Estimating Overwater Turbulence Intensity from Routine Gust-Factor Measurements

S. A. HSU AND BRIAN W. BLANCHARD

Coastal Studies Institute, Louisiana State University, Baton Rouge, Louisiana

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ABSTRACT

For overwater diffusion estimates the Offshore and Coastal Dispersion (OCD) model is preferred by the U.S. Environmental Protection Agency. The U.S. Minerals Management Service has recommended that the OCD model be used for emissions located on the outer continental shelf. During southerly winds over the Gulf of Mexico, for example, the pollutants from hundreds of offshore platforms may affect the gulf coasts. In the OCD model, the overwater plume is described by the Gaussian equation, which requires the computation of σ_y and σ_z, which are, in turn, related to the turbulence intensity, overwater trajectory, and atmospheric stability. On the basis of several air–sea interaction experiments [the Barbados Oceanographic and Meteorological Experiment (BOMEX), the Air-Mass Transformation Experiment (AMTEX), and, most recently, the Tropical Ocean and Global Atmosphere Coupled Ocean–Atmosphere Response Experiment (TOGA COARE)] and the extensive datasets from the National Data Buoy Center (NDBC), it is shown that under neutral and stable conditions the overwater turbulence intensities are linearly proportional to the gust factor (G), which is the ratio of the wind gust and mean wind speed at height z (U_z) as reported hourly by the NDBC buoys. Under unstable conditions, it is first shown that the popular formula relating the horizontal turbulence intensity ($\sigma_{u,v}/u_*$, where u_* is the friction velocity) to the ratio of the mixing height (h) and the buoyancy length (L) (i.e., h/L) suffers from a self-correlation problem and cannot be used in the marine environment. Then, alternative formulas to estimate the horizontal turbulence intensities ($\sigma_{u,v}/U_z$) using G are proposed for practical applications. Furthermore, formulas to estimate u_* and z/L are fundamentally needed in air–sea interaction studies, in addition to dispersion meteorology.

1. Introduction

According to Glickman (2000), the turbulence intensity is defined as the ratio of the root-mean-square of the eddy velocity to the mean wind speed; in general, it is a quantity that characterizes the intensity of gusts in the airflow. Mathematically, the turbulent intensities in horizontal and vertical directions are σ_u/U, σ_v/U, and σ_w/U, where σ_u, σ_v, and σ_w are the standard deviations of velocity fluctuations in the x, y, and z directions, respectively, and U is the mean wind speed. Note that the mean wind speed, rather than the particular component mean velocity, is used in the definition of turbulence intensities (Arya 1999). In air pollution meteorology and dispersion, these turbulence intensities are related to the particle dispersion parameters in the y and z directions (i.e., σ_y and σ_z), respectively (see e.g., Panofsky and Dutton 1984; Zannetti 1990; Arya 1999).

In the atmospheric surface (or constant flux) layer (i.e., within 100 m above the surface where the variation of vertical turbulent flux with altitude is less than 10% of its magnitude), the turbulence intensities can be estimated (see Arya 1999) for neutral and stable conditions as follows:

for longitudinal (downwind) turbulence intensity

$$\frac{\sigma_u}{U_z} \simeq 2.5\frac{u_*}{U_z}, \tag{1}$$

for crosswind turbulence intensity

$$\frac{\sigma_v}{U_z} \simeq 1.9\frac{u_*}{U_z}, \tag{2}$$

for vertical turbulence intensity

$$\frac{\sigma_w}{U_z} \simeq 1.3\frac{u_*}{U_z}, \quad \text{and} \tag{3}$$

for unstable and convective conditions,

for horizontal turbulence intensity

$$\frac{\sigma_{u,v}}{U_z} = \left(\frac{u_*}{U_z}\right)\left(12 - 0.5\frac{h}{L}\right)^{1/3}, \quad \text{for } \frac{h}{L} < 0, \quad \text{and} \tag{4}$$

for vertical turbulence intensity

$$\frac{\sigma_w}{U_z} = 1.3\left(\frac{u_*}{U_z}\right)\left(1 - 3\frac{z}{L}\right)^{1/3}, \quad \text{for } \frac{z}{L} \leq 0, \tag{5}$$

Corresponding author address: Dr. S. A. Hsu, Coastal Studies Institute, 308 Howe-Russell Geoscience Bldg., Louisiana State University, Baton Rouge, LA 70803.
E-mail: sahsu@lsu.edu

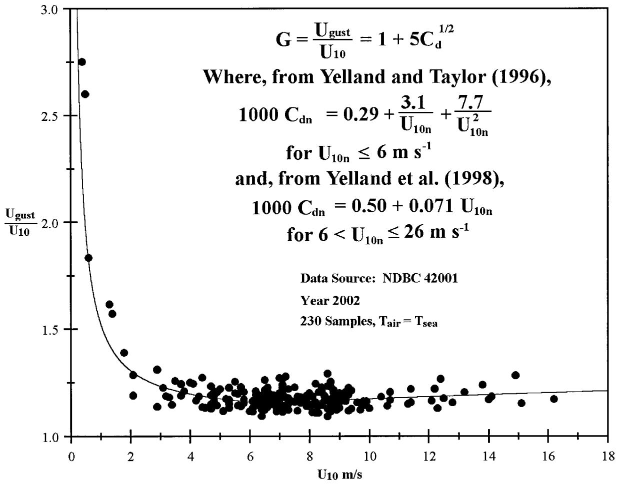

FIG. 1. A relationship between the gust factor and the drag coefficient under neutral conditions in the Gulf of Mexico.

where U_z is the wind speed at height z, u_* is the friction velocity, h is the boundary layer height, and L is the Obukhov (buoyancy) length. Note that the atmospheric stability classification by Pasquill categories using σ_w/U_z, including its range, is provided in Zannetti (1990, his Table 7-1, p. 148).

In boundary layer meteorology

$$C_d = \left(\frac{u_*}{U_z}\right)^2, \tag{6}$$

where C_d is the drag coefficient, which is related to the roughness length (Z_0). Because Z_0 over land is fixed for a given environment, C_d is also known. However, in the marine environment, Z_0 varies with the wind, sea, and swell characteristics, in addition to the atmospheric stability parameter (z/L) (see e.g., Hsu 1988). As stated previously, the turbulence intensity is related to the wind gust. Because the gust factor ($G = U_{gust}/U_z$) is measured routinely by the National Data Buoy Center (NDBC) buoys, it is the purpose of this study to find practical formulas to estimate turbulence intensities using G. For more detail about NDBC's measurement program, see their Web site (online at seaboard.ndbc.noaa.gov).

2. For neutral and stable conditions

At the air–sea interface when the temperature difference between the sea (T_{sea}) and air (T_{air}) is zero (i.e., $T_{sea} = T_{air}$), the stability parameter z/L also is zero (see e.g., Hsu and Blanchard 2003). Under these conditions, neutral stability prevails. With this criterion, we select NDBC buoy 42001 in the central Gulf of Mexico for our analysis. The anemometer was located at the standard 10 m above the sea surface so that $U_z = U_{10}$. When $T_{sea} = T_{air}$, the neutral wind is $U_{10} = U_{10n}$ at 10 m. In Fig. 1, the gust factor is plotted against U_{10} and the recent drag coefficient formulations based on extensive datasets in the open ocean by Yelland and Taylor (1996) and Yelland et al. (1998) are superimposed. Excellent agreement between the gust factor and the drag coefficient is reached, such that

$$G = 1 + 5C_d^{1/2}, \quad \text{or} \tag{7}$$

$$\frac{u_*}{U_{10}} = 0.2(G - 1). \tag{8}$$

Further verification of Eq. (8) is provided in Hsu (2003).

Now, substituting Eq. (8) into Eqs. (1)–(3), one obtains

$$\frac{\sigma_u}{U_z} = 0.50(G - 1), \tag{9}$$

$$\frac{\sigma_v}{U_z} = 0.38(G - 1), \quad \text{and} \tag{10}$$

$$\frac{\sigma_w}{U_z} = 0.26(G - 1). \tag{11}$$

Because hourly values of G are provided by NDBC buoys, Eqs. (9)–(11) are recommended for practical applications for neutral and stable conditions.

3. For unstable conditions

a. The self-correlation problem between ($\sigma_{u,v}/u_*$) and (h/L)

For unstable conditions when $T_{sea} > T_{air}$, Eq. (4) has been used extensively in air pollution meteorology (see, e.g., Panofsky and Dutton 1984; Zannetti 1990; Arya 1999). However, this equation suffers from self-correlation as discussed below.

In the convective boundary layer (see e.g., Kaimal and Finnigan 1994, p. 22),

$$\frac{w_*}{u_*} = \left(\frac{h}{\kappa|L|}\right)^{1/3}, \tag{12}$$

where w_* is the convective velocity and κ (=0.4) is the von Kármán constant. Notice that the same equation is provided in Arya [1999, p. 101, Eq. (4.64)], but that there is an error such that $\kappa^{1/3}$ should be $\kappa^{-1/3}$.

Now, by rearranging Eq. (4) we have

$$\left(\frac{\sigma_{u,v}}{u_*}\right)^3 = 12 + 0.5\left|\frac{h}{L}\right|, \tag{13}$$

and from Eq. (12)

$$\left|\frac{h}{L}\right| = \kappa\left(\frac{w_*}{u_*}\right)^3. \tag{14}$$

Substituting Eq. (14) into Eq. (13), one gets

$$\left(\frac{\sigma_{u,v}}{u_*}\right)^3 = 12 + 0.5\kappa\left(\frac{w_*}{u_*}\right)^3. \tag{15}$$

TABLE 1. Measured and derived boundary layer parameters during the Barbados Oceanographic and Meteorological Experiment (BOMEX) and pre-BOMEX cruises (Pond et al. 1971); $z = 8$ m and $A^* = (\sigma_u/U_z)/(1 + 3|z/L|)^{1/3}$.

Run	U_z (m s^{-1})	σ_u (cm s^{-1})	σ_v (cm s^{-1})	σ_w (cm s^{-1})	$-\frac{z}{L}$	$\frac{\sigma_u}{\sigma_v}$	$\frac{\sigma_w}{\sigma_u}$	A^*
Oregon State University								
1	5.78	69.1	70.7	29.0	0.20	0.98	0.42	0.102
2	6.60	75.7	80.0	32.3	0.12	0.95	0.43	0.104
3	4.78	55.1	60.1	25.8	0.18	0.92	0.47	0.100
4	6.05	109.0	129.0	38.1	0.16	0.84	0.35	0.158
5	4.65	58.5	67.1	25.2	0.27	0.87	0.43	0.103
6	5.55	59.5	72.1	26.4	0.23	0.83	0.44	0.090
7	7.22	81.5	79.0	32.6	0.20	1.03	0.40	0.097
8	5.92	79.3	96.5	35.1	0.14	0.82	0.44	0.119
9	7.21	71.7	100.0	42.1	0.11	0.72	0.59	0.090
10	6.79	67.3	68.0	29.6	0.11	0.99	0.44	0.090
11	5.31	56.3	57.1	24.6	0.22	0.99	0.44	0.090
12	6.51	67.5	63.0	32.8	0.17	1.07	0.49	0.090
13	5.86	63.4	69.6	30.4	0.22	0.91	0.48	0.091
14	4.97	61.7	71.4	27.0	0.33	0.86	0.44	0.099
15	6.78	73.1	74.5	33.8	0.15	0.98	0.46	0.095
University of British Columbia								
1	6.88	75.3	70.6	33.9	0.16	1.07	0.45	0.096
2	6.10	73.1	74.1	30.0	0.15	0.99	0.41	0.083
3	5.48	56.0	78.0	26.0	0.24	0.72	0.46	0.085
4	5.01	55.1	59.4	22.3	0.20	0.93	0.40	0.094
5	3.93	40.1	44.9	22.2	0.40	0.89	0.55	0.078
					Mean =	0.92	0.45	0.098
					Std dev =	0.10	0.05	0.017
					Coef of variation =	11%	11%	17%

From the Tropical Ocean and Global Atmosphere Coupled Ocean–Atmosphere Response Experiment (TOGA COARE) (see Fairall et al. 1996), it is given that

$$\beta^2 = \frac{\sigma_u^2 + \sigma_v^2}{w_*^2}, \tag{16}$$

where β (=1.25) is the gustiness parameter.

Under unstable conditions, $\sigma_u \simeq \sigma_v$ (see, e.g., Arya 1999 and our Table 1), Eq. (16) becomes

$$\beta = \frac{\sqrt{2}\sigma_{u,v}}{w_*}.$$

Setting $\beta = 1.25$, we have

$$w_* = 1.13\sigma_{u,v}. \tag{17}$$

Substituting Eq. (17) into Eq. (15) yields

$$\frac{\sigma_{u,v}}{u_*} = 2.56, \tag{18}$$

which is near the values under neutral conditions [see Eq. (1)]. Therefore, the self-correlation between $(\sigma_{u,v}/u_*)$ and (h/L) exists in Eq. (4). An alternative method is provided in the next section.

b. *An alternative relationship between $(\sigma_{u,v}/U_z)$ and (z/L)*

From Eq. (5),

$$\frac{\sigma_w}{u_*} \propto \left(1 + 3\left|\frac{z}{L}\right|\right)^{1/3}, \tag{19}$$

and from Table 1,

$$\frac{\sigma_w}{\sigma_u} = \text{constant}. \tag{20}$$

Therefore, σ_w may be replaced by σ_u numerically, so that

$$\frac{\sigma_u}{U_z} \propto \frac{u_*}{U_z}\left(1 + 3\left|\frac{z}{L}\right|\right)^{1/3}. \tag{21}$$

From Pond et al. (1971), $C_d = 1.52 \times 10^{-3}$, or

$$\frac{u_*}{U_z} = 0.039. \tag{22}$$

Thus,

$$\frac{\sigma_u}{U_z} = A\left(1 + 3\left|\frac{z}{L}\right|\right)^{1/3}, \tag{23}$$

and from Table 1

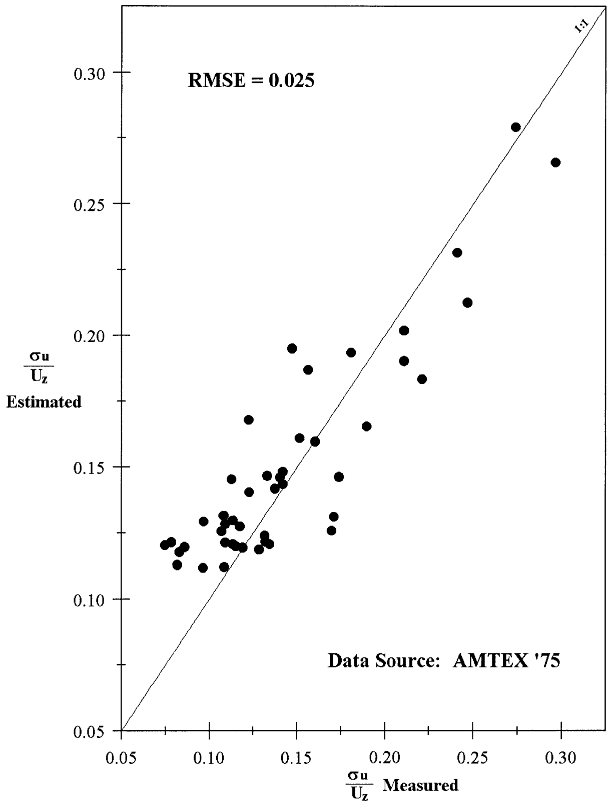

FIG. 2. An analysis of the rmse between Eq. (24) and the AMTEX '75 dataset. The anemometer was located at 18 m above the sea surface.

$$\frac{\sigma_u}{U_z} = 0.10\left(1 + 3\left|\frac{z}{L}\right|\right)^{1/3}. \tag{24}$$

Note that from Smith (1980), when $|z/L| = 0$, σ_u/U_z is also 0.10.

Equation (24) is further verified by another independent dataset. In order to obtain as large a variance of $|z/L|$ as possible, data from the 1975 Air-Mass Transformation Experiment (AMTEX '75) (see Fujitani and Hayashi 1975) are employed. This dataset includes overwater measurements of σ_u, σ_v, and σ_w in addition to wind speed and air and sea temperatures. Note that AMTEX '75 was conducted over the open East China Sea in February 1975 and that values of $|z/L|$ extended to nearly 7. [For the computation of $|z/L|$, see Hsu and Blanchard (2003).] Figure 2 is our result. Note that the vertical axis is the estimated (σ_u/U_z), based on Eq. (24), and the horizontal axis represents the measured (σ_u/U_z). Because the rmse is small in comparison with the data range, we conclude that Eq. (24) is a useful approximation between the longitudinal turbulence intensity (σ_u/U_z) and the stability parameter (z/L).

Now, analogous to the estimation of wind maxima, U_{max} (Panofsky and Dutton 1984),

$$U_{max} = U + 3\sigma_U,$$

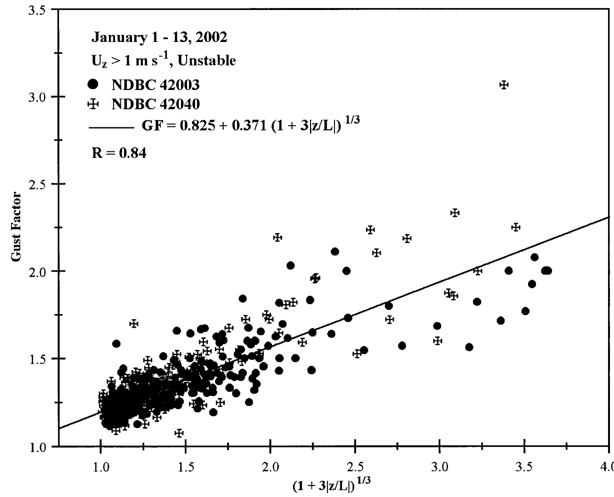

FIG. 3. A relationship between the gust factor and stability parameter over the eastern Gulf of Mexico in Jan 2002.

we postulate that, from buoy measurements, $U_{gust} = U \pm S + C\sigma_U$, or

$$G = \frac{U_{gust}}{U} = \left(1 \pm \frac{S}{U}\right) + C\frac{\sigma_U}{U}, \tag{25}$$

where S represents system accuracy of a buoy, such as instrument accuracy, sensor location, and angles of pitch and roll, and C is a coefficient.

Substituting Eq. (24) into Eq. (25), one gets the following statistical equation:

$$G = \alpha + \gamma\left(1 + 3\left|\frac{z}{L}\right|\right)^{1/3}, \tag{26}$$

where coefficients α and γ must be determined from field measurements. This is done in Fig. 3, which shows that

$$G = 0.825 + 0.371\left(1 + 3\left|\frac{z}{L}\right|\right)^{1/3}, \tag{27}$$

with the correlation coefficient $R = 0.84$.

Note that during the first half of January 2002, cold air moved over the eastern Gulf of Mexico, including the freezing temperature line, which eventually draped over the northern Gulf Coast. To further verify Eq. (27), Fig. 4 is provided. If one accepts these smaller rmses as compared with the gust factor analyzed, one can say that Eq. (27) is useful operationally.

From Eq. (27), we have

$$\left(1 + 3\left|\frac{z}{L}\right|\right)^{1/3} = 2.70(G - 0.825). \tag{28}$$

Substituting Eq. (28) into Eq. (24), one gets

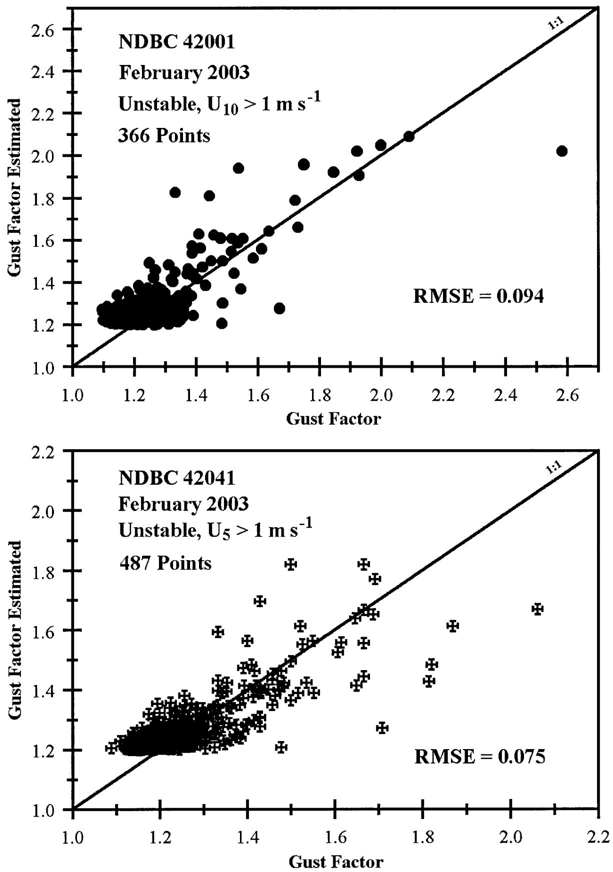

FIG. 4. Verifications of Eq. (27) over the central Gulf of Mexico in Feb 2003; (top) $U_{10} > 1$ m s^{-1} and (bottom) $U_5 > 1$ m s^{-1}.

$$\frac{\sigma_{u,v}}{U_z} = 0.27(G - 0.825). \tag{29}$$

Furthermore, from Eqs. (5) and (28) and Fig. 1, we have

$$\frac{\sigma_w}{U_z} = 0.70(G - 1)(G - 0.825). \tag{30}$$

Equations (29) and (30) are our proposed horizontal and vertical turbulence intensities for unstable conditions.

In addition, from Eqs. (17) and (29), we have

$$w_* = 0.31U_z(G - 0.825), \tag{31}$$

which means that the convective velocity parameter may be computed directly by the surface wind and gust measurements. Thus, Eq. (31) bypasses the utilization of mixing height h that was needed in Eq. (12).

4. Conclusions

Several conclusions may be drawn from this study:

1) Under neutral and stable conditions, the turbulence intensities in the horizontal and vertical directions are linearly related to the gust factor as derived in Eqs. (9)–(11).
2) Under unstable conditions, the ratio of the mixing height and buoyancy length cannot be related to the horizontal turbulence intensity because of the self-correlation problem. Therefore, the popular formula shown in Eq. (4) should not be used for overwater applications.
3) Because of the self-correlation problem stated above, an alternative formula is proposed in Eq. (24), which relates the horizontal turbulence intensity to the surface-based stability parameter (z/L).
4) For practical applications the gust factor is found to be related to (z/L), which was shown statistically in Eq. (27).
5) Using Eq. (27), variations in horizontal and vertical turbulence intensities with the gust factor are proposed for practical applications in Eqs. (29) and (30), respectively.
6) In order to further verify Eqs. (9), (10), (11), (29), and (30), turbulent intensity measurements on an NDBC buoy are necessary. Because U_{10}, G, and z/L are available at buoys 42001, 42002, and 42003, it is recommended that further experiments be conducted at these stations if logistically practical.

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