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On treatments of fetch and stability sensitivity in large-area estimates of sensible heat flux over sea ice

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Abstract. Bulk transfer coefficients estimated as a function of atmospheric stability and sea-ice lead width are combined with Arctic meteorological observations and ice thickness data to calculate the sensitivity of turbulent flux estimates to changes in lead width, wind speed, air temperature, and ice thickness for a high-concentration ice pack. These results are considered in terms of bulk transfer parameterizations that use a fixed transfer coefficient or that address atmospheric stability only. On the basis of the fetch-sensitive parameterizations considered here, differences in lead width for widths up to about 200 m can exert a substantial influence on sensible heat transfer coefficients and heat flux from leads under typical Arctic conditions. Fluxes from an open water lead decrease by 34% if fetch increases from 10 m to 100 m. This effect is greatest for open water leads, decreases considerably as leads refreeze, and is negligible for ice thicker than about 0.3 m. If open or newly refrozen leads make up 2% of the ice cover, then an increase in mean fetch from 10 m to 100 m yields a decrease of about 2 W m^{-2} in areally averaged flux from the ice pack. Calculations using observed and theoretical lead width distributions suggest that parameterizing lead widths in a sea ice model can be done effectively using a single, representative lead width rather than requiring a full distribution of widths. When coupled to the lower atmospheric boundary layer using a bulk similarity theory model, this sensitivity of heat transfer to fetch results in substantially higher near-surface air temperatures over narrow leads, with equilibrium air temperatures decreasing by about 50% as fetch increases from 10 to 100 m.

Introduction

The objective of this work is to address the relative importance of including information on ice lead width, width distributions, and atmospheric stability in calculations of turbulent fluxes relevant to large-scale sea ice and climate models. Since the polar regions act as heat sinks for the global energy system, factors that affect the energy balance in high latitudes are likely to be linked to the global climate. One such factor is the effect of sea ice on the turbulent exchange of energy between the atmosphere and underlying ocean. Within the ice pack the contribution of open water and thin ice areas (e.g., leads and polynyas) to heat transfer and new ice production is disproportionate to their size. Estimates by *Maykut* [1978] for the central Arctic in March show turbulent heat flux (sensible plus latent heat fluxes) of 765 W m^{-2} from open water compared with 44 W m^{-2} for 0.8-m ice. For a sea ice cover consisting of 95% thick ice with a turbulent flux near 0 W m^{-2} the areally weighted flux from the ice pack increases from 2 W m^{-2} to 38 W m^{-2} if the lead area consists of open water, rather than 0.8-m ice. In reality, since the lead area during winter is a mixture of open water and ice of different thicknesses, the true situation will fall somewhere between these two extremes.

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The contribution of lead area to atmospheric heating has been studied using climate models. For example, increasing mean lead fraction from 0 to 4.3% produced a 2K increase in air temperature in the central Arctic using an energy balance model [*Ledley*, 1988], while a similar perturbation in a general circulation model (GCM) changed sea level pressures substantially in the Antarctic [*Simmonds and Budd*, 1990]. In such models a single value is used per zone or grid cell to represent total open water fraction or the mixture of open water and thin ice within the grid cell. Given the potential contribution of leads to polar climate processes, it is worth considering how factors other than total lead-covered area might affect simulations of turbulent fluxes and changes in air temperatures.

In models where open water or thin ice fraction is included in heat flux calculations, bulk aerodynamic formulas; e.g.,

$$H_s = \rho C_p C_{Hr} U_r (T_s - T_r),$$
$$H_L = L_v C_{Er} U_r (Q_s - Q_r),$$

are typically used to represent surface turbulent heat transfer. H_s is sensible heat flux; H_L is latent heat flux; ρ is the air density; C_p is the specific heat of air at constant pressure; L_v is the latent heat of vaporization of water; U_r is the wind speed at reference height r; T_s and T_r are surface and surface air temperatures at height r, respectively; Q_s and Q_r are absolute humidities; and C_{Hr} and C_{Er} are bulk transfer coefficients at height r for sensible heat and latent heat, respectively. Here we use the convention of positive flux representing heat transfer from leads to the atmosphere. The transfer coefficients are determined empirically and parameterize complex interactions between forced convection (mechanical turbulence), free convection (convective turbulence), and scalar roughness lengths [*Deardorff*, 1968]. In large-scale sea ice models, values for C_H and C_E are often fixed and do not reflect changes in ice thickness, atmospheric stability, or lead width [e.g., *Parkinson and Washington*, 1979; *Flato and Hibler*, 1992; *Holland et al.*, 1993]. In other cases, atmospheric stability is considered, but possible effects of lead width are not [e.g., *Bauer and Martin*, 1983; *Ebert and Curry*, 1993].

Sensitivity studies that include simple scaling of transfer coefficients offer some insight into the effects on sea ice simulations of uncertainties in C_H and C_E [e.g., *Holland et al.*, 1993]. Linking changes in the transfer coefficients to physical variables can help relate these uncertainties to the broader question of relationships among sea ice and atmospheric conditions. The rate of turbulent heat transfer from leads and polynyas has been shown to be related to over-water distance of travel (fetch) as well as to boundary layer stability [e.g., *Andreas*, 1980]. While parameterizations of such effects in terms of bulk transfer coefficients are based on a fairly small sample of observations over a limited range of conditions, they provide a starting point to consider the importance of treating stability and lead width when estimating turbulent fluxes in climate models. The importance of knowing lead widths and width distributions, rather than mean lead area alone, can also help prioritize remote sensing activities and field programs. For example, mapping individual leads and lead width distributions requires much more effort than is needed to derive a reasonable estimate of total lead area for a region.

Rigorous studies of boundary layer processes over the ice pack point out the large uncertainties that still remain in terms of drag and heat transfer coefficients, as well as the complex nature of the polar atmosphere [*Overland*, 1985]. Results of applying simple relationships to represent these conditions thus must be interpreted with care, particularly for extreme cases where observations available to develop the relationships are sparse. However, the methods used here capture the first-order aspects of the problem and thus are meant to provide a useful overview of interrelationships among atmospheric conditions and lead characteristics. We also note that results achieved using these methods are particularly relevant for assessing the adequacy of the even simpler approximations currently used in large-scale dynamic-thermodynamic sea ice simulations.

Methodology

Starting with an existing parameterization of the effects of fetch and stability on bulk transfer coefficients, we assess the relative importance of including lead characteristics in estimating turbulent heat flux in the context of large-scale ice modeling. First, we consider the possible changes in bulk transfer coefficients under different atmospheric and surface conditions. Transfer coefficients and fluxes are estimated using a range of conditions, and are related to transfer coefficient parameterizations that use fixed coefficients or that treat atmospheric stability alone. These sensitivity studies are then extended to use observed atmospheric conditions and lead information, including observed and theoretical lead width and ice thickness distributions. Finally, since any change in heat transfer will affect air temperatures, which, in turn, will affect surface winds and the rate of heat exchange, we apply a simple bulk similarity theory model of the boundary layer to consider the change in air temperature when fetch is varied.

From observations, *Andreas and Murphy* [1986] and *Andreas* [1987] (referred to herein as AM) present a means to adjust neutral stability coefficients (C_{HN}, C_{EN}, drag coefficient C_{DN}) to include the effects of atmospheric stability and fetch X and to correct these transfer coefficients for reference height. The resulting coefficients are representative of the entire lead rather than a single point within the lead. AM suggest, as have others, that forced convection and free convection combine to affect the values of C_H and C_E. AM use the Obukhov length L as a means to represent fetch scaled by forced and free convection. As fetch increases and/or L decreases, C_{HN} tends toward open ocean values of about 1.0×10^{-3}. At small nondimensional fetch $-X/L$, free and forced convection combine to increase energy transfer. Under the same wind and temperature conditions and for leads narrower than about 200 m, narrow leads thus tend to yield a greater average flux (e.g., average heat loss per unit lead area) than do wider leads [*Andreas*, 1980; *McBean*, 1986].

To understand the applicability of L as a scaling parameter, it is important to note that the scaling assumes conditions where a cold, upwind air mass is advected over a warm lead. The scaling assumes this upwind temperature difference. Over very long fetches this large temperature contrast will have been reduced. The same scaling could be applied for such large leads or polynyas if a representative air mass temperature is used (E.L. Andreas, personal communication, 1994). In the AM approach, changes in C_{HN} for different fetches and air temperature-lead surface temperature gradients combine with stability-adjusted drag coefficients to yield appropriate C_H values. For a given set of air temperatures and winds, C_H always decreases with fetch. However, the effects of air temperature and winds on C_H can be different at different fetches, as seen in the results described later.

To define the critical combinations of atmospheric and surface conditions that might make lead width a significant factor in climate simulations, the AM equations are used here to calculate coefficients and fluxes under several combinations of wind speed, air temperature, fetch, and ice thickness. Bulk transfer coefficients are calculated for a reference height r of 10 m. Values are specified for U_r, T_s (for open water and thin ice), T_r and relative humidity RH. As noted above, U_r, T_r, and RH represent conditions over sea ice upwind of the lead. Salinity S at 34 parts per thousand (ppt) is used to estimate the freezing temperature of sea water. Derived conditions are air density, surface vapor pressures over ice and water, and vapor pressure at height r over water. For the examples given here, wind direction is taken as perpendicular to the long axis of the lead (e.g., fetch equals lead width).

Since leads refreeze rapidly during periods of large heat loss, it is also worthwhile to consider the effects of fetch and stability over leads consisting of thin and young ice. The only difference in the approach is that an ice surface temperature must be provided for the refrozen lead. Surface temperatures for different ice thicknesses are estimated using an energy balance equation coupled to the AM transfer coefficient model, with specified values for wind speed, air temperature, ice conductivities, ice albedo, oceanic heat flux, long-wave flux, and short-wave flux. The energy balance equation used is basically the single-layer model given by *Maykut* [1982]. Saturation vapor pressure over water and ice are calculated according to *Murray* [1967] and adjusted for salinity [*Andreas*, 1977]. A linear temperature profile and no internal heat storage are assumed for the thin ice (0.05 - 0.30 m) we are concerned with here. Since C_H is a function of surface temperature and vice versa, the energy balance equation is iterated to converge on C_H and T_s. Inputs for these flux calculations over different ice thicknesses are described in a subsequent section.

As the ice in the lead thickens, the temperature and height change between the lead and adjacent thick ice at the upwind edge of a lead will diminish (e.g., this "step effect" becomes smaller). However, for the experiments here we neglect the effect of this on forced and free convection. Also, as the lead ice becomes thicker, the T_s-T_r contrast over the lead decreases. The parameterizations used here, which assume unstable conditions, may thus be less reliable for old leads. Given these assumptions, care must be taken in interpreting the results for relatively thick ice.

Finally, we note that sensitivity studies that vary surface fluxes without any coupling to the atmosphere clearly cannot represent how the change in heat input to the lower boundary layer affects air temperature and surface wind speed and thus the rate of heat transfer. With few exceptions [e.g., *Koch*, 1988; *Stossel*, 1992; *Pollard and Thompson*, 1994], dynamic-thermodynamic ice models typically have no atmospheric coupling. Since parameterizations of heat transfer coefficients are based on observations, the parameterizations presumably include the effects of boundary layer modification taking place over the lead. For uncoupled models, this implies that a parameterized transfer coefficient is suitable for initial atmospheric conditions at the start of a model time step but that the coefficient is not adjusted to reflect a new energy balance between surface and atmosphere during the time step. Essentially, this lack of coupling means that all heat input to the atmosphere is removed from the system, since the prescribed air temperature supplied to the surface energy balance model at the next time step is independent of surface fluxes. In reality, an increase in the rate of heat loss from leads will warm the surface air layer over the lead, with subsequent advection and mixing with surrounding air. The rate of heat transfer and the appropriate transfer coefficients will thus change.

Here we attempt to represent some of the key elements of this influence of heat flux on air temperature and atmospheric stability using bulk similarity theory relationships described by *Koch* [1988] and applied by *Stossel* [1992]. Briefly, this vertically-integrated one dimensional model of the atmospheric boundary layer treats the air layer from the surface to a given reference height following Monin-Obukhov similarity theory. The relationships between turbulent fluxes at the surface to fluxes at the geostrophic level are determined using Rossby number similarity principles [*Koch*, 1988]. Winds and temperature at the geostrophic level provide the forcings for the model.

Detailed vertical profiles of the effects of changes in surface flux can be estimated using boundary layer models [e.g., *Overland*, 1988], and plume development models can be applied to define the specific effects of leads in terms of vertical development and vapor transport [e.g., *Pinto et al.*, this issue]. When confined to the unstable conditions over leads in particular (e.g., where buoyancy is less constrained by surface inversion), *Koch's* [1988] much simpler formulation provides the basic elements of the desired surface-atmosphere coupling and thus allows us to consider the negative feedback between surface air temperature and heat flux in a first-order sense.

Data

Forcing Fields

To apply the sensitivity calculations to actual conditions, meteorological data, lead widths, and ice thickness information are needed. For air temperatures and winds we use daily values and monthly means estimated from a 5-year set of gridded pressure and temperature fields provided by the Arctic Ocean Buoy Program (commencing with *Thorndike and Colony* [1980]). When used as a proxy for near-surface air temperatures, the internal buoy temperatures are expected to be biased toward overestimates of air temperatures during summer due to radiational heating and, perhaps, also during winter if the buoys are insulated by drifted snow. A comparison of buoy and station data [*Martin and Clark*, 1978] suggests that during the periods when turbulent fluxes are greatest, the daily averaged buoy temperatures are reasonable estimates of near-surface air temperatures. Daily air pressures and temperatures for 12 Greenwich Mean Time (GMT) were extracted for the period from January 1, 1979, to November 30, 1984, for a 200 x 200-km grid cell centered at approximately 80° N, 155° W. Geostrophic winds were calculated from the gridded pressure data and scaled to represent surface winds [*Albright*, 1980]. For the flux calculations these data are combined with radiative forcings and relative humidity from *Maykut* [1982] (Table 1).

Lead Width Parameterization and Lead Width Data

Given the difficulty of routinely collecting lead width data with sufficient coverage suitable for large-scale modeling, it is useful to consider the sensitivity of large-area estimates of mean heat flux to how lead width data are applied or parameterized. Fluxes could be calculated by integrating over actual or theoretical lead width distributions, or a representative flux could be estimated from a mean lead width. To address this question of how lead width distributions might best be used in turbulent flux estimates, a method is needed to compare fluxes using a single fetch (set equal to lead width) and fluxes integrated over a distribution of fetches. Specifically, let $f(X)$ be the distribution of leads widths X with mean λ. Since H (sensible or latent) is, in part, a function of the transfer coefficients and the transfer coefficient C_H is, in part, a function of lead width, then

$$H[C_H(\lambda,...),...] = \int H[C_H(X,...),...]\, f(X)\, dX$$

will only be true if $f(X)$ is linear. Otherwise, there will be a difference between heat flux estimates using a single lead width value or the distribution of lead widths. To examine these differences we use as $f(X)$ the negative exponential distribution characteristic of lead widths [*Key and Peckham*, 1991] as the model:

$$f(X) = \frac{1}{\lambda} e^{-X/\lambda}$$

This model implies that there are a finite number of small leads and that the field is characterized by a length scale λ.

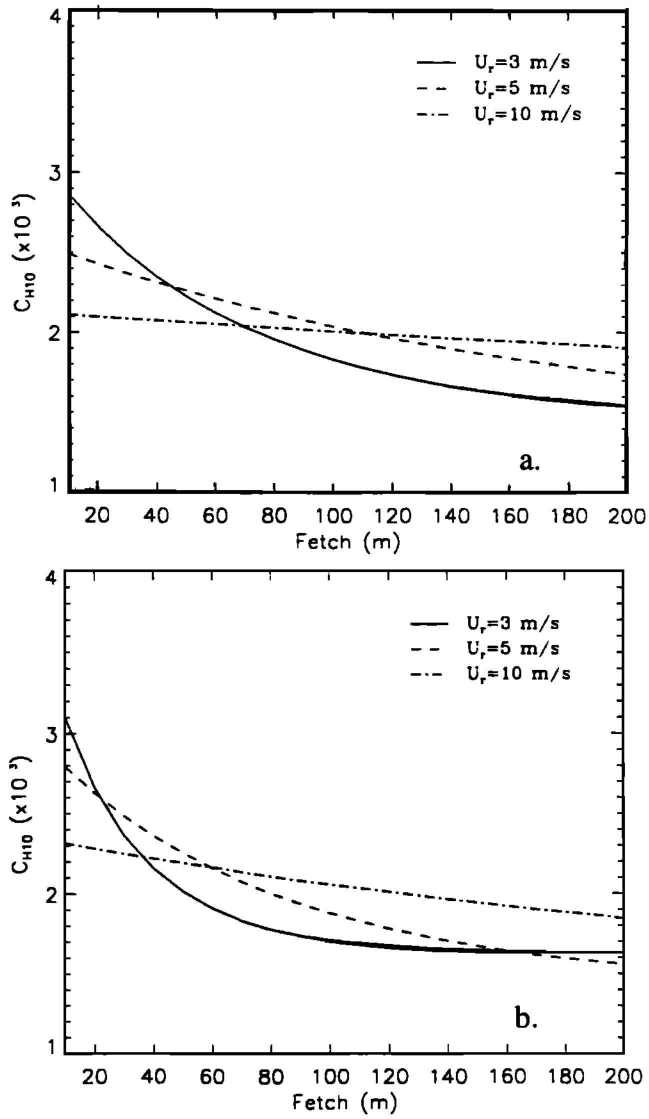

Figure 1. Sensible heat flux transfer coefficient C_{H10} calculated as a function of fetch, surface wind speed, and surface air temperature for surface air temperature T_r of (a) -10°C and (b) -20° C. Surface wind speed U_r = 3, 5, and 10 m s^{-1}.

In fact, the lead width distribution may be scale free, in which case a power law would be more appropriate. However, this distinction is not important here.

Distributions of lead widths and spacings have not been studied extensively, and statistics are available for only a few geographic areas and times. The statistics used here are based on submarine sonar data recorded in the Canada Basin in August 1970 [*McLaren*, 1989] and October 1978 (*A.S. McLaren*, unpublished data, 1989) with a narrow beam, upward looking acoustic profiler. Data were interpolated from analog records to 1.5-m intervals. In addition to this sample of observed ice thicknesses and lead width distributions and the theoretical exponential distribution function noted above, the modeled ice thickness distribution of *Maykut* [1982] is used as an additional comparison data set for calculating the area-weighted effect of modifying C_H over thin ice areas.

Results

Using these methods and data sets we first consider the sensitivity of turbulent flux estimates to fetch, wind speed, and temperature gradients for prescribed ranges of conditions. The calculated transfer coefficients are compared to other parameterized coefficients given in the literature. Estimates of turbulent fluxes are then presented for actual central Arctic conditions using the sea ice and meteorological observations noted above. Finally, we discuss the effects of stability and fetch on heat transfer from leads when the surface and atmosphere are coupled.

Sensitivity to Fetch, Wind Speed, and Temperature and Comparisons to Other Parameterizations

Figures 1 and 2 summarize how the mean transfer coefficient for sensible heat flux at a reference height of 10 m (C_{H10}) for an open water lead changes for different fetches and wind speeds at T_r = -10°C and -20°C, as determined using the AM approach. In Figure 2 the effect of fetch is presented as the percent decrease in C_{H10} relative to C_{H10} at a fetch of 10 m. This change relative to fetch (e.g., $d\ C_{H10}/dX$) is a function of stability and wind speed, as represented by X/L (AM's Figure 3). Higher wind speeds extend the range of fetch over which C_{H10} is affected, although the magnitude of the change decreases. At high wind speeds, L is large, so that for a given fetch the neutral stability coefficient C_{HN10} and C_{H10} remain high over longer fetches.

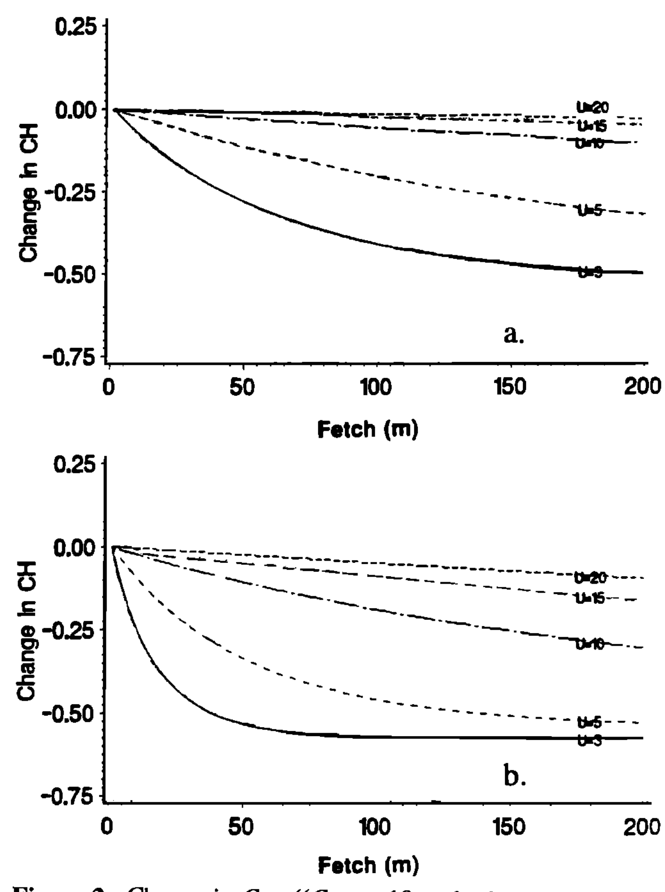

Figure 2. Change in C_{H10} ((C_{H10} at 10-m fetch minus C_{H10} at fetch X)/C_{H10} at 10-m fetch) with fetch at wind speeds of 3, 5, 10, 15, and 20 m s^{-1}. T_r = (a) -10 °C and (b) -30°C.

Increased instability increases the magnitude of dC_{H10}/dX, but the change is confined to a smaller range of fetches as the temperature difference becomes greater As stability decreases (such as between Figures 1a and 1b, with air temperatures of -10°C and -20°C), L decreases and C_{H10} falls off more rapidly with increasing fetch. For a larger temperature contrast the range of C_{H10} is about the same but with C_{H10} decreasing more rapidly with fetch. For example, at T_r = -30°C and U_r = 5 m s^{-1}, C_{H10} is 2.96x10^{-3} at X = 10 m, decreasing to 2.14x10^{-3} at a fetch of 50 m.

As noted earlier, thin ice in a lead decreases the surface temperature-air temperature difference by decreasing the lead surface temperature. This is essentially equivalent to the effect on transfer coefficients of an increase in air temperature over open water. As shown in Figure 3 for mean January radiative forcings (Table 1) and ice temperatures estimated using the energy balance model described earlier, the fetch dependency decreases substantially as the lower surface temperature of thicker ice decreases the temperature contrast between surface and the overlying near-surface air layer. At typical mean Arctic wind speeds (about 5 m s^{-1}), fetch dependency is minimal for ice thicker than about 0.30 m, or 0.10 m at high wind speeds (U_r = 15 m s^{-1}).

The dependence of flux on open water fetches as long as 200 m using the AM parameterization differs from results cited by *McBean* [1986] and *Smith et al.* [1990], who suggest that this dependence becomes minimal at a smaller fetch. These different findings are due to the choice of empirical function used to represent the relationship between heat flux and fetch. *Andreas* [1980] derived linear and nonlinear relationships between the fetch-dependent Reynolds number R_X and a Nusselt number N for sensible heat flux. Using the linear relationship, *McBean* [1986] and *Smith et al.* [1990] estimated that the average heat flux from a lead is independent of fetch when fetch exceeds about 30 m. Both the linear and nonlinear forms fit the data used to develop these relationships equally well. *Andreas* [1980] notes that the data used do not support choosing between the linear and exponential forms (the observations did not include fetches greater than 35 m), although he states that the exponential equation appears more physically reasonable over a large range of Reynolds numbers.

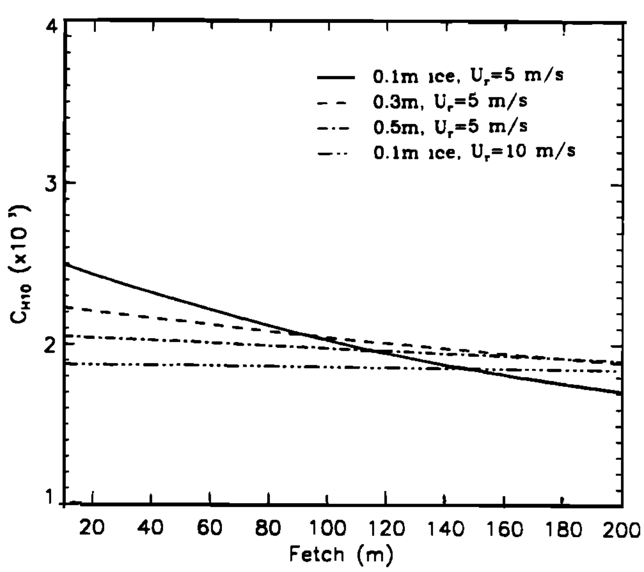

Figure 3. C_{H10} as a function of fetch and ice thickness for T_r = -20° C, U_r = 5 m s^{-1}, and ice thicknesses of 0.1 m, 0.3 m, and 0.5 m, and U_r = 10 m s^{-1} for 0.1-m ice .

Using this exponential form ($N=R_x^{0.76}$) in place of the linear relationship and integrating over a range of fetches yields a representative heat flux for a fetch X. This flux can be compared to flux estimated using AM-derived transfer coefficients, noting that the AM model is based on observations containing lead widths substantially greater than 35 m. As noted earlier, the AM value represents a single coefficient applicable to the entire lead. In an example given by *McBean* [1986] the average flux estimated using the linear equation decreases by only 4% as fetch increases from 33 m to 100 m for a wind speed of 3 m s^{-1} and a temperature difference between surface and air of 19°C. However, if the exponential form noted above is used, then the decrease is 17%. This compares to a 30% decrease in flux using the AM model.

These differences point out the uncertainties that exist among reasonable parameterizations of heat flux and transfer coefficients. Given the range of parameterizations that have been developed to fit observations in different ways, it is instructive to compare the AM results to some additional examples of transfer coefficients used in ice modeling. For example, a transfer coefficient of 3.0x10^{-3} was used by *Maykut* [1982] for open water leads and thin ice, while *Parkinson and Washington* [1979] and subsequent generations of their thermodynamic treatments (e.g., *Hibler* [1979] and derivatives) typically use a value of 1.75x10^{-3} for proportions of model grid cells covered by open water and ice less than 0.5 m thick. *Smith* [1988] suggests a value of 1.48x10^{-3} for a typical Arctic T_s-T_r difference of 20°C and a wind speed of 5 m s^{-1}. His values are representative of open ocean areas with a low, neutral stability transfer coefficient compared with the higher values observed for leads.

Two examples of methods to provide stability-adjusted coefficients that have been applied in one dimensional ice simulations include an approach that considers temperature differential only [*Rimsha and Donchenko,* 1957 (as applied by *Bauer and Martin* [1983]) and a second method that adjusts C_H based on a bulk Richardson number [*Ebert and Curry,* 1993; from *Louis,* 1979]. Using this latter method for T_r = -20°C and U_r=(3,5,10, and 15 m s^{-1}), C_{H10} is about 60% lower than the resulting AM coefficient at a 100 m fetch (1.19x10^{-3} compared with 1.98x10^{-3}). A relationship derived from lead observations and which is dependent on stability and winds [*Lindsay,* 1976] (as cited by *Smith et al.,* [1990]) yields C_H = 2.72x10^{3} at a 2-m reference height.

In most cases such differences can be attributed to whether data were acquired over open ocean or over leads. It is clear from these comparisons that variations in turbulent flux estimates due to the choice of parameterization can be as large as differences that arise due to uncertainties in fetch or ice thickness in leads. While the AM method allows inclusion of the additional physical characteristic of lead width within an ice model, these differences in available transfer coefficient parameterizations must be born in mind when choosing one parameterization scheme versus another. It is particularly important to consider whether the selected parameterizations are intended to apply to large, open water areas or to relatively small leads and polynyas.

Treatments of Lead Width Ensembles in Flux Estimates

The above comparisons address heat transfer as calculated for individual lead widths. However, the actual ice cover

Table 1. Monthly Averages of Downwelling Short-Wave Radiation F_r, Long-Wave Radiation F_L, Thick-Ice Turbulent Heat Flux H_i, Wind Speeds U_r, and Air Temperatures T_a used for Surface Energy Budget Calculations.

	Jan.	Feb.	Mar.	Apr.	May	Jun.	Jul.	Aug	Sep.	Oct.	Nov.	Dec.
F_L, W m^{-2}	168	166	166	191	244	289	309	301	267	223	184	174
F_r, W m^{-2}	0	0	39	162	281	302	223	142	60	8	0	0
H_i, W m^{-2}	-18.0	-13.2	-11.2	-4.5	6.0	6.4	5.2	6.0	2.9	-2.7	-8.6	-13.2
U_r, m s^{-1}	4.4	4.3	4.4	3.4	3.1	3.9	3.7	4.0	3.9	4.2	3.8	4.6
T_a, °C	-23.7	-24.9	-23.2	-18.6	-7.3	1.3	2.4	0.5	-5.0	-12.3	-18.0	-20.1

F_r, F_L, and H_i for thick ice (ice >= 0.8 m) are taken from *Maykut* [1982]. U_r are mean monthly geostrophic wind speeds scaled to approximate surface winds. T_a are mean monthly buoy temperatures estimated from daily gridded buoy data. U_r and T_a represent observations for January 21, 1979 to December 31, 1984 for a region centered at approximately 80° N, 155° W.

includes leads of varying width and ice thickness. As noted earlier, measuring these lead width distributions (e.g., identifying individual leads and lead widths) from remotely sensed data is considerably more difficult than estimating total lead-covered area for a model grid cell or region. Also, prescribing a single representative width to represent all open water/thin ice area within a grid element in a sea ice model is more efficient than calculating fluxes from a set of lead widths for each grid element. To help define the types of lead observations needed, we test methods of parameterizing width distributions. As discussed in the data section, this parameterization could take the form of a model distribution that approximates the true distribution of lead widths. The parameterization might also be some other measure such as the mean lead width.

To test parameterizations based on theoretical distributions, fluxes estimated using the observed lead width distributions from the sonar data were compared to fluxes estimated using negative exponential and lognormal models. In general, there are more small lead widths than expected with the negative exponential model, although approximately 25% of the observed distributions appear to fit this model at the 0.1 level of significance. Distributions for leads defined using different maximum drafts (e.g., different thresholds of ice thickness used to classify between refrozen leads and thick ice) vary significantly only when the number of leads in a category becomes too small to provide a reliable sample. Other studies provide similar descriptions of lead widths and distributions.

Fluxes estimated using the AM equations and the observed distributions, the observed distribution means, and distributions constructed using a theoretical negative exponential distribution and the observed means are compared in Table 2. Leads were treated as open water. In this comparison, two sets of observed distributions from the submarine sonar transects were used. First, an average flux for each of the two sets of observations was estimated by summing the fluxes from each of the individual leads (e.g., flux estimated using each individual lead width) in the observed distribution. Fluxes for each lead width category were weighted by the areal coverage of each category in the sonar observations. The resulting representative flux (given in the "Observed" column in Table 2) is taken to represent the "true" average flux, since it is estimated using the actual distribution of lead widths. In the second case a mean lead width was determined from each observed distribution, and a single flux value was calculated using this mean (the "Mean" column in Table 2). Finally, a negative exponential distribution was used to construct a lead width distribution from the observed mean width. From this constructed distribution, a representative flux (given in the "Distribution" column in Table 2) was then estimated as for the observed distribution. A flux was calculated at each lead width and then weighted by the number of leads in that width category as defined by the theoretical negative exponential distribution. Each set of calculations was done under a range of wind speeds and air temperatures to determine the sensitivity of the different lead treatments to meteorological conditions.

Table 2. Representative Sensible Heat Flux From Leads As Calculated Over All Lead Widths in Two Observed Lead-Width Distributions With Mean Widths of 11.7 m and 40.8 m, From Only the Means of the Two Observed Distributions, and From the Theoretical Negative-Exponential Distributions Constructed From the Observed Means.

		Observed Mean Lead Width of 11.7 m			Observed Mean Lead Width of 40.8 m		
Ts, °C	U_r, m s^{-1}	Observed	Mean	Distribution	Observed	Mean	Distribution
-10	5	133	133	133	126	124	125
-10	10	226	226	226	222	222	222
-20	5	328	327	328	296	278	289
-20	10	546	546	546	527	524	526

T_S is surface temperature; U_r is wind speed at reference height r. All other values are in watts per square meter.

As seen in Table 2, fluxes calculated from the model (negative exponential) distribution constructed from the observed mean, as well as fluxes estimated using only the mean, agree well with the fluxes summed over the actual distribution. The agreement would be closer if leads are refrozen, since as noted earlier, the change in C_H with changing fetch is less. Similar experiments done using other width distributions estimated using both the negative exponential and log normal distributions and a range of mean widths show that these results hold for different ice conditions. For fetch and stability adjustments in a large-scale model, it therefore appears acceptable to calculate fluxes with a mean lead width, rather than from the distribution. Since the availability of an observed lead width mean implies that a distribution exists from which to estimate the mean, calculations could be done using the observed distribution itself. However, time can be saved by calculating a single representative flux using a mean, rather than integrating over a full distribution at each time step.

Fetch Sensitivity Under Observed Conditions

The sensitivity tests discussed above shed light on the range of conditions where fetch and stability adjustments are likely to be most important and where parameterized width distributions are acceptable. It is also useful to test the relevance of these sensitivities using observed meteorological and sea ice conditions. Using the buoy temperatures and wind speeds described earlier, a series of daily adjusted bulk transfer coefficients and turbulent fluxes were calculated over a range of fetches for fluxes from open water and over thin ice. Figure 4 summarizes the change in monthly averaged H_s using a fixed C_{H10} of 3.0×10^{-3} compared to fetch and stability-adjusted C_{H10} values at fetches of 10, 50, and 100 m. Values are given for open water and 0.15-m ice. Monthly mean values for C_{H10} are given in Table 3. At a fetch of 10 m the difference in H_s due to the use of adjusted C_{H10} values, rather than a fixed C_{H10}, increases as ice thickness increases (i.e., as the decrease in instability with a lower

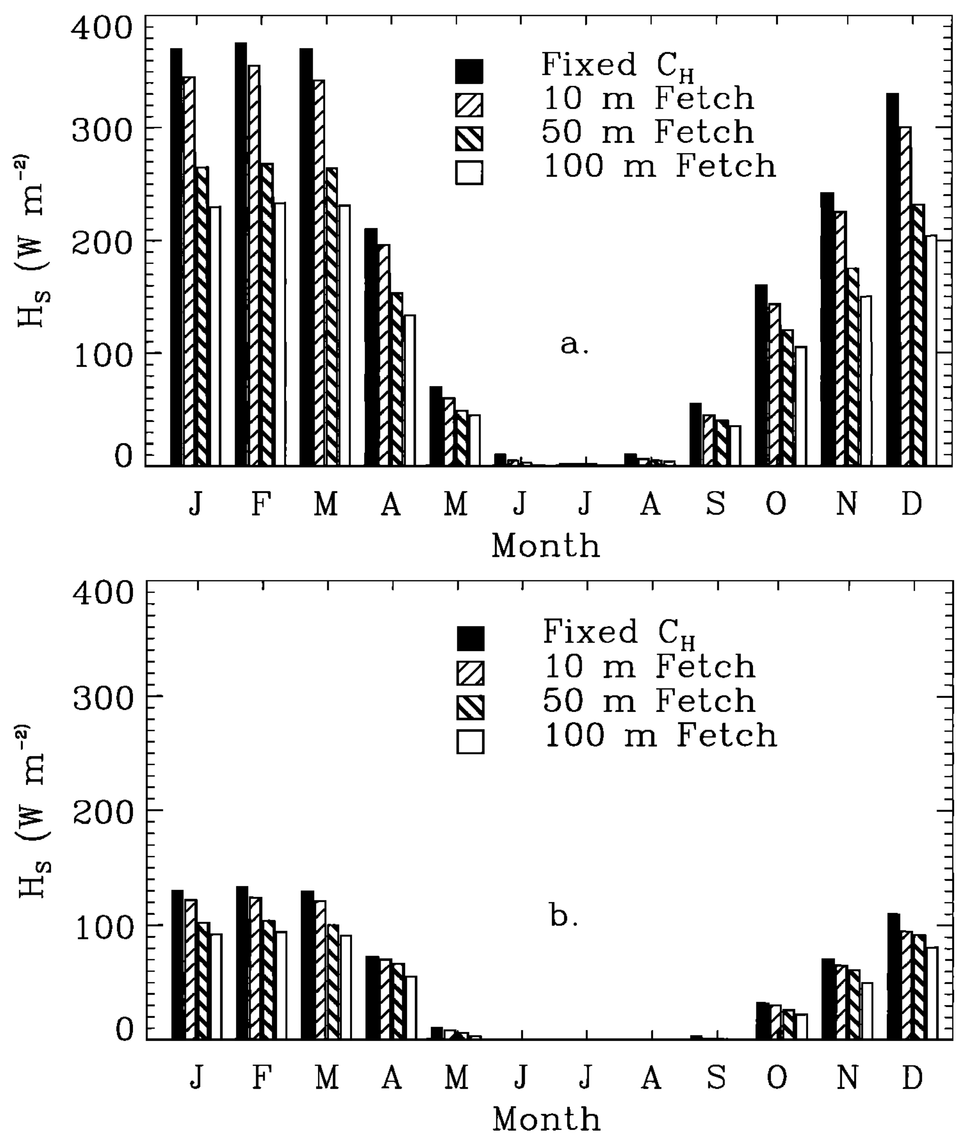

Figure 4. Change in monthly averaged H_s using a fixed C_{H10} of 3.0×10^{-3} compared to fetch- and stability-adjusted C_{H10} values at fetches of 10, 50, and 100 m, calculated using the daily buoy data and monthly mean radiative fluxes in Table 1. Values are given for (a) open water and (b) 0.15 m-thick ice.

Table 3. Mean Sensible Heat Transfer Coefficients by Month For Different Fetches and Lead Ice Thicknesses As Calculated Using Daily Buoy Temperatures and Wind Speeds Input Into an Energy Balance Model and the *Andreas and Murphy* [1986] Formulation for Fetch and Stability Correction

	Jan.	Feb.	Mar.	Apr.	May	Jun.	Jul.	Aug.	Sep.	Oct.	Nov.	Dec.
C_H for H_i=0 m; X=10 m	2.87	2.87	2.84	2.83	2.66	2.05	1.92	1.96	2.36	2.67	2.84	2.80
C_H for H_i=0 m; X=50 m	2.14	2.15	2.17	2.19	2.13	1.99	1.91	1.92	2.10	2.17	2.16	2.16
C_H for H_i=0 m; X=100 m	1.90	1.91	1.93	1.98	1.92	1.93	1.89	1.89	1.96	1.94	1.90	1.91
C_H for H_i=0.05; X=10	2.77	2.77	2.74	2.75	2.45	1.60	1.43	1.63	2.04	2.52	2.72	2.68
C_H for H_i=0.05; X=50	2.14	2.16	2.17	2.18	2.08	1.59	1.45	1.63	1.91	2.13	2.16	2.16
C_H for H_i=0.05; X=100	1.91	1.92	1.94	1.98	1.89	1.58	1.47	1.63	1.84	1.94	1.92	1.93
C_H for H_i =0.15; X=10	2.66	2.64	2.63	2.67	2.40	1.55	1.39	1.53	1.57	2.31	2.58	2.54
C_H for H_i=0.15; X=50	2.14	2.16	2.16	2.16	2.07	1.55	1.40	1.54	1.52	2.05	2.15	2.15
C_H for H_i=0.15; X=100	1.92	1.93	1.95	1.98	1.90	1.56	1.43	1.55	1.50	1.91	1.94	1.95
C_H for H_i=0.30; X=10	2.53	2.53	2.54	2.62	2.39	1.71	1.50	1.52	1.27	2.07	2.43	2.40
C_H for H_i=0.30; X=50	2.13	2.15	2.14	2.14	2.08	1.70	1.51	1.53	1.27	1.91	2.10	2.11
C_H for H_i=0.30; X=100	1.93	1.95	1.96	1.97	1.91	1.68	1.52	1.55	1.28	1.81	1.93	1.95

Coefficients C_H (x10^3) are given for fetches X of 10 m, 50 m, and 100 m, for ice thicknesses H_i of 0 m (open water), 0.05 m, 0.15 m, and 0.30 m. Monthly mean temperatures and wind speeds used are given in Table 1.

surface temperature at a fetch of 10 m and observed wind act to drop C_{H10} further below 3.0x10^{-3}). The change is a decrease in H_s of 7% for open water, 9% for 0.15-m ice, and 11% for 0.30-m ice in January. As the proportion of refrozen leads in a region increases relative to open water leads, using a fixed C_{H10} value will thus yield an increasing overestimation of H_s, although as shown below, the direction of this change reverses for thicker ice. At a thickness of 0.05 m, fluxes calculated using the fixed transfer coefficient and averaged over November-March are 8% greater than fluxes calculated with a fetch of 10 m, 21.5% greater than fluxes estimated with a 50-m fetch and about 29% greater than fluxes for a 100-m fetch.

Since the change in heat flux with increasing fetch is a function of ice thickness, it is instructive to extend these results to consider in more detail the actual ice thickness distributions within the pack. For this we again draw upon the thickness distribution data from the sonar transect described earlier, as well as the thickness distribution simulated by *Maykut* [1982] for the central Arctic, and apply the energy balance model and forcings in Table 1 (for January only). In the 150-km sonar transect, 2.6% of the transect length is made up of leads with ice thicknesses less than or equal to 0.8 m. The sonar data indicate that only 0.4% of this lead area is open water or very thin ice (0 to 0.05-m ice) while 87.6% of the ice in leads is between 0.6-0.8 m thick. Given this observed distribution of ice thicknesses within the leads of different widths, the average sensible flux from a lead was 30 W m^{-2} when a fixed coefficient of 3.0x10^{-3} was used, and 43 W m^{-2} using adjusted coefficients.

Since the stability-adjusted coefficients over a refrozen lead are typically less than 3.0x10^{-3}, this lower mean flux using the larger coefficient warrants some explanation. When a large transfer coefficient unadjusted for stability is used to estimate turbulent fluxes over relatively thick lead ice such as that shown in the sonar data, the surface temperature of lead ice is reduced to the point where the effect of the large transfer coefficient on the turbulent flux estimate is outweighed by the lower surface temperature. In other words, in the energy balance calculation the large transfer coefficient forces the lead surface temperature to be lower than is the case if a stability-adjusted coefficient is used. This decreases the temperature contrast between the lead and overlying colder air and thus reduces the estimated sensible heat flux.

The same results are found when the areal distribution of ice thicknesses estimated by *Maykut* [1982] are used to represent another sample of ice pack conditions. The energy balance calculations for thin ice are used as above with the forcings in Table 1, along with Maykut's sensible heat fluxes for ice thicker than 0.8 m, and applied to the full annual cycle. Comparison of fluxes estimated using a fixed coefficient of 3x10^{-3} and fetch and stability-adjusted coefficients (Table 4) show this same effect, since most of the thinner ice is between 0.4 and 0.8 m thick. In this case, adjusting C_{H10} for stability (e.g., using an adjusted coefficient instead of 3.0x10^{-3}) decreases the mean annual, areally averaged heat flux (summed over the mixture leads and thick ice) from -1.2 W m^{-2} to -1.6 W m^{-2} for a mean 10-m fetch. Increasing fetch from 10 m to 100 m reduces the areally averaged flux from -1.6 W m^{-2} to -1.9 W m^{-2}. In this example the overall effect of changing mean lead width from 10 to 100 m is small (about 0.3 W m^{-2}, below measurement accuracy for flux estimates), since most of the lead ice is relatively thick and the fetch dependency is thus less.

From Figure 4 and Tables 3 and 4 the effects of ice thickness in leads can be placed into a more general context for other ranges of ice thicknesses. For example, increasing mean fetch from 10 m to 100 m decreases open water flux by 34% during January (C_{H10} decreases by 34%) or about 110 W m^{-2}. For leads of 0.15-m thick ice, the increase in fetch decreases flux by 28% or about 30 W m^{-2}. The change in the areally-averaged flux (e.g., the combination of flux from leads and from thick ice, weighted by the fraction of leads and thick ice) can be estimated by multiplying this difference by the lead fractional coverage. If, for example, leads make up 2% of the mid winter ice pack and the leads consist of open water, then sensible heat flux averaged over the mixture of leads and thick ice decreases by about 2 W m^{-2} as fetch increases from 10 to 100 m. If leads are covered by 0.15-m ice, then the effect of a change in fetch drops substantially to about 0.6 W m^{-2}. Estimates of annual mean fluxes from an ice pack are considerably different when an ice thickness distribution is considered, rather than assuming, for example, a uniform ice thickness of 3.0 m [*Maykut*, 1982]. As shown here, the use of transfer coefficients more appropriate to refrozen leads accentuates this difference.

Table 4. Area-Weighted Sensible Heat Flux H_s and Transfer Coefficient C_{H10} Estimated Using 1979-1984 Buoy Wind Speeds and Temperatures Combined With Ice Thickness Distributions From *Maykut* [1982]

	Jan.	Feb.	Mar.	Apr.	May	Jun.	Jul.	Aug	Sep.	Oct.	Nov.	Dec.
H_S [a]	-11.6	-9.8	-6.4	0.3	7.3	6.0	4.0	5.7	2.9	-0.8	-4.3	-7.2
C_H [a]	1.83	1.80	1.80	1.83	1.90	1.87	1.96	2.04	2.01	1.86	1.85	1.87
H_S [b]	-12.3	-10.1	-6.8	-0.1	6.9	5.9	3.7	4.9	2.5	-1.2	-4.8	-8.0
C_H [b]	1.79	1.78	1.78	1.80	1.84	1.76	1.74	1.74	1.69	1.76	1.79	1.80
H_S [c]	-12.4	-10.3	-7.0	-0.5	6.7	5.9	3.7	4.8	2.5	-1.4	-5.1	-8.2
C_H [c]	1.77	1.77	1.77	1.77	1.79	1.76	1.74	1.73	1.68	1.75	1.77	1.78
H_S [d]	-12.6	-10.5	-7.2	-0.6	6.5	5.9	3.7	4.8	2.4	-1.4	-5.2	-8.4
C_H [d]	1.76	1.76	1.76	1.76	1.77	1.75	1.75	1.73	1.68	1.74	1.76	1.77

H_S values are in watts per square meter. C_{H10} are times 10^3. A fixed C_{H10} of 1.75x10^{-3} and *Maykut*'s [1982] thick ice fluxes were used for ice greater than 0.8 m.

[a] A fixed C_{H10} of 3.0x10^{-3} was used for ice less than or equal to 0.8 m (thin ice).

[b] An adjusted C_{H10} at 10 m fetch was used for thin ice.

[c] An adjusted C_{H10} at 50 m fetch was used for thin ice.

[d] An adjusted C_{H10} at 100 m fetch was used for thin ice.

Effects of Fetch Change in a Coupled, Near-Surface Atmosphere-Ice System

As pointed out in the introduction, sensitivity studies that vary surface fluxes should consider how changes in fluxes affect the overlying atmosphere. An increase in heat flux will warm the near-surface air and, in turn, reduce the heat loss from the lead. To assess the precise nature of such interactions requires detailed boundary layer modeling. However, we can gain an idea of the approximate effects of changing heat transfer through a change of fetch by employing the bulk similarity theory model introduced earlier [*Koch*, 1988]. This model was coupled with the AM parameterizations for C_H and applied to a range of fetches, wind speeds, and air temperatures as was done for the uncoupled AM calculations with January radiative forcings.

In these coupled calculations the equilibrium heat flux is essentially unchanged by a change in fetch; C_{H10} values increase with decreasing fetch, but with this higher transfer rate, air temperatures increase, so that the heat flux from lead to air is less. In this case then, it is more instructive to consider how the air temperature at the reference height relates to fetch (Figure 5). In these examples the increased heat transfer rate from narrower leads yields higher surface air temperatures over these narrow leads than is the case for wider leads. The effect is greatest at lower initial air temperatures and extends over longer fetches as wind speed increases, consistent with the earlier, uncoupled results. Where transfer coefficients are small (e.g., for longer fetches), the results indicate some error in the energy balance terms perhaps related to the fact that as the stability of the air mass changes, the assumptions inherent in the AM parameterization such as upwind temperature contrast may become less valid. Also, the range of wind speeds and temperatures used here to test the extent of fetch sensitivity is an extreme case. More comprehensive approaches are needed to test this further. However, the general pattern of higher air temperatures associated with narrower leads shown by this experiment is likely to be realistic. Since this warming will take place over time, inclusion of a boundary layer adjustment in a sea ice model ought to relate the change in air temperature to total heat loss in order to calculate ice growth in the lead.

Summary and Conclusions

In simulations of the polar ice cover, turbulent heat fluxes are typically estimated using bulk aerodynamic formulas. A variety of parameterizations suitable for large-scale ice models exist to relate these coefficients to atmospheric stability and to lead conditions, although in many cases a single fixed coefficient is chosen for fluxes from open water and thin ice. Here we investigate the sensitivity of parameterized transfer coefficients and sensible heat flux to atmospheric stability, lead width, lead width distribution, and ice thickness.

To calculate the sensitivity of turbulent fluxes to changes in fetch, wind speed, air temperature, and surface temperature, the formulations of *Andreas and Murphy* (AM) [1986] and *Andreas* [1987] were combined with an energy balance model, meteorological observations, and ice thickness data. Transfer coefficients for sensible heat typically decrease by about 50% (e.g., a 50% decrease in flux) as fetch increases from 10 m to 200 m for light to moderate winds (3 to 5 m s^{-1}) with a surface and air temperature difference of 20°C. The magnitude and rate of change are functions of temperature contrast and wind speed but the relationship decreases as wind speeds increase. In keeping with the decreased difference between surface temperature and air temperature as ice thickens in a lead, the change in rate of heat loss with fetch decreases substantially for ice thicker than about 0.3 m for typical Arctic winds and air temperatures. Comparisons of the AM-derived coefficients with other treatments used in ice modeling point out the large range in accepted parameterizations.

To test methods of parameterizing leads in ice models, fluxes estimated using approximations of observed lead width distributions were compared. Little difference is found between fluxes estimated using an observed lead width distribution, a mean lead width estimated from the observed distribution, and a theoretical lead width distribution. Thus for flux calculations within a large-scale model, where lead width distributions are not treated in the model, the mean alone is sufficient to parameterize transfer coefficients as a function of fetch for the conditions tested here. If lead distributions are simulated or available from observations, then it is sufficient to estimate a mean width and then calculate one representative flux, rather than calculating a flux for each width in the distribution.

To better assess the significance of the sensitivity calculations in terms of actual conditions in the Arctic, the parameterizations for fetch and stability were applied to lead statistics and a 5-year time series of buoy temperatures and winds. For leads consisting of open water or thin ice, fetch and stability-adjusted coefficients yield lower monthly mean fluxes than when a fixed coefficient of 3.0×10^{-3} is used. With adjusted coefficients this decrease in annual mean flux from lead-covered area ranges from 7% for open water leads to 11% for 0.3-m ice. For a central Arctic ice cover where the leads consist of relatively thick young ice this net change

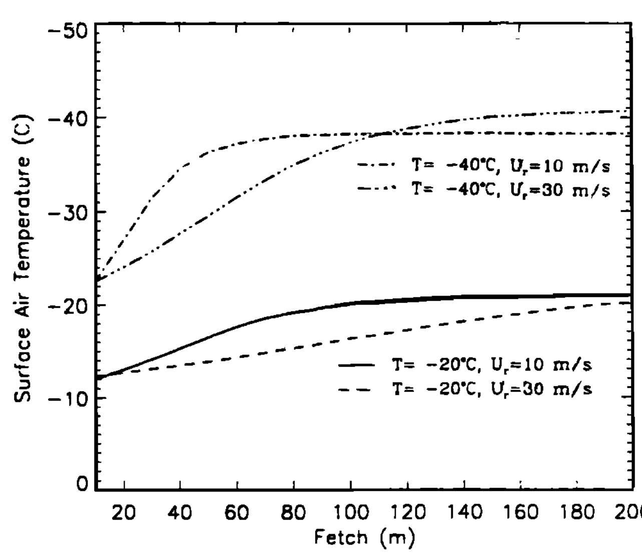

Figure 5. Simulated change in surface air temperature as a function of fetch, geostrophic air temperature T_g and geostrophic wind speed U_g. T_g = -20°C and U_g = 10 m s^1, T_g = -20°C and U_g = 30 m s^{-1}, T_g = -40°C and U_g = 10 m s^{-1}, T_g = -40°C and U_g = 30 m s^{-1} . Note that air temperature increases from bottom to top on the x axis.

in mean annual sensible heat flux from the pack (summed over leads and thick ice) is of the order of only 0.4 W m^{-2} when adjusted coefficients are used. The difference due to including a fetch dependency is about 0.3 W m^{-2} between fetches of 10 m and 100 m. If a greater proportion of the ice pack consists of open water or newly refrozen leads with thin ice (less than about 0.15-m thick), then the fetch dependency amounts to between 1 to 2 W m^{-2} averaged over a pack consisiting of 2% leads and 98% thick ice.

Calculations using the AM parameterizations coupled to a simple, bulk atmospheric boundary layer model point out the negative feedback between increased heat flux and increased air temperature. At equilibrium, air temperatures over narrower leads are considerably higher than those over wider leads, with the rate of change of temperature with fetch a strong function of wind speed. Narrower leads thus may have the potential to modify the overlying surface air layer more per unit area than will occur with wide leads.

In terms of applications to large-scale ice modeling, the sensitivity tests and calculations using observed data suggest that adjusting bulk transfer coefficients for stability is clearly warranted. Fetch has a smaller effect but can be significant under typical mid winter Arctic conditions of relatively light winds and low air temperatures, particularly when the model estimates fluxes separately for thick ice and open water leads. Fetch corrections become increasingly less important as the ice thickness in refrozen leads increases. In terms of prioritizing remote sensing and field programs, knowledge of factors such as radiative fluxes and percent lead area will likely outweigh uncertainties due to fetch, although a change in mean lead width might still affect areally averaged heat loss by about 1 to 2 W m^{-2}, depending on atmospheric conditions and distribution of ice thickness within the leads. Given the potential variability in heat flux from leads and since some rationale exists to suppose that a climatological change in ice thickness might affect lead width distributions, experiments are underway with coupled ice-atmosphere-ocean models to assess the effects of lead characteristics on ice growth and air temperatures.

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