Normalizing Air–Sea Flux Coefficients for Horizontal Homogeneity, Stationarity, and Neutral Stratification

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(Manuscript received 25 November 2009, in final form 19 April 2010)

ABSTRACT

Monin–Obukhov similarity (MOS) theory is routinely applied over the ocean to describe surface layer profiles of wind speed, temperature, and gas concentrations. Using this theory, fluxes are in turn estimated based on the best available parameterizations of normalized flux coefficients: for example, neutral flux coefficients. Flux coefficients can vary with environmental conditions. Because it is generally assumed that the domain of interest must be characterized by spatially homogeneous and steady-state conditions, systematic violations of the assumptions may lead to significant uncertainties in flux estimates. In this paper, the author has extended MOS theory to accommodate nonstationarity and spatial inhomogeneity in the representation of the normalized drag coefficient, Stanton number, and Dalton number. The author illustrates the importance of his theoretical extension, based on a reexamination of a historical air–sea interaction dataset obtained from the North Sea.

1. Introduction

For more than a half century, air–sea interaction experimentalists have employed the bulk aerodynamic method in compiling estimates of momentum, heat, and water vapor fluxes. Flux estimates are applied in turn to numerical models of oceanic and/or atmospheric circulations, wave state, waveguide prediction, and loads on engineered infrastructures. More recently, flux estimates have served to support major remote sensing programs: for example, where microwave signatures of the surface are inverted to produce wind fields and/or spatial patterns of heat exchange.

The more commonly used coefficients include the drag coefficient for momentum, Stanton number for sensible heat flux, and Dalton number for latent heat flux. These coefficients, in turn, must be parameterized based on individual field datasets and/or aggregated datasets representing widely varying environmental conditions. In spite of a large number of field studies, one finds large differences between flux coefficient parameterizations. Environmental variations associated with, for example, upwind fetch, sea state, spatial variability, and temporal variability have been hypothesized as causes of the differences between parameterizations (see, e.g., Geernaert et al. 1986).

With a goal to produce flux estimates with as small an uncertainty as possible, it is common practice to remove differences between flux parameterizations by setting reference conditions (e.g., standard height of 10-m altitude and von Kármán constant of 0.4) and normalizing the flux coefficients to idealized environmental conditions. Idealized conditions that may be part of a flux coefficient normalization process may include, for example, neutral stratifications, spatially homogeneous conditions, steady state, idealized wave state, etc. At the present, the only widely accepted normalization is for neutral stratifications: that is, the neutral drag coefficient, neutral Stanton number, and neutral Dalton number. Although normalization for neutral stratifications can significantly reduce uncertainties in the flux parameterizations, other environmental variabilities remain as possible sources of uncertainty and difference between flux coefficient parameterizations derived from different sites.

In this study, we build on the results reported in Geernaert (2002) and investigate the roles of both nonstationarity and inhomogeneity as potential sources of uncertainty and bias, in existing as flux coefficient

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DOI: 10.1175/2010JPO4407.1

parameterizations. We herein derive an extended representation of the normalized drag coefficient, normalized Stanton number, and normalized Dalton number, so that estimates can be adjusted for variations of not just atmospheric stratification but also spatial inhomogeneity and nonstationarity.

In the next section, a brief outline of the similarity theory and its fundamental assumptions is presented. This is followed in section 3 with a derivation of the flux profile relation for momentum and the normalized drag coefficient, where one allows for variations of atmospheric stratification, spatial inhomogeneity, and nonstationarity. In section 4, we derive similar equations for the Stanton and Dalton numbers. In section 5, we summarize the potential importance of these new results.

2. Surface layer profiles and fluxes: A brief review

Monin–Obukhov similarity (MOS) theory was conceived in the 1950s based on theoretical developments over the previous six decades that tackled the science of eddy diffusion, mixing length theory, and turbulence similarity (see review in Panofsky and Dutton 1984). Outlined by Monin and Obukhov (1954), MOS theory assumes a priori that for stationary conditions and for horizontal homogeneity of all state variables that act as indicators of surface layer and interfacial processes, one may apply an equation of the form

$$\partial X/\partial z = (\langle w'x'\rangle/u^*kz)\phi_x(z/L), \qquad (1)$$

where the quantities X and x' represent the mean and fluctuating parameter of interest, which in turn may be, for example, the horizontal component of the wind speed in the along-wind direction U and u'; temperature T and T'; specific humidity q and q'; or any other trace gas concentration c and c'. In (1), $(u^*)^2 = \langle -u'w'\rangle$, k is the von Kármán constant (=0.4), and $\phi_x(z/L)$ is the nondimensional vertical gradient expressed in terms of the measurement height z and Monin–Obukhov length L. The reader is referred to Panofsky and Dutton (1984) for a complete treatment of the theory leading to Eq. (1) and definition of L.

Equation (1) is assumed to be valid within the turbulent surface layer where the constant flux layer assumption is invoked, thus implying that all environmental quantities that are involved as indicators of governing processes must exhibit no spatial or temporal variability in either the downwind (x) or crosswind (y) directions: that is,

$$\partial/\partial(x, y, t)(U, T, q, z_0, c, c_0, z_c, z_i, T_0, U_0) = 0. \qquad (2)$$

Variations in only the vertical direction are allowed. In (2), the terms in brackets are the wind speed U, air temperature T, humidity q, roughness length z_0, gas concentration c, surface gas concentration c_0, roughness length associated with the scalar (temperature or gas) z_c, mixed layer depth z_i, surface temperature T_0, and surface current if on the ocean U_0.

The treatment of the surface layer as a "constant flux layer" and the a priori assumption of horizontal homogeneity allows one to integrate Eq. (1) into a form which is useful to a wide variety of applications: that is,

$$X_{(z)} - X_0 = (\langle w'x'\rangle/u^*k)[\ln(z/z_{ox}) - \Psi_x], \qquad (3)$$

where $X_{(z)}$ and X_0 are the bulk average values of a given quantity at height z and at the height of a roughness length z_0, respectively. The roughness length (with subscript x referring to mean quantity X) in (3) corresponds to the height immediately above the surface where $U = U_0$, $T = T_0$, and $c = c_0$. The roughness length has a value on the order of 10^{-4} to 10^{-2} m over the ocean. The stratification function Ψ_x is the integrated form of ϕ_x; this function is defined to be zero for neutral conditions and positive (negative) for unstable (stable) conditions (see Businger et al. 1971).

Because profiles of bulk quantities (e.g., wind speed, temperature, and trace gases) may be measured on towers, buoys, and masts, Eq. (3) may be rearranged to give estimates of surface fluxes, roughness lengths, and exchange coefficients for a variety of environmental conditions. For example, when $X = U$, the momentum flux normalized by density may be represented as

$$u^{*2} = (kz)^2(\partial U/\partial z) = C_D(U_z - U_0)^2, \qquad (4)$$

where C_D is the drag coefficient, which in turn is a function of a variety of site-dependent quantities that are either constants (such as the von Kármán constant) or associated with reference values (measurement height of 10 m),

$$C_D = k/(\ln z/z_0 - \Psi_U)^2. \qquad (5)$$

Analogous to the momentum flux, the fluxes of temperature and water vapor may be represented, respectively, with bulk aerodynamic relations,

$$\langle -w'T'\rangle = C_H U_z(T_z - T_0) \quad \text{and} \qquad (6)$$

$$\langle -w'q'\rangle = C_E U_z(q_z - q_0). \qquad (7)$$

In (6) and (7), the quantities T and q are temperature and specific humidity, respectively, and the temperature and humidity fluxes are by convention positively upward (in contrast to momentum flux, which is positive downward); C_H and C_E are referred to as the Stanton and Dalton numbers. In applying the bulk aerodynamic method, as shown in (4), (6), and (7), the scientific community generally uses a value of the von Kármán constant of 0.40, and a value for z is routinely chosen as 10 m above mean sea level.

To produce user-friendly parameterizations with greater certainty, experimentalists have been computing and parameterizing values of the neutral drag coefficient C_{DN}, representative of the correspondingly neutral conditions,

$$C_{\mathrm{DN}}^{-1/2} = C_D^{-1/2} + k^{-1}\Psi_U. \quad (8)$$

One will find similar expressions that relate the Stanton and Dalton numbers to their neutral stratification counterparts,

$$C_{\mathrm{HN}}^{-1/2} = C_H^{-1/2}[1 + \Psi_H/(kC_D^{1/2})] \quad \text{and} \quad (9)$$

$$C_{\mathrm{EN}}^{-1/2} = C_E^{-1/2}[1 + \Psi_E/(kC_D^{1/2})] \quad (10)$$

In the next section, the assumptions of horizontal homogeneity and stationarity are relaxed to explore the roles of spatial variability and nonstationarity on flux coefficient parameterizations. We begin with the wind speed profile and momentum flux in the next section and then extend the discussion to temperature and gas profile relations in section 4.

3. Derivation of the drag coefficient for quasi-inhomogeneous and quasi-stationary conditions

a. Flux profile relations with spatial inhomogeneity and nonstationarity

We begin by applying the momentum budget to the downwind direction,

$$\begin{aligned} dU/dt &= \partial U/\partial t + U\partial U/\partial x + W\partial U/\partial z + \partial\langle u'w'\rangle/\partial z \\ &= -\rho^{-1}\partial p/\partial x + fV, \end{aligned} \quad (11)$$

where ρ is air density, p is the atmospheric pressure, f is the Coriolis parameter ($= 2\Omega\sin\theta$, where Ω is the earth's rotation rate and θ is latitude), V is the crosswind velocity relative to surface layer coordinates, and W is the vertical velocity. We assume that the term $W\partial U/\partial z$ may be nonzero, particularly over the coastal ocean where, for example, sea breeze circulations can significantly influence the flow field. By defining the coordinate system such that the crosswind velocity in the surface layer is zero, the geostrophic balance may now be combined with (7), thus leading to

$$\partial U/\partial t + U\partial U/\partial x + W\partial U/\partial z + fV_g = \partial\langle -u'w'\rangle/\partial z. \quad (12)$$

In (12), V_g represents the crosswind geostrophic wind velocity, with respect to the coordinate system defined by the surface layer flow. Because the flux divergence denoted in (12) is related, in part, to the sum of the horizontal gradient of mean momentum, alternatively represented as $\partial(U^2/2)/\partial x$, and the local time rate of change of wind speed, one may hypothesize that horizontal along-wind gradients of wind speed, stability, and roughness, along with terms representing vertical advection, crosswind component of the geostrophic velocity, and $\partial U/\partial t$ are responsible for creating a local vertical flux divergence in the surface layer.

To explore the roles of systematic horizontal inhomogeneity, vertical advection, and nonstationarity of state variables on the parametric form of the wind profile and recalling that $u^{*2} = \langle -u'w'\rangle$, we rewrite Eq. (4) as

$$\langle -u'w'\rangle = k^2U^2[\ln(z/z_0) - \Psi_U]^{-2}. \quad (13)$$

Although Eq. (13) is defined to be relevant only for a constant flux surface layer, we know that, according to (12), the constant flux layer is at most an approximation. In fact, experimental data from many field campaigns have shown that the surface layer often exhibits substantial flux divergence and/or spatial variability of flux (see, e.g., Vogel and Crawford 1999; Mahrt 1999; Wilczak et al. 1999; Mahrt et al. 2001; Geernaert 2007); however, most experimentalists still opt to apply (13) to develop practical coefficients for numerous applications. We will proceed here with the assumption that weak violations of the constant flux layer hypothesis are permissible with MOS theory.

Letting the surface layer be a nearly constant flux layer, versus an absolutely constant flux layer as in classical MOS theory, we combine (12) and (13) to yield the following:

$$\begin{aligned} &\partial U/\partial t + U\partial U/\partial x + W\partial U/\partial z + fV_g \\ &\quad = \partial/\partial z[k^2U^2(\ln z/z_0 - \Psi_U)^{-2}]. \end{aligned} \quad (14)$$

In this regard, we have invoked the assumption that one can differentiate the rhs of Eq. (13) with spatially varying quantities, even though (13) was originally assumed to represent a surface layer with little or no vertical flux divergence. A physical justification of this step using flux footprint theory is described in detail in Geernaert (2002).

The combination of (13) and (12) has deeper implications on the applications and/or limitations of similarity theory. As a necessary but not sufficient condition, this assumption requires that nonstationarity and advection be sufficiently small that the turbulence remains in equilibrium with the mean vertical gradients. Given that there is a practical benefit with this assumption that allows a wider set of surface layer flux processes to be normalized, we will proceed with this analysis but strongly recommend that experimental studies be conducted in the future to determine limits on allowable degrees of nonstationarity and advection so that similarity theory remains appropriate for flux estimates.

Carrying out the differentiation in (14) where all parameters are variables and using simple algebra to rearrange the terms, we obtain the following:

$$\partial U/\partial z = u^*/kz(1 - R - S^*) + (U^2/2u^{*2})\partial U/\partial x + (WU/2u^{*2})\partial U/\partial z + (U/2u^{*2})\partial U/\partial t + (U/2u^{*2})fV_g, \quad (15)$$

where $R = (z/z_0)\partial z_0/\partial z$ and $S^* = z\partial\Psi_U/\partial z$. Given that Ψ_U may be approximated as $-\beta z/L$ for near-neutral stratifications (see Businger et al. 1971), the term S^* may be rewritten as

$$S^* = -\beta z/L + \beta(z/L)^2\partial L/\partial z. \quad (16)$$

In addition, one may write the approximation $\phi_u = 1 + \beta z/L$; using measurements reported in Businger et al. (1971), we make a simple approximation for β to be around 3 for unstable stratifications and 5 for stable stratifications.

Noting that W may be derived from the incompressible mass balance equation (i.e., such that $\partial U/\partial x + \partial V/\partial y + \partial W/\partial z = 0$), we will hereafter approximate W to be $-z(\partial U/\partial x + \partial V/\partial y)$. We recognize that in most oceanic environments $|\partial V/\partial y| \ll |\partial U/\partial x|$; however, in many sea breeze domains, there is often significant mesoscale divergence where $\partial V/\partial y$ may be important. Therefore, because W will at the least be dominated by $\partial U/\partial x$, we will simplify the derivation by rewriting the expression as $W = -\xi z\partial U/\partial x$, where $\xi = (\partial U/\partial x + \partial V/\partial y)/(\partial U/\partial x)$. The reader should note that we include the lateral velocity gradient in particular because many tower measurements of air–sea fluxes are within sea breeze domains, which in turn are influenced by complex coastal topography and a divergent wind field.

The combination of (15) and (16) yields

$$\partial U/\partial z = u^*/kz(\phi_u - R - S + H + J + G), \quad (17)$$

where

$$R = z/z_0\partial z_0/\partial z, \quad (17a)$$

$$S = \beta(z/L)^2\partial L/\partial z, \quad (17b)$$

$$H = (kzU^2/2u^{*3})[1 - (\xi z/U)(\partial U/\partial z)]\partial U/\partial x, \quad (17c)$$

$$J = (kzU/2u^{*3})\partial U/\partial t, \quad \text{and} \quad (17d)$$

$$G = kzfV_gU/2u^{*3}. \quad (17e)$$

The terms R and S in (17) may be converted to spatial derivatives by making an assumption that the slope γ, derived from upwind flux footprints to the local measurement point (e.g., on the downwind mast at 10-m altitude) is linear; thus, $\partial Y/\partial z = \gamma\partial Y/\partial x$, where Y may be either z_0 or L. It is assumed for the sake of simplicity that γ has a value around 60 for neutral conditions (Smedman et al. 1999); values of γ may be substantially larger or smaller than 60 for stably or unstably stratified conditions, respectively. An explanation of flux footprint theory and the range of typical values of γ associated with air–sea fluxes can be found in more detail in Smedman et al. (1999).

We note herein that $\partial U/\partial z$ appears not only on the lhs of (17) but also on the rhs within term H. Because the quantity $(\xi z/U)(\partial U/\partial z)$ inside the second term on the rhs of (17c) is on the order of 10^{-1}, by inserting $(u^*/kz)\phi_u$ for $\partial U/\partial z$, we obtain a simplified form of H as $H = (kzU^2/2u^{*3})[1 - (\xi u^*\phi_u/kU)]\partial U/\partial x$.

In (17a)–(17e), the quantities R, S, J, and H may be considered to be corrections to the wind speed profile, caused by: horizontal gradients of roughness, atmospheric stratification, wind speed, vertical mean advection, and time rate of change of the local wind speed, respectively. In (17), ϕ_u may be considered to be the only "local" parameter (on the order of unity for neutral stratifications), whereas R, S, J, and H involve spatially or temporally varying quantities. The quantity G represents a correction resulting from the geostrophic wind that in turn exhibits a latitudinal dependence. If any of the quantities R, S, J, H and G are substantially different from zero (i.e., with absolute values that deviate from zero with a value significantly greater than 10^{-2}), they will hereafter be considered to be important.

b. The stratification terms: ϕ_u, S, and G

In the application of classical surface layer theory, the function ϕ_u is based on field observations from major field campaigns conducted over a flat land surface: for example, in Kansas (Businger et al. 1971). In such terrestrial regions, variations of stratification, represented by z/L, are dominated by large diurnal variations of surface temperature, with local shear, thermal convection, and turbulence decay acting as local processes to describe

ϕ_u. In stark contrast, the marine atmosphere is dominated by a relatively constant surface temperature; variations of atmospheric stratification are in large part associated with thermal advection. Thermal advection, in turn, is closely related to the thermal wind (Holton 1979).

The quantity G depends on z/L, U, latitude, and veering or backing of the wind due to thermal advection. Based on typical values of the ratio V_g/U (see, e.g., Mendenhall 1967), it is easy to show that $|G|$ is typically much less than 0.1 for near-neutral conditions but can be as large as 0.4 for nonneutral conditions.

Typical values of the quantity S may be determined by simplifying L into a form that is proportional to $(U/\Delta T)$ (see Geernaert 2007). Referring to Geernaert (2007), the value of S may be significant in the coastal zone for fetches less than 10 km (i.e., where values of $|S|$ may reach 0.2); for fetches significantly beyond 10 km, the value of S approaches zero with increasing fetch.

Because the values of S and G over the ocean exhibit a significant dependence on z/L and because many atmospheric flux profile datasets have been collected on coastal oceanic midlatitude platforms, we will assume that stratification functions in the marine atmospheric surface layer (particularly over the coastal ocean) may be significantly influenced by S and G because of advection and the earth's rotation. We hereafter will treat the sum $(\phi_u + S + G)$ as the more appropriate stratification function to be considered for over-ocean wind profile corrections and we note herein that the stratification function will exhibit a latitudinal dependence through the function G. Given these arguments, we introduce the sum as a general form of the wind profile's stratification function: that is, $F(z/L, \theta) = (\phi_u + S + G)$, where θ is latitude.

c. Estimated values of terms R, H, and J

By applying flux footprint theory, the quantity R may be expressed as $\gamma(z/z_0)\partial z_0/\partial x$. The quantity R may be estimated by employing the Charnock relation (Charnock 1955), where $z_0 = \alpha u^{*2}/g$ and α is the Charnock coefficient. Recalling (4), the roughness length may also be expressed as $\alpha C_D U^2/g$. Letting $\partial U/\partial x$ for fetch-limited flow (e.g., for offshore blowing winds) be approximated by $U/10D$ (Astrup et al. 1999; Geernaert 2002), with D representing upwind fetch, the substitution of these simple expressions into (17a) leads to the approximation $R = \gamma z/5D$. For neutral stratifications where γ is approximately 60 and an upwind fetch of 10 km, R would have a value on the order of 0.01, and fetches beyond 10 km will lead to substantially smaller values of R. Even for more stably stratified conditions where γ can be substantially larger than 60, R is relatively small. We will hereinafter ignore the term R from further analysis.

The quantity H is much more complex, because one must consider both horizontal and vertical advection simultaneously. Typical values of the horizontal component of H may be estimated by employing the fetch-limited relation for $\partial U/\partial x$ (described in the previous paragraph), thus yielding a simple inverse dependence on upwind fetch and no dependence on wind speed. We find that an upwind fetch of 10 km leads to a value for the horizontal component of 0.15 and that a fetch of 50 km will lead to a value of 0.03. The vertical component is somewhat smaller; however, there is the potential for substantial variability in regions of mesoscale circulations (e.g., sea breezes) where the vertical velocity field may vary significantly. For two-dimensional flow (i.e., where $\xi = 1$), the vertical advection term for a height of 10 m will offset the horizontal advection term by approximately 10%. (The reader should note that, for altitudes greater than 10 m, the importance of vertical advection increases.) For strongly divergent flow as can be observed in sea breezes, the vertical advection term will clearly increase in importance. Recognizing that the horizontal component of advection is the dominant term in H and that vertical advection may be spatially inhomogeneous, we will consider H to be on average a quite important and potentially highly variable term. We will retain this term for further analysis.

The term J may be examined by first simplifying (17d) into the form $J = [(kz)/(2C_D^{3/2}U^2)]\partial U/\partial t$. Given typical values $k = 0.4$, $z = 10$ m, and $C_D = 0.0012$ and noting that an increase in mean wind speed of 20% h^{-1} is possible, a corresponding range of T that one could expect for a 5–15 m s^{-1} span of wind speeds will be from 0.6 to 0.2. The large values of T for this wind speed range suggest that the quantity T may be important. Like F and H, the term J will also be retained for further analysis.

A practical form for Eq. (16) now becomes

$$\partial U/\partial z = u^*/kz(F + H + J), \qquad (18)$$

where $F = (\phi_u + S + G)$; $H = (kzU^2/2u^{*3})[1 - (\xi u^*\phi_u/kU)]\partial U/\partial x$; and $J = (kzU/2u^{*3})\partial U/\partial t$.

d. Derivation of the normalized drag coefficient

Based on the analysis in the previous section, Eq. (18) may be simplified to the following expression:

$$\partial U/\partial z = (u^*/kz)F + (U^2/2u^{*2})[1 - (\xi u^*\phi_u/KU)]\partial U/\partial x + (U/2u^{*2})\partial U/\partial t. \qquad (19)$$

Integrating (19) between z_0 and z, one obtains the following:

$$U = (u^*/k)[\ln(z/z_0) - \Lambda] + (U^2 z/2u^{*2}) \times [1 - (\xi u^* \phi_u/kU)]\partial U/\partial x + (Uz/2u^{*2})\partial U/\partial t. \quad (20)$$

In (20), we have retained the ϕ_u term as part of a small correction to a small term. However, to simplify the derivation, we will force ϕ_u to unity only in the small correction on the rhs of the equation. This simplification requires the integration to be appropriate for stratifications that deviate not too substantially from neutral conditions. Noting that C_D is by definition $(u^*/\mathrm{U})^2$ and letting ϕ_u in the second term on the rhs of (20) be close to unity, Eq. (20) may be rewritten as

$$kC_D^{-1/2} = \ln(z/z_0) - \Lambda + (U^2 kz/2u^{*3}) \times [1 - (\xi u^*/kU)]\partial U/\partial x + (Ukz/2u^{*3})\partial U/\partial t. \quad (21)$$

In (20) and (21), the function Λ is assumed to follow systematic variations with z/L; we anticipate that variations of Λ will be similar to but not the same as Ψ_U because variations associated with G and S terms are also included. We have not attempted to derive an analytical form of Λ.

We herein define a normalized drag coefficient C_{DR} for reference conditions of neutral stratifications, horizontal homogeneity, and stationarity: that is, where $\Lambda = \partial U/\partial x = \partial U/\partial t = 0$. Substituting the reference conditions in (21), $kC_{\mathrm{DR}}^{-1/2} = \ln(z/z_0)$. Recalling that $C_D = (u^*/U)^2$, Eq. (21) may now be rewritten as

$$C_{\mathrm{DR}}^{-1/2} = [C_D^{-1/2} + (\Lambda/k)] - [(z/2U)C_D^{-3/2}] \times \{[1-(\xi u^*/kU)]\partial U/\partial x + U^{-1}\partial U/\partial t\}. \quad (22)$$

Substituting (8) and letting $\Psi_U = \Lambda$ for illustrative purposes only, Eq. (22) becomes

$$C_{\mathrm{DR}}^{-1/2} = C_{\mathrm{DN}}^{-1/2} - [(z/2U)C_D^{-3/2}]\{[1 - (\xi C_D^{1/2}/k)]\partial U/\partial x + U^{-1}\partial U/\partial t\}, \quad (23)$$

where C_{DN} is the commonly used neutral drag coefficient.

e. Is the difference between C_{DR} and C_{DN} significant and important?

To answer this question, we first ignore temporal variations and examine the importance of $\partial U/\partial x$ in creating significant differences between C_{DR} and C_{DN}. By rearranging Eq. (23), we find

$$C_{\mathrm{DR}}/C_{\mathrm{DN}} = [1 - (z/2U)C_D^{-1}(1 - \xi C_D^{1/2}/k)\partial U/\partial x]^{-2}. \quad (24)$$

Recalling that $\partial U/\partial x$ can be approximated by $U/10D$, we find that, for a typical offshore coastal region (e.g., with 10-km fetch), the value of the ratio $C_{\mathrm{DR}}/C_{\mathrm{DN}}$ will be approximately 0.90. For longer fetches, the value of the ratio rapidly approaches unity. This would imply that, for air–sea flux parameterizations derived from coastal tower datasets of widely different fetches (i.e., where many air–sea datasets have been collected), the term containing the quantity $\partial U/\partial x$ may potentially be a source of difference between flux parameterizations. Following this argument and the use of (24), measurements of C_{DN} in past datasets would likely be larger for short fetch conditions than for long fetch. Ironically, many analyses of flux parameterizations (e.g., Geernaert et al. 1986) have suggested that the higher drag coefficients observed at shorter fetch were attributed to steeper short waves and thus higher roughness. We suggest herein that both the higher roughness at short fetch and the role of advection may be used together to explain a significant part of the dependence of the drag coefficient on fetch, based on field observations.

To evaluate the importance of nonstationarity, we will for simplicity ignore spatial variability and write the analogous equation to (24): that is,

$$C_{\mathrm{DR}}/C_{\mathrm{DN}} = (1 - C_D^{-1}U^{-1}\partial U/\partial t)^{-2}. \quad (25)$$

Data from Geernaert et al. (1987) that document successive records of 30-min average wind speeds show that the quantity $(\mathrm{U}^{-1}\partial \mathrm{U}/\partial \mathrm{t})$ is often on the order of 10^{-5} to $10^{-4}\ \mathrm{s}^{-1}$. Thus, the ratio $C_{\mathrm{DR}}/C_{\mathrm{DN}}$ should be expected to be within the range 0.90–1.1. Combined with the analysis following Eq. (24), we find that a time series of the calculated ratio $C_{\mathrm{DR}}/C_{\mathrm{DN}}$ for a dataset that includes both spatial and temporal variations of mean quantities can easily vary from unity by 20%. Ironically and in the same spirit as the higher drag coefficients with shorter fetch (as discussed earlier), the neutral drag coefficients would be slightly higher for rising winds than for steady or falling winds according to (25). The drag coefficient variation with rising and falling winds as noted earlier is similar to observations in past datasets (see, e.g., Geernaert et al. 1986), where higher drag coefficients for rising winds were argued at the time to be associated with steeper short surface waves for rising winds.

4. Derivation of the Stanton and Dalton numbers for quasi-inhomogeneous and quasi-stationary conditions

a. Flux profile relations for temperature and humidity with spatial inhomogeneity and nonstationarity

In analogy to the wind speed profile, we start with the similarity law applicable to heat and trace gas fluxes for homogeneous conditions: that is,

$$\partial c/\partial z = (\langle w'c'\rangle/u^*kz)\phi_c, \quad (26)$$

where ϕ_c is a function of z/L. The quantity c may represent temperature, water vapor, or any trace gas. If we assume that the surface layer is a constant flux layer (as per MOS theory), Eq. (26) may be integrated and one quickly obtains

$$\langle w'c'\rangle = k^2 U\Delta c(\ln z/z_0 - \Psi_u)^{-1}(\ln z/z_c - \Psi_c)^{-1}, \quad (27)$$

where $\Delta c = c_0 - c$ is local on a vertical coordinate. As for momentum, the budget of an advecting concentration of the quantity c undergoing vertical turbulent flux will also be governed by the possibility for chemical reactions and phase transformations accordingly as

$$dc/dt = \partial c/\partial t + U\partial c/\partial x + W\partial c/\partial z + \partial\langle w'c'\rangle/\partial z + \Sigma = 0, \quad (28)$$

where Σ is a chemical source or sink of c (see, e.g., Kepert et al. 1999). As for the derivation of the normalized drag coefficient, we will assume herein that the vertical velocity may be allowed to be nonzero. Rearranging (28), one obtains

$$\partial\langle w'c'\rangle/\partial z = -\partial c/\partial t - U\partial c/\partial x - W\partial c/\partial z - \Sigma. \quad (29)$$

Combining (27) with the lhs of (29), one gets after manipulation the following:

$$k^{-2}\partial\langle w'c'\rangle/\partial z = (\langle w'c'\rangle/k^2 U\Delta c)[\partial(U\Delta c)/\partial z] - (U\Delta c/z)[\langle w'c'\rangle/(K^2 U\Delta c)] \times (\Phi_u A_u + \Phi_c A_c), \quad (30)$$

where

$$\Phi_u = 1 - (z/z_0)\partial z_0/\partial z - z\partial\psi_u/\partial z = \phi_u(z/L) - \gamma(z/z_0)\partial z_0/\partial x - \gamma\beta(z/L)^2\partial L/\partial x \quad \text{and} \quad (31a)$$

$$\Phi_c = \phi_c(z/L) - (\gamma z/z_c)\partial z_c/\partial x - \gamma\beta(z/L)^2\partial L/\partial x. \quad (31b)$$

The coefficient γ is the slope to upwind footprints as defined in the previous section, on the order of 60, for neutral conditions. The quantities A_u and A_c are the drag factors, introduced here only for convenience to simplify the derivation:

$$A_u = (\ln z/z_0 - \psi_u)^{-1} \quad \text{and} \quad (32a)$$

$$A_c = (\ln z/z_c - \psi_c)^{-1}. \quad (32b)$$

We hereafter invoke results of Geernaert (2002), such that $\partial z_0/\partial x$, $\partial z_c/\partial x$, and $\partial L/\partial x$ may be ignored. We furthermore constrain the derivation to stratifications that do not deviate significantly from neutral conditions: that is, $\phi_u = \phi_c$. Defining the Stanton or Dalton numbers $C_c = k^2 A_u A_c$ and the drag coefficient $C_D = k^2 A_u{}^2$ and using Eq. (19) for the wind profile, Eq. (30) may be algebraically rearranged to produce the following expression:

$$\partial c/\partial z = (\langle w'c'\rangle/u^*kz)[\phi_c - (C_D/C_c)(H + J)] - C_c^* + Y_c^* - U_c^* - T^* - \Sigma^*, \quad (33)$$

where

$$C_c^* = [U(c - c_0)/\langle w'c'\rangle]\partial c/\partial x, \quad (34)$$

$$Y_c^* = \gamma\partial c_0/\partial x, \quad (35)$$

$$U_c^* = (\Delta c/\langle w'c'\rangle)\partial U/\partial x(1 - \xi\partial c/\partial z), \quad (36)$$

$$\Sigma^* = (\Delta c/\langle w'c'\rangle)\Sigma, \quad \text{and} \quad (37)$$

$$T^* = (\Delta c/\langle w'c'\rangle)\partial c/\partial t. \quad (38)$$

Equation (36) may be simplified by introducing the quantity $A = (C_C/C_D^{1/2})$, where C_C is the exchange coefficient for, for example, temperature and humidity, and confining the domain to near-neutral conditions, such that $\xi\partial c/\partial z = \xi A\Delta c/kz$. Note that, if the conditions are horizontally homogeneous and steady state and if the chemical is slowly reacting, Eq. (33) reduces to the more popular form shown in (1).

b. *Derivation of the normalized Stanton and Dalton numbers for nonstationarity and inhomogeneity*

The terms in Eq. (33) involve a wide variety of spatially varying quantities. If we confine ourselves to compounds and/or conditions where chemical reactions are relatively unimportant and if we simplify the domain to weakly varying surface concentrations or temperatures, we may reduce (33) to the following more practical form for the vertical temperature gradient:

$$\partial T/\partial z = (\langle w'T'\rangle/u^*kz)[\phi_T - (C_D/C_H)(H + J)] - C_c^* - U_c^* - T^*. \quad (39)$$

Integrating both sides of (39) from z_0 to z and noting the logarithmic behavior of temperature with height as shown in (3) as well as the definition of the Stanton number in (6), the following relationship emerges:

$$1 + [z/(C_H U\Delta T)]B_T = C_H/(kC_D^{1/2})[\ln(z/z_{ot}) - \Psi_H] - (zC_H/C_D^2)[\partial U/\partial t + U\partial U/\partial x \times (1 - \xi C_D^{1/2}/K)], \quad (40)$$

where $B_T = \partial T/\partial t + U\partial T/\partial x - U\partial T_0/\partial x + \Sigma + (C_H\Delta T/2C_D)(1 - \xi A\Delta T/kz)\partial U/\partial x$. Defining $C_H = C_{HR}$ for neutrally stratified, spatially homogeneous, and stationary conditions [i.e., where (40) reduces to $C_{HR} = (kC_D^{1/2})/(\ln z/z_{ot})$], the following relationship emerges:

$$C_{HR} = C_H \left\langle \begin{matrix} 1 + (C_H\Psi_H/kC_D^{1/2}) + \{z/[C_H U(T - T_0)]\}B_T \\ + [z(C_H/C_D^2)][\partial U/\partial t + U(\partial U/\partial x)(1 - \xi C_D^{1/2}k)] \end{matrix} \right\rangle^{-1}. \tag{41}$$

If the domain is spatially homogeneous, stationary, and with no vertical advection, Eq. (41) conveniently reduces to the expression for the neutral Stanton number, found in (9). In analogy to (41), the normalized Dalton number is represented as

$$C_{ER} = C_E \left\{ \begin{matrix} 1 + (C_E\Psi_E/kC_D^{1/2}) + \{z/[C_E U(q - q_0)]\}B_q \\ + [z(C_E/C_D^2)][\partial U/\partial t + U(\partial U/\partial x)(1 - \xi C_D^{1/2}k)] \end{matrix} \right\}^{-1}, \tag{42}$$

where $B_q = \partial q/\partial t + U\partial q/\partial x - U\partial q_0/\partial x + \Sigma + (C_E\Delta q/2C_D)(1 - \xi A\Delta q/kz)\partial U/\partial x$ and q is specific humidity.

Expressions (41) and (42) are clearly more complicated than that for the drag coefficient, insofar that spatial and temporal variations of wind speed must be considered in addition to variations of temperature or water vapor flux.

c. *Are the ratios (C_{HR}/C_{HN}) and (C_{ER}/C_{EN}) noticeably different from unity?*

Substituting (9) and (10) into (41) and (42) leads to expressions that relate C_{HR} to C_{HN}, and similar expressions may be found for C_{ER} and C_{EN}. For illustration, we will assume neutral stratifications and apply the same typical values of $\partial U/\partial x$ for limited fetch coastal zones as for the normalized drag coefficient (e.g., where $D = 10$ km, the value of $\partial U/\partial x$ may easily be on the order of 10^{-3} s^{-1}); values of $\partial U/\partial x$ are typically smaller over the open ocean. Similarly, typical variations of $\partial T/\partial x$ and $U^{-1}\partial T/\partial t$ are on the order of 10^{-5}. In addition, we assume that the vertical advection is at most 10% of the value of horizontal advection. Given these typical variations, one finds that (C_{HR}/C_{HN}) could potentially vary by 10% or more. We assume herein that variations of (C_{ER}/C_{EN}) may also be of the same order of magnitude.

5. Illustration with historical field data

Normalizing the drag coefficient, Stanton number, and Dalton number for more environmental variabilities than just stratification has the potential benefit of providing parameterizations based on a larger set of normalized bulk quantities. We would furthermore hypothesize that flux coefficients normalized with more quantities than just stratification could lead to smaller differences between flux coefficient parameterizations: that is, where different parameterizations were developed from past datasets originating from very different geographic and/or environmental conditions.

To test the importance of a multiprocess normalization versus just neutral stratification as the normalization procedure, we would ideally desire access to field observations containing spatial variations of wind speed, temperature, humidity, and roughness, as well as local temporal variabilities including fluxes, to demonstrate a reduced variance of C_{DR}, C_{HR}, and C_{ER}: that is, compared to the variance associated with simply C_{DN}, C_{HN}, and C_{EN}. In addition, an accurate measure of W would be highly desirable. Such an analysis would require a dataset that contains all of the environmental information to quantify all terms in (18), (41), and (42). Unfortunately, although numerous datasets exist that include measurements of the air–sea fluxes of momentum, heat, and moisture flux along with their local temporal variations of bulk quantities, there are no known datasets that contain this full suite of information.

For this study, we perform a simple analysis to demonstrate the use of additional processes beyond just stratification. However, because of the complexity of obtaining a time series of advection combined with a time series of local fluxes, we will restrict this demonstration to temporal variations of U and T: that is, in addition to variations of stratification. This simplification allows the computation of normalized drag coefficients and Stanton numbers, by forcing all terms with spatial variations to zero. We use (25) for the normalized drag coefficient; for the Stanton number, we simplify (41) to the following form:

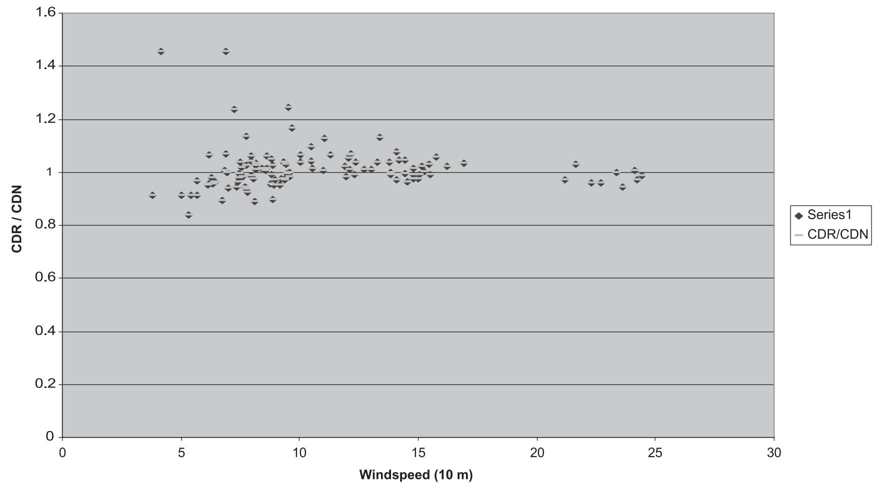

FIG. 1. Distribution of the ratio of the normalized drag coefficient for stratification and stationarity to the normalized drag coefficient for stratification only. Data are based on field observations from the North Sea Platform field experiment, reported in Geernaert et al. (1987).

$$C_{\mathrm{HR}} = C_H\langle 1 + (C_H\Psi_H/kC_D^{1/2}) + \{z/[C_H U(T - T_0)]\}\partial T/\partial T + [z(C_H/C_D^2)]\partial U/\partial t\rangle^{-1}. \quad (43)$$

For this analysis, we have resurrected the North Sea platform data reported in Geernaert et al. (1987). We compute $\partial U/\partial t$ as $[U(r + 1) - U(r - 1)]/2\Delta t$, where r is a given data record in a block and t is the temporal length of the record; for data representing the first or last record of a data block, the value of $\partial U/\partial t$ is estimated by taking the difference between two sequential records within the same block. The same process is used for estimating $\partial T/\partial t$. Except for wind speeds exceeding 20 m s^{-1}, the length of most records is 20 or 30 min; refer to Table 1 of Geernaert et al. (1987) for complete details.

Referring to the time series of C_{DN} and C_{HN} reported in Geernaert et al. (1987) and comparing to newly computed flux coefficients that take into account temporal variability (i.e., C_{DR} and C_{HR}), we report the ratios ($C_{\mathrm{DR}}/C_{\mathrm{DN}}$) and ($C_{\mathrm{HR}}/C_{\mathrm{HN}}$) in Figs. 1 and 2, respectively. As can be observed, variations of these ratios are typically on the order of 5%–10%.

In a reanalysis of the normalized flux coefficient data that extended from 6 to 24 m s^{-1}, we find slightly different expressions that relate the normalized drag coefficient with wind speed: that is,

$$10^3 C_{\mathrm{DN}} = 0.085U_{10} + 0.57; \quad \text{standard deviation} = 0.73 \quad \text{and} \quad (44)$$

$$10^3 C_{\mathrm{DR}} = 0.092U_{10} + 0.61; \text{ standard deviation} = 0.71. \quad (45)$$

As can be seen, the variance is slightly reduced when one extends the set of normalizations from just neutral stratification to include steady state (no temporal variability).

Heat fluxes were observed over a narrower range of environmental conditions than momentum flux in the Geernaert et al. (1987) dataset. The range of observed wind speeds for heat fluxes spanned 6–15 m s^{-1}, and stratifications were generally near neutral. Stanton numbers exhibited little dependence on either wind speed or stratification. We found no significant dependence on wind speed, and we therefore computed means for the normalized Stanton numbers. We found means for 10^3C_{HN} and 10^3C_{HR} to be 0.565 and 0.553, respectively. Standard deviations of the two quantities are 0.68 and 0.66, respectively. Refer to Fig. 2, where we plot the distribution of the ratio $C_{\mathrm{HR}}/C_{\mathrm{HN}}$ with wind speed.

6. Discussion and summary

In this paper, we have extended the concept of normalized flux coefficients from simple neutral drag coefficients and neutral Stanton and Dalton numbers to expressions that also accommodate nonstationarity and spatial variability of bulk quantities. We introduced the quantities C_{DR}, C_{HR}, and C_{ER} to represent the broader set of normalized quantities.

We also demonstrated that the stratification functions for surface layer wind speed and other profiles should not be based simply on the nondimensional profile expressions found in, for example, Businger et al. (1971) but that there are additional terms associated with spatial

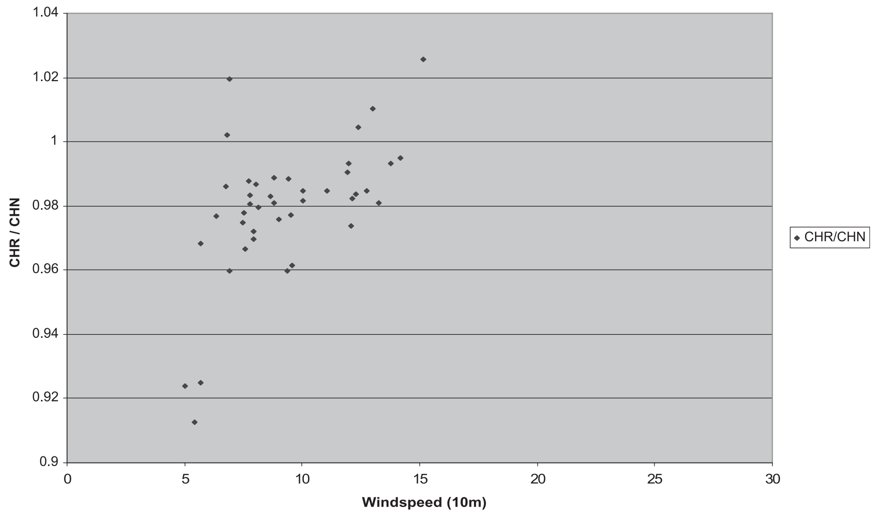

FIG. 2. Distribution of the ratio of the normalized Stanton number for stratification and stationarity to the normalized Stanton number for stratification only. Data are based on field observations from the North Sea Platform field experiment, reported in Geernaert et al. (1987).

gradients of stratification and the earth's rotation. We speculate that the additional terms are small, but they may have added to the data scatter in previous datasets that were designed to derive expressions for dimensionless flux profiles as a function of stratification.

To illustrate the importance of these results, we chose a dataset that was compiled during a North Sea field experiment, spanning a wide range of wind speeds and with upwind fetches on the order of 100 km or greater. The reanalysis of the historical North Sea dataset to produce normalized flux coefficients revealed small but noticeable differences in the parameterization of the drag coefficient versus wind speed, and only slight differences in the Stanton number were observed. A slight reduction of variance is reported for normalized drag coefficient and Stanton number, when stratification and steady state are included in the normalization procedure. The reader is reminded that the use of the North Sea dataset was able to illustrate normalization using stratification and temporal variability to normalize flux coefficients; other variabilities associated with spatial variability of wind speed and other quantities were ignored because of lack of information.

A surprising result of our analyses suggests that the fetch-dependent drag coefficient, which in turn is related to a dependence of the roughness length on wave state, could in part be an artifact of the influence that horizontal wind speed gradients have on the normalized drag coefficient [Eq. (24)]. Similarly, past relationships between the larger (smaller) drag coefficient and rising (falling) winds could be an artifact of the drag coefficient dependence on nonstationarity [Eq. (25)]. This result suggests that a reexamination of the expressions that relate roughness length to wave state should be carried out in the context of multiprocess normalization of drag coefficients, particularly when expressions were based on fetch-dependent wave state.

Given that spatial and temporal variability of bulk quantities is often larger in coastal zones than in the open ocean and given that many air–sea flux parameterizations are based on coastal towers and platforms, we would anticipate that our new approach to producing normalized flux coefficient parameterizations may be significantly important. However, given that only one dataset was analyzed in our study (i.e., with an objective to illustrate the method), we anticipate that its application over a wider set of field datasets may lead to composite parameterizations with reduced uncertainty.

The derivations in this manuscript were initiated with the observation that advection and nonstationarity can act as independent external forcings that locally lead to deviations from similarity theory. One should recognize that the marine boundary layer system is fully integrated with interdependencies: for example, advection and nonstationarity are in fact influenced by the surface fluxes. Therefore, the application of this extended theory in local and regional modeling studies must recognize that the turbulent fluxes, advection, and nonstationary aspects of the flow field are fully coupled and highly complex.

The reader is reminded that there was a basic assumption involved in the derivation of C_{DR}, C_{HR}, and C_{ER}: that is, that Eq. (13) could be safely inserted into Eq. (11) and that Eq. (27) may be inserted into (28). We are unable to test the validity of this assumption over a wide range of environmental conditions without existing field observations. Furthermore, we are unable to test the allowable range of values for horizontal advection, vertical advection, and nonstationarity, so that similarity theory remains valid for wide applications. We therefore recommend that future field studies of air–sea fluxes be designed to include information on spatial variability of bulk quantities and accurate measurements of vertical velocity. We furthermore recommend that future flux observations be normalized not just for neutral stratification but also for homogeneous and stationary conditions. Finally, we suggest that a more rigorous investigation of stratification functions for marine atmospheric profiles be pursued, both theoretically and experimentally, to resolve the contributions to the term $F(z/L)$ found in Eq. (19).

Acknowledgments. The author acknowledges support from the Laboratory Directed Research and Development Program of Los Alamos National Laboratory, in performing this research. Valuable suggestions for improvement were provided by two anonymous reviewers, and they are gratefully acknowledged.

REFERENCES

Astrup, P., N. O. Jensen, and T. Mikkelsen, 1996: Surface roughness model for LINCOM. Risø National Laboratory Rep. Risø-R-900(EN), 31 pp.

Businger, J. A., J. C. Wynaard, and Y. Izumi, 1971: Flux profile relationships in the atmospheric surface layer. *J. Atmos. Sci.,* **28,** 181–189.

Charnock, H., 1955: Wind stress on a water surface. *Quart. J. Roy. Meteor. Soc.,* **81,** 639–640.

Geernaert, G. L., 2002: On extending the flux-profile similarity theory to include quasi-homogeneous conditions in the marine atmospheric surface layer. *Bound.-Layer Meteor.,* **105,** 433–450.

——, 2007: On the evaporation duct for inhomogeneous conditions in coastal regions. *J. Appl. Meteor. Climatol.,* **46,** 538–543.

——, K. B. Katsaros, and K. Richter, 1986: Variation of the drag coefficient and its dependence on sea state. *J. Geophys. Res.,* **91,** 7667–7679.

——, S. E. Larsen, and F. Hansen, 1987: Measurements of the wind stress, heat flux, and turbulence intensity during storm conditions over the North Sea. *J. Geophys. Res.,* **92,** 13 127–13 139.

Holton, J., 1979: *An Introduction to Dynamic Meteorology*. 2nd ed., Academic Press, 391 pp.

Kepert, J., C. Fairall, and J.-W. Bao, 1999: Modelling the interaction between the atmospheric boundary layer and evaporating sea spray droplets. *Air-Sea Exchange: Physics, Chemistry, and Dynamics,* G. L. Geernaert, Ed., 363–410.

Mahrt, L., 1999: The coastal zone. *Air-Sea Exchange: Physics, Chemistry, and Dynamics,* G. L. Geernaert, Ed., Kluwer, 247–268.

——, D. Vickers, J. Sun, T. L. Crawford, G. Crescenti, and P. Frederickson, 2001: Surface stress in offshore flow and quasi-frictional decoupling. *J. Geophys. Res.,* **106** (D18), 20 629–20 639.

Mendenhall, B. R., 1967: A statistical study of frictional wind veering in the planetary boundary layer. Colorado State University Department of Atmospheric Sciences Tech. Paper 116, 56 pp.

Monin, A. S., and A. M. Obukhov, 1954: Basic turbulent mixing laws in the atmospheric surface layer. *Tr. Akad. Nauk. SSSR Geofys. Inst.,* **24,** 163–187.

Panofsky, H. A., and J. A. Dutton, 1984: *Atmospheric Turbulence*. John Wiley and Sons, 397 pp.

Smedman, A., U. Hogstrom, H. Bergstrom, H. Rutgersson, K. K. Kahma, and H. Pettersson, 1999: A case study of air-sea interaction during swell conditions. *J. Geophys. Res.,* **104** (C11), 25 833–25 851.

Vogel, C., and T. L. Crawford, 1999: Exchange measurements above the air-sea interface using an aircraft. *Air-Sea Exchange: Physics, Chemistry, and Dynamics,* G. L. Geernaert, Ed., Kluwer, 231–246.

Wilczak, J., J. Edson, J. Højstrup, and T. Hara, 1999: The budget of turbulence kinetic energy in the marine atmospheric surface layer. *Air-Sea Exchange: Physics, Chemistry, and Dynamics,* G. L. Geernaert, Ed., Kluwer, 153–174.