Constraining the inertial dissipation method using the vertical velocity variance

G. Dardier,[1,2] A. Weill,[1] H. Dupuis,[3] C. Guerin,[1] W. M. Drennan,[4] S. Brachet,[1,4]
F. Lohou,[1] and R. Pedreros[3]

Received 30 September 2002; revised 20 November 2002; accepted 23 January 2003; published 13 March 2003.

[1] The inertial dissipation method (IDM) is commonly used to measure turbulent fluxes over the ocean. It has the advantage over more direct methods in that it depends on the turbulent fluctuations only in the high frequencies of the so-called inertial subrange. These frequencies are above those of typical ship motions and are considered to be relatively unaffected by flow distortion. However, a drawback in applying the method is that the problem is underdetermined: estimation of the fluxes requires knowledge of the Obukhov length L, which is itself a function of the fluxes. The problem is typically solved by iteration, using an initial L estimated from bulk formulae. This introduces a possible dependency on the initial bulk estimate along with problems of convergence. Recently, several authors have proposed improvements to the basic algorithm. For instance, *Dupuis et al.* [1997] proposed a parameterization of the "imbalance term" in the budget of turbulent kinetic energy (TKE). We explore an alternative approach to the problem. In order to constrain the equations resulting from the IDM we use the vertical velocity variance, σ_w, measured from the research vessel *L'Atalante* and an ASIS buoy, both deployed during the 1998 FETCH experiment. These data are compared to several parameterizations of σ_w on stability derived in experiments. For unstable cases, the data are found to be well described by the *Panofsky and Dutton* [1984] parameterization, although the scatter of the data is higher for swell conditions than for pure wind sea, indicating a likely sea state effect. Using measured values of σ_w along with this parameterization, the inertial dissipation problem is fully specified. The convergence of the method is satisfactory, and it offers u_* estimates independent of bulk formulae. *INDEX TERMS:* 4504 Oceanography: Physical: Air/sea interactions (0312); 0312 Atmospheric Composition and Structure: Air/sea constituent fluxes (3339, 4504); 3339 Meteorology and Atmospheric Dynamics: Ocean/atmosphere interactions (0312, 4504); *KEYWORDS:* turbulent fluxes, inertial dissipation method, air-sea interaction

Citation: Dardier, G., A. Weill, H. Dupuis, C. Guerin, W. M. Drennan, S. Brachet, F. Lohou, and R. Pedreros, Constraining the inertial dissipation method using the vertical velocity variance, *J. Geophys. Res.*, *108*(C3), 8063, doi:10.1029/2002JC001667, 2003.

1. Introduction

[2] During the 1998 FETCH experiment in the Mediterranean Sea [*Hauser et al.*, 2000, 2003], air-sea fluxes were measured from the R/V *L'Atalante*, a large research vessel owned by IFREMER and from an ASIS spar buoy [*Graber et al.*, 2000]. Flux estimates were made using both the Eddy Correlation Method (ECM) and the Inertial Dissipation Method (IDM). The ECM provides direct measurements of the fluxes, but its application at sea is difficult due to the contamination of the velocity signals by platform motion. During FETCH, these platform motions were measured and used to correct the measured wind velocity components to a stationary reference frame, following *Anctil et al.* [1994].

[3] Even with these corrections, there remain questions about the accuracy of ECM fluxes from large ships. *Edson et al.* [1991] and R. Pedreros et al. (The eddy correlation method on large structure ships, submitted to Journal of Geophysical Research, 2002) (hereinafter referred to as Pedreros et al., submitted manuscript, 2002) showed that the ECM covariances for momentum flux are significantly affected by the turbulent air flow distortion around the hull and superstructure of large ships; scalar fluxes are less affected. Since the IDM appears to be unaffected by turbulent flow distortion, it remains an attractive method for use onboard large vessels. We note here that distortion of the *mean* flow around the ship's hull and superstructure must be accounted for with either method [*Dupuis et al.*, 2003].

[1]Centre d'étude des Environnements Terrestre et Planétaires (CETP), Velizy, France.
[2]Now at Davidson Laboratory, Stevens Institute of Technology, Hoboken, New Jersey, USA.
[3]Département de Géologie et d'Océanographie (DGO), Université Bordeaux I, UMR58, Talence, France.
[4]Rosenstiel School of Marine and Atmospheric Science (RSMAS), University of Miami, Florida, USA.

0148-0227/03/2002JC001667

[4] However, it is well known that the IDM involves solving an underdetermined system. In the case of momentum flux, a single equation, the turbulent kinetic energy (TKE) budget, must be solved for two unknowns: the friction velocity u*, and the Obukhov length L. A summary of the IDM method is given below. Detailed descriptions are provided by *Yelland et al.* [1994], *Yelland and Taylor* [1996], *Edson et al.* [1991] and *Dupuis et al.* [1995, 1997]. To avoid this problem, IDM algorithms use an iterative approach, using an estimate of L as a "first guess". This initial estimate is typically derived from bulk relations or using the bulk Richardson number [*Launiainen*, 1995; *Grachev and Fairall*, 1997; *De Bruyn et al.*, 2000]. The latter approach has been found to improve convergence but there remains the problem that for a significant number of cases the method does not converge. Also, the solution can be dependent on the initial estimate.

[5] Here we take a different approach to solving the IDM problem. As eddy correlation fluxes require the measurement of the vertical velocity fluctuations, a by-product of the method is the vertical velocity variance, σ_w. The vertical velocity variance has been intensively studied in the continental boundary layer. See, for example, *Merry and Panofsky* [1974], *Panofsky* [1972], *Panofsky et al.* [1977], *Wyngaard et al.* [1971, 1974], and *Högström* [1990] for the surface layer and *Weill et al.* [1980] for the boundary layer. Upon the sea, the behavior of σ_w has been investigated by *Smith and Anderson* [1984], *Smedman et al.* [1999], and *Rutgersson et al.* [2001]. The main results are that: the standard deviation of vertical velocity fluctuations can be parameterized in the form $\sigma_w/u^* = f(z/L)$, where z is the height of the measurement; and that for very unstable stratifications, σ_w can be expressed as a function of the virtual heat flux alone, with no dependence on the friction velocity, u_*.

[6] It is proposed here to use the measured σ_w in order to constrain the IDM system. Essentially we combine the normalized TKE budget and the σ_w parameterization to yield a system of two equations, which can then be solved for the two unknowns, u* and z/L.

[7] This paper is organized as follows: in the next section, a brief description of the FETCH experiment is presented and methods to derive the vertical velocity from two platforms are summarized. Then, σ_w and its dependence on z/L are analyzed, and several parameterizations are tested. In the next section, we present a modification to the IDM which uses a combination of the turbulent kinetic energy budget and σ_w parameterization, to be solved for the two unknowns z/L and u_*. Only unstable cases are considered due to the uncertainty in the parameterization for the stable cases. The conclusion emphasizes the advantages and limitations of the proposed method.

2. Wind Velocity Estimates During FETCH

[8] We present here a brief description of the FETCH experiment. For details, see *Hauser et al.* [2003]. The FETCH (Flux, Etat de mer et Télédétection en Condition de fetcH variable) experiment took place in March–April 1998 in the Gulf of Lions of the Mediterranean Sea. The campaign's objectives revolve around the study of the exchanges at the air-sea interface, oceanic circulation and the improvement in the use of remote sensing to estimate wind, waves and fluxes at the air-sea interface. The principal objectives regarding the turbulent fluxes are related to improving flux parameterizations used in both atmospheric and circulation models. For instance, the dependence of the turbulent fluxes on the wind and the state of wave development remains poorly known.

[9] Measurements of fluxes and vertical velocity fluctuations are performed both on the 85m research vessel, *L'Atalante*, and on an ASIS (air-sea interaction spar) buoy moored 60 km from the coast. See *Dupuis et al.* [2003] and *Drennan et al.* [2003] for detailed descriptions of the respective platforms, their instrumentation, and the computation methodology used on each. In each case, measurements are based on the direct eddy correlation method, where the momentum flux vector τ is given by the following expression:

$$\vec{\tau} = -\rho\left(\overline{u'w'}i + \overline{v'w'}j\right), \tag{1}$$

where ρ is the air density, and u', v', and w' are the detrended turbulent components (horizontal downwind, horizontal cross-wind and vertical, respectively) of the wind velocity. The overbar represents a time average of order 30 minutes.

[10] On the R/V *L'Atalante*, the wind vector was measured using an ultrasonic Gill R3HS anemometer, mounted on the top of a mast on the foredeck, at a height of 17.8 m above mean sea level. On a large vessel such as *L'Atalante*, two corrections must be performed on the measured velocities: one to compute the velocities in a stationary, ground-based reference system, accounting for the motion of the ship (and anemometer), and a second one to account for flow distortion. A motion package, measuring the three angles of rotation (pitch, roll and yaw), along with vertical acceleration (heave), was located about 1 m below the anemometer. The motion correction procedure, described in detail by Pedreros et al. (submitted manuscript, 2002), essentially follows that of *Anctil et al.* [1994].

[11] Once the velocities have been calculated in the fixed reference frame (ground based velocity), one needs to estimate the effects of airflow distortion around the ship. Airflow distortion effects can be separated into mean flow and turbulent flow components. The distortion of the turbulent components is difficult to model. It is generally considered to be negligible at the small scales used for the IDM [*Oost et al.*, 1994], but may significantly affect those relevant for the ECM (Pedreros et al.,submitted manuscript, 2002). Based on the numerical simulations of *Nacass* [2001], the primary effect of mean flow distortion is the tilting of the streamlines to follow the ship superstructure. The mean streamline tilt angles estimated from the model are consistent with those determined from the data. The tilting affects the horizontal mean wind speed, as well as the effective measurement height, which are corrected as explained in *Dupuis et al.* [2003]. The mean flow distortion is found to vary with anemometer position, and wind direction with respect to the bow. For example, for bow-on flow, the wind speed at the anemometer position is found to be decelerated by 6% and the flow to be vertically elevated by 1.21m. Also evident from the modeling is that flow distortion increases significantly as the angle of the wind with respect to the ship increases beyond 30 degrees from the

bow. Thus in the following, data from R/V *L'Atalante* are restricted to angles ranging from −30 to 30 degrees.

[12] On the ASIS buoy, the wind velocity vector is measured using a Gill 1012R2A sonic anemometer, mounted on a mast at 7m above mean sea level. The motion of ASIS is measured using a complete motion package, and these motion signals are used to correct the wind vector to a fixed reference frame using an algorithm based on *Anctil et al.* [1994]. See *Drennan et al.* [2003] for further details regarding the ASIS deployment during FETCH.

3. A Modified Inertial Dissipation Method

3.1. TKE Budget

[13] We present here a brief summary of the theoretical background of the IDM. See *Edson et al.* [1991], *Fairall et al.* [1996], *Yelland et al.* [1994], or *Dupuis et al.* [1995, 1997] for further details. Based on the Kolmogorov hypothesis, the power spectral density $S_{uu}(n)$ of the downstream wind component u can, in the inertial subrange, be related to the TKE dissipation rate, ε, via the wave number n:

$$S_{uu}(n) = k\varepsilon^{2/3}n^{-5/3}, \tag{2}$$

where k is the one dimensional Kolmogorov constant, here assumed to be 0.55. This relationship assumes that the turbulence is isotropic, which may not be justified over the ocean or at least should be verified.

[14] Using Taylor's hypothesis (frozen turbulence), (2) becomes

$$S_{uu}(f) = k\varepsilon^{2/3}f^{-5/3}\left(\frac{u_{rel}}{2}\pi\right)^{2/3}, \tag{3}$$

where u_{rel} is the mean wind speed measured by the anemometer and f is the measurement frequency. Hence, the dissipation rate can be obtained by calculating the mean value of $S_{uu}(f) \times f^{5/3}$ over an appropriate frequency range.

[15] The wind stress is derived from the dissipation rate using the TKE budget [*Busch*, 1972], which, for steady state horizontally homogeneous turbulence, can be written as

$$M + B + D_t + D_p = \varepsilon, \tag{4}$$

where M is the mechanical production of momentum by the wind shear, B the buoyant production, D_t the transport term corresponding to a vertical divergence of TKE transport, and D_p the vertical divergence of pressure (p) transport.

[16] More explicitly:

$$M = -\overline{u'w'}\frac{\partial U}{\partial z},$$
$$B = \frac{g\overline{\theta_v' w'}}{\Theta_v},$$
$$D_t = \frac{\partial \overline{w'e'}}{\partial z},$$
$$D_p = -(1/\rho)\frac{\partial \overline{p'w'}}{\partial z},$$

where e = TKE and θ_v is virtual temperature.

[17] Following Monin-Obukhov (MO) similarity theory, (4) can be made dimensionless through multiplication by the surface layer scaling parameter, $\frac{\kappa z}{u_*^3}$, where $\kappa = 0.4$ is the von Kármán constant. The resulting expression is:

$$\varphi_m - \frac{z}{L} - \varphi_t + \varphi_p = \varepsilon\frac{\kappa z}{u_*^3} = \varphi_\varepsilon, \tag{5}$$

where each of the dimensionless profile functions, the φ's, are expected to be universal functions of z/L. The indices m, t, p stand respectively for momentum, transport and pressure. Equation (5) defines the dimensionless dissipation function φ_ε. If the terms on the left hand side are known, the friction velocity can be calculated from an estimate of the dissipation rate, such as that given by (3).

[18] Unfortunately, the exact forms of the φ's are not well known [*Fairall and Larsen*, 1986]. Such profile measurements over the sea are rare, or in the case of φ_p nonexistent. As a result, previous authors have made various assumptions as to their magnitude for turbulence over the sea. *Dardier et al.* [1999] present a detailed review of the expressions found in the literature. In that report, one can find values for the Kolmogorov and von Kármán constants, the stability functions, and the imbalance term ($\varphi_{imb} = \varphi_t - \varphi_p$).

[19] An exact calculation of the Obukhov length L,

$$L = -T_0 u_*^3 / \left(g\kappa\left(\overline{w'\theta'} + 0.61T_0\overline{w'q'}\right)\right), \tag{6}$$

requires estimates of the momentum, sensible heat and moisture fluxes. In (6), T_o and θ' represent the mean and turbulent components of potential temperature, and q' the turbulent component of specific humidity. Since measurements of the heat and moisture fluxes are not usually available, bulk estimates are used in their place. Bulk values of L are computed from mean surface parameters (mean wind speed, air temperature and humidity, along with the mean sea surface temperature), and empirical bulk coefficients. Here the bulk heat transfer coefficients of *Large and Pond* [1982] and the drag coefficient of *Smith* [1980] are used as a first guess but many other bulk estimates have been proposed; see *Fairall et al.* [1996], *Zeng et al.* [1998]. Since both L and u* are unknown, (5) is underdetermined. It is therefore common to solve (5) iteratively. In many cases, however, there is no convergent solution, resulting in the loss of data (up to 50% in some studies).

[20] In many applications of the IDM, the pressure and flux divergence terms in equation (5) are neglected, since early observations have shown the two terms to be roughly in balance [*Large and Pond*, 1981]. In analyzing their IDM results, *Dupuis et al.* [1997] noted a stability dependence of the "neutral" drag coefficients. In order to remove this dependence, *Dupuis et al.* [1997] parameterized the imbalance term, φ_{imb}, as a function of z/L ($\varphi_{imb} = -0.5z/L$). Although this imbalance term was found to significantly improve the convergence of the IDM iterations, other suggestions to solve the convergence and interpretation of the imbalance term have been proposed in the literature [*Taylor and Yelland*, 2000]. It has also been suggested that the imbalance term is associated with sea state effects, and not stability [*Edson and Fairall*, 1998].

3.2. The σ_w Parameterizations

[21] Our approach here to solving the underdetermined IDM problem is different. We seek a parameterization of

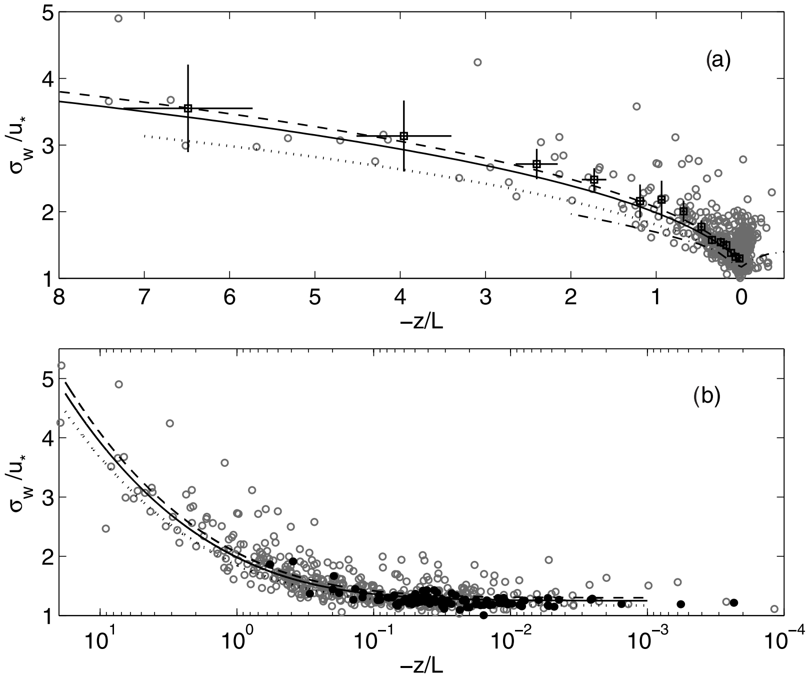

Figure 1. Standard deviation of vertical velocity fluctuations, σ_w, normalized by u_* as a function of z/L bulk. Data are from an ASIS buoy, using eddy correlation fluxes. The curves are $\sigma_w/u_* = 1.25(1-3z/L)^{1/3}$, the "Panofsky parameterization" (solid), and $\sigma_w/u_* = 1.3(1-3z/L)^{1/3}$ (dashed). (a) Averages of z/L groups are indicated as boxes, with lines showing two standard errors in both σ_w/u_* and z/L. Additional lines are the parameterizations of *Högström* [1990] (dash-dotted) and *Rutgersson et al.* [2001] (dotted). (b) The unstable data in the range $-20 < z/L < -10^{-4}$ are plotted. The wind sea data are indicated by solid circles. The dotted line is the wind sea parameterization.

σ_w/u^* as a function of z/L - i.e. in the same form as terms in the dimensionless TKE budget, equation 5. Several parameterizations of σ_w have been proposed in the literature for surface layers over land. Based on the asymptotic limit of $z/L \rightarrow -\infty$ (shear-free, convective conditions, when the dependence of σ_w on u^* must disappear), most parameterizations present a similar $(-z/L)^{1/3}$ asymptotic behavior for large $-z/L$ values [e.g., *Merry and Panofsky*, 1974; *Panofsky et al.*, 1977; *Kaimal and Finningan*, 1994; *Högström*, 1990].

[22] We restrict our attention here to unstable cases, since scatter in both the data and parameterizations for stable cases remains very large. We choose to use and to discuss at first the parameterization of *Panofsky and Dutton* [1984],

$$\sigma_w/u_* = 1.25(1-3z/L)^{1/3} = \varphi_w(z/L), \tag{7}$$

since it has been extensively used in the literature. Also it has been found to compare well with mixed layer parameterizations [see *Weill et al.*, 1980]. We note here that earlier versions of (7) [e.g., *Merry and Panofsky*, 1974; *Panofsky et al.*, 1977] used a factor of 1.3 instead of 1.25.

[23] As (7) has yet to be validated over the ocean, we compare the predicted σ_w/u_* with data obtained from the ASIS buoy. ASIS is taken as the reference platform here, because flow distortion is minimal, and direct friction velocity measurements via ECM are available. Figure 1 shows the ASIS data plotted along with several parameterizations: (7), along with those of *Panofsky et al.* [1977], *Högström* [1990], and *Rutgersson et al.* [2001] (Figure 1a). The crosses, showing 2 standard errors in σ_w/u_* and z/L, represent the ASIS data averaged in bins of z/L in order to reduce the scatter. On Figure 1b, the same data are plotted in semilogarithmic axes, in order to emphasize the behavior near neutral stability ($z/L \rightarrow 0^-$). Here we use solid circles to indicate pure wind sea cases, which are identified using directional wave spectra and the wave age criterion of *Donelan et al.* [1985], see *Drennan et al.* [2003] for details. Open circles identify cases where swell is present. It is evident that the wind sea data are much less scattered than the swell data, and also the neutral limit of σ_w/u_* for the wind sea data is somewhat lower than the full data set. This behavior was confirmed using two other data sets: tower measurements over Lake Ontario collected during the "Water-Air Vertical Exchange Study" (WAVES) [*Drennan et al.*, 1999] and measurements from a small ship in the coastal Atlantic during SWADE [*Donelan et al.*, 1997]. The wind sea data from the three experiments are consistent (not shown), and show a neutral value of 1.17. We denote the parameterization of equation (7) with a neutral value of 1.17 as the wind sea relation (WS). The WS curve is shown in Figure 1b.

[24] Although the ASIS data lie significantly above the *Högström* [1990] and *Rutgersson et al.* [2001] curves, it is not possible to distinguish between the two Panofsky curves. *Smedman et al.* [1999] noted a dependence of σ_w/u_* on sea state, with wind sea values around 1.17 at 10 and 18 m;

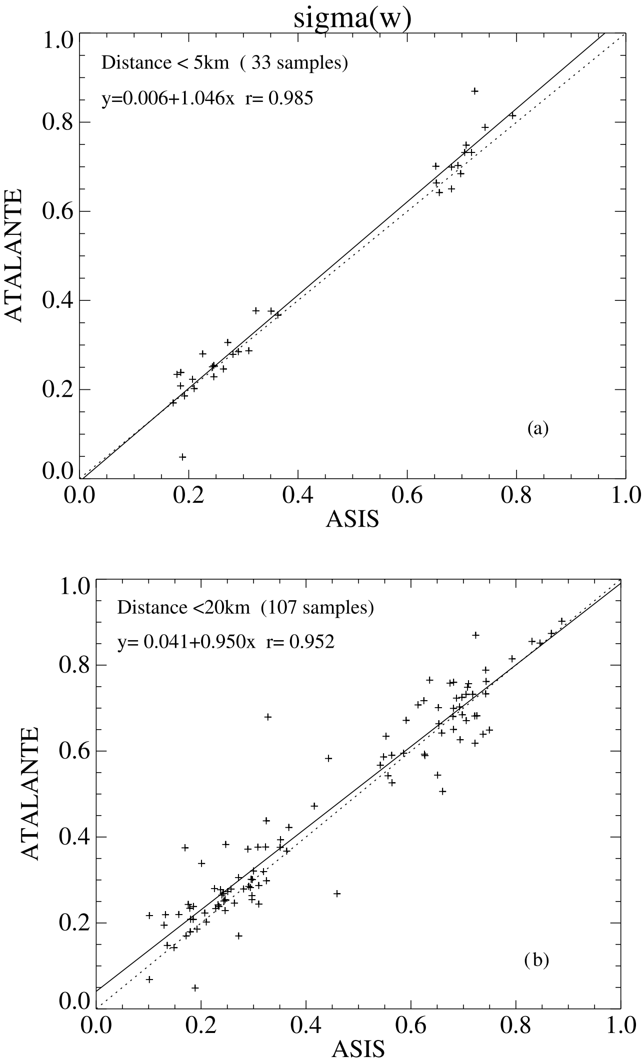

Figure 2. Comparison between the standard deviation of vertical velocity fluctuations σ_w, measured on the R/V *L'Atalante* and ASIS when the two platforms were within (a) 5 km and (b) 20 km of each other.

slightly higher (1.26) at 26 m, in agreement with the height dependence found by *Högström* [1990]. Most of the wind sea data *of Smedman et al.* were collected during a gale, with wind speeds over 12 m/s. During this time conditions were near neutral, and wind sea values agree well with the ASIS data (and WS).

[25] In the presence of swell, the *Smedman et al.* [1999] σ_w/u_*'s were significantly higher, typically between 1.5 and 2.5. Although Smedman et al. interpreted this increase in σ_w/u_* as a wave age effect, neither the *Rutgersson et al.* [2001] data nor the present ASIS data show such an abrupt transition. If we consider the near neutral data, $z/L > -0.1$ (Figure 1b), the higher values of σ_w/u_* are associated with swell, but the median value is not significantly above that of the wind sea: there are few near-neutral values of σ_w/u_* in the range reported by Smedman et al. The high σ_w/u_* values reported by Smedman et al. may be attributed in part to stability, and not only wave age. Their swell data were collected during a single event following a gale, with wind speeds around 4 m/s. At these light winds, conditions may no longer be neutral. There remains the need for further investigations into this topic. In conclusion, although there is evidence for an additional sea state dependence, the ASIS data are described well by equation (7) for unstable cases.

[26] In Figure 2 we compare estimates of σ_w measured on ASIS and R/V *L'Atalante* when the two platforms were close to each other. The two panels show different thresholds for distance between the platforms: 5 km in Figure 2a and 20 km in Figure 2b. The correlation is high for both plots. Also, the two regressions are close to unity, indicating that σ_w estimates on *L'Atalante* do not appear to be significantly affected by turbulent flow distortion around the ship. Though the number of points available when the distance was less than 5 km is small, the scatter among these data is very small which suggests that part of the increase in scatter for larger distances is due to spatial heterogeneity. However the bias at the origin in Figure 2b warrants further analysis. For that purpose, the vertical velocity standard deviation is divided by the stability and height dependence $(1-3z/L)^{1/3}$ of the Panofsky parameterization to account for

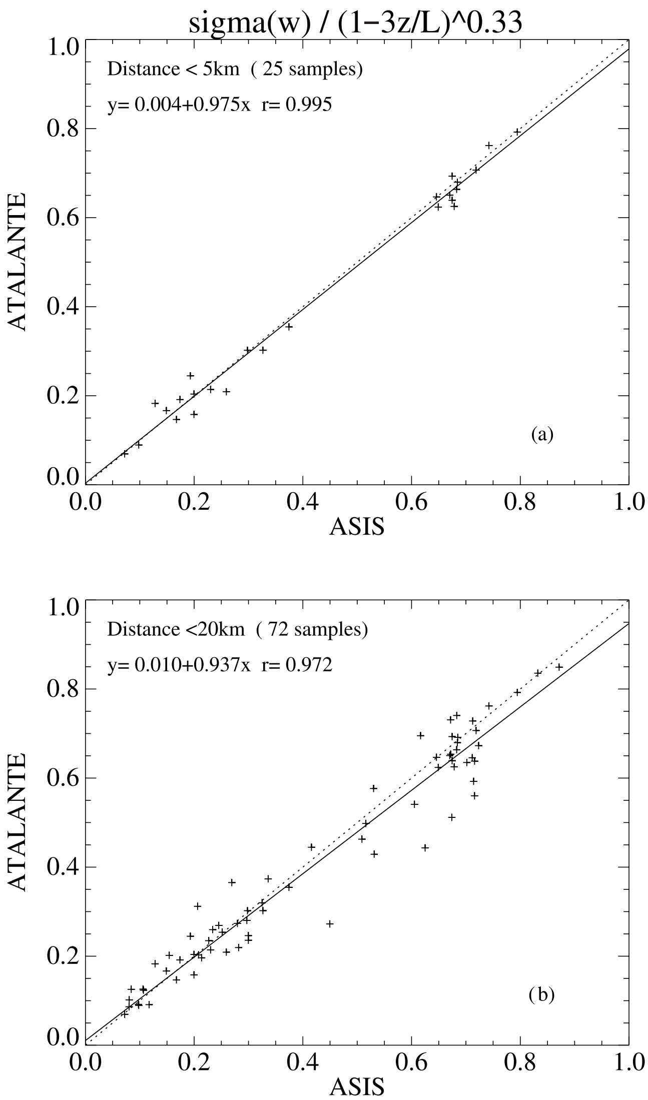

Figure 3. As in Figure 2 but for $\sigma_w/(1-3z/L)^{1/3}$.

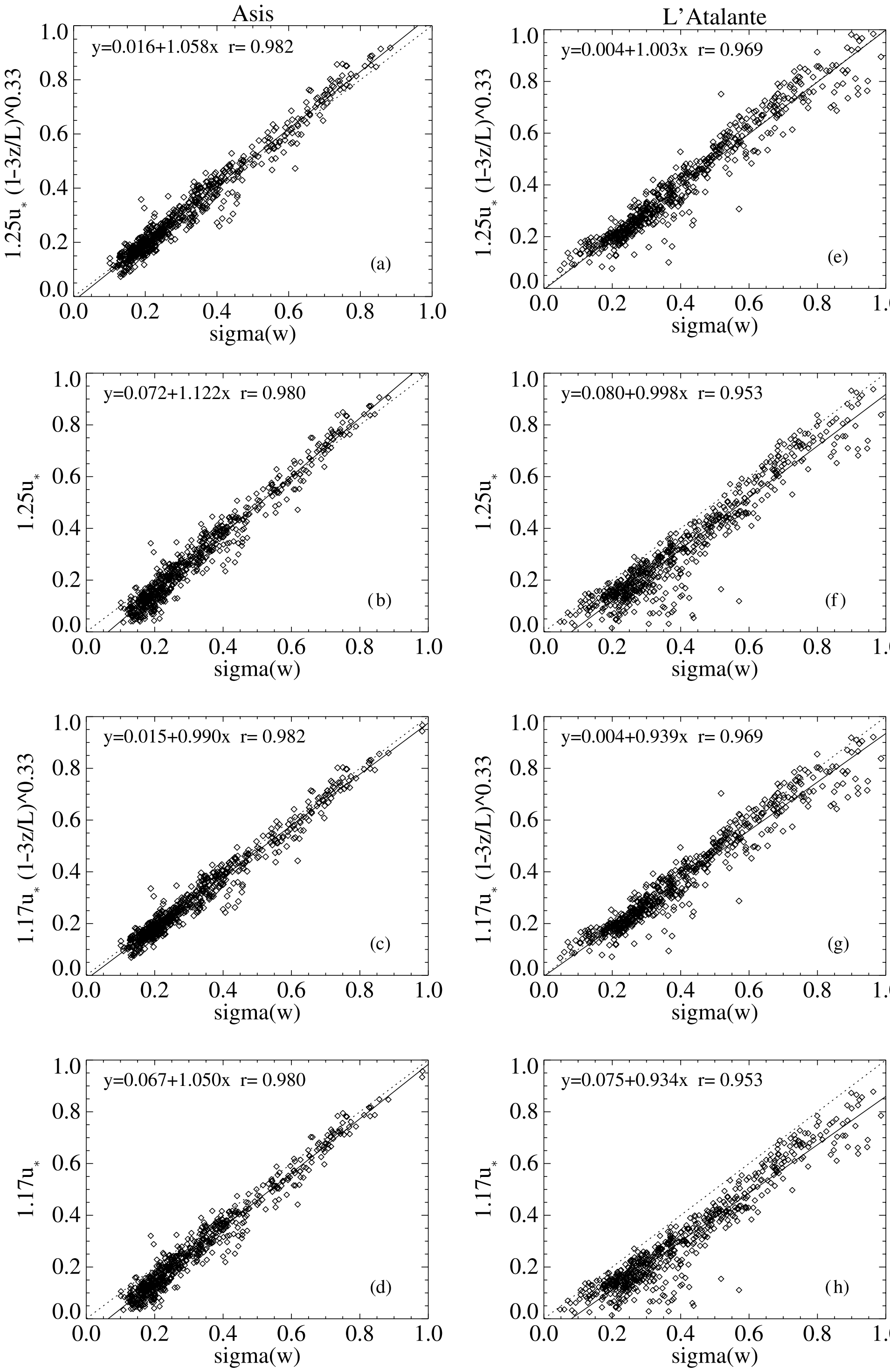

Figure 4. Vertical velocity variance parameterizations (a–d) for ASIS (eddy correlation fluxes) and (e–h) for R/V *L'Atalante* (inertial dissipation fluxes). For ASIS/*L'Atalante* data the plots show measured σ_w versus estimates using the Panofsky parameterization (Figures 4a and 4e), 1.25 u_* (Figures 4b and 4f), the WS parameterization (Figures 4c and 4g), and 1.17 u_* (Figures 4d and 4h).

different measurement heights of the two platforms (17.8 m for *L'Atalante*; 7 m for ASIS). These data are plotted in Figure 3. It is evident that the scatter is decreased and the bias at the origin is reduced (Figure 3b). This supports the view that this bias was due to stability and height effects [cf. *Högström*, 1990].

[27] In Figures 4a–4d are four scatterplots of σ_w for the ASIS buoy compared with: the Panofsky parameterization (Figure 4a), denoted PA, with 1.25 u_*, to test the z/L dependence (Figure 4b), with WS (Figure 4c), and with 1.17 u_*, to test the z/L stability dependence of WS (Figure 4d). The PA and WS parameterizations are equivalent, with

Table 1. Synthesis of the Different Regressions of Figure 4[a]

Parameterization	*L'Atalante* (543)	ASIS (633)
$1.25u_*(1-3z/L)^{1/3}$	a = 1.003, b = −0.004, r = 0.97	a = 1.058, b = −0.016, r = 0.98
$1.25u_*$	a = 0.998, b = −0.08, r = 0.95	a = 1.122, b = −0.072, r = 0.98
$1.17u_*(1-3z/L)^{1/3}$	a = 0.939, b = −0.004, r = 0.97	a = 0.990, b = −0.015, r = 0.98
$1.17u_*$	a = 0.934, b = −0.075, r = 0.95	a = 1.050, b = −0.067, r = 0.98

[a]Here a, slope; b, bias at the origin; r, correlation.

the same correlation coefficients, and slopes statistically indistinguishable from identity. It should also be noted (Figures 4b and 4d) that the dependence on stability is small for the ASIS data, since a change (relative to Figures 4a and 4c) is observed only for small σ_w values. Figures 4e–4f show the same representations for the R/V *L'Atalante* data. Here the Panofsky parameterization (Figure 4e) yields the minimum bias, with a slope closer to identity, compared with the WS parameterization (Figure 4g). For the *L'Atalante* data, the regressions improve significantly when the stability term is taken into account: compare Figures 4f and 4h with Figures 4e and 4g. As pointed out above, for the ASIS data this effect is very small.

[28] The fact that the stability plays a more important role on R/V *L'Atalante* than on ASIS is due to the different measurement heights, respectively 18 m and 7 m. The results shown in Figure 4 can be synthesized for each platform and the different parameterizations in the Table 1 with the number of points used for the analysis indicated for each platform.

[29] The significant dependence of the R/V *L'Atalante* σ_w data on stability justifies the use of the σ_w parameterization to constrain the IDM method. For the ASIS buoy with measurements at 7 m, u_* can be directly estimated from σ_w.

3.3. Combining the TKE Budget and σ_w Parameterization

[30] Hereafter the normalized TKE budget and the normalized vertical velocity standard deviation parameterization are combined to yield two equations for the two unknown quantities u_* and z/L.

$$\varepsilon\kappa z/u_*^3 = \varphi_m(z/L) - z/L, \tag{8}$$

$$\sigma_w/u_* = \varphi_w(z/L), \tag{9}$$

where φ_m is the universal wind shear function parameterized with $\varphi_m = (1-16z/L)^{-1/4}$ [*Dyer and Hicks*, 1970] and φ_w is given by (7).

[31] Using (9) to substitute for u_* in (8), we first solve

$$P(z/L) \equiv \left[\varepsilon\kappa z/\sigma_w^3\right]\varphi_w^3(z/L) - \varphi_m(z/L) + z/L = 0, \tag{10}$$

for z/L, and then determine u_* from (9). Note here that solutions correspond to the intersection between the function $Q(z/L) = [(\varphi_m(z/L) - z/L)/\varphi_w^3(z/L)]$ and experimental values of $[\varepsilon\kappa z/\sigma_w^3]$.

[32] The nonlinear equation (10) is numerically solved by Newton's method, initiated with a bulk estimate for z/L. Of the 657 unstable runs considered, 459 have a solution for P(z/L) = 0. The majority of cases which do not satisfy P = 0 correspond to large initial values of −z/L, for which u_* is not an appropriate scaling parameter. For large −z/L values, σ_w becomes more dependent on the virtual temperature flux which can be directly estimated from σ_w and then be substituted directly into (5) to get u_*. See *Kader and Yaglom* [1990] for further details.

[33] Figure 5a presents the comparison between u_* calculated from the iterative IDM and from the system of equations. Figure 5 shows that the proposed method is relevant and does not present a large bias when compared to the iterative IDM method though there is some scatter. We also choose to use the same system (9) and (10) but to

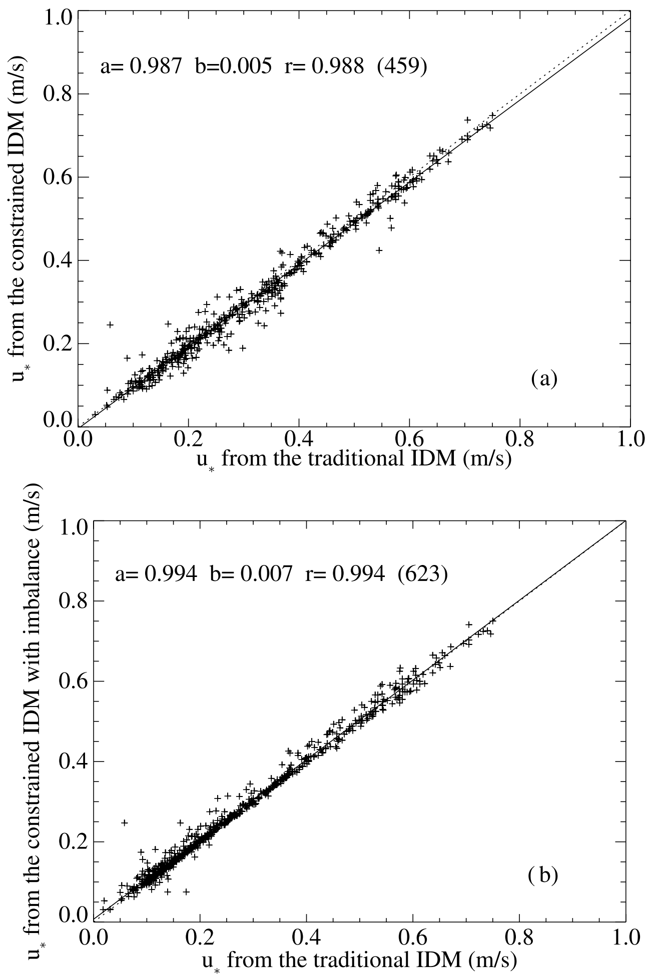

Figure 5. (a) Friction velocity u_* from the constrained IDM versus u_* from the traditional IDM. (b) Friction velocity u_* from the constrained IDM versus u_* from the traditional IDM with the *Dupuis et al.* [1997] imbalance parameterization.

introduce in (10) an imbalance equal to $-0.5z/L$ as proposed by *Dupuis et al.* [1997]. Figure 5b shows the output of the system of equations with the parameterized imbalance. Several facts warrant mentioning. The correlation is a little better with the imbalance term, with a slope closer to identity. The scatter is reduced particularly for friction velocities smaller than 0.4 m/s. Also, more points are available as solutions of the system of equations. For the 657 cases used, 623 have solutions, i.e., 164 points more than for the case without imbalance. However, as the friction velocities from the IDM are obtained with the imbalance parameterization, following *Dupuis et al.* [1997], these results do not give a new validation of the imbalance term but rather indicate a consistency of the IDM results.

4. Conclusions

[34] On both a large vessel, the *L'Atalante*, and a research buoy, the variance of the vertical velocity fluctuations, σ_w, obtained after corrections for platform motion, are well described by the Panofsky parameterization for unstable stratification. It has been found that a modified inertial dissipation method, combining the TKE budget and the Panofsky parameterization for σ_w/u_*, can be used to calculate friction velocity and stability (z/L) on board typical research vessels. The new method has an advantage over the traditional IDM in that it avoids the convergence problems often encountered in the past. For large $-z/L$ values the new method cannot be used since it requires a precision on σ_w which cannot be experimentally achieved. However at these large $-z/L$ values, the Panofsky parameterization provides an estimate of the virtual heat flux. A suggestion is then to use this estimate to initiate an iteration in L. However in these highly unstable conditions, *Yelland et al.* [1998] have expressed doubts that the IDM itself is applicable. Indeed at low winds (usually associated with high $|z/L|$) there is evidence that sea state effects, in particular swell, can cause significant errors in the ID estimates of u_* compared to EC values. When swell dominates the wave field, the IDM method can not be used [see, e.g., *Donelan et al.*, 1997; *Drennan et al.*, 1999; *Grachev and Fairall*, 2001]. With the ASIS data here, collected much closer to the surface than those of R/V *L'Atalante*, a sea state effect is observed in the vertical velocity variance. This is currently an area of active research [e.g., *Edson and Fairall*, 1998].

[35] At this stage, the validation of the constrained inertial dissipation method has not been fully achieved. The originality of this study concerns more a methodology. Further validations of the performance of the conventional and constrained IDM should be undertaken using direct ECM measurements of u_*. This was not possible here due to the lack of stability dependence found with the ASIS data collected at 7m above sea level. ECM data from a higher level (or with larger $|z/L|$) are needed.

[36] **Acknowledgments.** The authors thank INSU (PATOM), PNTS, and the European community (MATER programme) for their support during the FETCH 1998 experiment. This work was done in the context of the AUTOFLUX project from the MAST programme of the European community: their support is acknowledged. We are also indebted to our colleagues at CETP, Météo-France, RSMAS and NWRI who participated in the campaign.

References

Anctil, F., M. A. Donelan, W. M. Drennan, and H. C. Graber, Eddy correlation measurements of air-sea from a discus buoy, *J. Atmos. Oceanic Technol.*, *11*, 1144–1150, 1994.

Busch, N. E., On the mechanics of atmospheric turbulence, in *Workshop on micrometeorology*, edited by D. A. Haugen, pp. 1–65, Am. Meteorol. Soc., Boston, Mass., 1972.

Dardier, G., H. Dupuis, and A. Weill, AUTOFLUX annual report, Southampton Oceanogr. Inst., Southampton, UK, Nov. 1999.

De Bruyn, H. A. R., R. J. Ronda, and B. J. H. Van de Wiel, Approximate solutions for the Obukhov length and the surface fluxes in terms of bulk Richardson numbers, *Boundary Layer Meteorol.*, *95*, 145–157, 2000.

Donelan, M. A., J. Hamilton, and W. H. Hui, Directional spectra of wind generated waves, *Philos. Trans. R. Soc. London, Ser. A*, *315*, 509–562, 1985.

Donelan, M. A., W. M. Drennan, and K. Katsaros, The air-sea momentum flux in conditions of wind sea and swell, *J. Phys. Oceanogr.*, *27*, 2087–2099, 1997.

Drennan, W. M., K. K. Kahma, and M. A. Donelan, On momentum flux and velocity spectra over waves, *Boundary Layer Meteorol.*, *92*, 489–515, 1999.

Drennan, W. M., H. C. Graber, D. Hauser, and C. Quentin, On the wave age dependence of wind stress over pure wind seas, *J. Geophys. Res.*, *108*, doi:10.1029/2000JC000715, in press, 2003.

Dupuis, H., A. Weill, P. K. Taylor, and K. B. Katsaros, Turbulent heat flues by profile and inertial dissipation methods: Analysis of the atmospheric surface layer from shipboard measurement during the SOFIA/ASTEX and SEMAPHORE experiments, *Ann. Geophys.*, *13*, 1065–1074, 1995.

Dupuis, H., P. K. Taylor, A. Weill, and K. B. Katsaros, Inertial dissipation method applied to derive turbulent fluxes over the ocean during SOFIA/ASTEX and SEMAPHORE experiments with low to moderate wind speeds, *J. Geophys. Res.*, *102*, 21,115–21,129, 1997.

Dupuis, H., C. Guerin, D. Hauser, A. Weill, P. Nacass, W. M. Drennan, S. Cloché, and H. C. Graber, Impact of flow distortion corrections on turbulent fluxes estimated by the inertial dissipation method during the FETCH experiment on R/V *L'Atalante*, *J. Geophys. Res.*, *108*, doi:10.1029/2001JC001075, in press, 2003.

Dyer, A. J., and B. B. Hicks, Flux-gradient relationships in the constant flux layer, *Q. J. R. Meteorol. Soc.*, *96*, 715–721, 1970.

Edson, J. B., and C. W. Fairall, Similarity relationships in the marine atmospheric surface layer for terms in the TKE and scalar variance budget, *J. Atmos. Sci.*, *55*, 2311–2328, 1998.

Edson, J. B., C. W. Fairall, P. G. Mestayer, and S. E. Larsen, A study of the inertial-dissipation method for computing air-sea fluxes, *J. Geophys. Res.*, *96*, 10,689–10,711, 1991.

Fairall, C. W., and S. E. Larsen, Inertial-dissipation methods and turbulent fluxes at the air-ocean interface, *Boundary Layer Meteorol.*, *34*, 287–301, 1986.

Fairall, C. W., E. F. Bradley, D. P. Rogers, J. B. Edson, and G. S. Young, Bulk parameterization of air-sea fluxes for Tropical Ocean-Global atmosphere Coupled-Ocean Atmosphere Response Experiment, *J. Geophys. Res.*, *101*, 3747–3764, 1996.

Graber, H. C., E. A. Terray, M. A. Donelan, W. M. Drennan, J. C. Van Leer, and D. B. Peters, ASIS: A new air-sea interaction spar buoy: Design and performance at sea, *J. Atmos. Oceanic Technol.*, *17*, 708–720, 2000.

Grachev, A. A., and C. W. Fairall, Dependence of the Monin-Obukhov stability parameter on the bulk Richardson number over the ocean, *J. Appl. Meteorol.*, *36*, 406–414, 1997.

Grachev, A. A., and C. W. Fairall, Upward momentum transfer in the marine boundary layer, *J. Phys. Oceanogr.*, *31*, 1698–1711, 2001.

Hauser, D., H. Dupuis, X. Durrieu de Madron, C. Estournel, C. Flamant, J. Pelon, P. Queffelou, and J. M. Lefevre, La campagne FETCH, Etude des échanges océan-atmosphère dans le golfe du Lion, *Meteorologie*, *8*(29), 14–31, 2000.

Hauser, D., et al., The FETCH experiment: An overview, *J. Geophys. Res.*, *108*(C3), 8053, doi:10.1029/2001JC001202, 2003.

Högström, U., Analysis of Turbulence structure in the surface layer with a modified similarity formulation for near neutral conditions, *J. Atmos. Sci.*, *90*, 1849–1972, 1990.

Kader, B. A., and A. M. Yaglom, Mean fields and fluctuation moments in unstably stratified turbulent boundary layers, *J. Fluid. Mech.*, *212*, 637–662, 1990.

Kaimal, J. C., and J. J. Finningan, *Atmospheric Boundary Layer Flows, Their Structure and Measurement*, 289 pp., Oxford Univ. Press, New York, 1994.

Launiainen, J., Derivation of the relationship between the Obukhov stability parameter and the bulk Richardson number over the ocean, *Boundary Layer Meteorol.*, *76*, 165–179, 1995.

Large, W. G., and S. Pond, Open ocean momentum flux measurements in moderate to strong winds, *J. Phys. Oceanogr.*, *11*, 324–336, 1981.

Large, W. G., and S. Pond, Sensible and latent heat flux measurements over the ocean, *J. Phys. Oceanogr.*, *12*, 464–482, 1982.

Merry, M., and H. A. Panofsky, Statistics of vertical motion over land and water, *Q. J. R. Meteorol. Soc.*, *102*, 255–260, 1974.

Nacass, P., Shipborne wind measurements corrected for airflow distortion by computational fluid dynamics, paper presented at 11th Conference Symposium on Meteorological Instrumentation and Observation, Am. Meteorol. Soc, Albuquerque, 2001.

Oost, W. A., C. W. Fairall, J. B. Edson, S. D. Smith, R. J. Anderson, J. A. B. Wills, K. B. Katsaros, and J. DeCosmo, Flow distortion calculations and their application in HEXMAX, *J. Atmos. Oceanic Technol.*, *11*, 366–386, 1994.

Panofsky, H. A., Tower micrometeorology, in *Workshop in Micrometeorology*, edited by D. Haugen, pp. 151–176, Am. Meteorol. Soc., Boston, Mass., 1972.

Panofsky, H. A., and J. Dutton, *Atmospheric Turbulence: Models And Methods for Engineering Applications*, 161 pp., John Wiley, New York, 1984.

Panofsky, H. A., H. Tennekes, D. H. Lenschow, and J. C. Wyngaard, The characteristics of turbulent velocity components in the surface layer under convective conditions, *Boundary Layer Meteorol.*, *11*, 355–361, 1977.

Rutgersson, A., A. S. Smedman, and U. Högström, Use of conventional stability parameters during swell, *J. Geophys. Res.*, *106*, 27,117–27,134, 2001.

Smedman, A. S., U. Högström, H. Bergström, A. Rutgersson, K. Kahma, and H. Pettersson, A case study of air-sea interaction during swell conditions, *J. Geophys. Res.*, *104*, 25,833–25,852, 1999.

Smith, S. D., *Instruments and Methods in air-sea interactions*, pp. 65–80, Plenum, New York, 1980.

Smith, S. D., and R. J. Anderson, Spectra of humidity, temperature and wind over the sea at Sable Island, *J. Geophys. Res.*, *89*, 2029–2040, 1984.

Taylor, P. K., and M. J. Yelland, On the apparent "imbalance" term in the turbulent kinetic energy budget, *J. Atmos. Oceanic Technol.*, *17*, 82–89, 2000.

Weill, A., C. Klapisz, B. Strauss, F. Baudin, C. Jaupart, P. Van Grunderbeeck, and J. P. Goutorbe, Measuring heat flux and structure functions of temperature fluctuations with an acoustic Doppler sodar, *J. Appl. Meteorol.*, *19*, 337–346, 1980.

Wyngaard, J. C., O. R. Coté, and Y. Izumi, Local free convection, similarity and the budget of shear stress and heat flux, *J. Atmos. Sci.*, *28*, 1171–1182, 1971.

Wyngaard, J. C., O. R. Coté, and K. S. Rao, Modelling the atmospheric boundary layer, *Adv. Geophys.*, *18A*, 193–211, 1974.

Yelland, M. J., and P. K. Taylor, Wind stress measurements from the open ocean, *J. Phys. Oceanogr.*, *26*, 541–558, 1996.

Yelland, M. J., P. K. Taylor, I. E. Consterdine, and M. H. Smith, The use of the inertial dissipation technique for shipboard wind stress determination, *J. Atmos. Oceanic Technol.*, *11*, 1083–1108, 1994.

Yelland, M. J., B. I. Moat, P. K. Taylor, R. W. Pascal, J. Hutchings, and V. C. Cornell, Wind stress measurements from the open ocean corrected for airflow distortion by the ship, *J. Phys. Oceanogr.*, *28*, 1511–1526, 1998.

Zeng, X., M. Zhao, and R. E. Dickinson, Intercomparison of bulk aerodynamic algorithms for the computation of sea surface fluxes using TOGA COARE and TAO data, *J. Clim.*, *11*, 2628–2644, 1998.

S. Brachet, C. Guerin, F. Lohou, and A. Weill, CETP, 10–12 Avenue de l'Europe, 78140 Velizy, France.

G. Dardier, Davidson Laboratory, Stevens Institute of Technology, Hoboken, NJ 07030, USA. (gdardier@stevens-tech.edu)

H. Dupuis and R. Pedreros, DGO, Université Bordeaux I, UMR58, Avenue des Facultés, 33405 Talence, France.

W. M. Drennan, RSMAS, University of Miami, 4600 Rickenbacker Causeway, Miami, FL 33149, USA.