Bulk Transfer Coefficients for Heat and Momentum over Leads and Polynyas

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ABSTRACT

Leads and polynyas are areas of open water surrounded by pack ice. In winter, when the polar oceans have extensive ice covers and the water–air temperature difference is typically 20°–40°C, they allow enormous amounts of sensible and latent heat to escape from the ocean to the atmosphere. Parameterizing these fluxes accurately is thus an important part of modeling the growth and decay of sea ice.

To develop a unified method for parameterizing the turbulent transfer from open water surrounded by pack ice, we have reanalyzed data reported in the literature on momentum and heat transfer over Arctic leads and polynyas. The neutral stability value of the 10-m drag coefficient, $C_{DN10} = 1.49 \times 10^{-3}$, is independent of wind speed and open-water fetch for winds from 1 to 10 m s^{-1} and fetches from 7 to 500 m. That value is slightly higher than values typical of the open ocean at these wind speeds, probably because of the form drag over the upwind ice or at the ice edges and because the wave field is still actively growing. We parameterize the neutral stability value of the 10-m transfer coefficient for sensible heat, C_{HN10}, with the nondimensional fetch X/L, where X is the over-water fetch and L is the Obukhov length. The C_{HN10} has a maximum value of 1.8×10^{-3} at very small $-X/L$ but decreases rapidly with increasing $-X/L$ to 1.0×10^{-3}—essentially the open ocean value. We find no compelling reason to believe that the bulk transfer coefficient for latent heat is different from C_{HN10}. The approach of C_{HN10} to its open ocean value at relatively short fetches (roughly 200 m) implies that horizontal homogeneity may not be as severe a constraint for evaluating scalar transfer coefficients as it is for some other applications of the flux-gradient relations.

The bulk transfer coefficients actually used in modeling turbulent transfer over leads and polynyas are derivable from our reported C_{DN10} and C_{HN10} values if the atmospheric stability is known. Lastly, we therefore develop a simple formula for estimating L from an easily obtainable bulk Richardson number.

1. Introduction

Leads are linear channels of open water that occur in an otherwise continuous cover of pack ice. Polynyas are broader, irregularly shaped areas of open water within pack ice. In winter, when the water–air temperature difference can be 20–40°C, leads and polynyas allow enormous amounts of heat to escape from the relatively warm ocean to the cold atmosphere. They are thus essential elements in any study of the polar heat budget and must be modeled accurately. Miyake (1965), Badgley (1966) and Shreffler (1975) did some of the early, primarily theoretical, work on turbulent transfer over leads.

Figure 1 shows a schematic diagram of a winter wind blowing over open water surrounded by sea ice. The large water–air temperature difference fosters a turbulent loss of sensible (H_s) and latent (H_L) heat from the water surface that can easily total 500 W m^{-2} (Andreas et al., 1979; Andreas, 1980; S. D. Smith et al., 1983). As this heat and moisture mix upward, they modify the vertical profiles of wind speed, temperature, and humidity over the water, and an internal boundary layer (IBL) develops. The warm surface has modified the air within the IBL; above the IBL, upwind profile and flux conditions persist. The height of the IBL, δ, depends on both upwind stability and stability conditions over the lead or polynya but is roughly proportional to $x^{0.4}$, where x is the over-water distance (Andreas et al., 1979). Because the IBL is shallow—δ is typically 2.5 m at the downwind side of a 20-m wide lead—the temperature gradient through it is severe. Andreas et al. (1979) measured numerous temperature profiles over relatively narrow leads and reported temperature differences of 20°C between the water surface and a height of 0.1 m. This temperature gradient, of course, decreases as the IBL deepens with increasing over-water distance. S. D. Smith et al. (1983) estimated that 400 m from the upwind edge of a polynya, the temperature gradient has weakened enough that the heat flux 4 m above the surface will be only 13% lower than the surface flux.

In trying to account for the fetch-limited nature of the flow over the narrow leads that they studied, Andreas et al. (1979) and Andreas (1980) parameterized the sensible heat flux from them as a Nusselt number,

$$N_H = \frac{H_s X}{D(T_s - T_2)}. \tag{1.1}$$

They then related this to a flow parameter, the fetch Reynolds number,

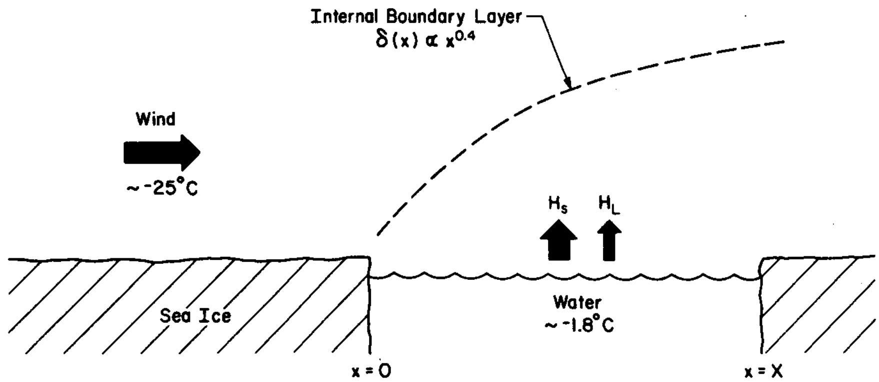

FIG. 1. Typical conditions over a lead or polynya in winter.

$$R_x = U_2X/\nu, \tag{1.2}$$

by

$$N_H = 0.08R_x^{0.76}. \tag{1.3}$$

In (1.1) and (1.2), X is the over-water fetch; D, the thermal diffusivity; T_s, the surface water temperature; T_2 and U_2, the upwind air temperature and wind speed 2 m above the ice; and ν, the kinematic viscosity of air. S. D. Smith et al. (1983) measured H_s over a polynya—which was at times as narrow as 210 m—and parameterized it with a bulk transfer coefficient. The neutral stability value of that coefficient, C_{HN}, was little different from its open ocean value and, consequently, indicated no fetch dependence.

We here try to resolve these differences in the lead and polynya datasets and to develop a unified model with which to predict the sensible and latent heat fluxes from any open water within pack ice. The model we devise predicts C_{HN} as a function of the fetch and the atmospheric stability; we then show how to estimate the stability easily from a bulk Richardson number.

We also investigate the drag coefficient over leads and polynyas. Measurements over the open ocean have shown that C_{DN10}, the neutral-stability drag coefficient referenced to 10 m, is a constant near 1.2×10^{-3} for wind speeds less than 11 m s^{-1} (Large and Pond, 1981)—speeds typical of ones we will discuss. For fetch-limited oceanic measurements, however, C_{DN10} is larger than this because the wave field is still actively developing and thus extracting additional momentum from the wind (Smith, 1980; Graf et al., 1984; Wu, 1985; Hsu, 1986). Generally, C_{DN10} is also larger than 1.2×10^{-3} over sea ice (Overland, 1985). Banke et al. (1980) found that because of the form drag due to pressure ridges, C_{DN10} could be as large as 2.1×10^{-3} over compact sea ice. In the marginal ice zone, where there is a high percentage of open water and many vertical faces to enhance the form drag, Andreas et al. (1984) observed C_{DN10} values as large as 4.0×10^{-3}. Because both these limited-fetch and form-drag effects are at work, we were not surprised to find that $C_{DN10} = 1.49 \times 10^{-3}$ over leads and polynyas.

2. Mathematical foundation

Because of the large water–air temperature difference, the air over leads and polynyas is quite unstable. Andreas et al. (1979) reported Obukhov lengths of −0.2 to −4 m over leads from 7 to 34 m in width; S. D. Smith et al. (1983) reported Obukhov lengths of −1 to −10 m for a polynya with a nominal fetch of 500 m. Thus, we will try to eliminate the effects of stability so we can compare the lead and polynya data, but also so we can compare both sets with measurements over the open ocean.

In horizontally homogeneous conditions, the vertical profiles of wind speed (U), potential temperature (T) and water vapor density (Q) in the atmospheric surface layer are (e.g., Large and Pond, 1982; Andreas and Makshtas, 1985)

$$U(z) = (u_*/k)[\ln(z/z_0) - \psi_m(z/L)], \tag{2.1}$$

$$T(z) = T_s + (t_*/k)[\ln(z/z_T) - \psi_h(z/L)], \tag{2.2}$$

$$Q(z) = Q_s + (q_*/k)[\ln(z/z_Q) - \psi_h(z/L)]. \tag{2.3}$$

Here, z is the height; k (=0.4), von Kármán's constant; z_0, the roughness length for velocity; z_T and z_Q, the so-called scalar roughness lengths; T_s and Q_s, the temperature and water vapor density at the surface; L, the Obukhov length; and ψ_m and ψ_h, semi-empirical stability functions. For unstable conditions ($\zeta = z/L < 0$), the only situation we need consider over leads and polynyas in winter, these stability functions are

$$\psi_m(\zeta) = 2\ln[(1+x)/2] + \ln[(1+x^2)/2] - 2\arctan(x) + \pi/2, \tag{2.4}$$

$$\psi_h(\zeta) = 2\ln[(1+x^2)/2], \tag{2.5}$$

where

$$x = (1 - 16\zeta)^{1/4}. \tag{2.6}$$

The u_*, t_*, and q_* in (2.1)–(2.3) are related to the surface stress (τ) and to the sensible (H_s) and latent (H_L) heat fluxes by

$$\tau = -\rho u_*^2, \quad (2.7)$$

$$H_s = -\rho c_p u_* t_*, \quad (2.8)$$

$$H_L = -L_v u_* q_*, \quad (2.9)$$

where ρ is the air density, c_p is the specific heat of air at constant pressure, and L_v is the latent heat of vaporization of water. These fluxes, in turn, yield the Obukhov length,

$$L = -[(g/\bar{T})(kH_s/\rho c_p u_*^3)]^{-1} = [(g/\bar{T})(kt_*/u_*^2)]^{-1}, \quad (2.10)$$

where g is the acceleration of gravity, and $\bar{T}$ is a kelvin temperature representative of the surface layer. A rigorous definition of L would include a term related to the latent heat flux (Zilitinkevich, 1966; Busch, 1973). We have estimated, however, that the term contributes only 1%–3% to (2.10) over leads and polynyas and is thus negligible.

Using the flux-gradient relations (2.1)–(2.3) to describe the flow over leads and polynyas may seem a dubious practice, because the surface is certainly not horizontally homogeneous. But Andreas et al. (1979) and S. D. Smith et al. (1983) derived believable and useful results with flux-gradient methods. Thus, perhaps the effects of the horizontal nonhomogeneity on measurements made near the surface are minimal. The analysis we report tends to substantiate this hypothesis and, therefore, suggests that the constraints of horizontal homogeneity on the flux-gradient relations are not always as severe as micrometeorologists have believed.

In modeling turbulent surface transfer, it is expedient to define bulk transfer coefficients as

$$\tau = -\rho C_{Dr} U_r^2, \quad (2.11)$$

$$H_s = \rho c_p C_{Hr} U_r (T_s - T_r), \quad (2.12)$$

$$H_L = L_v C_{Er} U_r (Q_s - Q_r). \quad (2.13)$$

Here, C_D is the drag coefficient, and C_H and C_E are the bulk transfer coefficients for sensible and latent heat, respectively. All three coefficients are functions of the height r at which the wind speed, air temperature, and humidity are measured; the subscript r denotes coefficients or measurements at this arbitrary reference height.

From (2.1)–(2.3), (2.7)–(2.9) and (2.11)–(2.13), C_{Dr}, C_{Hr} and C_{Er} are clearly stability dependent:

$$C_{Dr} = \frac{k^2}{[\ln(r/z_0) - \psi_m(r/L)]^2}, \quad (2.14)$$

$$C_{Hr} = \frac{k^2}{[\ln(r/z_0) - \psi_m(r/L)][\ln(r/z_T) - \psi_h(r/L)]}. \quad (2.15)$$

Because C_{Er} is identical to C_{Hr} when z_Q replaces z_T, we will henceforth ignore C_E and focus on C_D and C_H.

At neutral stability $r/L = 0$, the stability functions are both zero, and (2.14) and (2.15) reduce to

$$C_{\mathrm{DN}r} = \frac{k^2}{[\ln(r/z_0)]^2}, \quad (2.16)$$

$$C_{\mathrm{HN}r} = \frac{k^2}{[\ln(r/z_0)][\ln(r/z_T)]} = \frac{kC_{\mathrm{DN}r}^{1/2}}{\ln(r/z_T)},$$

$$= \frac{C_{\mathrm{DN}r}}{1 - k^{-1}C_{\mathrm{DN}r}^{1/2}\ln(z_T/z_0)}. \quad (2.17)$$

That is, knowing z_0 and z_T is equivalent to knowing the bulk transfer coefficients at neutral stability for any r.

From (2.14)–(2.17) we can also convert transfer coefficients obtained at height r to some other height, say, the standard reference height 10 m:

$$C_{\mathrm{DN}10} = \frac{k^2}{[kC_{Dr}^{-1/2} - \ln(r/10) + \psi_m(r/L)]^2}$$

$$= \frac{C_{Dr}}{\{1 - k^{-1}C_{Dr}^{1/2}[\ln(r/10) - \psi_m(r/L)]\}^2}, \quad (2.18)$$

$$C_{\mathrm{HN}10} = \frac{kC_{\mathrm{DN}10}^{1/2}}{kC_{Dr}^{1/2}C_{Hr}^{-1} - \ln(r/10) + \psi_h(r/L)}$$

$$= \frac{C_{Hr}}{(C_{Dr}/C_{\mathrm{DN}10})^{1/2} - k^{-1}C_{Hr}C_{\mathrm{DN}10}^{-1/2}[\ln(r/10) - \psi_h(r/L)]}. \quad (2.19)$$

3. Data

Andreas et al. (1979) tabulated values of the sensible heat flux measured over natural and artificial leads with fetches from 6.8 to 34 m. They used an integral or energy conservation method to obtain the fluxes. That is, by measuring the vertical profiles of wind speed and temperature at the upwind and downwind sides of leads, they determined the average surface heat flux from the difference in the heat content of the upwind and downwind air. Andreas and Paulson (1979) reported simultaneous eddy correlation measurements of u_* at heights generally 10–20 cm above the water surface. When a flow experiences a step change in surface roughness or in surface heat flux, the stress profile (u_*^2) changes with height through the internal boundary layer that develops over the new surface. But Taylor's (1970) numerical model for the development of an IBL following a step change in surface heat flux shows that, for their low measurement heights, the u_* values measured by Andreas and Paulson (1979) should be within a few percent of the surface value. Therefore, the lead study yields 29 pairs of simultaneous H_s and u_* values from which we can calculate C_{DN} and C_{HN}.

S. D. Smith et al. (1983) made eddy correlation measurements of H_s and u_* at a height of 4.4 m at the downwind side of a polynya in the Canadian Archipelago. They did not report u_*, but from their tabulated data—the wind speed at 4.4 m (U_4) and C_{D4}—we could compute it. Thus, from them we have 17 sets of simultaneous H_s and u_* values with which to compute C_{DN} and C_{HN} using the equations presented in the previous section.

Notice, the interpretation of H_s is slightly different for the two studies. For Andreas et al. (1979), H_s is the surface-averaged heat flux. In contrast, since S. D. Smith et al. (1983) measured H_s by eddy correlation techniques, theirs are, in a strict sense, point values. They argued, however, that because the surface heat flux decreases little with fetch after the first few meters of open water, their H_s values can also be thought of as surface averages.

4. Results

a. Drag coefficient

Generally, computing C_{Dr} from $(u_*/U_r)^2$ is not a statistically sound practice. Both u_* and U_r will have measurement errors; computing $(u_*/U_r)^2$ simply compounds the uncertainty in the C_{Dr} estimate. We have therefore chosen to compare u_* with U_{N2} (Fig. 2); U_{N2} is the neutral stability value of the 2-m wind speed, which from (2.1) is

$$U_{N2} = U_r - (u_*/k)[\ln(r/2) - \psi_m(r/L)]. \quad (4.1)$$

Andreas et al. (1979) tabulated U at 2 m, and S. D. Smith et al. (1983) tabulated U at 4.4 m. We choose here a 2-m rather than the standard 10-m reference level because, for the lead data, U_2 is given and, for the polynya data, interpolating to U_{N2} is more accurate than extrapolating to U_{N10}.

Figure 2 shows that the lead and polynya data agree well. No essential difference in the trends of the two sets is evident, nor is any fetch dependence obvious.

Comparing (4.1) with (2.18), we see that the slope of the data in Fig. 2 is $C_{DN2}^{1/2}$. When both dependent and independent variables have measurement errors, as u_* and U_{N2} do, least-squares linear regression underestimates the true slope of the data and overestimates its intercept (M. V. Smith et al., 1983). Hence, we computed two least-squares lines for the data in Fig. 2, one with U_{N2} as the independent variable and the other with u_* as the independent variable. The bisector of the two lines had an intercept that was virtually zero. Therefore, we recomputed the two least-squares lines, forcing the intercept to be zero. The bisector of these lines,

$$u_* = 0.0457\, U_{N2}, \quad (4.2)$$

is the line in Fig. 2. The neutral stability drag coefficient at 2 m is, therefore, $C_{DN2} = 2.09 \times 10^{-3}$, and the neutral stability drag coefficient at 10 m is $C_{DN10} = 1.49 \times 10^{-3}$ (from (2.18)).

This is somewhat larger than the value 1.0–1.3 ($\times 10^{-3}$) typically reported for the open ocean at these relatively low wind speeds (Smith and Banke, 1975; Smith, 1980; Wu, 1980; Large and Pond, 1981). As we discussed in the Introduction, however, higher C_{DN10} values are frequently reported in fetch-limited flows where the wave field is still developing. When the wave age becomes important to the dynamics, the dimen-

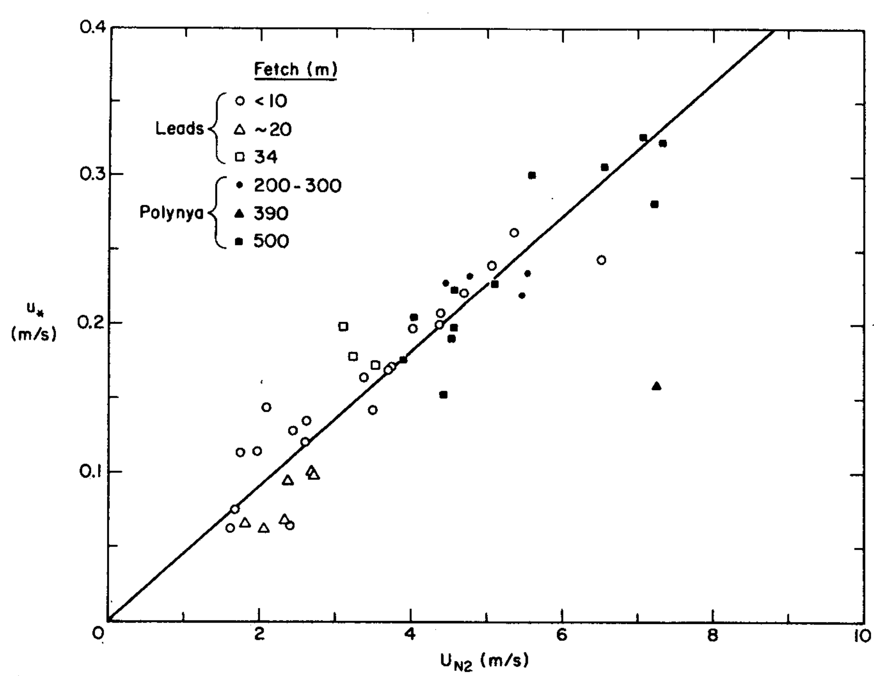

FIG. 2. The friction velocity u_* compared to the neutral stability wind speed at 2 m, U_{N2}. The line is (4.2); its slope is $C_{DN2}^{1/2}$.

sionless fetch $X_* = gX/u_*^2$ can be used to parameterize C_{DN10} (Charnock, 1958; Graf et al., 1984). A plot of C_{DN10} versus X_* (not shown), however, shows no trend in C_{DN10} with X_*—not surprising in view of Fig. 2. Our X_* range, 1.2×10^3 to 2.1×10^5, is probably too small to resolve any fetch effect on C_{DN10}.

Our high C_{DN10} value may also reflect the influence of the upwind ice or the step change in surface height at the upwind and downwind ice edges. Because of the form drag on pressure ridges, compact sea ice has been found to have C_{DN10} values as high as 2.2×10^{-3} (Banke et al., 1980; Shirasawa, 1981). But the highest 10-m drag coefficients measured anywhere over the ocean—values as high as 3–4 $\times 10^{-3}$—have been found in marginal ice zones, where there is a mix of water and ice (Smith et al., 1970; Andreas et al., 1984; Overland, 1985). Either the persistence of enhanced mixing due to the roughness of the upwind ice or mixing induced by the ice edges may, thus, also contribute to our relatively high over-water value of C_{DN10}. The absence of any discernible fetch dependence in C_{DN10} suggests that more than one process is controlling its value.

b. Sensible heat transfer coefficient

Over the ocean, in all but the lightest winds, heat and moisture are transferred at the interface by forced convection (Grachev and Panin, 1984); and the neutral stability transfer coefficients for sensible and latent heat, C_{HN10} and C_{EN10}, are near 1.0×10^{-3} (Friehe and Schmitt, 1976; Smith, 1980; Large and Pond, 1982). In contrast, over leads and polynyas, the surface temperature and humidity steps at the upwind edge can be extraordinarily large—typically 25°C for $T_s - T_r$ and 4×10^{-3} kg m^{-3} for $Q_s - Q_r$. The consequent instability will undoubtedly foster heat transfer by free convection until turbulent mixing erodes the vertical temperature gradient as the IBL develops downwind. Hence, because of this combined free and forced convection at short fetch, we would intuitively expect higher values of C_{HN10} and C_{EN10} over narrow open water areas. The comparison S. D. Smith et al. (1983) made of their polynya results with the lead data of Andreas et al. (1979) supports this conceptual picture. As the over-water fetch increases, the free-convection contribution to the heat flux will wane, and forced convection will become the dominant transfer mechanism. Consequently, we expect C_{HN10} and C_{EN10} to approach their oceanic values at long fetch.

Parameterizing the fetch effects is thus the key to modeling heat transfer from leads and polynyas. We need a length scale that fits our conceptual picture. The Obukhov length L fits. When H_s is large, $-L$ is small, and a lot of heat must be mixed through the IBL; free-convection transfer would therefore persist for relatively long fetches. On the other hand, when u_*^3—which is proportional to the production of turbulent kinetic energy—is large, $-L$ is large, and the turbulent mixing can rapidly diminish the temperature gradient driving the free convection. Free convection transfer would thus be confined to relatively short fetches. Such reasoning suggests scaling the fetch X with L.

Figure 3 shows that this scaling is successful. In it we plotted the neutral-stability transfer coefficients for sensible heat referenced to 10 m, derived from (2.19), versus $-X/L$. The lead and polynya data demonstrate the same trend; C_{HN10} is large for relatively short fetches—the effects of combined free and forced convection—but decreases to a constant value for longer

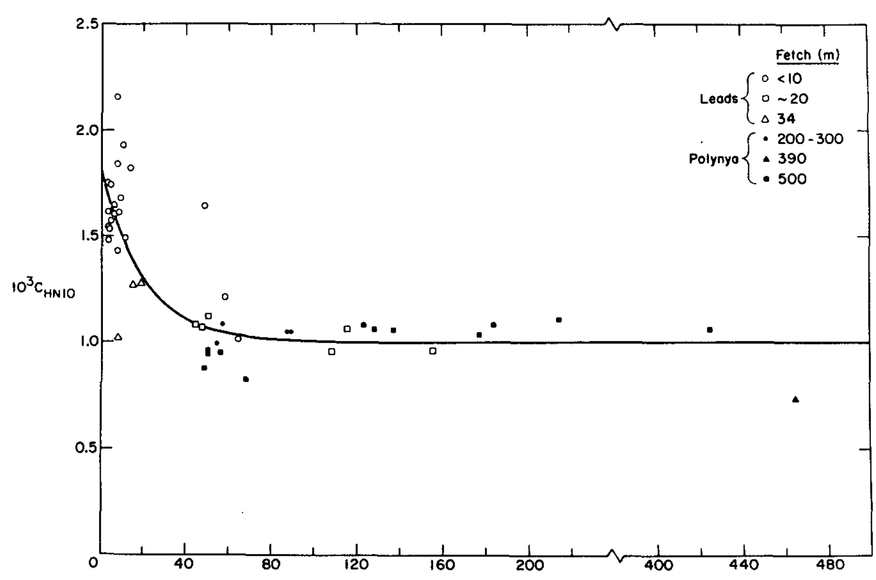

FIG. 3. The neutral stability transfer coefficient for sensible heat at 10 m plotted against the nondimensional fetch. The line is (4.3).

fetches—the forced convection regime. The line in Fig. 3, which we fitted by eye, is

$$10^3 C_{HN10} = 1.0 + 0.8 \exp[0.05(X/L)]. \quad (4.3)$$

This result is quite important. It shows that at a relatively short fetch—of order 200 m—C_{HN10} falls to a value typical of long fetches over the open ocean, 1.0×10^{-3}. Results from other observational studies (Dyer and Crawford, 1965) and from numerical models (Taylor, 1970, 1971; Rao et al., 1974) tend to corroborate this finding: When the roughness change is small, surface fluxes of sensible and latent heat following a step change in the surface value of the flux approach equilibrium at fairly short fetch. Therefore, parameterizing these fluxes in terms of bulk transfer coefficients is meaningful.

Alternatives to nondimensionalizing the fetch as X/L, of course, exist. For example, we could try the parameter $X_* = gX/u_*^2$ that we already mentioned. Not surprisingly, a plot of C_{HN10} versus X_* (not shown) looks a lot like Fig. 3, because $-X/L = (kt_*/\bar{T})X_*$. But X_* does not have the ready physical interpretation that X/L does. The X_* is a parameter tied to wave growth. Since the momentum flux is more closely related to the wave field than are the heat fluxes, and since for our data C_{DN10} shows no correlation with X_*, we see no justification for parameterizing C_{HN10} with X_*.

5. Estimating C_D, C_H and C_E

We must assume that models for momentum and heat transfer over leads and polynyas have as minimum input data U_r, T_r, Q_r, T_s and X, the open-water fetch. Q_s can then be obtained by assuming that air in contact with the water surface is saturated with vapor. The simplest way to estimate the momentum and sensible and latent heat fluxes is thus with (2.11)–(2.13), first estimating C_{Dr}, C_{Hr} and C_{Er}. Although C_{DN10} is a constant, C_{Dr} depends on the stability parameter L (see (2.18)]. Both C_{HN10} and C_{Hr} likewise depend on stability. Hence, we must first estimate L from what we know.

Deardorff (1968) and Irwin and Binkowski (1981) discussed the benefits of defining a bulk Richardson number referenced to height r,

$$\mathrm{Ri}_B = -(rg/\bar{T})[(T_s - T_r)/U_r^2]. \quad (5.1)$$

From (2.10) and (2.1)–(2.2) we can relate this to the stability parameter $\zeta = r/L$,

$$\zeta = \mathrm{Ri}_B \frac{[\ln(r/z_0) - \psi_m(\zeta)]^2}{\ln(r/z_T) - \psi_h(\zeta)}, \quad (5.2)$$

or, on substituting (2.16),

$$\zeta = \mathrm{Ri}_B \frac{[kC_{DNr}^{-1/2} - \psi_m(\zeta)]^2}{kC_{DNr}^{-1/2} - \ln(z_T/z_0) - \psi_h(\zeta)}. \quad (5.3)$$

Consequently, using (2.18) to compute C_{DNr} from the measured value of C_{DN10}, we can relate ζ and Ri_B through the function

$$f(r, \zeta, z_T/z_0) = \frac{[kC_{DNr}^{-1/2} - \psi_m(\zeta)]^2}{kC_{DNr}^{-1/2} - \ln(z_T/z_0) - \psi_h(\zeta)} \quad (5.4)$$

for arbitrary stability and reference height.

Figure 4 shows f for $C_{DN10} = 1.49 \times 10^{-3}$, $r = 10$ m, and for ψ_m and ψ_h given by (2.4)–(2.6). According to the figure, f depends only weakly on stability but strongly on z_T/z_0. With (2.17) and our result for C_{HN10}, (4.3), we can estimate $\ln(z_T/z_0)$; values are usually between -5 and 0. This range is also consistent with $\ln(z_T/z_0)$ values estimated over the ocean (Garrett and Hicks, 1973; Liu et al., 1979), over sea ice (Andreas, 1986), and over an arbitrary surface (Brutsaert, 1975b). Figure 4 therefore shows

$$\zeta \approx 8.0\,\mathrm{Ri}_B \quad (5.5)$$

for a 10-m reference height over leads and polynyas. The relationships between ζ and Ri_B given by Deardorff (1968) and Barker and Baxter (1975), both based on the assumption that $z_T = z_0$, would yield values for the constant in (5.5) of 10.4 and 11.2, respectively, for $C_{DN10} = 1.49 \times 10^{-3}$. Figure 4 shows that we would also predict a value near 10 if $z_T = z_0$.

Because C_{Dr} is a function of the reference height, f is also. Hence, the constant, 8.0, in (5.5) decreases as the reference height decreases. For reference heights less than 10 m and values of $\ln(z_T/z_0)$ near -2,

$$\zeta \approx 8.0(0.65 + 0.079r - 0.0043r^2)\,\mathrm{Ri}_B, \quad (5.6)$$

where r is in meters. We see that the height-dependent polynomial in (5.6) represents only about a 30% change in the multiplicative constant when r decreases from 10 m to 1 m. Thus, in some applications (5.5) might be accurate enough even for heights other than 10 m.

With (5.6), (4.3) and $C_{DN10} = 1.49 \times 10^{-3}$, it is now relatively easy to compute C_{Dr} and C_{Hr} for arbitrary reference height, stability and open-water fetch. From (2.14)–(2.17), the appropriate equations are

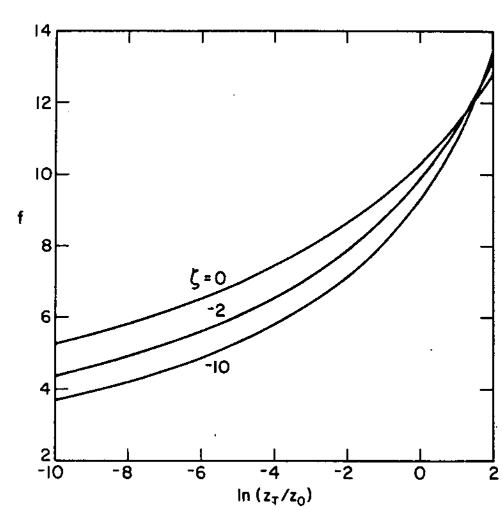

FIG. 4. A plot of the dimensionless function f, (5.4), that relates the stability parameters ζ and Ri_B. Since $\zeta = 10/L$, the plot is for $L = -1, -5$ and $-\infty$ m.

$$C_{Dr}=\frac{C_{DN10}}{\{1+k^{-1}C_{DN10}^{1/2}[\ln(r/10)-\psi_m(r/L)]\}^2}, \quad (5.7)$$

$$C_{Hr}=\frac{C_{HN10}(C_{Dr}/C_{DN10})^{1/2}}{1+k^{-1}C_{HN10}C_{DN10}^{-1/2}[\ln(r/10)-\psi_h(r/L)]}. \quad (5.8)$$

Measurements of the bulk transfer coefficient for latent heat, C_E, and of C_H over water are rarely precise enough to establish a difference between the two values (e.g., Large and Pond, 1982; Smith and Anderson, 1984; Blanc, 1985). Hence, as a first model, we could take $C_E = C_H$. Andreas' (1980) analysis, however, suggested that over leads the sensible and latent heat flux Nusselt numbers are equal; the bulk transfer coefficients thus would obey

$$C_E=(\mathrm{Pr}/\mathrm{Sc})C_H=(D_w/D)C_H, \quad (5.9)$$

where Pr (=0.71) is the Prandtl number, Sc (=0.63) is the Schmidt number, D (=0.183 cm^2 s^{-1}) is the thermal diffusivity, and D_w (=0.208 cm^2 s^{-1}) is the molecular diffusivity of water vapor. Consequently, C_E would be about 13% larger than C_H. Friehe and Schmitt (1976) similarly suggested that over the ocean C_E should be 16% larger than C_H, again because of the difference in molecular diffusivities. Theoretical models also predict that C_E should be larger than C_H regardless of the surface (Brutsaert, 1975a; Liu et al., 1979; Andreas, 1986); that predicted difference, however, is only a few percent, not 13%–16%. In light of the experimental uncertainties involved in measuring C_H and C_E and the minor differences predicted by theoretical models, we feel that

$$C_{Er}=C_{Hr} \quad (5.10)$$

is currently the best model for the latent heat transfer coefficient over leads and polynyas.

We have focused on bulk transfer coefficients referenced to 10 m primarily because we wanted to compare the values over leads and polynyas with values obtained over the open ocean and over sea ice. In practice, however, a 10-m reference height may not be the best choice for estimating fluxes from open water within pack ice. Because the internal boundary layer that develops will generally be quite shallow—typically 2 m over narrow leads—(2.1)–(2.3) may not accurately model the U, T and Q profiles for all heights between the surface and 10 m. Hence, although we can formally compute the 10-m transfer coefficients, measuring U_r, T_r and Q_r at a lower reference height will usually lead to more accurate flux estimates.

The Appendix demonstrates a sample calculation of the momentum, heat and moisture transfer at the surface of a 100-m wide lead based on our results and using the equations developed in this section.

6. Conclusions

We have reanalyzed momentum and heat transfer data tabulated in the literature in an attempt to derive a unified model for the turbulent transfer from open water surrounded by pack ice. When we corrected measured bulk transfer coefficients for stability, the data from narrow leads (7 to 34 m across) and from a much broader polynya (200 to 500 m fetch) were in good agreement.

We could therefore derive the drag coefficient for open water surrounded by ice. The value for neutral stability, referenced to 10 m, $C_{DN10} = 1.49 \times 10^{-3}$, had no obvious wind speed or fetch dependence for winds of 1–10 m s^{-1} and fetches from 7 to 500 m. This value is somewhat higher than values measured over the open ocean at similar wind speeds, probably because of the combined effects of wave growth and the form drag associated with the upwind ice or the ice edges. This drag coefficient—rather than one derived from measurements over the open ocean—is thus the appropriate one to use for modeling processes in leads and polynyas that depend on the surface stress, such as grease ice formation (Bauer and Martin, 1983).

For small values of the dimensionless fetch $-X/L$, the neutral stability value of the sensible heat transfer coefficient decreases rapidly with increasing $-X/L$ from a maximum value of $C_{HN10} = 1.8 \times 10^{-3}$. As $-X/L$ continues increasing, C_{HN10} levels out at 1.0×10^{-3}—essentially the open ocean value—at a surprisingly short fetch of about 200 m. We explain this behavior in terms of the turbulent transfer regime. Free convection, driven by the large surface temperature jump, combines with forced convection at short nondimensional fetch to enhance the heat transfer. At longer nondimensional fetch, the temperature gradient erodes, and forced convection—the primary transfer mode over the ocean—dominates. The behavior of C_{HN10} implies two conclusions. First, for similar meteorological conditions, narrow leads can lose much more heat per unit surface area than more expansive open water areas, as S. D. Smith et al. (1983) found when they compared their polynya results with measurements over leads. Second, for some applications of the flux-gradient relations, the requirement for horizontal homogeneity may not be as crucial as micrometeorologists have believed.

The bulk transfer coefficients required for modeling turbulent fluxes over leads and polynyas necessarily depend on stability. We, thus, developed a method for estimating the Obukhov length L from an easily obtainable bulk Richardson number. With this stability information, it is possible to convert C_{DN10} and C_{HN10} to their corresponding stability-dependent values for any arbitrary reference height. The Appendix shows a sample calculation.

Finally, we reviewed experimental and theoretical values for C_E, the latent heat transfer coefficient, and concluded that no compelling evidence refuted the simplest model, $C_E = C_H$, for leads and polynyas.

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APPENDIX

Sample Calculation of the Turbulent Fluxes

Suppose we have, from observations or from a numerical model, the velocity, temperature and relative humidity (RH) at a 2-m reference height upwind of a 100-m wide lead or polynya. Suppose also that we know the surface water temperature and salinity (S). Table A1 shows the steps in calculating the bulk transfer coefficients and the turbulent fluxes from this information.

To find Q_s, we assumed that the air at the water surface was in saturation with water of temperature T_s and salinity S. The air density and the constants c_p and L_v are functions of temperature; we evaluated them at T_s; ρ also depends on the atmospheric pressure, which we took as 1000 hPa.

TABLE A1. Steps in computing the turbulent fluxes over a lead or polynya, given standard meteorological data.

Given or Observed Conditions:	
$X = 100\,\mathrm{m}$	
$U_2 = 5.0\,\mathrm{m\,s^{-1}}$	
$T_2 = -25.0°\mathrm{C}$	
$\mathrm{RH}^2 = 75\%$	
$T_s = -1.8°\mathrm{C}$	
$S = 35‰$	
Derived Conditions:	
$\bar{T} = \frac{1}{2}(T_s + T_2) = 259.75\,\mathrm{K}$	
$\rho = 1.284\,\mathrm{kg\,m^{-3}}$	
$Q_2 = 0.416 \times 10^{-3}\,\mathrm{kg\,m^{-3}}$	
$Q_s = 4.215 \times 10^{-3}\,\mathrm{kg\,m^{-3}}$	
Constants:	
$g = 9.827\,\mathrm{m\,s^{-2}}$	
$c_p = 1.005 \times 10^3\,\mathrm{J\,kg^{-1}\,K^{-1}}$	
$L_v = 25.04 \times 10^5\,\mathrm{J\,kg^{-1}}$	
Stability Effects:	
$\mathrm{Ri}_B = -0.070$	from (5.1)
$\zeta = 2/L = -0.444$	from (5.6)
$L = -4.5\,\mathrm{m}$	
$X/L = -22.2$	
$\psi_m(\zeta) = 0.744$	from (2.6) and (2.4)
$\psi_h(\zeta) = 1.308$	from (2.6) and (2.5)
Transfer Coefficients:	
$C_{\mathrm{DN10}} = 1.49 \times 10^{-3}$	
$C_{D2} = 2.49 \times 10^{-3}$	from (5.7)
$C_{\mathrm{HN10}} = 1.26 \times 10^{-3}$	from (4.3)
$C_{H2} = C_{E2} = 2.14 \times 10^{-3}$	from (5.8) and (5.10)
Fluxes:	
$\tau = 0.080\,\mathrm{N\,m^{-2}}$	from (2.11)
$u_* = 0.250\,\mathrm{m\,s^{-1}}$	from (2.7)
$H_s = 320\,\mathrm{W\,m^{-2}}$	from (2.12)
$H_L = 102\,\mathrm{W\,m^{-2}}$	from (2.13)

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