Determination of surface turbulent fluxes over leads in Arctic sea ice

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Abstract. We present a new model to compute turbulent surface heat and momentum fluxes over leads in the Arctic sea ice. The momentum roughness length uses a 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. The surface roughness length for heat is determined from an application of surface renewal theory to the air-sea interface. The modeled fluxes are compared with in situ observations of lead fluxes. We also compare our model results with a bulk flux algorithm which has been commonly used to evaluate surface heat fluxes from leads. We perform sensitivity studies to examine the role of the surface renewal timescale, the importance of the cool skin, and impact of wave age dependence on the momentum roughness length. We have computed integral heat fluxes as a function of lead width/fetch for various atmospheric states to determine the magnitude of heat flux in a mesoscale model grid in which a lead is present.

1. Introduction

Leads are quasi-rectilinear cracks in the sea ice resulting from dynamic motions within the ice pack. Leads can be kilometers to tens of kilometers long and meters to kilometers wide. Even in the central Arctic ice pack in winter, leads cover 1-2% of the area [Wittmann and Schule, 1966; Asselin, 1977; Barry et al., 1993]. Typically, leads tend to remain open for a day or less during winter [Makita, 1991]. The occurrence of open water areas in the sea ice is of major significance for the ocean-atmosphere exchange of heat and moisture, particularly during winter [e.g., Maykut, 1986], when air-sea temperature differences may be 20o-40oC over leads. Leads are thus essential elements in any study of the Arctic Ocean heat budget.

Miyake [1965], Badgley [1966], and Schreifler [1975] conducted some of the early work on turbulent transfer over leads. The AIDJEX (Arctic Ice Dynamics Joint Experiment) Lead Experiment (ALEX) in 1974 was particularly successful in determining the turbulent flux of sensible heat over natural and artificial leads. Field measurements taken during ALEX [Andreas et al. 1979] show that the turbulent heat fluxes from leads can exceed 400 W m-2 for sensible heat and 130 W m-2 for the latent heat flux. Raffieux et al. [1995] measured time series of heat flux over two refreezing leads in the early spring of 1992 in the Beaufort Sea during the Lead Experiment (LEADEX) and found that heat fluxes remain large even for a thin ice cover. Since the annual average sensible heat flux from mature ice to the atmosphere is about 3 W m-2 [e.g., Untersteiner, 1964], leads have an effect which is greater by about 2 orders of magnitude. Andreas et al. [1979] obtained empirical formulas describing heat flux variance with fetch over a lead, and Smith et al. [1983] computed heat transfer coefficients over larger polynyas for which the heat flux was not limited by fetch. Andreas and Murphy [1986] and Andreas [1988] combined the ALEX eddy flux data with polynya data from Smith et al. [1983] to recommend a fetch-dependent bulk transfer coefficient for neutral stability conditions. Maslanik and Key [1995] used the Andreas and Murphy [1986] empirically determined fetch-dependent bulk transfer coefficient to investigate the sensitivity of the turbulent heat fluxes over leads to changes in fetch, wind speed, air temperature, and surface temperature in the Arctic.

It has been difficult to model or parameterize the fluxes from the leads because of the strong horizontal inhomogeneities caused by the presence of the lead. Inhomogeneous surfaces cause special problems because it is difficult to relate the point measurements used to characterize the surface to the larger-scale mean fluxes [e.g., Schuep et al., 1990]. In spite of the complications introduced by fetch in determining the surface turbulent fluxes from leads, simulations of lead-induced atmospheric circulations [e.g., Glendenning and Burk, 1992; Glendenning, 1995; Alam and Curry, 1995] have used simple surface flux parameterizations that do not include the fetch dependence.

In this paper we develop a new bulk model to determine surface turbulent fluxes over leads. This model is a physically based scheme which uses surface renewal theory along with a momentum roughness parameterization that includes a fetch-dependent sea state determination that is consistent with the atmospheric state. The effects of the surface “cool skin” are included. At the upwind edge of the lead there is both forced and buoyant convection. The modeled fluxes decrease as the air is modified by passing over the relatively warm water, until the forced convection dominates and the heat flux does not decrease any further with increasing fetch. We compare the modeled fluxes with the AIDJEX observations and bulk flux models that have been used previously to evaluate fluxes from leads. We perform sensitivity studies to examine the role of the surface renewal timescale, the importance of the cool skin, and impact of wave age dependence on the momentum roughness length. Lead integral fluxes are evaluated for different temperature and wind conditions and lead widths.

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2. Model Description

We use a surface flux algorithm that computes the momentum and heat fluxes as a function of the atmospheric and oceanic variables by applying surface renewal theory to the air-sea interface. The surface fluxes are defined in terms of the surface scales of velocity, temperature, and moisture (\( u_s, T_s, \) and \( q_s \)) respectively, which appear in the Monin-Obukhov similarity theory as functions of the surface roughness length and atmospheric profiles. These equations can then be solved iteratively once we know the surface roughness lengths for momentum and heat. The self-consistent surface roughness model modified to include fetch dependence is used to obtain the momentum roughness length, and surface renewal theory is applied thereafter to obtain the roughness length for temperature.

2.1. Turbulence Flux Model

The turbulence flux model used in this study is described by Clayson et al. [1996]; only a brief summary of the model is given here. This model utilizes both Monin-Obukhov similarity theory and the surface renewal theory as described by Brutsaert [1975], Liu et al. [1979] (hereinafter referred to as LKB) made partial use of surface renewal theory, and further improvements to the LKB parameterization have been made by Fairall et al. [1996]. The Clayson et al. model includes the following improvements relative to the LKB and Fairall et al. models: incorporation of a new timescale parameterization for surface renewal, inclusion of a more sophisticated surface roughness model, and derivation of the surface roughness scales of water vapor and heat based solely upon surface renewal theory.

It would appear that the assumption of horizontal homogeneity violated in the Monin-Obukhov similarity theory is violated under the conditions typical of leads, which include strong horizontal gradients in the surface roughness, the surface temperature, and the temperature of the atmospheric surface layer. Over a horizontally homogeneous surface, Siegel [1996] has shown that surface fluxes can be determined with sufficient accuracy by the Monin-Obukhov similarity theory using local velocities and temperatures, provided that the horizontal grid spacing exceeds several tens of meters. The precise lower limit to the resolution over which the Monin-Obukhov similarity theory can be applied remains undetermined. In view of this uncertainty, we will attempt to examine the horizontal variability of surface fluxes over a horizontal resolution of less than 10 m to resolve the surface fluxes for narrow leads and near the edge of leads. A fundamental assumption in this study is that the surface inhomogeneities associated with horizontal variations in surface roughness length dominate the other inhomogeneities and that Monin-Obukhov similarity theory can be applied over a locally varying surface roughness length.

Using Monin Obukhov similarity theory, the turbulent fluxes of momentum (\( \tau \)) and sensible heat (\( H \)) are defined as

\[
\tau = -\rho_s u^2_s \\
H = -\rho_s c_p u^2_s T_s
\]

(1)

where \( T_s \) and \( u_s \) are the Monin-Obukhov similarity scaling parameters for temperature and horizontal wind and \( \rho_s \) is the density of surface air. In the surface layer, the following profiles for velocity and temperature have been determined empirically:

\[
\frac{T_s - T_e}{T_e} = \frac{Pr_t}{k} \left[ \ln \left( \frac{z}{z_{*e}} \right) - \Psi_T \right]
\]

(2)

\[
\frac{u_s - u_e}{u_e} = \frac{1}{k} \ln \left( \frac{z}{z_{*u}} - \Psi_u \right)
\]

(3)

where \( k \) is the von Karman constant (= 0.4) and \( Pr_t \) is the turbulent Prandtl number (= 0.85). The subscript \( s \) denotes values at the surface skin, and the subscript \( e \) denotes a value at the atmospheric surface layer at height \( z \). The terms \( z_{*e} \) and \( z_{*e} \) represent the surface roughness lengths for momentum and heat, respectively. The terms \( \Psi_u \) and \( \Psi_T \) 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 Beljaars and Holtslag [1991] for stable conditions and Benoit [1977] for unstable conditions.

Using the value of \( z_{*e} \) determined in section 2.2, the value of \( z_{*u} \) is determined by Clayson et al. [1996] using surface renewal theory to be

\[
z_{*e} = z_{*u} \exp \left( k \frac{u_e - u_s}{Pr_t S_{*u}} \right)
\]

(4)

where \( h \) is the depth of the interface layer such that \( u_e = 5u_s \). The value of the interface Stanton number \( (S_{*u}) \) is determined by Clayson et al. [1996] to be

\[
S_{*u} = \frac{H}{\rho \ c_p \ u_s \ (T_e - T_s) = \left( \frac{\kappa}{u^2_s T_s} \right)^{0.5}}
\]

(5)

where \( \kappa = 2 \times 10^5 \text{ m}^2\text{s}^{-1} \) is the coefficient of thermal diffusivity.

The variable \( t_s \) is the surface renewal timescale described by Wick et al. [1996] to include both shear and convective effects

\[
t_s = t_{*\text{shear}} \left( t_{*\text{conv}} - t_{*\text{shear}} \right) \exp \left( -\frac{\text{RF}_s}{\text{RF}_c} \right)
\]

(6)

where \( t_{*\text{shear}} \) is the shear-driven timescale, \( t_{*\text{conv}} \) is the convective timescale, \( \text{RF}_s \) is a critical value of the surface Richardson number, and \( \text{RF}_c \) is the surface Richardson number [Soloviev and Schluesssel, 1994] as given by

\[
\text{RF}_s = \frac{\alpha g Q_n \nu}{\rho_s c_p u^3_s}
\]

(7)

The shear-driven timescale is given by

\[
t_{*\text{shear}} = c_{\text{shear}} \left( \frac{\nu^3}{u^2_s} \right)^{0.5}
\]

(8)

and the convective timescale is given by

\[
t_{*\text{conv}} = c_{\text{conv}} \left( \frac{\nu^3}{\alpha Q_n} \right)^{0.5}
\]

(9)

where \( c_{\text{shear}} \) and \( c_{\text{conv}} \) are empirically determined constants described below, \( \nu = 1.5 \times 10^3 \text{ m}^2\text{s}^{-1} \) is the viscosity, \( \rho_s \) is the density of the atmosphere, \( c_p \) is the specific heat capacity, and \( \alpha = -1/(T_s + 273.16) \text{ K}^2 \) is the coefficient of thermal expansion. The variable \( Q_n \) represents the sum of the latent and sensible heat fluxes plus the radiative fluxes and is the net heat flux
which is positive for the atmosphere (for details see Wick et al. 1996). It is noted that the solar insolation is zero in the Arctic winter, which is the scenario for the simulations in the present paper.

The fetch dependence of the lead fluxes has an implication for the convective timescale. As the air travels over the warm surface of the lead, the surface air temperature rises so that with increase of fetch, the air surface air temperature approaches the water temperature and the convective timescale increases as the interfacial layers have more similar temperatures. So the following formulation is used for the constant of proportionality in the convective timescale

$$c_{\text{conv}} = 11X^{0.8}$$  \(\text{(9)}\)

where X is the nondimensional fetch [e.g., Andreas and Murphy, 1986]

$$X = Xg / u^2_s$$  \(\text{(10)}\)

X is the fetch across which the air has traveled, and g is the acceleration due to gravity. The fetch dependence of \(c_{\text{conv}}\) in equation (9) (exponent = 0.8) is the same as the fetch dependence of the internal boundary layer height resulting from a step change in surface temperature [Elliott 1958]. This is because the fetch dependence of the convective term arises from the warming of the surface air and the formation of the internal boundary layer as the air flows over the lead. The constant of proportionality, 11, in equation (9) is determined empirically from comparison with the ALEX data. Beyond a nondimensional fetch of 37.54, \(c_{\text{conv}}\) is considered constant at 200. The constant of proportionality in equation (7), \(c_{\text{struc}}\), is defined [Soloviev and Schlüssel, 1994] as follows:

$$c_{\text{struc}} = \left( \frac{c_{\text{conv}}^{0.5}}{\text{abs}(RF_c)} \right)$$  \(\text{(11)}\)

where the critical surface Richardson number is unity for the atmosphere.

The variable \(T_s\) in (2) refers to the "skin" temperature. Although the skin temperature is measured from satellite, skin temperatures are rarely available from ship or buoy measurements, and may not be available from an ocean mixed layer model. Thus corrections to the modeled or observed bulk temperature to obtain a surface skin temperature measurement need to be performed. 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

$$\Delta T = \frac{Q_n}{\rho_s c_p k^{0.5} u^2_s}$$  \(\text{(12)}\)

where the various terms are defined as follows:

$$c_{\text{conv}} = \left( \frac{V_s}{u^2_s} \right)^{0.5} + \left( c_{\text{conv}} \frac{\psi D}{\rho_c u^2_s} \right) - c_{\text{struc}} \left( \frac{V_s}{u^2_s} \right)^{0.5} \exp \left( \frac{RF_c}{RF_c} \right)$$

where all of the variables refer to the ocean. \(Q_n\) 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, \(K = 1.4 \times 10^{-7}\) m$^2$ s$^{-1}$ is the coefficient of thermal diffusivity, \(v = 1.0 \times 10^{-6}\) m$^2$ s$^{-1}$ is the viscosity, and \(\alpha = 3196 \times 10^{-5}\) K$^{-1}$ is the coefficient of thermal expansion. For the ocean, \(c_{\text{conv}} = 209, c_{\text{struc}} = 3.13\) and \(RF_c = -2.9 \times 10^4\) have been empirically determined [Wick et al., 1996].

In equation (12), \(u_s\) for water is given by \(\frac{\rho_s}{\rho_{\text{water}}} u^*\). The difference between the mixed layer temperature and surface skin temperature (surface radiation temperature) varies with wind speed and the surface heat flux but can reach 1°C under conditions examined in this study, with the skin (thickness of the order of millimeters) cooler than the ocean mixed layer.

2.2. Surface Roughness Model

The surface roughness length is a function of the state of the sea surface. The surface is smooth at zero wind speed. As the wind speed increases beyond the minimum required to generate waves, the surface roughness with the formation of small scale capillary waves (wavelength of the order of centimeters). With further increase in wind speed, gravity waves form, with wavelength increasing with wind speed. In the presence of effectively infinite fetch, the sea surface eventually comes into equilibrium with the wind field. Limited fetch results in growing waves, whereby the wave structure on the water surface changes as the distance from the upwind edge of the lead increases. The waves closest to the upwind edge of the lead are capillary waves, changing over to gravity waves as the fetch increases.

The surface roughness model used here is based on the model described by Bourassa et al. [1997] (hereinafter referred to as BVW). The model includes a nonarbitrary wave age which depends on the shape and size of oceanic waves. Capillary and gravity waves are included in the model. There is a criterion for capillary cutoff below which the sea surface is smooth. To account for the fetch-dependence of surface fluxes wherein the waves are not in a state of wind-wave equilibrium, we have incorporated a sea in a state of fetch-dependent growth, as the fetch limited growth does not allow the waves to come into equilibrium with the wind.

The roughness length parameterization of BVW is given by

$$z_0 = \beta \frac{0.11 V}{u_s} + \frac{\beta b a}{u^2_s} + \frac{\beta}{w_g} \frac{0.48 u^2_s}{g}$$  \(\text{(13)}\)

where \(z_0\) is the surface roughness length; \(u_s\) is the friction velocity; \(V\) is the viscosity of air; \(b\) is an empirically derived constant equal to 0.019; \(g = \sigma / \rho_c\), where \(\sigma = 0.075\) J m$^{-3}$ is the surface tension of water and \(\rho_c\) is the density of water; \(g\) is the gravitational acceleration; and \(w_g = c / u_s\) is the wave age of the sea water, where \(c\) is the phase speed of the dominant waves.

The first term on the right-hand side of (13) gives the contribution to the roughness length from the aerodynamically smooth surface, based on Kondo [1975]. The second term on the right-hand side of (13) gives the contribution to the momentum roughness length associated with capillary waves, and the last term on the right-hand side of (13) is the roughness length due to gravity waves and is given by Charnock's relation,

$$z_{\text{rg}} = c u_s^2 / g$$  \(\text{(14)}\)

where \(z_{\text{rg}}\) is the contribution of the gravity waves to the surface roughness and \(c\) is the Charnock's constant, whose dependence on the sea state is given by [Smith et al., 1992]

$$c = 0.48 / w_g$$  \(\text{(15)}\)
The coefficients $\beta_1$, $\beta_2$, and $\beta_3$ are the weighting variables for smooth, capillary, and gravity waves on the ocean surface. A smooth sea surface implies $\beta_2=1$, $\beta_1=0$, and $\beta_3=0$ so that the roughness length equals that for a smooth surface (only the first term in (13) is nonzero). This occurs for a phase speed $c < c_{\text{min}}$ the minimum phase speed required for wave generation following BVV. Once waves form, BVV specifies $\beta_3=0$, $\beta_1=1$, and $\beta_2=1$, which assumes that the surface is completely covered with both capillary and gravity waves once the phase speed exceeds the minimum phase speed. The last two terms dominate the roughness of the sea surface once waves form, so that the first term can be ignored for rough sea surfaces by using $\beta_3=0$. For a smooth ocean surface, as the wind speed increases, the phase speed of dominant waves increases beyond the minimum required for waves to form. and capillary and gravity waves are generated. This can be achieved by using the following dispersion relationship for $\beta_1$, $\beta_2$, and $\beta_3$:

$$\beta_2 = 0, \beta_3 = 0, \beta_1 = 1 \quad c < c_{\text{min}}$$

$$\beta_2 = 1, \beta_3 = 1, \beta_1 = 0 \quad c \geq c_{\text{min}}$$

(16)

The capillary cutoff phase speed, i.e., $c_{\text{min}}$, determines whether waves exist on the ocean surface and hence the respective values of $\beta_1, \beta_2,$ and $\beta_3$. The corresponding wavelength is determined by using the following dispersion relationship for gravity and capillary waves

$$\frac{g}{2\pi} \lambda^2 - c^2 \lambda + 2\pi \gamma = 0$$

(17)

where $\lambda$ is the wavelength of the dominant waves and $\gamma = \sigma / \rho_s$, where $\sigma$ is the surface tension of water and $\rho_s$ is the density. The solution of the quadratic equation (17) gives the wavelength as

$$\lambda = \frac{4\pi}{g \left( c^2 - 4 \right)^{0.5}}$$

(18)

where $c_{\text{min}} = (4\pi)^{0.5} / g$ is the minimum phase speed of gravity waves and also defines the phase speed for the capillary cutoff limit for the formation of waves on the ocean surface. It may be noted that the above wavelength reduces to the wave length for gravity waves $\lambda = 2\pi c^2 / g$, for the gravity wave limit, i.e., when $c^2 > c_{\text{min}}$.

Equation (13) for $\zeta_e$ requires a value of the wave age $w_\omega$, which is determined as a function of wind speed and the sea state using the following formulation from BVV. Significant wave height is the average height of the highest one third of the waves. The mean wind speed at the significant wave height, $h_s$, is composed of an equilibrium (EQ) component (which is the speed needed for local wind wave equilibrium) and an additional nonequilibrium (NE) component as follows:

$$u(h_s) = u_{\text{EQ}}(h_s) + u_{\text{NE}}(h_s)$$

(19)

Subtracting the mean surface current $u_\zeta$ and dividing by the phase speed $c$ of dominant waves results in the following equation:

$$\frac{u(h_s) - u_\zeta}{c} = \frac{u_{\text{EQ}}(h_s) - u_\zeta}{c} \left[ 1 + \frac{u_{\text{NE}}(h_s)}{u_{\text{EQ}}(h_s) - u_\zeta} \right]$$

(20)

The first term on the right-hand side of (20) is the wind-wave equilibrium term, $G = 0.81$, and is constant to first order [Kitagorodskii, 1970]. The second term on the right-hand side of (20) is the wind-wave stability term, $W$. A value of $W > 1$ implies that $u(h) > u_{\text{EQ}}(h)$ and hence a wave growth condition exists. Using the above definitions for $G$ and $W$, (20) can be written as

$$\frac{u(h_s) - u_\zeta}{c} = GW$$

(21)

Since the fetch limitation implies that local wave equilibrium is not reached, the waves are in a state of growth for the case of leads. The smaller the fetch, the farther the sea surface is from coming to equilibrium with the imposed atmospheric wind. Also, the larger the wind speed, the more difficult it is for the sea surface to come to equilibrium with the atmosphere. Hence, $W - 1$, which is proportional to wave growth, has an inverse dependence on fetch and a direct dependence on the wind speed. Since $W$ is nondimensional, we adopt the following relationship between $W$ and the variables which influence it, where the coefficients are determined as "tuning" parameters in a comparison between the modeled and observed values of $u_\zeta$:

$$\Delta T = W - 1 = \exp \left\{ -0.0002 \left( X_\omega / u_\zeta \right)^2 \right\} = \exp \left( -0.0002X_\omega^2 \right)$$

(22)

Combining (21) with the logarithmic wind profile (2) yields the following equation for wave age

$$w_\omega = \frac{\ln \left( h / h_s \right) - \psi(\omega)}{kGW}$$

(23)

where $\psi(\omega)$ is the dimensionless stability function for momentum at $h / h_s$, $L$ being the Obukhov length. According to (23), knowing the significant wave height is a requirement for determining $w_\omega$. Once $w_\omega$ is known, the phase speed $c$ is determined as $w_\omega c_\omega$ from the earlier definition of $w_\omega$ as $c \zeta_e$.

BVV used the Toba et al. [1990] formula for $h_s$

$$h_s = B \left( \frac{X_\omega \cdot T}{10^5} \right)^{0.5}$$

(24)

where $B = 0.062$ and $T = \lambda/c$ is the time period of the dominant waves. This is an empirical formulation for growing waves whose heights are limited by fetch considerations. The fetch limitation implies that the significant wave height is lower than it would be for infinite fetch where the ocean surface is able to realize its equilibrium wave height. The significant wave heights obtained in the present model using the above relation (22) for $W$ are similar to those observed by Toba et al. [1990] for varying fetches and wind speeds. We have applied the breaking wave criterion to the significant height, which means that $h_s$ does not exceed $\lambda / c$.

The equations for $\beta_1$, $\beta_2$, and $\beta_3$ (equation (16)); $\lambda$ (equation (18)); $W$ (equation (22)); $w_\omega$ (equation (23)); and $h_s$ (equation (24)) can be solved iteratively. If the values of $w_\omega$ and $\beta_1$, $\beta_2$, and $\beta_3$ are determined, which means that the variables defining the sea state to the atmosphere are known, then (13) can be used to calculate a value of $c_\omega$. The details of this iterative process are as follows. Initially, a value of wave age is assumed which gives the phase speed from $w_\omega c_\omega$ and hence the state of the
3. Comparison With Observations

In this section we compare the model results with the 1974 ALEX observations for the small lead withs (6.8 m to 34 m) and the Dundas Island Polynya measurements for larger fetches of several hundred meters [Smith et al., 1983]. In situ observations of fetch dependent surface fluxes from leads have been tabulated by Andreas et al. [1979] and Andreas and Paulson [1979]. A profile tower and a flux tower were set up to sample velocity, temperature, and humidity profiles and fluxes of momentum and heat. The velocity sensor was a hot-film probe and the temperature sensors were thermocouples. The velocity profiles were corrected for sampling error. Two crossed hot-film velocity sensors were used to obtain the momentum flux at two levels [Andreas and Paulson, 1979]. The high-frequency response of the thermocouples was inadequate to directly measure heat flux, so the principal estimate of the surface heat flux comes from the profiles. ALEX sampled upwind and downwind of natural and artificial leads. By pairwise upwind and the respective downwind profiles, they could obtain an estimate of the turbulent heat flux at the surface of the lead on the basis of energy conservation. Andreas et al. [1979] tabulated 57 cases of integral heat fluxes (overall heat flux averaged over the entire lead) and the 2-m wind speed and air temperature over leads.

Andreas and Paulson [1979] derived measurements of integral statistics of the momentum flux. The results for surface momentum flux were presented for 39 cases with the observation height ranging from 10 cm to 50 cm. The momentum flux was observed directly and not inferred in this case. The observations from ALEX have fetches ranging from 6.8 to 34 m. Wind speeds at 2 m ranged from 1.06 to 5.92 m s⁻¹. Air temperatures ranged from -18.7°C to -32.8°C.

Andreas and Murphy [1986] proposed a bulk transfer coefficient method to calculate the momentum and heat fluxes over a lead. The fluxes are formulated in terms of bulk transfer coefficients as follows

\[ \tau = - \rho C_D \frac{k^2}{\ln (r/L)} \]  
(25)

\[ H = - \rho C_H \frac{k^2}{\ln (r/L)} \]  
(26)

where \( r \) is a reference height. Using the definition of surface scales (equation (1)) and the similarity relations in the surface layer (equations (2)), these bulk transfer coefficients can be written as

\[ C_D^n = \frac{k^2}{\ln\left(\frac{r}{L}\right)} \]  
(27)

\[ C_H^n = \frac{k^2}{\ln\left(\frac{r}{L}\right)} \]  
(28)

At neutral stability the stability functions are zero so that the above equations reduce to

\[ C_{DN} = \frac{k^2}{\ln\left(\frac{r}{L_0}\right)^2} \]  
(29)

\[ C_{HN} = \frac{k^2}{\ln\left(\frac{r}{L_0}\right)^2} \]  
(30)

such that knowing the roughness lengths for heat and momentum in the Monin-Obukhov similarity laws is equivalent to knowing the neutral transfer coefficients at any height. These neutral transfer coefficients are then specified empirically. The stability functions are related to the Obukhov length, which is inferred in terms of the bulk Richardson number. It is noted here that the Andreas and Murphy [1986] parameterization used the ALEX data for tuning.

Comparisons of our model and the bulk parameterization of Andreas and Murphy with the observations obtained during ALEX are shown in Figures 1 and 2, and the comparisons of the polynya observations with the two models are shown in Figures 3 and 4. Andreas et al. [1979] tabulated several heat flux results, some of which are inferred from the observed data assuming Monin-Obukhov similarity theory. We have used their integral heat flux data for comparison (Figure 1) because this is the fundamental observed estimate of the surface heat flux based on applying energy conservation to the upwind and downwind profiles. We do not use the 52, 68 and 85 m fetches in the comparison, as there are no integral heat fluxes reported for these cases. For the \( \zeta \), comparison (Figure 2) we used the lower, \( \left( \frac{z \omega}{u^*} \right)^{0.5} \), which is close to the friction velocity since all these measurements are close to the surface (heights of 9 to 50 cm), which came from eddy correlation measurements with hot-film anemometers. Table 1 summarizes the bias, slope, correlation coefficient and the root-mean-square error (RMSE) of the two models when compared with the observations. For a perfect fit to observations, the bias would be zero, the slope of the best fit regression line would be 1, as would be the correlation coefficient; and the RMSE would be zero. A slope significantly far from 1 and a bias very different from zero indicate that the trend of the model results is quite different from the trend of the observations. The correlation coefficient is a measure of the scatter about the best fit regression line and hence a measure of the significance of the regression line. The RMSE is a good overall measure of the difference between the observations and the model results.

In Figure 1 we compare the upwind and downwind profile derived integral values of heat flux [Andreas et al., 1979] with the results from our model and with the bulk parameterization of Andreas and Murphy [1986]. When compared with the measured heat flux (Figure 1), the present model and the Andreas and Murphy model results are clustered around the 1:1 line which represents the observations. For the lower values of the heat fluxes, both models underestimate the heat flux. This is because most of the observations (except the three cases for 34 m fetch which are denoted by open circles and triangles) are for artificial leads. These were semicircular perimeters laid out on the sea ice of Elson lagoon. A hole was augured in one corner of the perimeter for water to drain out and water was pumped in at the other corner to fill the lead. The pumping was providing its own roughness so that the roughness of surface is more than just the wind acting upon it. Since the models take account of only the surface wind, they underestimate the roughness, and consequently \( z_m \) and the heat flux as seen in Figure 1. Another interesting feature is seen when we consider the natural lead.
Table 1. Bias, Slope of the Best Fit Regression Line, Correlation Coefficient and the Root-Mean-Square Error (RMSE) of the Two Models with the Observations

<table>
<thead>
<tr>
<th></th>
<th>Heat Flux W m⁻²</th>
<th>u* m s⁻¹</th>
<th>-ρu'w' N m⁻³</th>
<th>Heat Flux W m⁻²</th>
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<td></td>
<td>[Andreas et al., 1979]</td>
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<tr>
<td><strong>Bias</strong></td>
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<tr>
<td>Alam and Curry</td>
<td>-39.1</td>
<td>-0.019</td>
<td>-0.007</td>
<td>20.4</td>
</tr>
<tr>
<td>Andreas and Murphy</td>
<td>8.2</td>
<td>0.029</td>
<td>0.009</td>
<td>21.1</td>
</tr>
<tr>
<td><strong>Slope</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Alam and Curry</td>
<td>0.96</td>
<td>0.96</td>
<td>1.03</td>
<td>0.94</td>
</tr>
<tr>
<td>Andreas and Murphy</td>
<td>0.92</td>
<td>0.67</td>
<td>0.81</td>
<td>0.86</td>
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<tr>
<td><strong>Correlation Coefficient</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
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<td>0.92</td>
<td>0.92</td>
<td>0.95</td>
</tr>
<tr>
<td>Andreas and Murphy</td>
<td>0.86</td>
<td>0.92</td>
<td>0.93</td>
<td>0.94</td>
</tr>
<tr>
<td><strong>RMSE</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
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<td>77.9</td>
<td>0.036</td>
<td>0.017</td>
<td>33.4</td>
</tr>
<tr>
<td>Andreas and Murphy</td>
<td>53</td>
<td>0.034</td>
<td>0.015</td>
<td>37.1</td>
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</table>

cases. Two of them are very close to the 1:1 line, but one of them is much higher than the observations. This could be because the natural lead was freezing, which would lead to the lower observed surface heat flux. Table 1 shows the results of this comparison quantitatively. The present model’s best fit regression line has a bias of -39.1 W m⁻² and a slope of 0.96 while the Andreas and Murphy model has a bias of 8.2 W m⁻² and a slope of 0.92. The present model has a correlation coefficient of 0.83, while the Andreas and Murphy model has a correlation coefficient of 0.86. The RMSE for the present model is 77.9 W m⁻² while it is 53 W m⁻² for the Andreas and Murphy model. So we see that the Andreas and Murphy model is closest to the observations. This is not surprising, as this model was empirically based on the results of the AFX data. However, given problems with the data, particularly for the artificial leads, it remains unclear which model performs best.

Figure 2 compares the friction velocity $u*$ (square root of momentum covariance) obtained from observations [Andreas and Paulson, 1979] with the corresponding results for the two models. When compared with the observations, our model has a RMSE of 0.036 m s⁻¹ while the Andreas and Murphy parameterization has a RMSE of 0.034 m s⁻¹. Both models have a correlation coefficient of 0.92. The present model has a low bias of -0.019 m s⁻¹ and a slope of the regression line of 0.96 which is very close to 1. The Andreas and Murphy parameterization has a higher γ - intercept of 0.029 m s⁻¹ and also has a slope that is significantly less than 1 (0.67). There were five natural leads in this data set, which are shown by open circles and triangles. Four of these appear to be close to the observations line, while the case which is far removed from the observations corresponds to the case in Figure 1 which appeared to be freezing. The rest of the observations are for artificial leads, which are beset by the problems of artificial surface roughness from pumping water into these artificially formed and maintained

![Figure 1](image1.png)  
**Figure 1.** Comparison of modeled heat fluxes with observations [Andreas et al., 1979] and the Andreas and Murphy [1986] model.

![Figure 2](image2.png)  
**Figure 2.** Comparison of $u*$ with observations [Andreas and Paulson, 1979] and the Andreas and Murphy [1986] model.
leads (as described earlier). We see a considerable underestimate of the $u^+$ values, as the model is considering only the wind-induced surface roughness.

For the larger fetches we have compared the results for the two models with the Smith et al. [1983] polynya observations (Figures 3 and 4) wherein a sonic anemometer-thermometer measured wind and temperature fluctuations at a height of 4.4 m at the downwind edge. Profiles of wind, temperature, and temperature fluctuations were also measured. In Figure 3 we compare the downward momentum flux for the Danila Island polynya data as presented by Smith et al. [1983], with the two surface flux algorithms. The present model has a bias of -0.007 N m$^{-2}$, while the Andreas and Murphy model has a bias of 0.009 N m$^{-2}$. The present model has a slope of the best fit regression line of 1.03, while the Andreas and Murphy parameterization has a slope of 0.81. Both models have comparable low scatter as evidenced by the equivalent high correlation coefficients. The RMSE is 0.017 N m$^{-2}$ for the present model and 0.015 N m$^{-2}$ for the Andreas and Murphy model.

Figure 4 shows the results of the polynya heat flux comparison. The present model gives a better correlation with the observations as shown in Table 1. The bias is 20.4 W m$^{-2}$, while it is 21.1 W m$^{-2}$ for the Andreas and Murphy model. The slope of the best fit regression line is 0.94 for the present model and 0.86 for the Andreas and Murphy model. The present model has a slightly higher correlation coefficient and a lower RMSE than the Andreas and Murphy model. The comparison with the heat flux observations from the polynya indicates that both the present model and the Andreas and Murphy parameterization give reasonable results. However, the Andreas and Murphy model is an empirical one based on observed neutral transfer coefficients, whereas the present model is based on taking into account the physical processes occurring at the surface of the lead.

To examine further the differences between the observations and the modeled fluxes, we study the difference in heat flux and friction velocity as a function of fetch, wind speed, and air-sea temperature difference when compared with the Andreas et al. [1979] observations of the integral heat flux and the Andreas and Paulson [1979] observations of the friction velocity. Figures 5 and 6 demonstrate the bias in the results. The fractional bias is nondimensional and is defined as

$$ \text{fractional bias} = \frac{\text{modeled value} - \text{observed value}}{\text{observed value}} \quad (31) $$

Figure 5 shows the fractional bias in friction velocity as a function of fetch, wind speed, and temperature difference between the ocean and atmosphere for the present model. The fractional bias in $u^+$ is seen to be high for low fetches. This could be indicative of the Monin-Obukhov similarity theory not being applicable at the highly inhomogeneous region very close to the lead edge as occurs for very small fetches. Alternatively, it could reflect the artificial currents induced by the pumping for the artificial leads (fetches below 20.5 m). There appears to be a large bias for low wind speeds less than 1.3 m s$^{-1}$ with the model underestimating the momentum flux. At low wind speeds the pumping in the artificial leads would most clearly be reflected in the low wind speed cases, biasing the observations. The bias in friction velocity is independent of the air-sea temperature difference. The variations in the fractional bias in heat flux with changes in lead width, wind, and air-sea temperature difference are shown in Figure 6. The results appear to mirror the friction velocity bias. There is an indication of the overestimate from the possible freezing of the natural lead case. The Andreas and Murphy model flux biases show a dependence similar to that of the present model, so they are not shown here.

4. Sensitivity Studies

Sensitivity studies are conducted to give insight into the physics of the model. Specifically, we examine the influence of the shear and convective contributions to the surface renewal time, the effect of excluding the cool surface skin, and the effect of neglecting the wave age dependence on the turbulent flux model.
Figure 5. Fractional bias in friction velocity as a function of fetch, wind speed, and air-sea temperature difference for the present model.

Figure 6. Fractional bias in heat flux as a function of lead width, wind speed, and air-sea temperature difference for the present model.
4.1. Surface Renewal Timescale

In this section we examine the importance of the shear versus convective contribution to the surface renewal time. Figure 7 shows that omitting the convective contribution does not substantially change the results, as the model results with renewal time which only considers the shear contribution match quite closely with the model results where the surface renewal time includes both shear and convective effects (the 1:1 line). The atmosphere above the lead is in the unstable regime with both buoyant convection (since air is colder than the lead surface) and shear-generated turbulence (resulting from the airflow over the lead). The dominance of convection or shear in the surface renewal timescale is determined by the critical Richardson number, which defines the turbulence regime (see equation (5)). For a value of surface Richardson number greater than 1 (the critical surface Richardson number for the atmosphere), convective turbulence dominates, and for values of Richardson number much smaller than the critical Richardson number, shear dominates the turbulence. Since the surface Richardson number for the data used here [Andreas et al., 1979] is much less than the critical surface Richardson number, the shear contribution dominates the convective contribution to the renewal timescale, and hence neglecting the convective contribution does not make much difference to the model results.

4.2. Cool Skin

The model accounts for the cooling of the sea surface skin resulting from the loss of energy via surface turbulent fluxes and net radiative flux loss by the ocean surface (equation (12)). The Andreas et al. 1979] data set used here has 57 cases with the 2-m air-sea temperature difference ranging from 16.5 to -30.6 K and the wind speed ranging from 1.06 to 5.92 m s^-1 and the fetch varying from 6.8 to 34 m. The ALEX data [Andreas et al., 1979] measured the temperature in near-surface water (and thus were not skin temperatures). For all leads, the measured surface temperature was between -2°C and -2.4°C; it is assumed here that radiometric skin temperature would be slightly cooler than the measured subsurface temperature. Since the freezing temperature at a salinity of 31.83 is about -1.7°C, the measured surface temperatures reflect the cooling of the surface skin resulting from the surface flux. The observations thus indicate a surface supercooling of at least -0.3°C to -0.7°C. The modeled difference between the surface skin temperature and the bulk ocean mixed layer temperature is determined from (12) to range from -0.62 to -1.35 K (for the Andreas et al. 1979] cases) which is 2.5% to 7% of the temperature difference between the ocean and the atmosphere and hence not negligible. It is also noted here that the measured sea surface temperature (SST) in the artificial leads was likely to have been influenced by the pumping. Although the Wick et al. 1996] skin surface temperature parameterization has not been evaluated previously under Arctic wintertime conditions, the modeled results are in a plausible range.

To investigate the importance of the skin cooling on the surface heat flux, we ran the model for the Andreas et al. 1979] profiles without considering the cooling of the skin with a surface temperature of -1.7°C, which corresponds to a climatological ocean mixed layer temperature during winter. When this is compared with the flux obtained when the cooling of the skin is considered (given a bulk SST of -1.7°C, so that the skin temperature found from using the surface renewal theory varies from -3.05°C to -2.32°C for the various cases), the results for this sensitivity study are shown in Figure 8. It is seen that not including the cooling of the surface skin results in larger heat fluxes. This is because the cool skin gives a smaller temperature difference relative to the cold atmosphere, which results in reduced heat fluxes. This indicates that the cooling of the skin acts as a buffer on the interface by allowing less energy to be lost from the surface.
Katsovos [1973] observed a supercooling of several tenths of a degree on the surface of a lead. For salt water with a salinity above 25% (as occurs in the Arctic Ocean), the temperature of maximum density is lower than the freezing point, so that the supercooled water on the surface is heavier than water of the same salinity at its freezing point. This heavy supercooled water, which forms on the surface, sinks to a lower level with no nucleation because it is denser than the underlying seawater at its freezing point. The buoyant sinking may be enhanced by wind stirring. Frazil then forms in the water below, the ice particles then rising to the surface because of buoyancy effects. Therefore sinking of the supercooled water leads to the maintenance of open water [Hishio and Wakatsuchi, 1993]. Continued freezing results in the production of grease ice, a mixture of unconsolidated frazil crystals. The surface wind waves induce turbulence, which inhibits the development of a solid cover on the lead surface. Wind and wave action advect the frazil crystals to the downwind edge, leaving the lead surface clear of ice. With the continuation of ice formation and advection, the surface ice accumulation zone upwind with time [Bauer and Martin, 1983]. A supercooling of more than 2°C (which is beyond the surface cooling for the given lead observations) results in pancake ice cover over the surface [Weeks and Ackley, 1982], which is harder to advect than frazil ice, so the ocean surface does not remain ice-free at high supercoolings. Presence of ice on the surface would affect the surface roughness [Guest and Davidson, 1991] but Guest and Davidson [1991] also observed ice-free regions in the marginal ice zone with complicated surface wave fields due to highly variable fetches and sea surface temperatures similar to some coastal areas. We have considered an ice-free lead surface, and the production of frazil ice is not accounted for in the present study; future studies will address this issue.

4.3. Wave Age Dependence

This model uses a surface roughness length (equation (13)) which depends on a nonarbitrary wave age that is a function of atmospheric and oceanic variables. If we ignore the wave age dependence of the surface roughness length, i.e., assume that the sea is saturated with waves that are in equilibrium with the wind under neutral atmospheric conditions, the last term on the right-hand side of (13) can be written as $\mathcal{E}u^*/lg$. The value of $\mathcal{E}$, Charnock's constant, remains a constant when we do not consider wave age dependence. When we replace the wave age dependence of the last term on the right-hand side of (13) by $0.017u^*/lg$, i.e., $\mathcal{E} = 0.017$, the results obtained are shown in Figure 9. The transition to a wavy surface when we do not consider wave age dependence is taken to occur at a roughness Reynolds number of 1.065, since the transition from smooth sea surface to wave-covered surface occurs between roughness Reynolds numbers 0.13 and 2. This result matches quite well with the results for the model considered to be at wind-wave equilibrium by taking the wind-wave stability term $W$ to be 1, which implies that the sea surface wave field is in equilibrium with the imposed atmospheric wind. $\mathcal{E} = 0.017$ (no wave age dependence with the sea surface in wind-wave equilibrium) represents just such an equilibrium condition, causing the two results to practically coincide as is observed. These results with an effective $\mathcal{E}$ of 0.017 agree with the present wave-age-dependent model results for low wind speeds but increasingly diverge from the present model as the wind speed increases or as the sea surface goes further from the state of wind-wave equilibrium.

5. Lead Integral Fluxes

In this section we examine the effect of atmospheric winds and stability on the fetch dependence of the lead sensible heat fluxes. Initially, as the cold incoming air encounters the lead, it experiences a strong contrast at the surface resulting in buoyant convection. As the air flows over the lead, it is modified, i.e., becomes warmer, so that the convection is dominated more and more by forced convection. This results in fluxes that decrease with increasing fetch until the flow is no longer fetch-limited for large enough fetches.

We examine the heat flux as a function of fetch for varying atmospheric conditions (ocean-atmosphere temperature difference ranging from -10 to -40 K and wind speeds ranging from 3 to 7 m s$^{-1}$). The results for integral/overall heat flux as a function of lead width for an air-sea temperature difference of -30 K and a 10 m wind speed of 4 m s$^{-1}$ (typical conditions in the Arctic winter) are shown in Figure 10. The integral fluxes (as defined below) are shown both for the present model and the Andreae and Murphy model. We see an initial sharp decrease in heat flux with lead width up to 100 m, beyond which the heat flux decreases more slowly and is practically constant beyond a fetch of 200 m. Figure 10 also shows the heat flux obtained by using the Andreae and Murphy [1986] model, which shows a similar dependence on lead wind. It is seen that the two models give close agreement for short lead widths, with a small, constant difference for large lead widths of hundreds of meters. As shown in Table 2 and Figure 4, the present model shows better agreement with the Smith et al. [1983] observations for large fetch (greater than 210 m). So the present model is closer to the truth for large fetches. It would be helpful to have observations at 100 m, since the two models differ strongly thereafter in their variation of heat flux with fetch.
To assess the integral flux contribution of leads in a mesoscale model grid, Table 2 shows the integral lead fluxes $\Pi$ as a function of the lead width, temperature difference between the atmosphere and the ocean, and the 10 m wind speed

$$\Pi = \sum \frac{X_i H(X_i)}{\sum X_i}$$

(32)

where $H(X_i)$ is the heat flux at fetch $X_i$ and the denominator in equation (32) gives the total lead width. The lead integral flux $\Pi$ gives the average flux over the lead. In the calculations shown in Table 2, we have assumed that the wind direction is perpendicular to the lead (i.e., in the across-lead direction). Incoming wind with an along-lead component will result in an effectively longer lead width. The flux from the lead for different lead widths under different wind speeds and stability regimes is shown in Table 2. We see that the integral fluxes increase with wind speed and with increasing atmospheric instability, i.e., increasing air-sea temperature difference for a given lead width. For a given atmospheric state, the fluxes decrease with increasing lead width, sharply at first and then gradually until there is no significant change in heat flux for further increase in lead width.

The dependence on lead width of integral heat fluxes is seen in Table 2 to extend to farther distances with increasing wind speed. The initial strong lead width dependence of the lead turbulent fluxes extends to 50 m with a 3 m s$^{-1}$ wind whereas it extends to 125 m and 250 m with a 5 m s$^{-1}$ and 7 m s$^{-1}$ wind speed, respectively. This is because a low enough wind is equivalent to infinite fetch, and as the wind speed increases, the upwind effect is felt over much longer fetches because the air is able to go farther without being modified significantly. Hence the buoyant convection effect can last over a longer distance with increasing wind speed, resulting in increasing lead width dependence of lead integral surface fluxes with increasing wind speeds. The aforementioned characteristics of the surface flux variations can be seen qualitatively in Figure 11, which shows the integral surface flux variation with lead width for varying atmospheric conditions.

### Table 2. Lead Integral/Overall Heat Fluxes as a Function of Lead Width for Varying Air-Sea Temperature Difference and 10-m Wind

<table>
<thead>
<tr>
<th>$dT$</th>
<th>$u_a$ (m s$^{-1}$)</th>
<th>1 m</th>
<th>5 m</th>
<th>10 m</th>
<th>20 m</th>
<th>30 m</th>
<th>50 m</th>
<th>75 m</th>
<th>100 m</th>
<th>200 m</th>
<th>500 m</th>
<th>1 km</th>
</tr>
</thead>
<tbody>
<tr>
<td>10 K</td>
<td>3</td>
<td>103.0</td>
<td>89.7</td>
<td>81.9</td>
<td>73.3</td>
<td>68.1</td>
<td>63.0</td>
<td>61.2</td>
<td>60.5</td>
<td>59.7</td>
<td>59.5</td>
<td>59.5</td>
</tr>
<tr>
<td>15 K</td>
<td>3</td>
<td>173.0</td>
<td>153.9</td>
<td>142.4</td>
<td>129.4</td>
<td>121.3</td>
<td>110.8</td>
<td>102.5</td>
<td>96.6</td>
<td>89.6</td>
<td>87.1</td>
<td>86.7</td>
</tr>
<tr>
<td>20 K</td>
<td>3</td>
<td>266.9</td>
<td>216.9</td>
<td>218.7</td>
<td>198.3</td>
<td>185.6</td>
<td>169.7</td>
<td>156.1</td>
<td>147.0</td>
<td>125.9</td>
<td>115.8</td>
<td>114.6</td>
</tr>
<tr>
<td>25 K</td>
<td>3</td>
<td>250.2</td>
<td>214.8</td>
<td>194.4</td>
<td>172.4</td>
<td>159.3</td>
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<td>140.3</td>
<td>138.4</td>
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<td>137.8</td>
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<tr>
<td>30 K</td>
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<td>413.3</td>
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<td>330.1</td>
<td>306.7</td>
<td>276.7</td>
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<td>230.7</td>
<td>216.9</td>
<td>200.4</td>
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<td>194.1</td>
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<tr>
<td>35 K</td>
<td>3</td>
<td>616.5</td>
<td>540.7</td>
<td>495.9</td>
<td>446.5</td>
<td>416.4</td>
<td>377.9</td>
<td>347.4</td>
<td>326.2</td>
<td>277.8</td>
<td>254.5</td>
<td>251.5</td>
</tr>
<tr>
<td>40 K</td>
<td>3</td>
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<td>329.1</td>
<td>290.0</td>
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<td>229.7</td>
<td>229.6</td>
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<td>610.1</td>
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<td>414.7</td>
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<td>319.6</td>
<td>318.2</td>
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<tr>
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<td>3</td>
<td>1038.6</td>
<td>900.3</td>
<td>820.4</td>
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<td>681.4</td>
<td>615.3</td>
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<td>357.3</td>
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<td>331.5</td>
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<td>905.1</td>
<td>818.7</td>
<td>726.3</td>
<td>671.4</td>
<td>602.9</td>
<td>550.2</td>
<td>514.6</td>
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<td>459.8</td>
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<td>1321.8</td>
<td>1197.4</td>
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<td>756.1</td>
<td>638.1</td>
<td>583.6</td>
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6. Conclusions

A physically based turbulent flux model has been used to determine fluxes from Arctic leads. We use a surface flux algorithm that computes the momentum and heat fluxes as a function of the atmospheric and oceanic variables by applying surface renewal theory to the air-sea interface. The surface fluxes are defined in term s of the surface scales of velocity, temperature, and moisture, which appear in the Monin-Obukhov similarity theory as functions of the surface roughness length and atmospheric profiles. A self-consistent surface roughness model modified to include fetch dependence is used to obtain the momentum roughness length, and surface renewal theory is applied thereafter to obtain the roughness length for temperature.

The momentum and heat fluxes from leads with varying fetches obtained during ALEX in 1974 were compared with our model and a bulk-transfer coefficient based model [Andreas and Murphy, 1986], where the neutral stability coefficients were determined empirically from the ALEX data. The present model is seen to give a similar correlation with all the described ob-
Figure 11. Integral lead fluxes as a function of lead width for different atmospheric states.

...shown that the heat fluxes averaged over the entire lead width increase both with increasing imposed wind speeds and with increasing air-sea temperature difference. The integral lead fluxes are highly sensitive to the lead width. For a given atmospheric state it is seen that as the lead width increases, initially for narrow lead widths the integral lead flux decreases significantly, and eventually for large enough lead widths the integral surface flux does not change much with further increase in lead width. The range of this strong lead width dependence of the surface fluxes changes significantly with varying wind speed. The lead width dependence exists for longer distances as wind speed increases. This is because with increasing wind speed the air is able to go farther across the lead without being modified significantly by the surface, so that the buoyant convection dominates for longer distances for higher wind speeds.

In the absence of sufficient observations at low fetch in natural leads, it is difficult say whether the present model is more accurate than the Andreas and Murphy empirically derived parameterization. Nevertheless, the present model is more firmly based on the fundamental physical processes and provides insights into the importance of different physical processes. Additionally, a model that is more physically based can be trusted outside the parameter range from which empirically derived parameterizations were determined. More observations of surface heat and momentum fluxes over leads under a variety of environmental conditions are needed to further validate the model. It is anticipated that the forthcoming Surface Heat Budget of the Arctic Ocean (SHExA) experiment will provide additional observations to validate the model [SheBA SWG, 1994]. In particular, the details of the surface freezing process in leads as a function of the surface heat flux needs further investigation.

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