Standardization of Offshore Surface Wind Speeds

Y. C. HE
Department of Architecture and Civil Engineering, City University of Hong Kong, Kowloon, Hong Kong, China

P. W. CHAN
Hong Kong Observatory, Kowloon, Hong Kong, China

Q. S. LI
Department of Architecture and Civil Engineering, City University of Hong Kong, Kowloon, Hong Kong, China

(Manuscript received 14 October 2015, in final form 25 January 2016)

ABSTRACT

Wind measurement offers an essential data source for a wide range of practices in the fields of meteorology and wind engineering. However, records of surface winds are usually influenced by terrain/topographic effects, and direct usage of raw data may bring in nonignorable errors for follow-up applications. A data-driven standardization scheme was recently proposed by the authors to convert the surface wind measurements over rugged terrain into their potential values corresponding to reference conditions, that is, for neutral winds at a height of 10 m above open flat terrain ($z_o = 0.03$ m). As a complementary part of the preceding work, this study focuses on the standardization of surface wind speeds with marine exposures. The effect of wind strength on the roughness of the sea surface is further taken into account, with emphasis on the difference between deep-ocean and shallow-water cases. As an application example, wind measurements at a buoy site near the coastal line (water depth is 14 m) are adjusted to their potential values, which are then compared with those at a nearby station. The good agreement between the two sets of results demonstrates the accuracy and effectiveness of the standardization method. It is also found that the behavior of roughness length scale over shallow water may differ noticeably from that over deep ocean, especially under strong wind conditions, and an inappropriate usage of marine roughness predictors may result in significant estimation errors.

1. Introduction

Wind measurement offers the essential data source for a wide range of practices in the fields of meteorology and wind engineering, such as weather forecasting (Stauffer et al. 1991), wind energy assessment (Shu et al. 2015), wind-disaster mitigation, and wind-resistant design (Irwin 2006). Unfortunately, because of the terrain/topographic/height effects (He et al. 2013, 2014a) as well as the influences caused by measurement systems (Beljaars 1987), in situ observations usually deviate from the “truth,” and the results collected from one station may show remarkable inconsistency with those from others, especially for surface winds. These discrepancies introduce much inconvenience and uncertainty for the follow-up applications. Previous studies show that direct usage of raw data may introduce errors of surface wind speed as large as 40% (Powell et al. 1996), which may become significantly larger (up to 300%) under complex terrain conditions (He et al. 2014b). Even at the same site, it has been shown that wind records during varied periods may demonstrate noticeable discrepancies because of the evolution of terrain features (Tamura and Suda 1989) or replacement of measurement systems (Masters et al. 2010). These errors/uncertainty can be further amplified in the subsequent analysis of relevant effects of the wind, such as wind energy, wind loads, and wind-induced structural responses (Li et al. 2005), which are
proportional to a high order power of wind velocity, resulting in an exponentially increased inaccuracy of the obtained results.

Given the importance of wind measurements and the nonignorable distortion effects involved in such records, there is an urgent need to correct the raw wind speed data to the potential values that correspond to a reference condition, that is, for neutral winds at a height of 10 m above open terrain with a roughness length scale \( z_0 = 0.03 \text{ m} \) (WMO 2008).

Several methods have been established for the standardization of surface wind speeds. Under flat terrain conditions, there are five steps involved in a typical correction process (Powell et al. 1996): 1) calculate the mean wind speed by averaging the time series of measurement records, 2) estimate the exposure type, 3) adjust the speed to the reference height, 4) adjust the speed to the reference exposure, and 5) adjust the speed to the peak gust associated with the reference gust duration. Among the above procedures, step 3 relies on a mathematical predictor of mean wind profiles. The log law or power law is usually adopted, with the inputting of a specified exposure type as indicated either by the power exponent \( \alpha \) or by the roughness length scale \( z_0 \). Step 4 is generally based on the hypothesis that there exists a special height (e.g., the gradient height) where the mean wind speed stays unchanged over varied terrains. Thus, the mean wind speed at this height above a concerned site equals the one above reference terrain. Given the exposure information at the concerned site (i.e., \( \alpha \), or \( z_0 \)), the Engineering Science Data Unit (ESDU) method (ESDU 1982; Irwin 2006), the power-law-based method, the drag-law-based method (Simiu and Scanlan 1996), or the blending-height-based method (Wieringa 1986; Verkaik 2000) could be adopted for this adjustment. The last step is concerned with the gust-duration dependence of peak speed at a 10-m height above the reference terrain, and the Durst curve (Durst 1960) or other gust factor models, such as the one presented in ESDU (1983), can be utilized.

Under rugged terrain conditions, wind tunnel tests (Chock and Cochran 2005) or numerical simulations (Walmsley et al. 1990; Yan et al. 2016) are commonly employed to eliminate the topographic effects, although for some special cases (e.g., with regular hills/escarpments), empirical correction equations as stipulated in wind load codes may be resorted to.

Recently, the authors proposed a data-driven standardization method, based on a combined usage of surface wind measurements and wind profile records detected by a boundary layer–type Doppler radar profiler (He et al. 2014b). The terrain/topographic effects would be determined and eliminated by converting the surface wind speed to the corresponding gradient wind speed that is recognized as the maximum of the mean wind speed profile. Compared to wind tunnel testing and numerical simulations, this method has the merits of low cost and high efficiency. Thus, it is desirable and suitable to be applied to cases with complex terrain as long as sufficient wind measurements are available.

Although the validity of this method has been verified through wind tunnel testing and numerical simulations (He et al. 2014b; Yan et al. 2016), our later comparison analysis reveals that the computed potential wind speeds (i.e., at 10-m height above a reference terrain with \( z_0 = 0.03 \text{ m} \)) at an island-based weather station via this data-driven method are about 20% smaller than the synchronous measurements at 8 m above mean sea level (MSL) from a nearby buoy site. This finding causes us to recognize the significance of standardization of surface wind speeds over sea waters.

As a complementary part of our preceding work (He et al. 2014b), this study focuses on the standardization of surface wind speeds with marine exposures. The effect of wind strength on the roughness of the sea surface is further taken into account. Special attention is paid to the difference in marine roughness between deep-ocean and shallow-water cases. As an application example, wind measurements at a buoy site near the coastal line (water depth is 14 m) are adjusted to their potential values, which are then compared against those at a nearby station.

The layout of this paper is arranged as follows: Section 2 introduces the method adopted in this study, and section 3 presents the standardization results at the concerned site. The uncertainty involved in the presented method is discussed in section 4, and the main findings and concluding remarks are summarized in section 5.

2. Method statement

A flowchart of the standardization scheme is shown in Fig. 1. The aim is to determine the correction factor CF; by multiplying it, the measured mean wind speed \( U(T_r, z_m, z_{0r}) \) at the site in question can be converted into its potential value \( \bar{u}(T_r, \tau, z_r, z_{or}) \), which corresponds to the specified reference condition. Here, \( T \) stands for the average duration of the mean wind speed; \( \tau \) is the gust duration with \( \tau \approx T \); \( z \) represents the height above the terrain; \( z_0 \) is the roughness length scale; and the subscripts \( m \) and \( r \) denote the associated quantities that account for the measurement and reference conditions, respectively. The potential speed value \( \bar{u}(T_r, \tau, z_r, z_{or}) \) is defined as the peak gust at 10 m (i.e., \( z_r = 10 \text{ m} \)) above an open flat terrain with \( z_{0r} = 0.03 \text{ m} \) in this study. Note
that if \( \tau = T \), then \( \bar{u} = U \); that is, the mean wind speed can be treated as a special form of the peak gust.

Usually, the measured mean wind speed \( U(T_r, z_m, z_{0m}) \) is computed by averaging the time series of measurement records over a period \( T_r \). In practice, there are two kinds of average methods: the scalar average and the vector average. Although the computed mean speed values via the two methods tend to be identical for stationary and/or strong winds, they may vary considerably under nonstationary and/or weak wind conditions. It has been shown that the use of the scalar mean wind speed can add convenience in the analysis of air–sea fluxes (Fairall et al. 2006; Mahrt et al. 2016). Thus, throughout this study, the scalar average method is adopted to compute the mean speed values.

**a. Correction for height, exposure, and gust duration**

As demonstrated in Fig. 1, there are three kinds of adjustments for wind speed involved in the correction process, which are related to height, exposure, and gust duration.

1) **HEIGHT ADJUSTMENT**

The height-relevant adjustment relies on vertical wind profile models. It is well known that wind structures at the surface layer (which typically occupies the lowest 10% depth of the atmospheric boundary layer) are dominated by two kinds of effects: the mechanical mixing effect due to ground friction caused by surface roughness, and the thermal effect, which is related to the stratification condition. According to the Monin–Obukhov (M–O) similarity theory, mean wind profiles can be expressed as (Garratt 1994)

\[
U(z) = \frac{u_0}{\kappa} \ln \left( \frac{z}{z_0} \right),
\]

where \( U(z) \) is the wind speed at height \( z \); \( \kappa \approx 0.4 \) is the von Kármán constant; \( u_0 \) is the friction velocity; \( \zeta = z/L \) is the normalized height according to the Obukhov stability length scale \( L = u_0^2 / (\kappa g \theta_v \kappa) \), with \( g \), \( \theta_v \), and \( \theta_v \kappa \) being the gravity acceleration, mean virtual potential temperature, and surface virtual potential temperature, respectively. The term \( \Psi(\zeta) \) is the stability function, which reflects the thermal effects. The results from measurements over land suggest that (Dyer 1974; Garratt 1994)

\[
\Psi(\zeta) \approx 2 \ln \left( \frac{1 + x}{2} \right) + \ln \left( \frac{1 + x^2}{2} \right)
\]

\[
-2 \tan^{-1} x + \pi/2, \quad \text{for} \quad \zeta < 0, \quad \text{and} \quad (a2)
\]

\[
\Psi(\zeta) \approx -c_2 \zeta, \quad \text{for} \quad \zeta > 0, \quad \text{and} \quad (b2)
\]

with \( x = (1 - c_1 \zeta)^{1/4}, \) \( c_1 \approx 16, \) and \( c_2 \approx 5. \)

However, measurements over the sea are still insufficient to establish a well-recognized form of the stability function for marine winds, and the above similarity models have also been used to predict the behavior of marine wind speed profiles (Lange et al. 2004; Peña et al. 2008). It must be noted that the marine boundary layer differs from that over land (Mahrt et al. 1998), and the observed marine wind speed profiles may deviate from those predicted by the existing M–O models (Lange et al. 2004; Peña et al. 2008). Typically, it has been shown that stable boundary layers over the sea differ from clear-sky, nocturnal boundary layers over land (Fairall et al. 2006; Mahrt et al. 2016). The stable stratification over sea is generally generated by warm-air advection from the coast, and a developing internal boundary layer tends to influence the air–sea fluxes and the boundary layer depth in coastal zones (Mahrt et al. 1998). But such discrepancies become less evident as wind strengthens. For moderate to strong winds, the existing M–O theory is regarded as adequate for marine surface layer.

Under near-neutral conditions (\( \zeta \approx 0 \)), Eq. (1) degrades into the logarithmic law

\[
U(z) = \frac{u_0}{\kappa} \ln \left( \frac{z}{z_0} \right).
\]

The above equation may be mostly effective within the surface layer. For equilibrium neutral wind flows in the whole atmospheric boundary layer (ABL), ESDU (1982) recommends using the Deaves–Harris (D–H) model:

\[
U(z) = \frac{u_0}{\kappa} \left[ \ln \left( \frac{z}{z_0} \right) + a_1 \left( \frac{z}{h} \right) + a_2 \left( \frac{z}{h} \right)^2 + a_3 \left( \frac{z}{h} \right)^3 + a_4 \left( \frac{z}{h} \right)^4 \right],
\]

\[
(3b)
\]
where \(a = (23/4, -15/8, -4/3, 1/4)\) are coefficients and \(h\) is the ABL depth, usually with the form \(h = u_*/6f_c\) (\(f_c\) is the Coriolis parameter).

2) EXPOSURE ADJUSTMENT

Two kinds of methods for the adjustment of mean wind speed among varied terrains are considered herein: the ESDU method and the drag-law-based method. As mentioned previously, both of these methods assume that there exists a special height where the mean wind speed stays unchanged over varied terrain in the vicinity of the concerned site.

The ESDU method (ESDU 1982) is deduced and simplified (to a close approximation with a \(\pm 3\%\) deviation) from the D–H wind profile model, based on the hypothesis of unchanged gradient wind speed:

\[
\frac{u_{*m}}{u_{*r}} = \frac{\ln(10^5/z_{0m})}{\ln(10^5/z_{0r})}, \quad \text{or} \quad \frac{U(T_r, z_r, z_{0r})}{U(T_r, z_m, z_{0m})} = \frac{\ln(10^5/z_{0m})}{\ln(10^5/z_{0r})}.
\]

The drag-law-based method assumes that the geostrophic wind stays unchanged in the vicinity of the concerned site over varied terrain. According to the Rossby-number similarity theory (Garratt 1994), for barotropic and neutral ABL, the three parameters of geostrophic wind \(G\), Coriolis parameter \(f_c\), and surface roughness length scale \(z_0\) can be expressed by the Rossby number in the form of the drag law:

\[
C_g \frac{u_0^2}{G^2} = \frac{\kappa^2}{\ln(u_*/f_c z_0)} - A_4^2 + B_0^2.
\]

where \(C_gn\) is the geostrophic drag coefficient and \(A_0 \approx 2\) and \(B_0 \approx 4.5\) are two universal constants. Thus,

\[
\frac{u_{*r}^2}{u_{*m}^2} = \frac{C_gn_r}{C_gn_m} = \frac{\left[\ln(u_*/f_c z_{0r})\right] - A_4^2 + B_0^2}{\left[\ln(u_*/f_c z_{0m})\right] - A_4^2 + B_0^2}.
\]

Given the values of \(u_{*m}\) and \(z_{0m}\), \(u_{*r}\) can be determined via iterative calculations. Then, \(U(T_r, z_r, z_{0r})\) can be obtained by substituting \(u_{*r}\) into the log law [Eq. (3a)].

3) ADJUSTMENT FOR GUST DURATION

From Fig. 1, this step aims to calculate the peak gust \(u(T_r, \tau, z_r, z_{0r})\) from the associated mean wind speed \(U(T_r, z_r, z_{0r})\), which can be readily accomplished with the help of the gust factor models.

The gust factor GF is conventionally defined as a ratio of the peak gust value \(u(T, \tau)\) over a relatively shorter gust duration \(\tau\) in a given observation period \(T\) to the corresponding mean wind speed \(U(T)\). It depends on \(\tau, T, z, z_0,\) and thermal conditions. In reality, GF measurements may be attenuated because of the filtering effects of measurement systems (Beljaars 1987). But as far as this standardization step is concerned (i.e., under reference condition), we merely focus on the \(\tau\) dependence of GFs. ESDU (ESDU 1983) recommends the following close form:

\[
GF = 1 + g_c \frac{\sigma_u}{U},
\]

\[
g_c = \left[ \frac{1}{2 \ln(T\tau)} + \frac{0.5772}{2 \ln(T\tau)} \right] \frac{\sigma_u(T, \tau)}{\sigma_u},
\]

\[
\sigma_u = \frac{u_0}{7.5} \left[ 0.538 + 0.09 \ln(z/z_0) \right]^\eta,
\]

\[
\eta = 1 - 1.6 f_c z_r / u_0,
\]

\[
\frac{\sigma_u^2(T, \tau)}{\sigma_u^2} = \int_0^\infty \frac{S_u(n, z, z_0) \chi^2(n)}{S_u(n, z, z_0) \chi^2(n)} dn,
\]

\[
\chi^2(n) = \left[ \frac{\sin(\pi n \tau)}{\pi n \tau} \right]^2 - \left[ \frac{\sin(\pi n T)}{\pi n T} \right]^2, \quad \text{and}
\]

\[
S_u(n) = \frac{n S_n(n)}{\sigma_u^2} = \frac{4f}{\left[1 + 70.8f^2\right]^{5/6}},
\]

\[
f = n T_u, \quad T_u = 3.13 z_0^{0.2},
\]

in which \(g_c\) is the peak factor; \(\sigma_u\) is the theoretical standard deviation of longitudinal wind turbulence, compared to \(\sigma_u(T, \tau)\), which is filtered with a cutoff frequency of 1/Hz because of the average of instantaneous speed over \(\tau\); \(\chi^2\) indexes the filtering effects on the power spectrum; \(v\) is the up-crossing rate; \(S_n(n)\) is the von Kármán spectrum; and \(T_u\) is the integral time scale.

b. Estimation of roughness length scale over sea

As demonstrated in Fig. 1, the terrain roughness length scale \(z_0\) drives both the height conversion and the exposure conversion. Thus, this parameter plays an essential role in the standardization scheme. For on-land cases, \(z_0\) is usually regarded as constant for a given exposure, whose value can be determined via various estimation models that are driven either by wind turbulence records or by vertical wind speed profiles. However, because of the strong air–water interactions, the values of \(z_0\) demonstrate strong dependence on sea states. In general,
Table 1. Model coefficients for the determination of marine roughness in the literature.

<table>
<thead>
<tr>
<th>Model coefficient</th>
<th>$z_0 = a a_d^2/g$</th>
<th>$10^3 C_{d,10} = a_1 + a_2 U_{10}$</th>
<th>Remark (speed range, sea type)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Garratt (1977)</td>
<td>0.017 ($k = 0.4$)</td>
<td>0.75 0.067</td>
<td>$10 &lt; U_{10} &lt; 52$ m s$^{-1}$</td>
</tr>
<tr>
<td>Powell (1980)</td>
<td>1.024 0.054</td>
<td></td>
<td>$5 &lt; U_{10} &lt; 61$ m s$^{-1}$</td>
</tr>
<tr>
<td>Wu (1980)</td>
<td>0.0185</td>
<td>0.8 0.065</td>
<td>Literature review, $1 &lt; U_{10} &lt; 22$ m s$^{-1}$</td>
</tr>
<tr>
<td>Large and Pond</td>
<td>0.011–0.035</td>
<td>0.49 0.065</td>
<td>$10 &lt; U_{10} &lt; 25$ m s$^{-1}$</td>
</tr>
<tr>
<td>ESDU (1982)</td>
<td>1/60 = 0.0167</td>
<td>0.8 0.065</td>
<td>$10 &lt; U_{10} &lt; 25$ m s$^{-1}$</td>
</tr>
<tr>
<td>Smith et al. (1992)</td>
<td></td>
<td>0.27 0.116</td>
<td>$3 &lt; U_{10} &lt; 16$ m s$^{-1}$, 18-m depth</td>
</tr>
<tr>
<td>Makin et al. (1995)</td>
<td>0.018–0.030</td>
<td>— —</td>
<td>$0 &lt; U_{10} &lt; 20$ m s$^{-1}$</td>
</tr>
<tr>
<td>Yelland and Taylor (1996)</td>
<td>0.011–0.017</td>
<td>0.60 0.070</td>
<td>$6 &lt; U_{10} &lt; 26$ m s$^{-1}$, open ocean</td>
</tr>
<tr>
<td>Janssen (1997)</td>
<td>0.026–0.036</td>
<td>— —</td>
<td>$7 &lt; U_{10} &lt; 20$ m s$^{-1}$, 18-m depth</td>
</tr>
<tr>
<td>Vickery and Skerlj (2000)</td>
<td></td>
<td>4.5 $&lt; 10^3 C_{d,10} &lt; 7.5$</td>
<td>20 $&lt; U_{10} &lt; 30$ m s$^{-1}$, TC, 13-m depth</td>
</tr>
<tr>
<td></td>
<td></td>
<td>2.9 $&lt; 10^3 C_{d,10} &lt; 6.1$</td>
<td>20 $&lt; U_{10} &lt; 31$ m s$^{-1}$, TC, 48-m depth</td>
</tr>
<tr>
<td></td>
<td></td>
<td>2.8 $&lt; 10^3 C_{d,10} &lt; 8.1$</td>
<td>20 $&lt; U_{10} &lt; 45$ m s$^{-1}$, TC, 3.2-km depth</td>
</tr>
<tr>
<td></td>
<td></td>
<td>$1.4 &lt; 10^3 C_{d,10} &lt; 2.6$</td>
<td>$26 &lt; U_{10} &lt; 52$ m s$^{-1}$, deep ocean; reduced $C_{d,10}$ for $U_{10} &gt; 33$ m s$^{-1}$</td>
</tr>
<tr>
<td>Powell et al. (2003)</td>
<td>$\alpha &lt; 0.018$</td>
<td>$0.001 &lt; z_0 &lt; 0.004$</td>
<td>$18 &lt; U_{10} &lt; 54$ m s$^{-1}$, deep ocean; capped $C_{d,10}$ for $U_{10} &gt; 25–30$ m s$^{-1}$</td>
</tr>
<tr>
<td>Vickery et al. (2009)</td>
<td>0.0001$&lt; z_0 &lt; 0.01$</td>
<td>$1.0 &lt; 10^3 C_{d,10} &lt; 3.8$</td>
<td>$25–30$ m s$^{-1}$</td>
</tr>
<tr>
<td>Petersen and Renfrew (2009)</td>
<td>—</td>
<td>$10^3 C_{d,10} = 2.04$</td>
<td>Mean value for $15 &lt; U_{10} &lt; 19$ m s$^{-1}$, open water</td>
</tr>
</tbody>
</table>

as wind speed increases, so does water wave height. Hence, marine roughness normally increases with wind strength (ESDU 1982).

The Charnock relation is often adopted to estimate marine roughness (Charnock 1955):

$$z_0 = \alpha \frac{u^2}{g},$$

where $\alpha$ is the Charnock constant (values recommended in some publications are summarized in Table 1). Given the value of the mean wind speed, $z_0$ can be estimated via iterative calculation from Eqs. (7) and (3a).

Another widely adopted way of depicting the behavior of the air-sea momentum flux is by modeling the surface drag coefficient (e.g., Large and Pond 1981):

$$10^3 C_{d,10} = a_1 + a_2 U_{10},$$

in which $C_{d,10}$ is the neutral drag coefficient (Grachev et al. 1998) at 10 m MSL and $a_1$ and $a_2$ are two coefficients. Table 1 lists the recommended values for the two coefficients in some publications.

Substituting Eq. (8a) into the log law gives

$$z_0 = 10 \exp(-\kappa/\sqrt{C_{d,10}}).$$

Although the above two kinds of $z_0$ estimation methods are considered to lead to consistent results (Harper et al. 2010), the coefficient values for these models, as shown in Table 1, are found to vary noticeably in the literature. Wu (1980) further argued that the Charnock relation suffers from intrinsic errors and limitations. This is understandable since it has long been recognized that marine roughness and the surface drag coefficient do not solely depend on wind speed (e.g., Geernaert et al. 1986; Smith et al. 1992; Taylor and Yelland 2001).

First, water depth has a significant effect on many characteristics of sea waves. According to the wave theory, waves can propagate freely through water that is deeper than half their wavelength. But the orbits of water molecules in waves propagating through shallow water are flattened by the proximity of the sea surface bottom, and the phase speed (proportional to the root square of the shallow-water depth) decreases as sea waves approach the coastline. Under strong wind conditions, as the waves in shallow water can never match the wind speed as do those in deep water, they begin to act as roughness elements and result in increased wind drag above the shallow water. In a comparison study, Smith et al. (1992) found that wind stress at a relatively lower wind speed ($\sim 10$ m s$^{-1}$) was similar at two off-shore stations with varied water depths. But when wind speed exceeded 10–13 m s$^{-1}$, wind drag at the shallower site became larger. Previous studies (e.g., Garratt 1977; Geernaert et al. 1986; Smith et al. 1992; Vickery and Skerlj 2000) showed that the drag coefficient in open-sea conditions was 10%–60% lower than that in coastal or shallow-water situations.

Besides, it has been demonstrated that detailed sea states, such as wave height and wave steepness, are...
closely related to the behavior of marine flux (e.g., Geernaert et al. 1986; Smith et al. 1992; Taylor and Yelland 2001). Typically, the marine surface becomes smoother as waves mature from young steep waves to the older and less steep, albeit much higher, waves near full development. Thus, open-ocean data under strong wind conditions generally correspond to a narrow range of wave maturity near full development. Meanwhile, swells and wind waves from varied directions and fetches can modulate the local roughness independently of wind strength. Individual waves may distort the overlying wind field by setting up speedup and separation zones over and to the leeward of the wave crest (Powell et al. 2003). The storm-relative kinematic/thermodynamic structures of tropical cyclones (TCs) further complicate the situation in TC-attack areas where the sea states become even more inhomogeneous and disordered, and the storm-generated waves are believed to be fetch limited and generally young. These initially young and steeper waves and their shoaling as they approach the coastline produce a much rougher marine surface than that under normal wind conditions even for high wind speeds (Vickery and Skerlj 2000).

Furthermore, the turbulent air–sea exchange of momentum, heat, and moisture in the coastal zone generally differs from that under open-ocean conditions because of the influences of advection from land, internal boundary layers, and other coastal effects (Mahrt et al. 1998; Vickers and Mahrt 2010; Grachev et al. 2011). As is known, the heat conduction and capacity of land differ from those of water, so often the atmosphere over land gets warmer than the sea surface temperature. In such cases, the warm air is advected over the cold sea, which results in significant influences on the air–sea flux exchanges. Meanwhile, for offshore winds in coastal areas, when they are blowing from land over sea they must carry along exposure-branded characteristics that have been shown to differ significantly from those over open oceans. In a comprehensive field study, Grachev et al. (2011) compared the COARE 3.0 algorithm (Fairall et al. 2003) derived for open-ocean conditions with the direct measurements over bays and lakes and near coastal harbor areas. The momentum fluxes were observed to vary over different terrains.

Given the complex characteristics of the marine momentum flux, continuous efforts have been made to improve the predictions of \( z_0 \) and \( C_d \). On the one hand, more available measurements, which are of high quality and also cover a wider range of situations, help to refine the values of the coefficients involved in traditional models. Specifically, Fairall et al. (2003) examined the measurement results from a number of field programs, and proposed the following wind-strength-dependent model for the Charnock constant:

\[
\alpha = \begin{cases} 
0.011 & \text{for } U_{10} \leq 10 \text{ m s}^{-1} \\
0.011 + 0.000875(U_{10} - 10) & \text{for } 10 < U_{10} < 18 \text{ m s}^{-1} \\
0.018 & \text{for } 18 \leq U_{10} \leq 25 \text{ m s}^{-1}.
\end{cases}
\] (9)

On the other hand, new prediction models have been proposed to take into account the sea-state effects. For example, Wu (1980) recommended that marine roughness be modeled as a function of gravitational acceleration, surface tension, and viscosity, while Anctil and Donelan (1996) focused on the normalized roughness by wave height, and utilized the inverse wave age and wave slope to model this nondimensional parameter. By contrast, Taylor and Yelland (2001) modeled marine roughness in a function of the significant wave height and peak wavelength, while Oost et al. (2002), among others, presented an exponential relationship between the Charnock constant and the wave age. The latter two models have been incorporated into COARE 3.0 (Fairall et al. 2003):

\[ z_0 = 1200H_s (H_s/L_p)^4, \quad L_p = gT_p^2/(2\pi), \quad \text{and} \]
\[ \alpha = 50(C_p^2 u_{*})^{-2.5}, \quad C_p = gT_p/(2\pi). \] (11, 12)
where $H_s$ is the significant wave height and $L_p$ is the wavelength associated with the dominant wave period $T_p$ (thus, $H_s/L_p$ is approximately the slope of the dominant wave), $C_p$ is the phase speed of the dominant wave, and $C_p/U_m$ is a measure of the wave age. For fully developed waves in deep water, the following equations can be used:

$$H_s = 0.0248U_{10n}^2 \text{ and } T_p = 0.729U_{10n},$$

in which $U_{10n}$ stands for neutral wind speed at 10-m height.

Despite great efforts that have been made regarding estimation of marine roughness and significant advancements that have been achieved, no consensus on the most suitable prediction model has emerged yet for coastal areas. Thus, the performance of the different prediction models presented above will be examined and compared.

3. Application and verification

a. Sites and datasets

A case study is presented herein as an application of the introduced standardization method, based on time series of 1-min scalar mean surface wind measurements at a buoy meteorological site, that is, WB9, in Hong Kong, during the period from 14 September 2012 to 31 December 2014. For comparison purposes, we also analyze synchronous measurements that were collected from the nearby Cheung Chau station (CCH), located at the zenith of Cheung Chau Island (71.9 m MSL). It is equipped with a cup anemometer at 98.6 m MSL. The WB9 site is in the vicinity of CCH where the local water depth is 14 m. The buoy is equipped with a propeller anemometer at 8 m MSL. The locations of the two sites and the situations of the surrounding topography are shown in Fig. 2.

Because of the lack of detailed temperature information at WB9, it is difficult to quantify the local thermal effects on vertical wind profiles. Thus, this case study is only concerned with the neutral stratification condition. According to Hsu (1992), the stability class in offshore and coastal areas is dependent on two parameters: wind speed $U_{10}$ (mean speed at 10 m) and the air-sea temperature difference $\Delta T = T_{air} - T_{sea}$. Neutral stratification can be reasonably assumed in case $U_{10} > 8 \text{ m s}^{-1}$ and $|\Delta T| < 6^\circ \text{C}$, or $U_{10} > 10 \text{ m s}^{-1}$ and $|\Delta T| < 10^\circ \text{C}$. To take into account the situation of the relatively lower level of the anemometer (8 m MSL) at WB9 and the normally slight air-sea temperature difference at CCH, it is assumed that a neutral condition is achieved when the measured mean wind speed exceeds 8 m s$^{-1}$.

As shown in Fig. 2, the terrain in the north semiplane of the concerned site is dominated by topographic features, for example, Lantau Island, Tai Mo Shan, and Hong Kong Island, which may result in undesirable topographic effects on the sea state and local wind structure. To simplify the situation, this case study further focuses on winds blowing from the south semiplane, that is, in the azimuthal section from 90° clockwise to 270°. A total of 1428 runs of 10-min mean wind measurements at each of the two sites are selected, as shown in Fig. 3.

b. Results

For WB9, the introduced standardization scheme for marine winds is adopted. Both the ESDU method [Eq. (4)] and the drag-law-based method [Eq. (5)] are utilized for the exposure-relevant adjustment, while four predictors of marine roughness are examined for the estimation of $z_0$ values, which are depicted by Eq. (8) (with $a_1 = 0.27$ and $a_2 = 0.116$; Smith et al. 1992), Eqs. (7) and (9), Eq. (11), and Eqs. (7) and (12), respectively. For CCH, the data-driven standardization method proposed in our previous study (He et al. 2014b) is applied.

1) MARINE ROUGHNESS AND CORRECTION FACTOR

The estimations of marine roughness $z_0$ via different methods are shown in Fig. 4. For Eq. (8), values of the model coefficients recommended in Smith et al. (1992) are adopted for two reasons: 1) the water depth at the offshore site in Smith et al. (1992) is 18 m, which is similar to 14 m at WB9; and 2) the results presented in Smith et al. (1992) belong to a series of outputs from a long-term project conducted by an international multi-member team, and the accuracy of the presented results has been systematically checked and verified through cross comparisons. However, the results of Smith et al. (1992) only covered a wind speed range of 3–16 m s$^{-1}$. In this study, it is assumed that Eq. (11) with $a_1 = 0.27$ and $a_2 = 0.116$ works for $8 \leq U_{10} \leq 30 \text{ m s}^{-1}$, beyond which $z_0$ is capped by its maximum value.

The results in Fig. 4 reveal that for $U_m < 15 \text{ m s}^{-1}$ ($U_m$ stands for mean wind speed at the measurement height, or 8 m MSL) the differences among the estimations from the different methods are small. As wind strengthens, the results from Eq. (8) become significantly larger than those from the other methods, which predict similar results for $U_m < 25 \text{ m s}^{-1}$. But for even stronger winds, Eqs. (7) and (12) predict a much faster increase of $z_0$ with $U_m$ than Eq. (11). Within the whole studied speed range, the predictions from Eqs. (7) and (9) lie between those estimated by Eqs. (7) and (12) and Eq. (11). It is noted that Eq. (8) with $a_1 = 0.27$ and $a_2 = 0.116$ accounts...
for shallow-water cases, while the other three models are suitable for deep-water cases with fully developed sea waves.

Figure 5 shows the variations of calculated CF values using both the ESDU and drag-law-based exposure-adjustment methods, on the basis of estimations of $z_0$ via the four mentioned models. The results from the two exposure-adjustment methods are consistent, especially for $U_w \geq 15$ m s$^{-1}$. For weaker winds, the ESDU method predicts a bit smaller CFs than the drag-law-based method, with the maximum relative difference less than 5%. Relatively larger differences of CF are owing to the estimations of $z_0$, for example, the maximum differences at $U_w = 8$ and 32 m s$^{-1}$ are 7.3% and 8.4%, respectively. In consideration of the noticeable prediction differences caused by the estimation models for $z_0$, the CF values determined by Eq. (8) with $a_1 = 0.27$ and $a_2 = 0.116$ are utilized at WB9.

2) CORRECTED WIND SPEED

The calculated CF($T_r = 600$ s, $\tau_r = 600$ s) values are then utilized to generate the standardized 10-min mean surface wind speeds at the concerned site. Figure 6 compares the results before and after the implementation of such a standardization process in which the CF values are calculated using the drag-law-based exposure adjustment method. The corresponding results at CCH are also presented. As reflected in the figure, the differences between the raw wind measurements at WB9 and CCH are remarkable, especially for comparatively stronger winds. By contrast, the corrected wind speed values agree well between the two sets of data. The best-fitting correlation
equation for the corrected mean speeds at CCH and WB9 is 
\( y = 1.007x \), with the goodness of fit \( R^2 = 0.797 \). If the CF 
values calculated via the ESDU exposure adjustment 
method are adopted, the correlation equation turns to 
\( y = 1.02x \), with \( R^2 = 0.798 \).

In addition, the standardized peak gust speeds can 
also be calculated based on the CF\((T_r, \tau_r)\) values, which 
are determined by multiplying CF\((T_r, \tau_r = T_r)\) and 
GF\((T_r, \tau_r)\). The GF values can be computed using Eq. (6). 

Figure 7 depicts the dependence of GF\((T_r, \tau_r)\) on \( \tau_r \) 
(varying from 1 to 100 s) and \( T_r \) (600 and 3600 s). It is 
clear that the GF values increase with decreasing \( \tau_r \) and 
increasing \( T_r \) with GF(600, 3) and GF(3600, 3) being 
1.43 and 1.52, respectively. The dependence of CF(600, 
3) and CF(3600, 3) on the mean wind speeds is also 
demonstrated in Fig. 7. As reflected, almost all the 
depicted CF values are larger than unity. Meanwhile, 
the CF values also increase with wind strength, which 
suggests that the surface wind above the sea gets 
“gustier” under stronger wind conditions.

4. Uncertainty analysis

In accordance with the flowchart shown in Fig. 1, there 
may be three uncertainty sources involved in the stan-
dardization process, which are related to the height ad-
justment, estimation of marine roughness, and exposure 
adjustment, respectively.

a. Height adjustment

The log law [Eq. (3a)] is adopted in this study for 
the height adjustment of neutral winds with marine
exposures. Although the wind profile form is deduced based on a reasonable theoretical analysis and has been widely utilized in the literature, recent observations of TCs over deep oceans with dropsonde (Vickery et al. 2009) have revealed that the measured mean wind speeds below about 20 m MSL decrease less toward the sea surface than what the log law predicts, and the lower the height is, the larger the difference in modeled and observed wind speeds becomes. This discrepancy may be caused by the slip surface produced by sea spray or bubbles, which help to inhibit momentum transfer at the air–sea interface.

In this study, assuming that a 4% relative difference exists in the modeled and observed speeds at WB9 for $U_m = 15 \text{ m s}^{-1}$, then the modeled value turns out to be $14.4 \text{ m s}^{-1}$. From Fig. 5, the CF values for $U_m = 15$ and $14.4 \text{ m s}^{-1}$ are 0.8142 and 0.8080, respectively. Thus, the calculated potential values are $12.213 \text{ m s}^{-1}$ (in reality) and $11.635 \text{ m s}^{-1}$ (ideal value), with the difference increasing slightly to 4.7%. This error magnification is due to the increased CF for larger $U_m$. The sensitivity of such errors depends on the adopted $z_0$ predictor. If Eq. (8) (with $a_1 = 0.27$ and $a_2 = 0.116$) is adopted for shallow-water cases, as the CF gets more sensitive to $U_m$ for lower winds (Fig. 5), so does the error. But at the same time, because of the weakened air–sea interactions under weak wind conditions, the deviations between the modeled and observed wind speeds may decrease, which tends to restrain the prediction error. On the contrary, if Eq. (12) is adopted to estimate $z_0$ over deep waters, following the above analysis, the overestimation error would be magnified for stronger winds.

Another potential uncertain aspect lies in the thermal stratification conditions. Although neutral atmospheres are usually assumed under cases of strong winds, there is no widely accepted speed threshold exceeding which a true neutral condition would be achieved. The ESDU (1982) adopts a value of $10 \text{ m s}^{-1}$, compared to $5–5.5 \text{ m s}^{-1}$ suggested by Wieringa (1973). Even if a widely accepted speed threshold is recognized, there is no...
guarantee that the atmosphere, especially in TC cases, is always neutrally stratified. It has been observed that intense convections exist in the inner regions of a TC (e.g., Vickery and Skerlj 2005; Song et al. 2012). Under such nonneutral conditions, the speed profiles would not abide by the log law in the surface layer. According to the analysis in Verhaik (2000), this error is estimated to be lower than 10% for $U_{10} > 5 \text{ m s}^{-1}$.

![Figure 6](image6.png)

**FIG. 6.** Comparison of 10-min scalar mean surface wind speeds at WB9 and CCH before and after standardization.

![Figure 7](image7.png)

**FIG. 7.** Dependence of GF($T_r, \tau_r$) on the gust duration $\tau_r$ and the mean speed duration $T_r$ determined by Eq. (6), and the CF($T_r, \tau_r$) values calculated by multiplying GF($T_r, \tau_r$) and CF($T_r, \tau_r = T_r$). In this figure, the CF($T_r, \tau_r = T_r$) values are computed via Eqs. (5) and (8), as shown in Fig. 5.
b. Roughness length scale

To analyze the uncertainty related to the estimated marine roughness values, the \( z_0 \)-dependent CF \((\tau = T = 600\text{ s})\) values are computed via the drag-law-based exposure adjustment method. Also computed are the marine roughness values estimated using the model presented in Large and Pond (1981), that is, Eq. (8) with \( a_1 = 0.49 \) and \( a_2 = 0.065 \) and that adopted in this study \([\text{i.e., proposed in Smith et al. 1992;}\) Eq. (8) with \( a_1 = 0.27 \) and \( a_2 = 0.116 \)]. The relationship presented in Large and Pond (1981) is considered herein because it has been widely utilized in literature in cases of either deep ocean \((\text{Powell et al. 2003})\) or shallow water \((\text{Masters et al. 2010})\). To be consistent with the previous analysis, both models are capped at \( U_{10} = 30\text{ m s}^{-1} \).

Assuming a 100% estimation error of \( z_0 \) for \( z_0 = 8.35 \times 10^{-3} \) and \( 1.175 \times 10^{-3} \text{ m} \) (corresponding to \( U_m = 8 \) and \( 15 \text{ m s}^{-1} \) in the case study), then the calculated CF values for the two cases turn out to be 0.725 and 0.705 (with error and ideal value), and 0.835 and 0.800, respectively, which means that the relative errors involved in the computed potential speeds are 2.8% and 4.2%. If the relationship in Large and Pond (1981) is applied to calculate marine roughness, then the computed values are \( 2.9 \times 10^{-2} \text{ m} \) (for \( U_m = 8 \text{ m s}^{-1} \)) and \( 2.49 \times 10^{-1} \text{ m} \) (\( U_m = 15 \text{ m s}^{-1} \)). Accordingly, the CF values are 0.678 and 0.738. Thus, the relative differences of CFs by using these two estimation methods for \( z_0 \) become 3.8% (for \( U_m = 8 \text{ m s}^{-1} \)) and 7.8% (\( U_m = 15 \text{ m s}^{-1} \)). This difference will increase for larger \( U_m \), for example, 11.4% for \( U_m = 25 \text{ m s}^{-1} \).

It is clear that inappropriate usage of marine roughness predictors may result in noticeable errors in the obtained CF values or the potential wind speeds. Typically, if estimation models for open sea exposures are adopted for shallow-water cases, the CFs would be smaller than the ideal values, resulting in an underestimation of the potential wind speeds.

c. Exposure adjustment

In the present study, two exposure adjustment methods have been considered: the ESDU method and the drag-law-based method. The related uncertainty comes from the adopted relationship between the gradient wind (or geostrophic wind) and surface wind, as reflected by the coefficient of \( 10^5 \) (hereafter referred to as \( z_h \)) in Eq. (4) for the ESDU method and \( A_0 \) and \( B_0 \) in Eq. (5) for the drag-law-based method, respectively. As stated in ESDU (1982), Eq. (4) can provide estimations with ±3% deviation. For the drag-law-based method, estimations of these two coefficients differ slightly in the literature. Zilitinkevich (1989) suggested \( A_0 = 1.7 \) and \( B_0 = 4.5 \). Simiu (1973) recommended \( A_0 = 1.4 \) and \( B_0 = 4.7 \), whereas Garratt (1994) advised \( A_0 = 2 \) and \( B_0 = 4.5 \).

To estimate the uncertainty due to these parameters, Fig. 8 depicts the CF values calculated via both the ESDU and the drag-law-based exposure adjustment methods with varied inputting coefficient values. As reflected, different \( A_0 \) and \( B_0 \) values for the drag-law-based method correspond to almost identical CF predictions (the largest relative difference of 1.7% occurs at \( U_m = 8 \text{ m s}^{-1} \)). This is consistent with the discussion in Simiu and Scanlan (1996) who revealed that the inaccuracy of the two coefficients would have little effect on the calculated results of surface wind speeds for the case in which \( z_0 < 0.3 \text{ m} \). By contrast, \( z_h \) may bring in a relatively more evident influence on the estimation results. Again, the largest difference occurs at \( U_m = 8 \text{ m s}^{-1} \), and the CF\((T = 600\text{ s}, \tau = 600\text{ s})\) values for \( z_h = 10^4, 5 \times 10^4, 10^5, \) and \( 5 \times 10^5 \text{ m} \) are 0.754, 0.715, 0.678, and 0.639, respectively.
0.705, and 0.686, respectively. The ESDU method (ESDU 1982) is based on the D–H profile model, which predicts that the gradient height increases with the surface wind speed. However, for TC wind cases, observations show that the maximum wind (which is usually regarded as the gradient wind) in the inner region of a TC storm is stronger and located lower than that in the more outer area. Thus, the efficiency of the ESDU method for TCs may need further verification. Meanwhile, the drag-law-based method is deduced for geostrophic winds. For TC winds, as the centripetal forces may become evident, values of $A_0$ and $B_0$ may be different from those for monsoon winds. However, this effect is expected to be little for low-to-moderate winds.

It is noted that different uncertainty sources may counterbalance each other to some extent, and the error involved in the obtained results of this study are expected to be marginal. This can be reflected by the good agreement between the CFs calculated via different methods as shown in Fig. 5. While all the above-mentioned uncertainty sources should be minimized as far as possible, the main error involved in the calculated CF results is expected to come from the inappropriate use of the marine roughness predictor. Especially, the marine roughness values over shallow water and deep ocean may differ with each other remarkably under strong wind conditions.

5. Concluding remarks

An accurate estimation of surface wind strength is of great importance for many applications in the fields of both meteorology and wind engineering. Thus, it is necessary to standardize the raw wind records into their potential values so as to eliminate the undesirable distortion effects associated with topography, terrain, and height.

This study focused on the standardization of raw surface wind speed records collected at an offshore site. A standardization scheme for marine surface winds was proposed based on a combined application of the exposure adjustment methods and marine roughness estimation techniques. The validity of this method was verified through comparing the corrected mean wind speeds at a buoy site against those at a nearby station. Three uncertainty sources were discussed. The obtained results demonstrated that marine roughness for shallow waters may differ from that over deep oceans markedly, especially under strong wind conditions. This is consistent with previous field measurements as summarized in Table 1. Thus, an inappropriate use of the marine roughness predictor may result in significant estimation errors. Unfortunately, insufficient attention has been paid to shallow-water conditions, which are sometimes treated equally with deep-ocean conditions for simplification.

It is noted that because of the lack of detailed sea-state information, the efficiency of sea-state-based marine roughness estimators for coastal situations has not been examined in this study. To further verify/improve the results, more datasets, especially those collected during strong wind events ($U_m > 20 \text{ m s}^{-1}$), are desirable.

Acknowledgments. The authors express their gratitude to the Hong Kong Observatory for the provision of the wind data records and the permission to use the data for this study. The work described by this study was supported by grants from the National Natural Science Foundation of China (Project 51408520 and 51278439) and from the Research Grants Council of the Hong Kong Special Administrative Region, China (Project CityU 118213).

REFERENCES


——, ——, and ——, 2004: Im—


