice thickness has reached about 1 m, there is little difference in the exchanges relative to multiyear ice. To address these issues, this paper presents a physically based lead freezing model focusing on polar surface thermodynamics which includes frazil ice growth to study the evolution of the surface characteristics and turbulent surface fluxes, as a function of lead width and atmospheric conditions, as the lead freezes.

A brief description is given here of the evolution of the initial ice cover. Heterogeneous nucleation of sea ice occurs on solid impurities in sea water and on snow crystals deposited on the surface [Weeks and Ackley, 1986]. Ice forms as minute spheres of pure ice which rapidly grow into thin circular discs of up to 2-3 mm in diameter, developing into dendritic crystals which grow rapidly across the calm sea surface until they overlap and freeze together forming a continuous thin ice skin...
called congelation ice. Commonly, however, the open ocean
has some wave-induced turbulence during the initial ice forma-
tion so that the ice crystals, termed frazil, may be stirred
throughout a depth of up to several meters [Martin and Kauff-
man, 1981]. A threshold surface wind velocity is necessary un-
der freezing conditions to allow the onset of frazil formation
since sufficient mechanical shear is necessary to mix the flow
and carry the newly formed ice crystals rapidly away before a
complete skim of ice can form. Wind action herds the frazil
crystals into grease ice, which is a soupy layer with low re-
flexivity, matte-like appearance, and viscous fluid-like properties.
Martin and Kaufman [1981] describe a wave-damping experi-
ment, indicating that a concentration of about 40% by volume
of frazil ice crystals is necessary for the mechanical properties
to transition to more solid characteristics. Field experiments
[Martin, 1981] show that the frazil crystals are advected by
wave action and wind-driven surface drift currents to pile up
against the downwind lead edge to depths of the order of
0.05-0.3 m. Bonding between frazil particles occurs when they
are brought into contact and can form pancake ice, which con-
sists of roughly circular pieces of new ice with upturned edges
as constant abrasion by other pancakes removes sharp corners.
As the ice formation and advection process continues, the ice
derge extends upward with time.

During the formation of the initial ice cover, complex inter-
actions occur between the ice, atmosphere, and ocean. Accurate
treatment of new ice production and the associated fluxes is es-
ential for accurate determination of the sea ice mass balance
[e.g., Schramm et al., 1997]. While we have a qualitative un-
derstanding of new ice growth, time series observations docu-
menting concurrently the new ice growth and fluxes with the
atmosphere and ocean surface layer variables do not exist. Our
understanding of the new ice growth and its interactions with
atmospheric and oceanic surface layers is therefore lacking. As
a result, new ice growth and its turbulent exchanges with the
ocean and atmosphere is treated very crudely in climate models.

To put this modeling study in a historical perspective, some
previous process studies of the modeling of ice formation are
described below. The congelation ice growth, which occurs
during relatively calm conditions, has been extensively consid-
ered by Maykut [1978, 1986]. Under turbulent conditions,
Bauer and Martin [1983] have used surface dynamics to compute
the downwind grease ice pileup depth which, combined with the
ice production rate, gives the advance of ice over the lead sur-
face with time. Pease [1987] scaled the ice production rate by
the collection depth of the grease ice, which was considered as
an independent variable, to determine the maximum size at-
tained by a polynya. Ou [1988] incorporated finite surface drift
to examine the temporal behavior of a coastal polynya.
Omstedt and Svensson [1984] have studied the ice formation
process in detail by including supercooling, freezing, and ris-
ing of the buoyant frazil ice crystals to the surface.

The present model computes the evolution of the lead surface
characteristics and the corresponding surface fluxes by consid-
ering the ice formation and growth under various atmospheric
conditions for different lead widths. It extends the Maykut
[1986] model for young ice growth by including frazil ice for-
mation. The present model extends the Bauer and Martin
[1983] grease ice model by considering heat loss from the
frazil ice which leads to ice consolidation and growth with time
thus affecting the surface fluxes. The present model differs
from Pease [1987] and Ou [1988] polynya models by formulat-
ing a downwind pileup depth based on the wind speed and lead
width. Also, unlike the polynyas considered in these two mod-
els, a lead does not have a steady offshore wind so that the open
water portion of a lead decreases with time. We have not con-
sidered the details of the supercooling, sinking, freezing, and
buoyant rising of the frazil ice crystals formed [e.g., Omstedt
and Svensson, 1984] as this model considers the ice concentra-
tion, extent and thickness of the frazil layer after the frazil ice
crystals have risen to the surface and have been advected to the
downwind edge. In a study based on the Bauer and Martin
[1983] grease ice model. The concentration throughout this
paper refers to the the relative ice volume or the volume con-
centration of ice.

Ice formation on the surface affects the interaction between
the ocean and the atmosphere by changing the surface rough-
ness and the surface temperature which is in contact with the
atmosphere. Ruffieux et al. [1995] have described observations
obtained in the spring of 1992 in the Beaufort sea during the
Lead Experiment (LEADEX) over some refreezing leads. They
obtained a roughness length of 2 mm for an ice thickness of
10 cm which indicates a substantially greater roughness for
new ice than over an open lead, for which z0 is of the order of
0.1 mm. This implies an increase in the 10-m neutral drone co-
efficient from 1.2x10^4 for open water to 2.2x10^5 for a lead
covered with 10-cm thick new ice. They also observed that
the ice formed on the lead was thicker on the downwind side. They
obtained a downwind flux of 170 W m^-2 with ice thickness of
10 cm for the 2-day-old, 1-km-wide lead with a wind speed of 6.6
m s^-1. The freezing ice interface also introduces a heat and salt
flux between the ice and ocean. From the springtime oceanogra-
phical measurements of LEADEX, McPhee and Stanton
[1996] found that new ice of thickness 5-10 cm, heat was
extracted from the mixed layer at a rate of 10 30 W m^-2 in
the nighttime between solar heating events. In an ice thickness
distribution model with a coupled ocean mixed layer model
whose annual cycle of mixed layer temperature, salinity, and
depth agreed well with the drifting Arctic Ice Dynamics Joint
Experiment (AIDJEX) ice station observations, Holland et al.
[1997] determined that the wintertime turbulent heat flux at the
ice/ocean interface underneath first-year ice averaged 10 W m^-2,
where first-year ice has a thickness of less than 80 cm. For
growth rates corresponding to newly formed ice, they obtained
a turbulent heat flux as high as 55 W m^-2 at the ice/ocean in-
terface.

Toward improving our understanding of the physical pro-
cesses that interact in the freezing of leads and developing im-
proved parameterizations for climate models, we have formu-
lated a one-dimensional model in the across-lead direction for
determining the evolution of the new ice formation and associ-
atud turbulent fluxes with the atmosphere and ocean, varying
atmospheric conditions and the lead width. The model com-
putes the evolution of surface freezing characteristics, (e.g., ice
concentration in the case of frazil growth, ice thickness, and
surface temperature) and the resulting surface turbulent heat
fluxes for various atmospheric conditions and lead widths. We
find that, in the congelation regime of small lead width or low
wind speed, the ice grows rapidly at first and then proceeds with
a decreasing growth rate, so that surface ice temperature and the
resultant turbulent heat flux decrease sharply initially and then
more slowly with time. In the regime of frazil ice growth, the
ice crystals are advected to the downwind edge allowing the up-
wind ice free portion of the lead to lose heat by turbulent re-
lease from the open water. Hence the frazil case has a strong
fetch dependence of the ice concentration, thickness, surface
temperature, and turbulent heat flux, especially in the initial stages, and this fetch dependence increases with increasing lead width and wind speed.

2. Model Description

The new ice model is one-dimensional, allowing for zero-dimensional vertical ice growth and its variation in the across-lead direction. The surface radiation fluxes are specified, as are the atmospheric air temperature, humidity, and wind speed at 10 m and the ocean mixed layer temperature and salinity. The surface turbulent fluxes at the atmospheric interface are calculated interactively in the model, reacting to the changing surface temperature and roughness. A numerical solution to the model equations is found using a horizontal resolution of 1 m across all leads and a time step of 3 min (although a longer time step is used in the extended integrations of congelation growth).

2.1. New Ice Growth

Depending on the ambient conditions, new ice in freezing leads can form as sheet ice through congelation growth or as frazil ice crystals which are advected to the downwind lead edge. From the results of the grease ice model of Bauer and Martin (1983), we have arrived at the threshold wind speed of 4.35 m s\(^{-1}\) for transition to the frazil regime. Bauer and Martin (1983) computed the downwind ice pileup depths for various lead widths with different 10-m wind speeds. This ice pileup depth or collection thickness in their model goes to zero for a wind speed of 4.35 m s\(^{-1}\), so we have chosen this value as the wind speed for transition from congelation ice to frazil formation in the model. The lead also needs to have sufficient width for frazil formation, since the wind- and wave-induced turbulence is extremely damped when a significant percentage of the sea surface is covered by ice. This threshold width for frazil formation, which is undoubtedly dependent on the wind speed, is uncertain. Following Bauer and Martin (1983), we use a lead width threshold of 50 m.

2.1.1. Congelation growth. Under conditions favorable for congelation ice growth, the growth rate of the initial congelation ice layer is based on Maykut (1986) and is given by

\[
\frac{dh_i}{dt} = \frac{Q'}{c_i L_i}
\]

where \(h_i\) is the ice thickness, \(dh_i/dt\) is the ice growth rate, \(Q'\) is the net heat flux at the surface which is at the freezing point, \(c_i\) is the ice concentration which is unity for the congelation layer, \(\rho_i\) is the ice density, and \(L_i\) is the latent heat of fusion of ice.

Once the initial ice skim is formed, further ice growth occurs from the energy deficit at the bottom of the ice. At the underside of ice of an ice layer with a thickness \(h_p\), the accretion/ablation is given by

\[
\frac{dh_p}{dt} = \frac{-(F_c + F_w)}{\rho L_i}
\]

where \(dh_p/dt\) is the basal freezing rate or the rate of ice accretion at the underside. \(F_c\) is the conductive heat flux at the bottom and is negative when directed upward or away from the bottom toward the surface which results in ice growth. \(F_w\) is the oceanic heat flux and is positive when directed toward the bottom of the ice, as is generally the case since freezing releases salt which results in a bulk ocean mixed layer temperature and that is colder than the underside of the ice. The amount of growth (or ablation) at the bottom of the ice is determined by the sum of \(F_c\) and \(F_w\). If the sum is negative (directed away from the bottom of the ice), ice will grow, releasing latent heat to balance the energy deficit.

For very thin ice (thickness less than 0.3 m), the temperature gradient in the ice is essentially linear, as young ice responds rapidly enough to changes in surface thermal forcing to maintain an essentially linear temperature profile through the ice [Maykut, 1978]. Maykut [1978] assumed that the maximum thickness for which ice can maintain a linear profile is 0.8 m. Guest and Davidson [1994] defined the depth scale to which an ice floe can thermally interact with the surface as a function of the timescale of atmospheric forcing. On the basis of their definition, we find that 34-cm-thick ice can respond to changes in the atmospheric scale of greater than a day; and for an ice thickness of 80 cm to respond, the variations in the atmospheric forcing would have to be on a timescale of longer than 5 days. Since the ice thicknesses we obtain in the newly freezing leads are generally less than 40 cm, we have assumed a linear temperature profile through the ice. This means that the conductive flux at the top of the ice \(F_c(z = 0)\) equals the conductive flux at the ice bottom \(F_c(z = h_i)\). The conductive heat flux in thin ice is given by Maykut [1982]:

\[
F_c = \gamma(T_f - T_0)
\]

where \(T_f = \gamma/k_i\) is the thermal conducance of the ice slab where \(k_i = 2.034\) W m\(^{-1}\) K\(^{-1}\) is the conductivity of ice, \(h_i\) is the ice depth, the temperature at the bottom of the ice \(T_f\) is specified to be the freezing point of seawater which equals -1.7°C for a salinity of 30.8 practical salinity units (psu), and the temperature at the surface \(T_0\) is free to vary in response to changes in the surface energy balance. Upon ice formation, the ice layer separates the melt (sea water) from the cold source (air). For thin enough ice, internal heat storage can be neglected, and the net heat loss at the surface is extracted from the accretion of ice at the underside of the ice, and the growth rate is determined from (2).

Growth of an ice cover from arbitrary initial conditions can be calculated using (1)-(3) for any time-dependent thermal forcing [Maykut, 1986] if \(T_f\) and \(F_c\) are known. Determination of \(T_0\) from the surface energy balance is described in section 2.2.2. In the baseline simulations, the value of \(F_c\) is specified to be 0 W m\(^{-2}\), while in section 2.3 we describe a parameterization for \(F_c\) and explore the model sensitivity to \(F_c\) in section 6.

2.1.2. Frazil growth. Under turbulent conditions, frazil ice crystals that are formed are advected to the downwind lead edge, leaving the upwind portion of the lead ice free. The combination of this ice formation and advection process results in the ice edge advancing upwind with time. Heat loss from the icy portions of the lead results in ice consolidation and growth. Figure 1 is a schematic of a snapshot across the lead surface near the downwind edge of the lead during freezing in the frazil regime. There is open water and surface waves in the ice-free upwind portion of the lead. At the ice edge, there is frazil which has formed at a concentration of 0.2, and further downwind the ice concentration increases. Damped waves exist on the surface until the ice concentration reaches 0.4, beyond which the grease ice behaves like a solid and completely damps
out the waves [Martin and Kauffman, 1981]. Beyond the fetch where the volume ice concentration reaches unity, the ice thickness starts to increase with further heat loss and ice formation.

For wind speeds higher than 4.35 m s⁻¹ and lead widths greater than 50 m, we assume that frazil ice formation occurs. Frazil ice crystals form on the surface of the lead and are transported to the downwind edge by the wind waves and wind driven surface current. We have used a parameterization based on Bauer and Martin [1983] to calculate the thickness of the surface frazil ice and the formulation of Peuse [1987] to compute the extent of coverage of the grease ice on the lead surface as a function of time. Bauer and Martin [1983] determine the ice production rate from the surface heat flux. They use the wave field and surface current, which are functions of wind speed and fetch, to compute the ice pileup depth and the ice cover advance rate from the downwind edge. As the ice formation and advection process continues, the ice accumulation zone on the lead surface advances to the upwind edge of the lead. The surface heat flux in the open water determines the advance of the ice edge across the lead. In the Arctic winter, the total heat loss at the surface consists of the surface turbulent sensible heat flux, latent heat flux, and the long wave radiation loss from the lead surface. Bauer and Martin [1983] assume that the total heat loss \(Q^*\) goes into ice production.

We compute the ice production rate in terms of the volume of ice produced per second so that we can scale it with the ice thickness later when we compute the coverage rate of ice on the lead surface. The energy required to produce 1 m³ of grease ice (20% ice concentration) is \(0.2\rho L_i\), where \(\rho\) is the density of ice. The initial ice concentration is 0.2 based on the work of Martin and Kauffman [1981], who found from a laboratory study that the ice concentration at the leading edge of newly formed grease ice varied from 18% to 22%. The ice production rate is

\[
P_i = \frac{Q^*}{0.2\rho L_i}
\]

where \(P_i\) is the ice produced in meters produced per second per square meter of the surface area of the lead. Given the ice production rate \(P_i\), the model of Bauer and Martin [1983] calculates the thickness to which the grease ice piles up at the downwind edge, which is the ice pileup depth, as a function of wind and fetch and the ice advance rate as a function of wind, fetch, and time. We have parameterized the pileup depth of the grease ice \(h_i\) computed by Bauer and Martin [1983] as a function of the wind speed \(u_c\) and the total lead width \(X\)

\[
h_i = 0.055418 \times 1.1592 \times 10^{-4} X + (0.012761 + 2.6693 \times 10^{-5} X) h_i
\]

so that the frazil ice which forms covers the downwind portion of the lead at a thickness of the ice pileup depth \(h_i\) and the ice-covered portion is calculated as shown below. The ice edge advances upwind with time based on the ice production rate due to heat loss from the open water portion of the lead. Since the pileup depth \(h_i\) is the initial ice thickness and determines the rate of advance of ice across the lead surface and the rate of ice consolidation and growth over the ice-covered portions, this parameterization (5) has a strong effect on the evolution of surface characteristics and the turbulent fluxes in the model.

Knowing the ice production rate \(P_i\) and the grease ice layer depth \(h_i\), the advance of the grease ice layer from the downwind edge with time is computed following Peuse [1987]. We assume that over the course of the frazil ice growth the lead does not change in width dynamically, which implies that we have neglected ice divergence during freezing. The change in open water lead width with time \((dx/dt)\) is given by the ice production rate \(P_i\) over the open water portion of the lead \(X_p\), scaled by the collection depth of the grease ice \(h_i\)

\[
\frac{dx}{dt} = \frac{-X_p P_i}{h_i}
\]

Equation (6) can be solved for \(X_p\) as a function of time if we know the initial open water lead width \(X\), so that

\[
X_p = X \exp \left[ -\frac{t P_i}{h_i} \right]
\]

This gives us the extent of lead which remains ice free \(X_p\), as a function of time.

We assume that frazil ice forms at an initial concentration of 0.2 [Martin and Kauffman, 1981]. Over the ice-covered part of the lead, the surface energy balance is applied at the top and bottom of the ice layer at every grid point to compute the increase in ice concentration. As discussed in section 2.1.1, the net upward heat loss at the bottom of the ice, which is a directional sum of heat lost by conduction and the heat gained from the ocean, results in a concentration increase given by

\[
\frac{dc_i}{dt} = \frac{-(F_s + F_w)}{\rho L_i h_i}
\]

In (8) a negative sum of \(F_s\) and \(F_w\) denotes net flux directed away from the ice bottom and an ice concentration increase. When the concentration at a grid point reaches unity, further heat loss results in increase of ice thickness by congelation growth as described in section 2.1.1.

Bauer and Martin [1983] used their grease ice model to obtain ice pileup depths for different wind speeds and lead widths and obtained the time for ice coverage of lead for different air-sea temperatures, wind speeds, and lead widths. We have used their model to parameterize the initial frazil ice thickness or pileup depth and have determined the coverage times based on (7) from this pileup depth and the net heat loss tabulated by Bauer and Martin [1983] for different atmospheric conditions.
2.2. Surface Energy Balance

During winter, in the absence of solar radiation and surface melting, the surface energy balance for sea ice can be written as

$$0 = F_{lw} + F_{SH} + F_{LH} + F_C$$

In (9) $F_{lw}$ is the incoming long wave radiative flux, $\varepsilon_0 T_o^4$ is the outgoing long wave radiation, $F_{SH}$ is the sensible heat transport, $F_{LH}$ is the latent heat flux at the surface, $F_C$ is the conductive flux, $\varepsilon_0$ is the long wave emissivity of the surface, $\sigma = 5.67 \times 10^{-8}$ W m$^{-2}$ K$^{-4}$ is the Stefan Boltzmann's constant, and $T_o$ is the surface skin temperature. We assume that incoming short wave radiation is zero and specify the downwelling long wave radiation $F_{lw}$ (171.5 W m$^{-2}$) using observed winter climatology in the central Arctic from Maykut [1978]. When the ice concentration is less than 0.4, the long wave emissivity $\varepsilon_0$ equals that of water (0.97), and when the ice concentration is 1, the long wave emissivity $\varepsilon_0$ equals that of ice (0.99). For ice concentration between 0.4 and 1, $\varepsilon_0$ varies linearly from $\varepsilon_0$ of water to that of ice based on the concentration. The calculations of sensible and latent heat fluxes are described in detail in section 2.2.1, and the determination of the conductive heat flux is given in section 2.1.1.

The energy balance at the surface determines the surface temperature, which is needed in the determination of all of the components of the surface fluxes, including the conductive flux at the surface. The temperature gradient through the ice is linear for thin ice, and hence the bottom conductive flux equals the surface conductive flux. This, when combined with the ocean flux to the underside of the ice, gives the ice accretion/ablation growth. When ice is growing, there is an upward flux of heat in the ice which supplies much of the energy needed to balance the energy loss (radiative and turbulent) at the surface. During winter, temperatures in the upper part of the ice are colder than those at the underside, resulting in an upward conduction of heat. Sensible heat and latent heat of evaporation are transferred across the lower part of the atmospheric boundary layer as a result of temperature and moisture differences between the ice and the atmosphere. The determination of these turbulent surface fluxes of sensible and latent heat is described in section 2.2.1. The net heat input to the atmosphere over very thin ice is controlled primarily by the rate at which turbulent heat is transferred. Over thick ice, $F_{SH}$, $F_{LH}$, and $F_C$ must supply heat to the surface to compensate for long wave losses. The surface heat balance becomes insensitive to thickness once $h_i$ exceeds 80-100 cm [Maykut, 1986].

2.2.1. Surface Turbulent Fluxes. The surface turbulent fluxes of sensible and latent heat are evaluated using surface renewal theory, following Clayson et al. [1996] and Alam and Curry [1997]. We use a surface flux algorithm that computes these fluxes as a function of the atmospheric and oceanic/ice surface variables by applying the surface renewal theory to the air-sea/air-ice interface. A brief description of the surface flux formulation follows.

Using Monin-Obukhov similarity theory, the turbulent fluxes of sensible heat ($F_{SH}$) and latent heat ($F_{LH}$) are defined as

$$F_{SH} = - \rho_e C_p u_*, T_o$$
$$F_{LH} = - \rho_l L u_*, q_o$$

where $\rho_e$ and $\rho_l$ are the densities of the air and liquid water, respectively, $C_p$ is the specific heat at constant pressure, $L$ is the latent heat of vaporization, and $u_*$ and $q_o$ are the friction velocity and air specific humidity at the surface, respectively.
where $T_*, q_*$, and $u_*$ are the Monin-Obukhov similarity scaling parameters for temperature, water vapor mixing ratio, and horizontal wind, respectively, $\rho_0$ is the density of surface air, $c_p$ is the specific heat capacity of air at constant pressure, and $L$ is the latent heat of vaporization. In the surface layer, the following profiles for velocity, temperature, and humidity have been determined empirically [e.g., Businger, 1973]:

$$
\frac{T_a - T_s}{T_\ast} = \frac{P_r}{k} \left[ \ln \left( \frac{z}{z_0} \right) - \Psi_T \right]
$$

$$
\frac{q_a - q_s}{q_\ast} = \frac{S_c}{k} \left[ \ln \left( \frac{z}{z_0} \right) - \Psi_q \right]
$$

$$
\frac{u_a - u_\ast}{u_\ast} = \frac{1}{k} \left[ \ln \left( \frac{z}{z_0} \right) - \Psi_u \right]
$$

(11)

where $k$ is the von Karman constant (0.4) and $P_r$ and $S_c$ are the turbulent Prandtl number (0.85) and the turbulent Schmidt number (0.85) respectively. The subscript $s$ denotes values at the surface skin so that $T_s$, in this section is the same variable as $T_\ast$ used elsewhere, and the subscript $a$ denotes a value at the atmospheric surface layer at height $z$. The terms $z_0$, $z_{0r}$, and $z_{0q}$ represent the surface roughness lengths for momentum, moisture, and heat, respectively. The terms $\Psi_T$, $\Psi_q$, and $\Psi_u$ represent the respective stability functions, which are nondimensional functions of $z/L$, $L$ being the Obukhov length. The form of the dimensionless stability functions used follows Belskjaars and Holtslag [1991] for stable conditions and Benoit [1977] for unstable conditions.

At the open water portion of the surface, the value of $z_0$ is determined from section 2.2 of Alam and Curry [1991]. The momentum roughness length $z_0$ over water is based on sea state parameterization which is fully consistent with the surface turbulent flux parameterization. The flux parameterization accounts for the fetch limitation of the airflow over a lead [Alam and Curry 1997]. Once any ice forms, over the ice-covered portion of the lead we use a modified surface roughness length $z_0$ to account for the physical characteristics of the new ice. Table 1 shows the specified surface momentum roughness length $z_0$ values as a function of the ice type which depends on the ice concentration, thickness, and whether frazil or congelation ice formation is occurring. It is based upon the data of Guest and Davidson [1991]. In frazil production, the initial ice forms at a concentration of 0.2, and the surface behaves as a damped wave between ice concentration of 0.2 to 0.4. We have used a linear damping of the amplitude of the surface waves by damping the gravity wave contribution in the surface roughness length for momentum in the Alam and Curry [1997] work. The damping factor goes from 1 to 0 as the ice concentration changes from 0.18 to 0.44 [Martin and Kauffmann, 1981]. This linear damping implies that the choice of 20% as the initial grease ice concentration does not affect the lead flux and ice parameters significantly as $z_0$ does not change drastically if we use an initial grease ice concentration different from 0.2. Beyond a concentration of 0.4, we use the surface roughness length for grease ice from Guest and Davidson [1991]. After the concentration reaches unity, the surface roughness length is based on the ice thickness and the ice growth regime, which together determine the ice type, as described in Table 1. In reality, ice does not change roughness suddenly unless there is dynamic deformation, for example, rafting or ridging, so that the transition of $z_0$ with ice thickness at the different ice types is gradual. Therefore we have used a linear combination of $z_0$ as the ice type changes from nilas to young ice, pancake ice to young ice, young ice to first-year ice, and first-year ice to multi-year ice. So instead of $z_0$ changing sharply at a given ice thickness which signifies a change in ice type as shown in Table 1, $z_0$ in the model changes linearly over a 4-cm-thickness range centered at the thickness which defines a change in ice type. This results in a more gradual change in $z_0$ with ice thickness.

Using the appropriate value of $z_0$, the values of $z_{0r}$ and $z_{0q}$ are determined by Clayson et al. [1996] using surface renewal theory to be

$$
z_{0r} = z_0 \exp \left( k \left( \frac{u_\ast}{u_\ast} - \frac{1}{P_r T_\ast} \right) \right)
$$

$$
z_{0q} = z_0 \exp \left( k \left( \frac{u_\ast}{u_\ast} - \frac{1}{S_c D_\ast} \right) \right)
$$

(12)

where $h$ is the depth of the interfacial layer such that $u_\ast = 5u_\ast$. The values of the interfacial Stanton number ($St_\ast$) the interfacial Dalton number ($Da_\ast$) are determined by Clayson et al. [1996] to be

$$
St_\ast = \frac{H_s}{\rho_a c_p u_\ast \left( T_s - T_\ast \right)} = \left( \frac{\kappa}{u_\ast \sigma} \right)^{0.5}
$$

$$
Da_\ast = \frac{H_a}{\rho c_p u_\ast ( \theta_s - \theta_s )} = \left( \frac{\varepsilon}{\sigma u_\ast} \right)^{0.5}
$$

(13)

respectively. where $\kappa = 2.0 \times 10^{-4}$ m$^2$ s$^{-1}$ is the coefficient of thermal diffusivity of air, $\varepsilon = 2.4 \times 10^{-5}$ m$^2$ s$^{-1}$ is the coefficient of molecular diffusivity of water vapor in air, and $t_s$ is the surface renewal timescale. Over open water and up to an ice concentration of 0.4, $t_s$ is described by (5) given by Alam and Curry [1997]. Once the ice concentration exceeds 0.4, the surface renewal time $t_s$ over ice is from Andrews [1987] and is given by

$$
t_s = 31.36 \left( \frac{\varepsilon_0}{u_\ast} \right)^{0.5}
$$

(14)
For open water and ice concentrations less than 0.4, the skin is cooler than the ocean mixed layer which is at the salinity-determined freezing point. An approach to determining this temperature difference based on surface renewal theory has been taken by Wick et al. [1996] that is consistent with the surface renewal timescale used in the present model. The Wick et al. [1996] formulation of the difference between the bulk temperature and the skin temperature at night is given by

\[
\Delta T = \frac{Q_N}{\rho_s c_p \kappa^{0.5}} \left[ c_{\text{heat}} \left( \frac{V e_0}{u^*} \right)^{0.5} + c_{\text{conv}} \left[ \frac{\nu \rho_s c_p}{\alpha g Q_N} \right]^{0.5} \left( \frac{V e_0}{u^*} \right)^{0.5} \exp \left( -\frac{R_{fc}}{\kappa} \right) \right]^{0.5}
\]

(15)

where all of the variables refer to the ocean. \( Q_N \) (equals \( Q \)) is the net heat flux at the surface and is negative for water, \( \rho_s \) is the density of water, \( c_p \) is the specific heat capacity, \( \kappa = 1.4 \times 10^{-7} \text{ m}^2 \text{s}^{-1} \) is the coefficient of thermal diffusivity, \( \nu = 1.0 \times 10^{-6} \text{ m}^2 \text{s}^{-1} \) is the viscosity, and \( \alpha = 3196 \times 10^{17} \text{ K}^{-1} \) is the coefficient of thermal expansion. For the ocean, \( c_{\text{heat}} = 209, c_{\text{conv}} = 3.13, \text{ and } R_{fc} = -2.0 \times 10^5 \) have been empirically determined [Wick et al., 1996]. In (15), \( u^* \) for water is given by \( \left( \rho_s / \rho_{\text{sat}} \right)^{0.3} u_{*w} \). For grid points which have ice concentration greater than 0.4, the skin surface temperature is the weighted temperature of the freezing point of water and the surface ice temperature where the weighting fraction is based on the ice concentration. The methodology for obtaining the surface ice temperature is described in section 2.2.2.

Equations (11)-(15) and the equation for \( e_0 \) give us a closed set of 10 equations for 10 unknowns \( T_s, q_s, u_s, z_{atm}, z_{im}, S_{atm}, D_{atm}, t_s, \) and \( T_s \), which are solved iteratively. Values of \( T_s, q_s, \) and \( u_s \) are then used in (10) to solve for the surface turbulent fluxes of sensible and latent heat. Inputs required in the model are lead width, \( T_s, q_s, u_s, T_s, q_s, \) and \( u_s \). Heat flux is computed for different specified lead widths, 10-m wind speed \( u_s \), and 10-m air temperature \( T_s \). The 10 m specific humidity \( q_s \) is specified to be 0.91 \( q_s(T_s) \), where \( q_s(T_s) \) is the saturation specific humidity with respect to ice for the temperature \( T_s \). Evaluation of the surface variables \( T_s, q_s, \) and \( u_s \) is based on the state of the surface and is shown below. The justification for using a 90\% relative humidity of the saturation with respect to ice for the atmosphere at 10 m is as follows. Oort [1983] tabulated the climatological values of specific humidity \( q_s \) and air temperature \( T_s \) for December at 80°N and 100 kPa atmospheric pressure. We calculated the saturation specific humidity \( q_s \) from the tabulated air temperature \( T_s \), and knowing \( q_s \) determined that \( q_s \) equals 0.91 \( q_s \) for the Oort [1983] data. Hence the relative humidity is 91\% of the saturation with respect to ice based on observed climatology. Another data source that gives \( q_s \) of around 0.91 \( q_s \) is the Russian North Pole Drifting Station Data of Arctic Ocean meteorological observations for 1937 and 1950-1991 (version 1.0) over multiyear ice floes obtained from the National Snow and Ice Data Center [1996], University of Colorado, Boulder. For station 30, which has an average latitude of 80°N, the 2-m specific humidity values for the months of January and February are 0.89 \( q_s \) and 0.91 \( q_s \), respectively. Therefore we have used \( q_s = 0.91 q_s \) for the model runs.

If the surface is water or has frazil ice up to a volume concentration of 0.4, the bulk ocean mixed layer temperature is at the salinity-determined freezing point of seawater. We have used a salinity of 30.8 psu from Holland et al. [1997] based on the observed winter salinity from the ADCIRC drifting station data. This leads to a bulk temperature of -1.7°C. The skin temperature is obtained from this bulk temperature by using (15) which gives the lowering of the bulk temperature at the skin. For a surface which is 100\% ice by volume, the surface ice temperature is determined by the energy balance at the surface and is described in section 2.2.2. For an ice concentration value between 0.4 and unity, the surface temperature is the linearly weighted-by-concentration value between the freezing water temperature and the ice temperature.

The specific humidity at the surface \( q_s \) depends upon the physical state of the surface and its temperature. For a surface with a volume ice concentration up to 0.4, the specific humidity is determined following Fairall et al. [1996]:

\[
q_s = 0.98 q_{\text{sat}}(T_s)
\]

(16)

where \( q_{\text{sat}}(T_s) \) is the saturation specific humidity for the freezing water temperature \( T_s \). This expression accounts for the reduction in vapor pressure associated with a salinity of 30.8 psu. For an ice concentration of unity, the surface specific humidity \( q_s \) is

\[
q_s = q_{s,\text{w}}(T_s)
\]

(17)

where \( q_{s,\text{w}}(T_s) \) is the saturation specific humidity with respect to ice at the surface ice temperature \( T_s \). For an ice concentration between 0.4 and 1, the surface specific humidity \( q_s \) varies linearly between the liquid (16) and ice (17) values.

The surface velocity \( u_s \) is the mean ocean surface current. For a wave-covered surface over water and for an ice concentration up to 0.4, \( u_s \) is parameterized as a second-order Stokes flow [Stokes, 1847]

\[
u_s = \frac{2 n h_s^2}{\lambda} c_p
\]

(18)

where \( h_s \) is the significant wave height and \( \lambda \) is the wavelength of the dominant waves. Values \( h_s \) and \( \lambda \) are computed using section 2.2.1 of Alam and Curry [1997].

For an acro dynamically smooth surface, the surface current \( u_s \) over water and for ice concentrations up to 0.4 is estimated from measurements of the wind-induced current [Wu, 1975]

\[
u_s = 0.55 u_*
\]

(19)

The surface velocity \( u_* \) at an ice surface with the volume ice concentration greater than 0.4 is zero as the surface starts to act like a solid beyond a concentration of 0.4.

2.2.2. Surface temperature. The surface temperature is needed to compute all the terms in the surface energy balance except for the incoming long wave radiation term \( F_{cw} \), which we obtained from observed climatology. The surface ice temperature during winter results from the energy balance at the surface and is dependent upon the ice thickness which affects the various terms in the energy balance. Solution of the surface energy balance equation (9) to determine the surface temperature is implicit since the upwelling long wave flux, the convective flux, and the surface sensible and latent heat fluxes are all a function of surface temperature. Therefore we solve for ice surface temperature using a straightforward interval-halving technique, consisting of successive guesses at the surface temperature and calculation of the temperature-dependent surface flux components.

In the case of frazil formation and growth, for volume ice concentrations up to 0.4 the ocean mixed layer is assumed to be at the freezing temperature of water. The surface skin tempera-
ture is determined from this freezing temperature by using (15). Beyond a concentration of 0.4, the surface temperature equals the weighted temperature between the freezing temperature of seawater and the surface temperature of ice where the weighting fraction is based on the ice concentration. The ice surface temperature for this is calculated as shown above. When the concentration reaches unity, the surface is at the ice temperature, and the surface temperature in the model equals the ice temperature at the surface.

2.3. Interactive Oceanic Flux

When ice forms, the growth rate is very high initially and then decreases with time as the ice grows thicker. The ice formation releases salt into the ocean, modifying the salinity of the surface, and the density increase causes convection under the lead. This affects the ocean dynamics and the ocean heat flux $F_o$ to the bottom of the sea ice. Variations in salinity in the surface layer beneath the ice result in variations in the temperature which is at the salinity-dependent freezing point of water. This gives rise to wintertime values of $F_o$ beneath newly forming sea ice that have been observed to be as large as 30 W m$^{-2}$ [McPhee and Stanton, 1996] and simulated to be as large as 55 W m$^{-2}$ [Holland et al., 1997].

Determination of the turbulent heat flux at the base of the ice (which during wintertime is equal to $F_o$) has been done using a coupled ice/ocean model by Holland et al. [1997], where the parameterization of the turbulent heat flux follows McPhee et al. [1987]. Holland et al. [1997] assume that the mean temperature and salinity are uniform in the upper ocean mixed layer, and thus the equations for these properties can be integrated across the mixed layer depth. The ocean heat flux $F_o$ is obtained from the heat balance at the ice/ocean interface. This balance consists of the turbulent heat exchange between the ice and the ocean ($F_e$), the conduction of heat at the ice base ($F_c$), and the latent heat flux at the interface due to the ice ablation or accretion ($dh/dt$). Holland et al. [1997] use a parameterization of McPhee [1992] which solves for $dh/dt$ at the ice base, and then $F_o$ can be obtained from the heat balance since $F_o$ is known. The Holland et al. [1997] model is integrated for 100 years and is forced at the surface by an annual cycle of the radiation fluxes, a constant surface wind speed, and the 2-m atmospheric temperature and humidity. It reproduces the annual cycle of mixed layer temperature, salinity, and depth of the AIJEX drifting station data. To avoid the complexity here of using an ocean mixed layer model to determine $F_o$, we use results from the Holland et al. model (see Figure 3) for wintertime to develop a parameterization of $F_o$ as a function of the ice accretion rate, $dh/dt$

$$F_o = 6.6462 \times 10^4 \frac{dh}{dt}$$

(20)

where $F_o$ is determined in W m$^{-2}$ and $dh/dt$ is in meters of ice growth per second. A simple linear relationship is therefore used between the new ice growth rate and the turbulent oceanic heat flux. It should be noted that during LEADEX, frost flowers with a salinity of 100 psu formed in the surface skin which resulted from an upward expulsion of brine [Perochvich and Richter-Menge, 1994]. The brine expulsion into the upper mixed layer which accompanies the freezing of ice is a complex process, and the salt is not released instantaneously. This is not accounted for in our ocean flux parameterization.

Since the ice growth rate $dh/dt$ in parameterization (20) is found before we know $F_o$ and is based on using $F_o = 0$ in (2), we can reformulate the above parameterization for the ocean heat flux $F_o$ in terms of the net turbulent and radiative heat flux loss at the surface as follows. Using $F_o = 0$ in the growth rate equation (2) gives us $dh/dt$ in (20) in terms of the conductive heat flux $F_c$ and using (9), we can formulate $F_o$ in terms of the net radiative and turbulent heat lost from the surface, so that parameterization (20) is equivalent to

$$F_o = 6.6462 \times 10^7 \frac{Q'}{\rho_e L_i} = 0.216 Q'$$

(21)

where $Q'$ is the net radiative and turbulent heat lost from the surface which is balanced by the conductive heat flux toward the surface. From the thin ice assumption of linear temperature profile through the ice, this conductive heat flux at the surface equals the conductive heat directed away from the underside of the ice resulting in the freezing, salt release into the ocean and the ocean heat flux $F_o$. In the sensitivity experiments in section 6 where we study the influence of including this ocean heat flux on the new ice growth, the ice growth rate $dh/dt$ is given by (2). The inclusion of ocean heat flux results in a smaller ice growth which affects the ice formation and the turbulent surface fluxes as shown in section 6.

2.4. Model Summary

In the model of the freezing lead, in the frazil cases the lead-average net surface heat loss over the open part of the lead gives us the frazil ice production from the open water (equation (4)). This frazil ice piles up at the downwind side with a pileup depth given by the wind speed and the lead width (equation (5)) with the extent of coverage dependent on the ice production rate (equation (7)). In the congelation case, at the initial time step, the net heat loss at every grid point gives the initial ice formation at every grid point across the surface of the lead (equation (1)). Once ice is formed, surface energy balance (equation (9)) is applied at every icy grid point to get the sur-
face temperature and the conductive heat flux. This gives us ice consolidation for ice concentration less than unity (equation (8)) and ice growth for grid points where the ice concentration is unity (equation (2)). If an interactive ocean flux is being considered as we do in section 6 cases, we compute $P_e$ using (20) for grid points which have ice concentration of unity and then use (2) to get the net ice growth. So this model of freezing leads results in ice concentration, ice thickness, surface temperature, and surface heat fluxes at every grid point across the lead surface as a function of time.

3. Evolution of the Congelation Ice Cover

Our modeled growth of congelation ice is identical to that described by Meykut [1986]. Results of the simulations are shown here for later comparison with the growth of frazil ice.

Figure 4 shows the time evolution of the modeled ice growth rate, ice thickness, surface temperature, and turbulent heat flux for congelation growth under wind speeds of 2 m s$^{-1}$ and 4 m s$^{-1}$ and air temperatures of -20°C and -40°C for a 1-km-wide lead. Figure 4d shows that the heat flux increases with increasing wind speed and increasing air-sea temperature difference. The changes in air temperature dominate the changes in the surface turbulent heat flux, since the wind speeds are restricted here to the small values associated with congelation growth. The heat flux approaches zero more slowly for the larger air-sea temperature difference as it takes a longer time for the surface temperature to reach the air temperature as seen in Figure 4c. Figure 4c shows a very rapid initial decrease in the surface temperature which results from the rapid ice formation initially as seen in Figures 4a and 4b, resulting from the high surface fluxes from the open lead and a lead covered with very thin ice.
4. Evolution of the Frazil Ice Cover

4.1. Fetch Dependence

Figure 5 shows the ice concentration, ice thickness, surface temperature, and surface turbulent heat flux as a function of fetch at different times for a 1-km wide lead with a wind speed of 10 m s\(^{-1}\) and air temperature of -30°C. Figure 5a shows the advance of ice toward the upwind edge of the ice with time and the evolution of the ice concentration. Initially, ice forms at 20% volume concentration, the concentration increasing with time as further heat loss occurs. Once the ice concentration reaches unity, the ice thickness increases upon further heat loss as shown in Figure 5b. After the formation of ice at the pileup depth of 0.22 m, the frazil thickness does not increase until the concentration reaches unity. After 96 hours, the ice thickness reaches 0.33 m at the downwind edge of the lead, with slightly thinner ice toward the upwind edge of the lead.

Figure 5c shows the evolution of the surface temperature. Once ice concentration exceeds 40%, the surface temperature begins to decrease from the freezing water temperature and is the weighted temperature of ice and water. Once the ice concentration reaches unity, the ice begins to thicken, and the surface temperature, which equals the surface temperature of ice for a volume ice concentration of unity, drops further. After 48 hours, the sharp drop in the surface temperature at 324 m occurs because the ice consolidates, and the surface roughness increases to that of young ice which results in an increase in the turbulent heat loss (Figure 5d), so that the surface energy balance gives a lower surface temperature. At the end of 96 hours, the surface temperature has cooled to within 5°C of the surface air temperature.

The time evolution of the state of the surface affects the surface turbulent heat flux, as shown in Figure 5d. Initially, the surface heat flux corresponds to that from an open lead [e.g., Alam and Curry, 1997]. As the ice begins to form and advance
on the surface, the surface flux changes because of the changes in surface roughness and surface temperature. After 1 hour, ice has extended 183 m from the downwind lead edge. Since the concentration of the initial frazil is only 20%, the surface has a damped wave with very little amplitude change, and so the surface flux at 1 hour is quite similar to the initial surface flux over open water. After 6 hours, the first 281 m at the upwind edge of the lead remains ice free, the ice concentration increasing from that point toward the downwind edge of the lead from 20% to 41.9% across the lead surface. The surface flux accordingly is unchanged from the initial surface flux up to 281 m of the upwind edge and has decreased from its initial value beyond that fetch. The heat flux reflects the damped wave surface from 282 m to 860 m of the upwind edge as the concentration reaches 40%, beyond which the surface is frazil with a very low surface roughness. The ice advance and growth continues so that at 12 hours there is open lead up to 52 m from the upwind edge, damped waves on the surface from 53 m to 202 m of the upwind edge, and frazil beyond that to the downwind edge. At 24 hours, the entire surface is covered with frazil at a concentration of 0.4 to 0.77, and the surface temperature and surface sensible heat flux are decreasing but still maintain a fetch dependence across the lead. After 48 hours, the surface heat flux beyond 324 m of the upwind edge shows an increase as the ice concentration reaches unity, and the surface roughness increases from the grease ice value to that of young ice. Beyond 48 hours, the turbulent heat flux decreases steadily with time because of a decrease in surface temperature.

4.2. Time Evolution
The time evolution of the lead-average ice concentration, ice thickness, surface temperature, and the surface sensible and latent heat fluxes under the same environmental conditions as in section 4.1 is shown in Figure 6. Figure 6a shows that the lead average ice concentration increases from 0 to 1 in 54 hours, sharply at first and then more slowly as the sensible...
Figure 7. In the frazil regime, time evolution of the lead-average ice (a) concentration and (b) thickness and surface (c) temperature and (d) turbulent heat flux for a 1-km lead under different atmospheric conditions.

heat decreases. Figure 6b shows that the lead-average ice thickness increases to the pileup depth of 22 cm, then remains constant while the concentration increases, and then increases like congelation ice growth.

We see that the lead-average surface temperature is constant until the downwind edge reaches 40% frazil concentration. A sharp decrease in surface temperature is seen thereafter until the entire lead is covered, beyond which the temperature decreases more slowly as the ice concentration increases to unity at 54 hours. After 54 hours, the lead-average ice temperature decreases very slowly with increase in ice thickness.

The lead-average surface heat fluxes in Figure 6d reflect all of these changes in ice concentration, ice thickness, and surface temperature. We see that the time evolution of sensible and latent heat fluxes shows qualitatively a similar behavior. Since the total turbulent heat loss results in the ice formation and evolution, we will therefore throughout this paper focus on the total turbulent heat flux instead of considering the sensible and latent heat fluxes separately to investigate the evolution of surface characteristics and fluxes during the freezing of a lead. The initial lead-average turbulent heat flux is at the lead-average open lead value. With freezing, the heat flux decreases very slowly initially as increasing portions of the lead have wave damping on the surface. There is a sharp decrease in the turbulent heat flux with the formation of grease ice once the downwind edge and subsequently larger portions of the lead have concentrations above 0.4. This decrease becomes less pronounced after 22 hours, when the entire lead becomes ice-covered and the concentration is increasing with a resultant decrease in the surface temperature. At 43 hours, the downwind edge has reached a concentration of unity and hence has the higher surface roughness of young ice, so the lead-average turbulent heat flux increases slightly. Beyond 54 hours, the entire lead surface has reached a concentration of unity, and the ice thickens and cools with time, decreasing the surface turbulent heat flux.
4.3. Dependence of Surface Properties on Atmospheric Conditions

Figure 7 shows the lead-average ice concentration, ice thickness, surface temperature, and turbulent heat flux evolution with time for different atmospheric conditions in the frazil regime. A very strong influence of wind speed on the ice concentration and thickness is seen, as the wind speed affects the initial depth of the frazil layer (pileup depth) and the ice advances. Figure 7b shows that the pileup depth is much smaller ($h_i = 2.6$ cm) for wind speed of 5 m s$^{-1}$, relative to the greater depth ($h_i = 22$ cm) associated with the 10 m s$^{-1}$ wind speed. Therefore a smaller depth of frazil needs to consolidate for the lower wind speed case in order to reach an ice concentration of unity. Also, for colder air temperatures, the ice concentration reaches unity faster as the sensible heat loss is greater. Once ice concentration has reached unity, the rate of this congelation growth is seen to depend on the air temperature which determines the heat loss and hence the ice growth.

The time evolution of lead-average surface temperature is shown in Figure 7c. We see that as time increases, the surface temperature tends toward the air temperature. For a given air temperature, the surface temperature does not depend on wind speed initially as only the concentration is increasing, but once the ice thickness starts to increase, there is a strong dependence on wind speed with the surface temperature much lower for stronger wind speed than that with the higher ice thickness.

Figure 7d shows that the surface heat flux has initially a very strong wind speed dependence, this dependence decreasing with time. For higher wind speeds, it takes a longer time to increase concentration given the larger initial depth of the frazil layer through which the concentration has to be increased. Hence it takes a longer time for the heat flux to decrease to the frazil value. After about 40 hours, variations in the surface sensible heat fluxes among the four cases are determined primarily by the surface air temperature.

5. Comparison of Frazil and Congelation Growth

Here we compare the ice thickness, surface temperature, and surface heat flux for congelation versus frazil growth for different atmospheric conditions (Figures 8 and 9) and with lead width change (Figure 10). To make a strict comparison, we use a common wind speed in the simulation of both the frazil and congelation regimes, in spite of the fact that the two regimes are separated by wind speed. Such a comparison allows us to assess the magnitude of error made in sea ice models that include only congelation growth independent of the surface wind speed [e.g., Schramm et al., 1997].

5.1. Dependence on Air-Sea Temperature Difference

Figure 8 shows the differences in ice thickness, surface temperature, and surface turbulent heat flux for frazil versus congelation growth for a lead width of 1 km and surface wind speed of 10 m s$^{-1}$. The results are compared for air-sea temperature differences of -20°C and -40°C.

Figure 8a shows that there is a significant difference in the ice thickness between the congelation and frazil cases, with the difference being greater for the warmer temperature. The frazil
Figure 9. Comparison of the time evolution of lead-average (a) ice thickness and surface (b) temperature and (c) turbulent heat flux for frazil and congelation ice formation for a 1-km lead with $T_s = -30^\circ C$ and wind speeds of 5 m s$^{-1}$ and 10 m s$^{-1}$.

Figure 10. Comparison of the time evolution of lead-average (a) ice thickness and surface (b) temperature and (c) turbulent heat flux for frazil and congelation ice formation with $T_s = -30^\circ C$, $u_*$ = 10 m s$^{-1}$, and lead widths of 100 m and 1 km.
case results in more ice formation because the ice which forms is advected and collects downwind with the ice thickness equal to the pileup depth, leaving the upward part of the lead ice free, whereas in the congelation case, the ice covers the entire lead surface and grows vertically starting with zero ice thickness. In the frazil case, the concentration of the newly formed ice rapidly increases to unity throughout the pileup depth of 22 cm (equation 8) and thereafter the ice thickness increases by congelation growth thus resulting in more overall ice production in the frazil regime than in the congelation case for the same ambient condition. The difference in the ice thickness for frazil and congelation cases is greater for the smaller air-sea temperature difference because the congelation growth rate is much smaller for warmer air, whereas the frazil ice deposits at the pileup depth.

Figure 8b shows the surface temperature evolution with time. In the congelation case, the surface temperature decreases rapidly, whereas in the frazil case, the lead average surface temperature remains at freezing water until the downwind edge exceeds an ice concentration of 0.4, after which the surface temperature decreases rapidly as the ice advances across the lead and consolidates; when the ice at the downwind edge starts to thicken, this decrease slows significantly. Toward the end of the model simulation, the lead-average surface temperatures are virtually identical for both ice regimes. The difference in surface temperature for the frazil versus congelation case is very large initially for the -40°C air temperature. However, the temperature difference between the two different ice growth regimes remains significant over a longer period of time for the warmer air temperature, reflecting the different ice thicknesses shown in Figure 8a.

In the frazil case, Figure 8c shows that the surface turbulent heat flux stays close to its initial open water value longer for the warmer air temperature as it takes longer for the ice concentration to go beyond 0.4 with a smaller heat flux for the smaller air-sea temperature difference. The congelation case in Figure 8c shows a rapid decrease in surface heat flux initially as the ice growth rate is very high and surface temperature decreases rapidly. The turbulent heat flux is thus very different for the congelation and frazil cases for the first 3 hours. In spite of relatively large surface temperature differences between the frazil and congelation cases during the period after 3 hours and up to about 40 hours, there is not such a large corresponding difference in the surface turbulent heat flux. The effects of warmer lead-average surface temperature for the frazil case would act to increase the lead-average turbulent heat flux relative to the congelation case. However, the reduced surface roughness associated with grease ice (frazil) relative to that of nilas (congelation) acts to decrease the surface turbulent heat flux. The net effect of the differences in surface temperature and roughness, which have counteracting effects on the heat flux, is that there is relatively little difference between the frazil and congelation cases in surface turbulent heat flux during the initial period when there are relatively large differences in lead-average surface temperature.

5.2. Dependence on Wind Speed

Figure 9 shows the differences in ice thickness, surface temperature, and surface sensible heat flux for frazil versus congelation growth for a lead width of 1 km and a surface air temperature of 30°C. The results are compared for wind speeds of 5 m s⁻¹ and 10 m s⁻¹. The results show that beyond the first few hours, there are virtually no differences between the frazil and congelation regimes for a wind speed of 5 m s⁻¹. For a wind speed of 10 m s⁻¹, however, the differences are significant.

The wind speed dependence of the difference in ice thickness between the frazil and congelation regimes is shown in Figure 9a. The lack of difference in ice thickness between the two regimes at a wind speed of 5 m s⁻¹ is due to the very small pileup depth (2.6 cm) of the frazil layer and the consequent rapid ice consolidation (equation 8), after which growth occurs by congelation growth. For the wind speed of 10 m s⁻¹, however, there is a great difference between the ice thickness obtained by frazil formation and by congelation ice formation. For wind speed of 10 m s⁻¹, the initial thickness of frazil ice is high (22 cm), so the lead freezes over much more slowly than in the small pileup depth 5 m s⁻¹ case as the thicker frazil layer advances more slowly toward the upwind edge of the lead. Also, the high initial frazil depth of 22 cm results in a smaller consolidation rate than in the case of the smaller wind speed case with a much smaller pileup depth (equation 8). This slow ice advance and consolidation results in very different ice thickness evolution for frazil and congelation regimes in the higher wind speed case as shown by Figure 9a.

Figure 9b shows that there is not much difference in surface temperature for a 5 m s⁻¹ wind speed between frazil and congelation cases. However, for a wind speed of 10 m s⁻¹, there is a very large difference between the frazil and the congelation cases, the instantaneous temperature difference reaching as high as 10°C at a time of 16 hours. This is explained by slower rate of ice coverage and consolidation in the frazil regime at a wind speed of 10 m s⁻¹ because of the greater initial ice thickness. Once the frazil is consolidated and enters into congelation growth, the slightly colder surface temperatures reflect the thicker ice for the frazil regime as the surface roughness is the same for both regimes for ice thicker than 10 cm.

Figure 9c compares the turbulent heat flux between the two ice growth regimes for different wind speeds. For a wind speed of 5 m s⁻¹, the initial frazil depth is 2.6 cm, and the lead freezes over very quickly, and the surface turbulent heat flux for both ice growth regimes becomes virtually identical, whereas for the 10 m s⁻¹ wind speed it takes longer for the downwind ice concentration to reach 0.4 and a longer time for the lead to freeze over. There is a smaller change in surface turbulent heat flux when the concentration reaches unity as the ice is colder for the higher ice thickness in this case. Unlike the congelation case, in the frazil case there is a distinct effect of wind speed on the sensible heat especially for higher wind speed. For a wind speed of 5 m s⁻¹, the ice pileup depth for a 1-km lead is 2.6 cm as seen by the very small kink at 4 hours in the lead-average ice thickness in Figure 9a. This is because for such a small ice pileup depth, the lead fills up and consolidates very fast, after which the ice grows by congelation growth. This explains why the congelation and frazil cases are not very different for the 5 m s⁻¹ wind speed especially after 8 hours when consolidation has occurred across the lead and the ice is thickening. For the 10 m s⁻¹ wind, in the frazil case the ice thickness is much higher than the congelation case ice thickness, as the frazil ice piles up downwind at a higher pileup depth for 10 m s⁻¹ wind, and the surface temperature is warmer as it is at the weighted temperature until the ice concentration reaches unity. The
lead-average turbulent heat flux in the frazil case stays close to the open water value for the first 6 hours and then drops to a value lower than the congelation case while the lead is consolidating as the frazil case has the much smaller surface roughness of grease ice. After 43 hours, the turbulent heat flux starts to rise as the consolidated ice has the much higher surface roughness of young ice (Table 1), and after 50 hours the surface heat flux in the frazil case becomes similar to the sensible heat flux in the congelation case as the surface temperature for the two cases becomes almost identical, and the surface roughness lengths for consolidated frazil and congelation ice are the same after the formation of young ice of thickness 10 cm (Table 1). We see that for the higher wind speed, the turbulent heat flux for frazil and congelation cases becomes close after a longer time than for a lower wind speed because of the higher pileup depth associated with the higher wind speed, so that it takes a longer time for the ice to form at the pileup depth and consolidate across the lead even with the higher surface turbulent heat loss for the higher wind speed case.

5.3. Dependence on Lead Width

Figure 10 shows the dependence of ice thickness, surface temperature, and turbulent heat flux on the lead width for frazil and congelation ice formation processes. The results are shown for a wind speed of 10 m s⁻¹, air temperature of -30°C, and lead widths of 100 m and 1 km. Only three curves are shown as there is no fetch dependence in the congelation cases. This is because the ice covers uniformly across the lead and grows vertically in the congelation regime. The frazil case has a significant fetch dependence as the ice forms and is advected downwind and slowly fills up across the lead. Figure 10a shows that for the 100-m lead, the frazil ice is completely consolidated after 13 hours, after which it grows like congelation growth. For a 1-km lead, the ice thickness for the frazil case is
always larger than the congelation case and is much larger until the entire lead is covered with a volume ice concentration of unity.

Figure 10b shows that the air temperature shows a similar behavior. The surface temperature for the frazil case for a 1-km lead remains much warmer than the congelation case because the surface temperature is the weighted temperature until the ice concentration reaches unity after 50 hours on account of the high ice pileup depth. So we see that the difference between congelation and frazil cases increases with lead width which increases the initial frazil pileup depth.

Figure 10c shows that for the smaller lead width, the initial heat flux, which is high because of the fetch-limited flow over leads [Alam and Curry, 1997], rapidly decreases to the frazil value because of the small ice pileup depth (Figure 10n), so it is not very different from the congelation cases initially, but it drops more slowly than the sensible heat flux for congelation regime as the surface temperature is at the weighted temperature of frazil ice and water. Beyond the ice thickness of 10 cm, the sensible heat flux is identical for the 100-m frazil case and the congelation cases as both the ice thickness and surface temperature are identical, and z1 is the same for congelation ice and frazil ice after consolidation for ice thickness greater than 10 cm. The 1-km frazil case drops to a lower sensible heat value than the congelation cases because of the greater ice thickness.

6. Influence of Ocean Heat Flux

In the calculations presented thus far, we have specified a value of the ice/ocean heat flux $F_c = 0$. The complex interactions between a freezing lead and the ocean mixed layer are poorly documented and understood. However, observations show that there can be a large flux of heat from the ocean to the ice in a freezing lead, and results from a coupled ice/ocean model suggest that the ice/ocean heat flux increases with increasing growth rate for new ice formation as shown in section 2.3. To assess the importance of the ice/ocean heat flux in the evolution of the newly formed ice in leads and the associated surface temperature and surface heat fluxes, we use the parameterization for $F_c$ described in section 2.3 interactively in the model for both congelation and frazil growth. Figures 11-13 show the results when the ocean flux to the underside of ice $F_c$ is included during congelation growth, frazil formation, and a comparison of congelation growth and frazil ice growth for the same atmospheric condition, respectively.

Figure 11 shows the effect of ice growth rate dependent nonzero $F_c$ on congelation growth for a 1-km lead, with wind speeds of 2 m s$^{-1}$ and 4 m s$^{-1}$ and air temperatures of -20°C and -40°C. For any given atmospheric condition, it is seen from Figure 11a that the rate of ice growth is slower when the turbulent flux at the ice/ocean interface is considered. This results in less ice formation (Figure 11b), and Figure 11c shows that these smaller ice thicknesses result in warmer surface temperatures when the ocean flux contribution is included in the surface flux model for freezing leads resulting in higher turbulent heat flux as shown in Figure 11d. Figure 11 shows the quantitative effect of including $F_c$ under different atmospheric conditions. It is seen that the atmospheric conditions which lead to the greatest ice formation, namely, high air-sea temperature differences and high wind speeds, show the greatest change in surface characteristics and sensible heat flux when $F_c$ is included. This is because the greater ice formation leads to larger salt rejection and large oceanic fluxes to the ice bottom.

Figure 12 shows the effect of including $F_c$ in the frazil regime for a 1-km lead, with wind speeds of 5 m s$^{-1}$ and 10 m s$^{-1}$ and air temperatures of -20°C and -40°C. The ice concentration behavior for any atmospheric condition is not affected at all by including $F_c$ as shown in Figure 12a. This is because the frazil ice piles up at the downwind edge and consolidates before we consider the ocean turbulent flux from growing ice, which occurs after the ice concentration reaches unity. Figure 12b shows that the downwind concentration reaches unity, the ice thickness is lower for the cases in which $F_c$ is included. The lowering of ice thickness on including $F_c$ increases for increasing air-sea temperature difference as that leads to higher sensible heat flux, more ice formation, more salt rejection, and consequently higher $F_c$. For a given air-sea temperature difference, a higher wind speed results in a smaller ice thickness difference between including and excluding $F_c$ since the ice thickness before congelation growth starts, which is the initial frazil pileup depth, is larger, resulting in a smaller growth rate and hence a lower $F_c$. Figure 12c shows that the surface temperature is warmer after the ice growth begins, when $F_c$ is included. The lower wind speeds show the greatest difference when $F_c$ is included because of the smaller pileup depths, which implies that the growth starts sooner and also implies greater ice growth rates which result in higher $F_c$ when congelation growth starts. Figure 12d shows that when variable $F_c$ is included, the turbulent heat flux is only affected after the congelation growth starts; the inclusion of $F_c$ has a greater effect for larger air-sea temperature difference and for smaller wind speeds for the same reasons as explained above.

Figure 13 shows the effect of including $F_c$ on congelation and frazil ice formation for the same atmospheric condition to make a strict comparison between the two ice formation regimes. The results are shown for a 1-km lead with a 10 m s$^{-1}$ wind speed and an air temperature of 30°C. Figure 13a shows that the ice thickness decreases on including $F_c$ for the congelation regime from the beginning, while for the frazil regime ice thickness changes after the downwind concentration reaches unity and ice growth begins. Figure 13b shows a similar result for the surface temperature with the congelation case showing warmer surface temperatures instantly on including $F_c$, while the frazil case shows warming at a much later time. The surface turbulent heat flux is thus affected a lot more for the congelation case when the oceanic flux is included as seen in Figure 13c.

7. Conclusions

We have presented a model for new ice growth in leads that distinguishes between the congelation and frazil regimes. The frazil ice growth varies with fetch across the lead. The model also includes a sophisticated parameterization of the surface sensible and latent heat fluxes. The lead-average turbulent heat flux in frazil ice formation changes from open water value to damped wave and then decreases to frazil at the surface. It then increases to the heat loss from consolidated ice and then decreases beyond that with time. Frazil and congelation ice have different surface roughness lengths, even for ice of the same thickness, up to an ice thickness of 10 cm.

The frazil ice formation results in a fetch dependence during the evolution of the ice cover. The sensible heat flux decreases with the formation of damped waves, and then as the ice concentration increases, there is a rapid decrease in sensible flux due to the very small roughness length of grease ice. After the
ice concentration reaches unity, the ice thickness dictates the surface roughness, and there is an increase in the sensible heat flux across the lead since the pancake ice has larger surface roughness length. Subsequently, sensible heat flux decreases with time with decrease in surface temperature.

In frazil production, the surface turbulent heat loss is a very strong function of the wind speed initially as the pileup depth depends significantly on the wind speed, and the ice thickness affects the surface roughness and the surface temperature and hence the surface turbulent heat flux. The heat loss increases with increasing wind speed and increasing air-sea temperature difference. The ice concentration increases slowly for large wind speed as the concentration of a thicker pileup depth needs to increase. For a given wind speed the ice concentration increases faster for a larger air-sea temperature difference as that results in larger heat loss. The lead-average ice thickness increases to pileup depth which is larger for large wind speeds, beyond which the ice thickness increases as in congelation growth.

When we compare the congelation growth to the frazil growth for the same atmospheric conditions, we find that in the frazil case, the ice thickness increases much faster to the ice pileup depth value and that the turbulent heat flux differs considerably initially from that in the congelation case. The difference between congelation and frazil results is not a strong function of the air-sea temperature difference but depends strongly on the wind speed and the lead width. The larger the lead width and the wind speed are, the greater is the difference between the frazil and congelation results. This is because the ice pileup depth increases with wind speed and lead width and strongly affects the ice advance rate, the surface temperature, the rate of increase of ice concentration, and the ice growth rate. For a 10 ms⁻¹ wind speed and a 1-km wide lead, there are significant differences in the frazil and congelation ice thick-
ness for the first 20 hours leading to the formation of a lot more ice when frazil formation is considered. Also, the surface turbulent fluxes show significant differences in the two ice formation regimes for the first 10 hours. So a climate model should distinguish between these two regimes of ice formation and should take into account the frazil ice as this distinctly affects the heat balance and especially the mass balance at the surface.

The inclusion of an ice growth rate-dependent oceanic flux does not lead to any qualitative changes in the time evolution of the surface freezing characteristics or the turbulent heat flux. The ice thickness decreases, the surface is warmer, and the turbulent heat flux increases on including the oceanic flux which is directed toward the ice bottom for the freezing Arctic leads. The effect of including $F_o$ is stronger for higher air-sea temperature difference and higher wind speed in the congelation regime, since more ice forms under these conditions resulting in greater salt rejection and higher oceanic heat flux. In the frazil regime, the effect of including $F_o$ is greater for higher air-sea temperature difference but decreases for higher wind speed as that results in higher pileup depths and hence lower ice growth rates when ice begins to grow by congelation growth. Also, the effect of including $F_o$ is greater for congelation growth than for the frazil cases where the ice growth starts after the ice at the pileup depth has consolidated and the ice growth rates for a given pileup depth are not as high as the initial growth rates for the congelation regime.

This model details the characteristics of the surface ice formation and the turbulent flux changes as the lead freezes over. It can be used to study the freezing leads in the Antarctic with the appropriate incoming long wave radiative flux and a different ocean flux parameterization. It should be noted that there is considerable uncertainty in the value of lead transfer coefficients, particularly during the transition from liquid water to solid ice. The model formulation is based on our current limited knowledge of lead processes. The congelation growth part is well documented [Maykut, 1978], but we need observations of pileup depth, rate of ice advance, and the ice concentration and thickness changes with time and fetch to test the frazil growth model and the $z_o$ values used in the model. We also need to determine the wind speed and lead width at which the transfer from congelation to frazil growth occurs. The data which will be obtained using an ultralight aircraft in the Surface Heat Budget of the Arctic Ocean (SHiBA) experiment in 1997-1998 should allow us to eliminate some of the above uncertainties in the theory. To avoid complexity and added uncertainty, this model does not take into account any dynamic leadwidth changes which will affect the turbulent surface fluxes by changing the ice distribution in the lead, for example, by rafting. Here we have only considered winter cases wherein the incoming solar radiation is zero. The results from LEADEX suggest that in the springtime, solar radiation resulted in diurnal cycles of surface fluxes and affected the stability and depth of the boundary layer above the lead [Ruffieux et al., 1995] and the fluxes in the oceanic boundary layer of refreezing leads [McPhee and Stanton, 1996]. Since there is sufficient air-sea temperature difference at a lead surface in spring and fall for turbulent fluxes from freezing leads to be significant, we plan to include the effect of solar radiation on lead refreezing in future. Currently, we have not taken into account the warming of the atmosphere which results from the high turbulent flux release from the surface. This can be done by using a two-dimensional (2-D) atmospheric model [Alam and Curry, 1995], and further

Figure 13. Comparison of the time evolution of lead-average (a) ice thickness and surface (b) temperature and (c) turbulent heat flux for frazil and congelation ice formation with and without oceanic flux for a 1-km lead with $T_s = -30^\circ C$ and $u_o = 10$ m s$^{-1}$. 
work is planned to use this surface flux model over freezing leads as the bottom boundary of the 2-D model to evaluate the changing of the surface fluxes with the evolution of the atmosphere over the lead. The complexity of the heat exchange between ice and ocean is not taken into account in the climate models, so a parameterization of the turbulent flux evolution based on the model described here has significant implications for climate modeling given the large role played by the Arctic in the global climate and climate change.

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