A New Drag Relation for Aerodynamically Rough Flow over the Ocean

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ABSTRACT

From almost 7000 near-surface eddy-covariance flux measurements over the sea, the authors deduce a new air–sea drag relation for aerodynamically rough flow:

\[ u^* = 0.0583U_{N10} - 0.243. \]

Here \( u^* \) is the measured friction velocity, and \( U_{N10} \) is the neutral-stability wind speed at a reference height of 10 m. This relation is fitted to \( U_{N10} \) values between 9 and 24 m s\(^{-1}\). A drag relation formulated as \( u^* \) versus \( U_{N10} \) has several advantages over one formulated in terms of \( C_{DN10} \) \( (u^*/U_{N10})^2 \). First, the multiplicative coefficient on \( U_{N10} \) has smaller experimental uncertainty than do determinations of \( C_{DN10} \). Second, scatterplots of \( u^* \) versus \( U_{N10} \) are not ill posed when \( U_{N10} \) is small, as plots of \( C_{DN10} \) are; \( u^* - U_{N10} \) plots presented here suggest aerodynamically smooth scaling for small \( U_{N10} \). Third, this relation depends only weakly on Monin–Obukhov similarity theory and, consequently, reduces the confounding effects of artificial correlation. Finally, with its negative intercept, the linear relation produces a \( C_{DN10} \) function that naturally rolls off at high wind speed and asymptotically approaches a constant value of 3.40 \( \times \) 10\(^{-3}\). Hurricane modelers and the air–sea interaction community have been trying to rationalize such behavior in the drag coefficient for at least 15 years. This paper suggests that this rolloff in \( C_{DN10} \) results simply from known processes that influence wind–wave coupling.

1. Introduction

Turbulence research in the atmosphere has built on the groundwork laid by fluid mechanics research in the laboratory. But, atmospheric research may also have suffered by assuming too much similarity with laboratory fluid mechanics. As an example, we consider here drag parameterizations at the air–sea interface.

In fluid mechanics texts, most discussions of fluid motion start with the Bernoulli equation (e.g., Batchelor 1970, p. 158; Faber 1995, p. 46):

\[ P + \frac{1}{2} \rho U^2 + \rho g z = \text{const.} \tag{1.1} \]

Here \( P \) is the fluid pressure, \( \rho \) is the fluid density, \( U \) is the flow speed, \( g \) is the acceleration of gravity, and \( z \) is the height above some arbitrary reference level. This equation essentially states that the energy per unit volume is constant along a streamline in a fluid flow.

We can change this constant, however, by placing an obstacle in the flow—a sphere, for example. The total drag on such an obstacle is generally expressed as

\[ D = \frac{1}{2} \rho AC_D U^2, \tag{1.2} \]

where \( A \) is the frontal area of the obstacle and \( C_D \) is its drag coefficient. In the context of (1.1), \( D/A \) can be
thought of as the change in the constant along a streamline that results from frictional losses to the obstacle; the $\frac{1}{2}$ in (1.2) emphasizes the concept that the drag is a change in kinetic energy per unit volume in the fluid. Because energy can be changed only by adding or subtracting momentum, $D/A$ can also be thought of as a momentum flux.

When the problem turned to understanding the coupling between air and sea, early oceanographers and atmospheric scientists parameterized the wind’s drag on the sea surface as in (1.2) (e.g., Sverdrup et al. 1942; Francis 1954; Neumann 1956; Wilson 1960; Roll 1965, p. 152; Neumann and Pierson 1966, 208–210, p. 414):

$$\tau = \rho_a C_{D_r} U_r^2.$$ (1.3)

Here $\tau$ is the drag per unit area of sea surface (also called the surface stress or the momentum flux), $\rho_a$ is the air density, $U_r$ is the wind speed at some reference height $r$ above the sea, and $C_{D_r}$ is the dimensionless drag coefficient appropriate for $r$. Although the $\frac{1}{2}$ appeared in some early atmospheric versions of (1.3) (e.g., Sutton 1953, p. 232; von Arx 1967, p. 113) to emphasize its derivation from (1.2), modern versions are like (1.3), with the $\frac{1}{2}$ absorbed into $C_{D_r}$.

With the advent of Monin–Obukhov similarity theory, $C_{D_r}$ became a theoretical—not just empirical—coefficient (e.g., Garratt 1992, 52–55):

$$C_{D_r} = \frac{k}{\ln(r/z_0) - \psi_m(r/L)}.$$ (1.4)

In this, $k$ ($=0.40$) is the von Kármán constant; $z_0$, the roughness length; and $\psi_m$, an empirical function of the stratification parameter $L$, the Obukhov length.

Equation (1.4) actually derives through (1.3) from the similarity equation for the wind speed profile in the atmospheric surface layer:

$$U(z) = \frac{u_*}{k} \left[ \ln(z/z_0) - \psi_m(z/L) \right],$$ (1.5)

where $z$ is the height above the surface and $u_*$ is the friction velocity such that

$$\tau = \rho_a u_*^2.$$ (1.6)

In (1.5), we see that $z_0$ is the artificial height at which the wind speed is zero and is presumably a fundamental aerodynamic property of the surface (e.g., Wieringa 1993). Hence, $z_0$ and $C_{D_r}$ imply the same information.

To emphasize this point and to make comparing measurements of $C_{D_r}$ more meaningful, we usually eliminate the stability dependence in (1.4) and choose 10 m as the standard reference height. Then (1.4) becomes an expression for the neutral-stability, 10-m drag coefficient:

$$C_{DN_{10}} = \frac{k}{\ln(10/z_0)}^2,$$ (1.7)

where $z_0$ is expressed in meters. Likewise, we often plot $C_{DN_{10}}$ versus the neutral-stability wind speed at 10 m, which derives from (1.5) with the stability term ignored:

$$U_{N_{10}} = \frac{u_*}{k} \ln(10/z_0).$$ (1.8)

Equations (1.7) and (1.8) also provide the useful result

$$C_{DN_{10}} = \left( \frac{u_*}{U_{N_{10}}} \right)^2.$$ (1.9)

Over 50 years of research to develop a unified parameterization for $C_{DN_{10}}$ has, however, not narrowed the range of reported $C_{DN_{10}}$ values or satisfactorily explained that range. Reviews repeatedly show plots of widely spread $C_{DN_{10}}$ values at any given wind speed (e.g., Kraus 1968; Garratt 1977; Blanc 1985; Geernaert 1990; Banner et al. 1999; Toba et al. 2001; Drennan et al. 2005). It has long been suspected that $C_{DN_{10}}$ responds to other variables than just wind speed, including to sea state; see Jones and Toba (2001) for a review. Ignoring these dependencies was presumed to explain the scatter.

We have some other ideas, however, on why $C_{DN_{10}}$ has been so hard to pin down. First, consider the fundamental uncertainty in $C_{DN_{10}}$ as computed from (1.9). Measurements of $u_*$ over the sea have a minimum uncertainty of about $\pm 10\%$. [See, for instance, Fairall et al. (1996, Table 1) for typical uncertainties.] When $u_*$ is small, this uncertainty can approach $\pm 100\%$. Although $U_r$ may be measured at sea on a ship, from a tower or buoy, or from an aircraft with an uncertainty of, say, $\pm 5\%$, obtaining $U_{N_{10}}$ necessitates a stability correction involving $L$. Because $L$ requires $u_*^3$ and measurements of the surface heat fluxes, it is probably uncertain by at least $\pm 30\%$. Hence, the minimum uncertainty in $C_{DN_{10}}$ (i.e., if we assume that $U_{N_{10}}$ has the same uncertainty as $U_r$) is $\pm 30\%$. Other phenomena that are presumed to affect $C_{DN_{10}}$—such as sea state, swell, and the relative wind direction between wind and waves—likely make contributions smaller than this in moderate and high winds (cf. Janssen 1997); these effects are thus hidden in the uncertainty of the basic measurements.

In addition to these fundamental considerations of uncertainty, in stable stratification, the magnitude of $\psi_m$ in (1.4) and (1.5) can become comparable to $\ln(z/z_0)$.
Hence, in light winds, \( C_{DN} \) can become unrealistically small and is dominated by the uncertainty in \( L \). At the same time, \( U_{N10} \) can also become very small—and, at times, negative [see (2.1) below]. Clearly \( C_{DN10} \) is problematic in such conditions.

Recently, Foreman and Emeis (2010) focused on yet another problem with \( C_{DN10} \) by suggesting that the definition of the drag coefficient [which has its roots in (1.2)] is fundamentally flawed. Namely, we could reasonably infer from (1.9) that \( u_* \) is proportional to \( U_{N10} \) with a proportionality constant of \( C_{DN10} \). When Foreman and Emeis plotted roughly a thousand points from the literature as \( u_* \) versus \( U_{N10} \), however, they obtained

\[
u_* = aU_{N10} + b \tag{1.10}\]

for \( U_{N10} \geq 8 \text{ m s}^{-1} \) (presumably aerodynamically rough flow). In (1.10), \( u_* \) and \( U_{N10} \) are in meters per second, \( a = 0.051 \), and \( b = -0.14 \text{ m s}^{-1} \). Equation (1.10) shows that \( u_* \) is linearly related to \( U_{N10} \) but is not proportional to it.

This absence of proportionality is interesting because (1.10) then implies

\[
C_{DN10} = \left( \frac{u_*}{U_{N10}} \right)^2 = a^2 \left( 1 + \frac{b}{aU_{N10}} \right)^2 . \tag{1.11}\]

That is, instead of increasing linearly with \( U_{N10} \), as in most formulations of \( C_{DN10} \) (e.g., Garratt 1977; Smith 1980; Geernaert 1990; Smith et al. 1992), here \( C_{DN10} \) is more complex. Moreover, because \( b \) is negative, \( C_{DN10} \) rises, rolls off, and asymptotes to \( a^2 \) at high wind speed.

The hurricane community has been searching for behavior like this in \( C_{DN10} \) since Emanuel (1995) reported that hurricane models could not produce storms with enough intensity if their drag parameterization was simply an extrapolation of results from moderate wind speeds, which had \( C_{DN10} \) increasing linearly with \( U_{N10} \), without bounds. Modern hurricane and ocean mixed-layer models, on the other hand, have had some success in predicting storm intensity and ocean response by limiting the value of \( C_{DN10} \) in high winds (Jarosz et al. 2007; Moon et al. 2007; Sanford et al. 2007; Chiang et al. 2011).

Equation (1.10) has features aligned with our own philosophy of air–sea interaction: 1) The experimental coefficient \( a \) has only half the experimental uncertainty of \( C_{DN10} \) and is, thus, more reliably measured; 2) a plot of \( u_* \) versus \( U_{N10} \) does not have pathological behavior when \( U_{N10} \) is near zero, as do plots of \( C_{DN10} \); 3) (1.10) minimizes reliance on Monin–Obukhov similarity theory and thereby suffers little from the fictitious correlation typical of these types of analyses (e.g., Mahrt et al. 2003; Klipp and Mahrt 2004; Grachev et al. 2007a,b; Andreas 2011b); and 4) (1.11) produces a natural limit to \( C_{DN10} \).

Because of these merits in (1.10), after reading Foreman and Emeis (2010), we quickly plotted \( u_* \) versus \( U_{N10} \) for data that we had on hand. Figure 1 shows that our results corroborate those of Foreman and Emeis. We find

\[
u_* = 0.0581U_{N10} - 0.214 \tag{1.12}\]

for data in the aerodynamically rough flow regime, \( U_{N10} \geq 9 \text{ m s}^{-1} \); the correlation coefficient of these data is 0.929. We will elaborate on this figure later; but, for now, it showed enough promise for us to commit to a full study of the drag parameterization that Foreman and Emeis suggested.

As such, we add over 6000 more values measured by low-flying aircraft in winds up to 27 m s\(^{-1}\) to the 778 points shown in Fig. 1. This aircraft set also shows a straight-line relation between \( u_* \) and \( U_{N10} \) in the aerodynamically rough regime, and the fitting coefficients are not statistically different from those in (1.12).

Both datasets also suggest that \( u_* \) follows the prediction for aerodynamically smooth flow for low \( U_{N10} \). Consequently, we devise a continuous drag relation for all \( U_{N10} \) by smoothly combining this aerodynamically smooth regime with (1.10) for the aerodynamically rough regime.

On extrapolating this relation to hurricane-strength winds, we find that it predicts the rolloff in \( C_{DN10} \) that hurricane models seem to require. Moreover, the
straight-line behavior of \( u_* \) with \( U_{N10} \)—even in high winds—and the rolloff in \( CD_{N10} \) are compatible with theoretical models by Moon et al. (2007) and Mueller and Veron (2009) that compute the air–sea drag as resulting from just skin friction and the form drag from flow separation over the waves. In other words, our analysis suggests that there is no need to invoke exotic processes, such as sea spray loading or the disintegration of the air–sea interface, to explain the roll off in \( CD_{N10} \) with increasing wind speed. Wind–wave coupling suffices.

2. Datasets

Table 1 summarizes the data with which we made our first test of the Foreman and Emeis (2010) approach (i.e., our Fig. 1). We will refer to this as the “original” dataset. Most of these sets are available as tabulations in the cited references. We obtained the Fronts and Atlantic Storm Track Experiment (FASTEX) and Greenland Flow Distortion Experiment (GFDex) data, however, as electronic files from the scientists referenced for these sets.

We did not include the Coupled Boundary Layers Air–Sea Transfer (CBLAST) hurricane dataset mentioned in Table 1 in this analysis because these data are not consistent with our other data: The \( u_* \) values tend to be low, as we will show later. We suspect that this bias resulted because these aircraft data were obtained at flight levels that were never below 70 m, were as high as 383 m, and had a median level of 193 m, while the depth of the boundary layer for these flights during Hurricanes Fabian and Isabel was 350–550 m (Zhang et al. 2009). That is, because the stress is known to decrease with height through the boundary layer (e.g., Caughey et al. 1979; Nicholls and Readings 1979; Zhang et al. 2009; Wyngaard 2010, 244–247, 286–287), the measured flight-level stress was less than the surface stress. Although French et al. (2007) tried to correct for this flux divergence, their reported values of \( u_* \) remain low.

Table 2 summarizes a second set of data that we use in this study. Because all of these data come from low-flying aircraft, we will refer to this as our “aircraft” dataset.

Four different aircraft collected these data: the National Oceanic and Atmospheric Administration’s Long-EZ, the C-130 and Electra from the National Center for Atmospheric Research (NCAR), and the Twin Otter from the Naval Postgraduate School’s Center for Interdisciplinary Remotely Piloted Aircraft Studies (CIRPAS) (Khelif et al. 2005).

To measure the turbulent wind vector required for computing the momentum flux, the Twin Otter, C-130, and Electra used five-port radomes on the nose of the aircraft. Each port had a pressure sensor that sampled at 20–25 Hz; Lenschow (1986) describes the principles of obtaining the wind vector from aircraft pressure measurements. Each of these three aircraft used the Global Positioning System (GPS) to correct the aircraft’s inertial navigation system to find true ground speed (Khelif et al. 1999).

The Long-EZ used the Best Atmospheric Turbulence Probe (BAT) for measuring the wind vector (Crawford and Dobosy 1992; Garman et al. 2006). This is a baseball-bat-shaped device with nine pressure ports in its thicker end; it protruded from the Long-EZ into the undisturbed free stream ahead of the aircraft. A pressure sensor in each port sampled at 50 Hz; again, the aircraft’s inertial navigation system was corrected with GPS positioning to obtain the true wind vector with respect to the ground.

Regardless of aircraft, each flux value in the aircraft set is the average over a 4-km flight segment that we processed ourselves from the raw data. We use the flight-level momentum flux as \( r u_*^2 \); that is, we made no adjustments for height because the aircraft were always below 50 m.
We did only minimal initial screening of the aircraft data for quality control (cf. Mahrt et al. 2012). In low winds, wave effects or uncertainty in the aircraft turbulence measurements can produce a stress that appears upward, contrary to boundary layer theory. We screened for such spurious stress measurements and eliminated 662 cases from the initial 6080 flight legs summarized in Table 2. Over 95% of these questionable measurements occurred with flight-level winds of less than 8 m s\(^{-1}\).

On the other hand, the original authors of the Janssen, Risø Air–Sea Experiment (RASEX), Southern Ocean Waves Experiment (SOWEX), and Surface Wave Dynamics Experiment (SWADE) in Table 1 probably screened these datasets more strictly before publishing them and reported only “high quality” fluxes. Although the FASTEX, GFDex, and Humidity Exchange over the Sea (HEXOS) sets had been processed when we received them, we suspect that these sets had been screened only for instrument malfunctions—not for stationarity, homogeneity, wave characteristics, or other relevant quality metrics.

We further screened the SWADE set ourselves. The original set has 126 records; and Donelan et al. (1997) had identified whether each record represented conditions of wind sea or whether there was swell with wind in the same direction, in the opposing direction, or at right angles. In Fig. 1, we include only the 27 SWADE cases with wind sea; later we will discuss the remaining SWADE data that had swell present.

A distinct feature of all the data represented in Tables 1 and 2 is that they come from eddy-covariance measurements of the momentum and heat fluxes. We eschewed available datasets that were based on inertial-dissipation estimates of the fluxes because such fluxes rely heavily on Monin–Obukhov similarity theory. We want to minimize our reliance on similarity theory.

Initially in our analysis, we will treat the original and aircraft datasets separately. In effect, we are using the aircraft dataset to validate our analysis of the original dataset, or vice versa.

The method used for estimating \(U_{N10}\) for this work is crucial. One approach, obviously, is to use (1.8) and the measured \(u^*\). The required \(z_0\) could come from the corresponding measured value, from a lookup table or a comparable standard (e.g., Panofsky and Dutton 1984, 121–123; Stull 1988, p. 380; Wieringa 1993), or from a parameterization such as the Charnock relation. With that approach, however, the dependent variable in the analysis, the measured \(u^*\), will be very well (and artificially) correlated with the independent variable \(U_{N10}\).

To avoid such tautology, we start instead with (1.5). When our interest is in the wind speed at 10 m, we can rewrite (1.5) as

\[
\frac{u_a}{k} \ln(10/z_0) = U_{N10} = U(z) - \frac{u_a}{k} \ln(z/10) + \frac{u_a}{k} \psi_m(z/L).
\]  

(2.1)

That is, if we use the right-hand side of this equation to obtain \(U_{N10}\), \(U_{N10}\) is most sensitive to the actual wind measurement, \(U(z)\), and has generally only a modest built-in dependence on the measured \(u_a\). Furthermore, if the measurement height is close to 10 m [when

### TABLE 2. Our “aircraft” dataset consists of 4-km flight segments that we processed ourselves; see Mahrt et al. (2012) for additional details. “Number of runs” here is the number of such 4-km legs. The “Altitude range” gives the aircraft flight level; the wind speed noted is the range of measured wind speeds at those levels. The cited references give more details on the measurements.

<table>
<thead>
<tr>
<th>Dataset</th>
<th>Number of runs</th>
<th>Altitude range (m)</th>
<th>Range in wind speed (m s(^{-1}))</th>
<th>Aircraft/location</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>CARMA4</td>
<td>650</td>
<td>27–40</td>
<td>0.5–18.1</td>
<td>CIRPAS Twin Otter, off coast of southern California</td>
<td>Edson et al. (2007)</td>
</tr>
<tr>
<td>CBLAST-weak</td>
<td>740</td>
<td>1–16</td>
<td>1.4–9.3</td>
<td>Long-EZ, Martha’s Vineyard, MA</td>
<td>Mahrt and Khelif (2010)</td>
</tr>
<tr>
<td>GOTEK</td>
<td>859</td>
<td>24–49</td>
<td>2.3–27.1</td>
<td>NCAR C-130, Gulf of Tehuantepec</td>
<td>Romero and Melville (2010)</td>
</tr>
<tr>
<td>Monterey</td>
<td>654</td>
<td>26–39</td>
<td>2.2–18.0</td>
<td>CIRPAS Twin Otter, off Monterey, CA</td>
<td>Anderson et al. (2004)</td>
</tr>
<tr>
<td>POST</td>
<td>189</td>
<td>22–40</td>
<td>2.6–13.9</td>
<td>CIRPAS Twin Otter, off Monterey, CA</td>
<td>Sun et al. (2001)</td>
</tr>
<tr>
<td>RED</td>
<td>373</td>
<td>23–49</td>
<td>1.4–19.9</td>
<td>CIRPAS Twin Otter, east of Oahu, Hawaii</td>
<td></td>
</tr>
<tr>
<td>SHOWEX</td>
<td>508</td>
<td>10–49</td>
<td>1.9–12.1</td>
<td>Long-EZ, off coast of Virginia and North Carolina</td>
<td></td>
</tr>
<tr>
<td>SHOWEX (November 1997)</td>
<td>199</td>
<td>8–48</td>
<td>3.4–17.3</td>
<td>Long-EZ, off coast of Virginia and North Carolina</td>
<td></td>
</tr>
<tr>
<td>SHOWEX (March 1999)</td>
<td>970</td>
<td>3–48</td>
<td>0.5–16.5</td>
<td>Long-EZ, off coast of Virginia and North Carolina</td>
<td></td>
</tr>
<tr>
<td>TOGA COARE</td>
<td>938</td>
<td>26–43</td>
<td>0.5–9.4</td>
<td>NCAR Electra, western equatorial Pacific Ocean</td>
<td></td>
</tr>
</tbody>
</table>
In (\(z/10 \sim 0\)) and if the stratification is near neutral [when \(\psi_m(z/L) \sim 0\)], the \(U_{N10}\) obtained from (2.1) has very weak built-in correlation with \(u_*\), and \(U_{N10}\) has nearly the same uncertainty as \(U(z)\). A \(U_{N10}\) estimated from (1.8), in contrast, always has an uncertainty no smaller than the uncertainty in \(u_*\).

When we could, we estimated \(U_{N10}\) from (2.1). For all the aircraft data in Table 2, this was the case. For \(\psi_m\), we used the function from Paulson (1970) in unstable stratification and the function from Grachev et al. (2007a) in stable stratification. For the FASTEX, GFDex, and RASEX data in Table 1, we had enough information to also calculate \(U_{N10}\) according to (2.1). In the CBLAST-hurricane set, \(U_{N10}\) was estimated with a stepped-frequency microwave radiometer (Drennan et al. 2007) and thus has no built-in dependence on \(u_*\). In the HEXOS set, DeCosmo (1991) reported only \(U_{N10}\) and did not explain how she obtained this value. In the SOWEX set, Banner et al. (1999) obtained \(U_{N10}\) from (1.8). In the SWADE set, Donelan et al. (1997) reported \(U_{N10}\) and explained that they obtained it from (2.1).

Finally, Janssen (1997) reported only a variable denoted \(U_{10}\) but did not explain how this was obtained or whether it is the neutral-stability value. He did, however, report two simultaneous, independent sets of measurements: \(u_*\) and \(U_{10}\) were measured with both a pressure anemometer and a sonic anemometer. Under the assumption that \(U_{10}\) is \(U_{N10}\) but to avoid the built-in correlation in case Janssen estimated \(U_{N10}\) from (1.8), we switched the pressure anemometer and sonic measurements of \(u_*\) in our analysis. In other words, in Fig. 1 and subsequent figures, we plot \(u_*\) from the sonic against the corresponding \(U_{N10}\) value from the pressure anemometer and \(u_*\) from the pressure anemometer against the sonic \(U_{N10}\). Now, in the Janssen set, the \(u_*\) and \(U_{N10}\) pairs have no direct built-in correlation from a shared \(u_*\).

3. Results

Figure 2 is a plot like Fig. 1 but for the aircraft data summarized in Table 2. Unlike the original set, a high percentage of the aircraft data were collected in stable stratification—1123 of the 5418 data records left after our screening the stress. While we are not concerned about the flux divergence in unstable stratification for fluxes measured at the aircraft altitudes in this set (up to 49 m), we worry about possible flux divergence in stable stratification because of the generally shallower boundary layer. (Remember, we obtain the \(u_*\) surface value from the uncorrected momentum flux measured at flight level.)

To avoid biasing our analysis with \(u_*\) values biased low because of vertical flux divergence, we made several plots and analyses as in Fig. 2. Table 3 summarizes the calculations. First, we considered the aircraft data collected in unstable stratification and, separately, the data collected in stable stratification. Admittedly, the stable cases constituted only 26% as much data as in the unstable cases; still, as expected because of the flux divergence, the \(a\) value for the data collected in stable stratification is significantly less than the \(a\) value for the unstable cases.

Moreover, when we further segregated the stable data into cases with \(0 \leq z/L \leq 0.1\) and with \(0 \leq z/L \leq 0.2\), where \(z\) is the aircraft altitude and \(L\) is the measured Obukhov length, the set with the more stable conditions had a smaller \(a\) value than for the weakly stable set (Table 3). Finally, when we included just this weakly stable set (i.e., \(0 \leq z/L \leq 0.1\)) with all the unstable data, the resulting \(a\) and \(b\) values were indistinguishable from the \(a\) and \(b\) values for just the unstable data. Hence, as our analyzed “aircraft” dataset here, we use all the aircraft data collected in unstable stratification and the data from weakly stable stratification when \(0 \leq z/L \leq 0.1\).

This screening and the previously mentioned screening for stress reduced the original 6080 data records shown in Table 2 to 4878 records.

In both Figs. 1 and 2, the data clouds change character in the \(U_{N10}\) range 8–10 m s\(^{-1}\). Below this range, the points have a shallower slope than above it. This tendency is compatible with aerodynamically smooth flow at low wind speeds and the transition to aerodynamically rough flow as the wind speed increases.

The roughness Reynolds number

\[
R_* = \frac{u_* z_0}{v}
\]
characterizes the roughness regime of the flow, where \( \nu \) is the kinematic viscosity of air. For \( R_\kappa \leq 0.135 \) [see Andreas and Trevisio (2000) for a discussion of this choice], the flow is aerodynamically smooth; for \( R_\kappa \geq 2.5 \) (e.g., Kraus and Businger 1994, p. 145), the flow is aerodynamically rough. In between these limits, the flow is in transition.

Previously, Wu (1969, 1980), Melville (1977), Kraus and Businger (1994, p. 145), and Foreman and Emeis (2010), for instance, discussed what wind speed or friction velocity is required for the sea surface to become aerodynamically rough. Wu (1980), Kraus and Businger, and Foreman and Emeis based their analyses, however, on the assumption that the Charnock relation

\[
z_0 = \frac{a u^2_b}{\nu} \tag{3.2}
\]

specifies the wave-related roughness length. Here, \( \alpha \) is the Charnock parameter, \( 0.01 \leq \alpha \leq 0.02 \). While Wu (1969) and Melville based their analyses on data, we revisit the discussion of aerodynamic regimes here because we have much more data than they had available.

Figures 3 and 4 show our evaluation of \( R_\kappa \) for the two datasets. Each figure includes the individual values, bin averages in 1 m s\(^{-1}\) bins in \( U_{N10} \), and bin medians. Because \( R_\kappa \) values are approximately lognormally distributed, the proper bin average is computed as the geometric mean—as opposed to the arithmetic mean. If the \( R_\kappa \) values were perfectly lognormal within a bin, the median would be the same as the geometric mean. Notice that in both figures the medians and geometric means are close.

Both figures show that, on average, the sea surface is not aerodynamically rough until \( U_{N10} \) is greater than 8 m s\(^{-1}\). As a conservative estimate, we therefore used only the data for which \( U_{N10} \geq 9 \) m s\(^{-1}\) to determine the fitting lines in Figs. 1 and 2 and in Table 3. Equation (1.12) already gave our fit for Fig. 1; for Fig. 2, least squares linear regression yields (Table 3)

\[
u = 0.0583U_{N10} - 0.243, \tag{3.3}
\]

where \( u_* \) and \( U_{N10} \) are in meters per second. The correlation coefficient is 0.835.

Bendat and Piersol (1971, p. 131) explain that both the slope and intercept in (1.12) and (3.3) follow Student \( t \) distributions. We therefore calculated 95% confidence intervals on the slopes and intercepts fitted to the original and aircraft data in Figs. 1 and 2 (Table 3). For the slope in (1.12), the 95% confidence interval is \([0.0564, 0.0599]\); for the intercept, \([-0.239, -0.189]\). For the slope in (3.3), the 95% confidence interval is \([0.0566, 0.0600]\); for the intercept, \([-0.266, -0.221]\).

---

### Table 3. Fits to the model \( u_* = a U_{N10} + b \) for the original dataset and various configurations of the aircraft data. All cases use only data pairs for which \( U_{N10} \geq 9 \) m s\(^{-1}\). The columns are the number of pairs in the fitting, the correlation coefficient \( r \), the parameter \( a \), the 95% confidence interval on \( a \) [from Bendat and Piersol (1971), p. 131], parameter \( b \), and the 95% confidence interval on \( b \).

<table>
<thead>
<tr>
<th>Data source</th>
<th>Number</th>
<th>( r )</th>
<th>( a ) 95% on ( a )</th>
<th>( b ) (m s(^{-1})) 95% on ( b ) (m s(^{-1}))</th>
</tr>
</thead>
<tbody>
<tr>
<td>Original set</td>
<td>658</td>
<td>0.929</td>
<td>0.0581 ([0.0564, 0.0599])</td>
<td>-0.214 ([-0.239, -0.189])</td>
</tr>
<tr>
<td>All aircraft</td>
<td>1988</td>
<td>0.826</td>
<td>0.0584 ([0.0567, 0.0602])</td>
<td>-0.252 ([-0.275, -0.229])</td>
</tr>
<tr>
<td>Aircraft, all unstable</td>
<td>1680</td>
<td>0.835</td>
<td>0.0588 ([0.0569, 0.0606])</td>
<td>-0.246 ([-0.270, -0.221])</td>
</tr>
<tr>
<td>Aircraft, all stable</td>
<td>308</td>
<td>0.788</td>
<td>0.0512 ([0.0467, 0.0557])</td>
<td>-0.223 ([-0.279, -0.167])</td>
</tr>
<tr>
<td>Aircraft, 0 ≤ ( z/L ) ≤ 0.1</td>
<td>215</td>
<td>0.846</td>
<td>0.0528 ([0.0483, 0.0572])</td>
<td>-0.205 ([-0.262, -0.148])</td>
</tr>
<tr>
<td>Aircraft, 0 ≤ ( z/L ) ≤ 0.2</td>
<td>263</td>
<td>0.820</td>
<td>0.0508 ([0.0465, 0.0551])</td>
<td>-0.198 ([-0.252, -0.143])</td>
</tr>
<tr>
<td>Aircraft, all unstable, stable with 0 ≤ ( z/L ) ≤ 0.1</td>
<td>1895</td>
<td>0.835</td>
<td>0.0583 ([0.0566, 0.0600])</td>
<td>-0.243 ([-0.266, -0.221])</td>
</tr>
</tbody>
</table>

---

**Fig. 3.** Individual values of the measured roughness Reynolds number \( R_\kappa \) (gray circles), from the original dataset (Table 1) are plotted against \( U_{N10} \). The horizontal lines show the aerodynamically smooth limit (at 0.135) and the aerodynamically rough limit (at 2.5). The black circles are geometric means of the individual values within \( U_{N10} \) bins 1 m s\(^{-1}\) wide; the error bars are ±2 standard deviations in these bin means. The red circles are medians of the points within a bin. The blue line summarizes our analysis and derives from (3.1), (3.4), and (4.3).
Because both slope and intercept intervals coincide well, Figs. 1 and 2 give us essentially the same result. In effect, we validate the fitting line in Fig. 1 with Fig. 2, and vice versa. Henceforth, we will use the coefficients in (3.3) as our main result because they come from the larger dataset.

The very large $R^*$ values at small $U_{N10}$ in Figs. 3 and 4 are related to the “pathological behavior” in $C_{DN10}$ for small $U_{N10}$ that we mentioned earlier. The $z_0$ values used to create these figures came from the left-hand side of (2.1) [or, alternatively, from (1.5)]:

$$z_0 = 10 \exp \left( -\frac{kU_{N10}}{u_*} \right) .$$

When both $U_{N10}$ and $u_*$ are small, their uncertainties often cause $kU_{N10}/u_*$ to be unrealistically small. Consequently, $z_0$ is unrealistically large, and so is $R^*$.

The slope in (1.10) that Foreman and Emeis (2010) reported (0.051) is smaller than our values, and their intercept ($-2.04$ m s$^{-1}$) is larger. We suspect that, because they used $U_{N10} = 8$ m s$^{-1}$ as the lower limit for aerodynamically rough flow in their analysis, they may have retained some data reflecting aerodynamic transition. Notice in Figs. 1 and 2 how the $u_*$ values at lower winds turn up and away from (1.12) and (3.3). By including such data in their calculations, Foreman and Emeis would have inadvertently decreased $a$ and increased $b$ from what the data in truly rough flow suggest.

At $U_{N10} = 9$ m s$^{-1}$, (3.3) gives $u_* = 0.28$ m s$^{-1}$. In our analysis, this is the friction velocity at the transition to aerodynamically rough flow. For comparison, Wu (1969) concluded that this transition is at $U_{N10} = 7$ m s$^{-1}$, whereas Wu (1980) obtained $u_* = 0.263$ m s$^{-1}$, although he assumed $R^* = 2.33$ at the transition to aerodynamically rough flow. From his data analysis, Melville (1977) concluded that $u_*$ was in the range 0.15–0.30 m s$^{-1}$ at the onset of aerodynamically rough flow, although he also used for the transition an $R^*$ limit (=2) lower than ours. On invoking the Charnock relation, Kraus and Businger (1994, p. 145) and Foreman and Emeis (2010) estimated $u_* = 0.29$ and 0.28 m s$^{-1}$, respectively, at the transition to aerodynamically rough flow.

Hence, our estimate that $u_* = 0.28$ m s$^{-1}$ at the transition to aerodynamically rough flow agrees with most previous estimates; but our result that $U_{N10} = 9$ m s$^{-1}$ at this transition is a bit higher than previous estimates.

4. Discussion

a. Consistency of the results

Thoughtful readers might suspect that the data clouds in Figs. 1 and 2 obscure differences in behavior among the different datasets that are typical in plots of $C_{DN10}$ versus $U_{N10}$. Then our fitting lines in Figs. 1 and 2 would just be average results that ignore true differences in drag relations among the sets. To allay these worries, we created Figs. 5 and 6.

These show the individual datasets that went into Figs. 1 and 2. Figure 5 shows our original data; Fig. 6, the aircraft data. Reassuringly, 17 of the 18 datasets individually either lie along our aircraft fit, (3.3), suggest aerodynamically smooth scaling at low wind speed, or do both. That is, the individual datasets are not biased high or low such that, when we fitted (1.10) to the two consolidated datasets, the fitting line simply split the difference between systematically high and systematically low values.

The one exception to this behavior is the CBLAST hurricane dataset (Fig. 5). These $u_*$ values seem to be too low—probably for the reasons we discussed earlier. We have therefore not included them in our least squares fittings.

b. A unified roughness parameterization

The green lines in Figs. 1, 2, 5, and 6 show aerodynamically smooth scaling, where the roughness length is

$$z_{0s} = 0.135 \frac{\nu}{u_*} .$$


Because these green lines are virtually linear from $U_{N10} = 0$ to where they intersect our aerodynamically rough results, (1.12) and (3.3), we fitted them with
FIG. 5. Each dataset listed in Table 1 is plotted individually as $u_*$ vs $U_{N10}$. In each panel, the red line is the fit to the aircraft data, (3.3). The green line shows the aerodynamically smooth limit, (4.1). The Janssen plot is different from the others because Janssen (1997) reported simultaneous measurements of $u_*$ and $U_{N10}$ with both a sonic anemometer and a pressure anemometer. As discussed in the text, to avoid artificial correlation in this dataset, we paired the sonic $U_{N10}$ with the pressure anemometer $u_*$ and vice versa.
FIG. 6. As in Fig. 5, but for the individual datasets in the aircraft set (Table 2).
This line is appropriate for $U_{N10}$ in $[0, 8.76 \text{ m s}^{-1}]$.

Two intersecting lines now describe our results, (3.3) and (4.2). We can therefore represent $u_*$ with a smooth, differentiable function of $U_{N10}$ by combining (3.3) and (4.2) in a hyperbola. The result that best fits our data for all $U_{N10}$ is

$$u_* = 0.239 + 0.0433\{(U_{N10} - 8.271) + [0.120(U_{N10} - 8.271)^2 + 0.181]^{1/2}\}, \quad (4.3)$$

where $u_*$ and $U_{N10}$ are both in meters per second.

Figure 7 shows how well this expression fits the bin-averaged $u_*$ values from the combined original and aircraft datasets. Only for $U_{N10} < 3 \text{ m s}^{-1}$ do the data in Fig. 7 deviate significantly from (4.3). Instead of highlighting missing physics, these three large $u_*$ values reveal how difficult measuring $u_*$ is in low winds.

Equation (4.3) now becomes a core component of a new bulk flux algorithm that we are developing. The appendix sketches how we use (4.3) in this algorithm.

For readers used to looking at flux algorithms in terms of $C_{D10}$, we can insert (4.3) into (1.9) to obtain an expression for $C_{D10}$ for all wind speeds. Figure 8 shows this result and how it fits the bin-averaged $C_{D10}$ values in our combined original and aircraft datasets.

Figures 7 and 8 also reiterate some of the advantages of a drag relation based on $u_*$ over one based on $C_{D10}$ that we described in the introduction. Although the averaged $u_*$ values in the three lowest $U_{N10}$ bins in Fig. 7 do not follow aerodynamically smooth scaling, at least they are well behaved and have some of the smallest error bars in the plot. The $C_{D10}$ values in the two lowest $U_{N10}$ bins in Fig. 8, in contrast, are above the upper limit of the plot and thus do not show up at all. Moreover, the errors bars on the $C_{D10}$ values for small $U_{N10}$ are generally the largest on the plot and even encompass negative $C_{D10}$ values.

Figure 8 also suggests that the distribution of individual $C_{D10}$ values within bins is skewed toward larger values: For $U_{N10} < 9 \text{ m s}^{-1}$, the averages are larger than the medians. In Fig. 7, (4.3) fits the bin-averaged $u_*$ values very well. In contrast, the bin-averaged $C_{D10}$ values for $U_{N10} < 9 \text{ m s}^{-1}$ in Fig. 8 are above the $C_{D10}$ curve derived from (4.3) although the same data as in Fig. 7 went into this plot. All of these features are evidence of what we termed pathological behavior in $C_{D10}$.

c. Drag relations in high winds

Figure 9 shows (4.3) extrapolated to hurricane-strength winds. The figure also shows both the original and aircraft datasets to emphasize how consistent they are. Furthermore, Fig. 9 includes the CBLAST hurricane data to demonstrate that they are generally below reliable data measured at similar wind speeds.

The main features of Fig. 9, however, are the curves attributed to Moon et al. (2007) and Mueller and Veron (2009). These are theoretical results in which both sets of authors modeled the total wind stress on the sea as a combination of the viscous stress (or skin friction or...
tangential stress), a wave-induced stress from form drag, and the reduction of the viscous and wave-induced stresses by sheltering (or flow separation).

Moon et al. (2007) obtained their results [summarized in their (4) and (5)] by simulating the surface stress in 10 Atlantic hurricanes with wind speeds up to 70 m s\(^{-1}\) using the hurricane model of Moon et al. (2004). They then inferred \(z_0\) from this modeled stress through similarity theory. Their \(z_0\) values are thus statistical averages in wind speed bins. The Mueller and Veron (2009) model, on the other hand, simply provides a deterministic prediction of the surface stress for the given input conditions. [F. Veron (2011, personal communication) reran the Mueller and Veron model for our specified conditions for winds speed up to 80 m s\(^{-1}\).] Neither group evidently realized that its model yielded a nearly straight-line relation between \(u^*\) and \(U_{10}\) for \(U_{10}\) above 20–25 m s\(^{-1}\) (Fig. 9). We use this theoretical behavior in these two models to justify extrapolating our own result, (4.3), to hurricane-strength winds.

We cast our results in the familiar form of a drag coefficient in Fig. 10. Remember, because of (1.11), our drag coefficient rolls off and approaches an asymptotic limit of \(a^2\) in high winds: 3.40 \(\times\) 10\(^{-3}\). Figure 10 also shows the Moon et al. (2007) and Mueller and Veron (2009) results and the “Charnock + Smooth” curve, which we obtain by adding (4.1) and the Charnock relation, (3.2) [with \(\alpha = 0.0185\), Andreas et al. (2008)], to get a unified expression for \(z_0\) (e.g., Zilitinkevich 1969; Smith 1988; Fairall et al. 1996).

In models of the oceanic mixed layer under hurricanes, both Sanford et al. (2007) and Chiang et al. (2011) based parameterizations for \(C_{DN10}\) on the observations reported by Powell et al. (2003). While we do not endorse the Powell et al. results for several reasons, we show in Fig. 10 the Sanford et al. and Chiang et al. \(C_{DN10}\) parameterizations because they are continuous functions, like the other curves in Fig. 10, and because, we feel, they represent the lowest reasonable drag coefficient possible in high winds.

Finally, Fig. 10 also includes the drag coefficients that Bell (2010) obtained by using dropsondes launched in Hurricanes Fabian and Isabel in 2003 to estimate the angular momentum budget under the assumption that the storms were axisymmetric. These are the most recent determinations of air–sea drag in high winds that are available. Bell computed drag coefficients for six separate aircraft missions and used 72 control volumes per mission for flux calculations. The error bars on his points in Fig. 10 are thus (probably) standard deviations around the means of the 72 samples per mission. That is, they are not uncertainty estimates but rather indicators of the scatter in the individual values.

Bell’s (2010) results, unfortunately, do not help us decide which of the candidate drag parameterizations in Fig. 10 is the most realistic. His data range from below the lowest realistic parameterization to above our parameterization, which we suggest gives the greatest upper bound.
Ingel (2011) refers to the rolloff in $C_{DN10}$ depicted in Fig. 10 as a “drag crisis,” invoking the terminology of classical fluid mechanics when laminar flow transitions to turbulence and the drag coefficients of cylinders and spheres drop suddenly by a factor of 5 with increasing Reynolds number (Monin and Yaglom 1971, 82–83; Faber 1995, 266–267). That is, according to Ingel, something fundamental about the air–sea coupling has changed. Kudryavtsev (2006) believes that this rolloff signals a saturation of the surface stress: the stress no longer increases with increasing wind speed. Neither of these inferences is true.

If $u_*$ is a linear function of $U_{10}$ for moderate and high wind speeds—as our data and the theories of Moon et al. (2007) and Mueller and Veron (2009) suggest—nothing fundamental changes about the way air and sea couple as the wind speed increases. There is certainly no drag crisis in the classical sense. Nor does the surface stress saturate: (3.3) confirms that $u_*$ and thus the surface stress increase with wind speed for all wind speeds.

The roughness length does saturate, however, as Donelan et al. (2004) suggest. From (3.3) and (3.4), we see that $z_0$ approaches a limiting value of $1.05 \times 10^{-2}$ m [$=10 \exp(-k\alpha)$] at large $U_{10}$.

Because Moon et al. (2007) and Mueller and Veron (2009) account for the behavior that we see in the data by modeling just wind–wave coupling, invoking more exotic processes to explain the rolloff in $C_{DN10}$ seems unnecessary. Many of these attempts to explain the rolloff in $C_{DN10}$ involve injecting sea spray into the near-surface air in copious amounts (e.g., Makin 2005; Kudryavtsev 2006; Soloviev and Lukas 2010; Ingel 2011; Bianco et al. 2011). Such spray loading may stabilize the near-surface air and thus reduce the momentum transfer somewhat in very high winds.

Shpund et al. (2011) recently suggested, however, that such spray loading may not be as important as has been estimated from one-dimensional, eddy-diffusivity models (e.g., Lighthill 1999; Makin 2005; Kudryavtsev 2006; Ingel 2011; Bianco et al. 2011). When Shpund et al. introduced large eddies into their two-dimensional Lagrangian model, these eddies carried spray that was generated at the sea surface to higher levels in the marine boundary layer, thereby reducing the spray loading near the surface.

Still, we cannot rule out the possibility that, with increasing wind speed, spray loading may cause the actual drag relation to fall slightly below our result in Fig. 10. Nevertheless, wind–wave coupling appears to be the dominant mechanism causing the drag coefficient to roll off.

d. Effects of sea state

The idea that the roughness of the sea surface $z_0$ may depend on wave age $C_p/\theta_*$, where $C_p$ is the phase speed of the dominant waves, goes back at least to Kitaigorodskii and Volkov (1965). Many authors have investigated whether this and other sea-state parameters influence air–sea drag (e.g., Smith et al. 1992; Janssen 1997; Bourassa et al. 2001; Oost et al. 2002; Drennan et al. 2005). Jones and Toba (2001) dedicated an entire book to this subject of how to parameterize the drag; in that book, Toba et al. (2001) highlighted the fact that “there is still lively debate” on this issue of how sea state influences the air–sea drag.

Briefly, the question is whether an air–sea drag parameterization can be successful as a function of just wind speed or whether it must also include sea-state variables such as wave age, wave height, wave period, or swell height and direction. Although our drag parameterization for aerodynamically rough flow requires only $U_{10}$, we cannot conclude that sea state is unimportant. After all, our extrapolated drag relation for aerodynamically rough flow, (4.3), corroborates the model predictions of Moon et al. (2007) and Mueller and Veron (2009) (Figs. 9 and 10), which involved integrating the ocean wave spectrum. Because these models depend implicitly on wave age, wave height, and wave steepness through the wave spectrum, our results, in turn, imply that these parameters are important for air–sea coupling.

The SWADE dataset (Table 1), which includes observations of swell, also has implications for our parameterization. Remember that, in the SWADE set, Donelan et al. (1997) noted whether their measurements of surface stress were made over a wind sea or in the presence of swell that was across wind, counter to the wind, or following the wind. Figure 11 shows the full SWADE dataset, plotted as $u_*$ versus $U_{10}$, with these four regimes identified.

Although we add only 99 points to our analysis, Fig. 11 suggests how robust our $u_*$-versus-$U_{10}$ analysis is and how weakly swell affects the drag. For most of the $U_{10}$ range, the swell cases on average have $u_*$ values only 10%–15% higher than the wind sea cases. At the upper end of the $U_{N10}$ range, however, the swell and wind sea cases are essentially coincident (cf. Drennan et al. 2003). Only for the lowest $U_{N10}$ values—well below the aerodynamically rough regime ($U_{N10} < 9$ m s$^{-1}$)—do swell cases deviate significantly from the wind sea cases and from the central tendency of our analysis.

We hypothesize that, in moderate and high winds, the sea surface is so strongly forced by the wind that swell has a low-order effect on the surface stress. In light winds, on the other hand, the relationship between wind stress and swell becomes significant.

Still, the observed erratic behavior of the drag in light winds—expressed as the drag coefficient $C_{DN10}$, the roughness length $z_0$, or $u_*$ (Figs. 7, 8, and 11; e.g.,
5. Conclusions

Despite many measurements, the drag coefficient formulated as \( C_{DN10} = (u_*/U_{N10})^2 \) —a legacy from laboratory fluid mechanics—still has wide variability at low and moderate wind speeds. For hurricane-strength winds, it is uncertain by a factor of 3 (Fig. 10). We discussed several reasons why \( C_{DN10} \) is naturally prone to such variability. Here, we therefore evaluated an alternative formulation of the air–sea drag relation, following the suggestion by Foreman and Emeis (2010).

Using seven times as much data as Foreman and Emeis (2010) used, we confirm their main conclusion that the friction velocity at which the sea surface becomes aerodynamically rough increases linearly with \( U_{N10} \), the 10-m, neutral-stability wind speed. We find

\[
    u_* = 0.0583 U_{N10} - 0.243, \tag{5.1}
\]

Not only do our two independent datasets, comprising 7 and 10 individual datasets, respectively, follow this relation, but each individual set that includes data for which \( U_{N10} \geq 9 \text{ m s}^{-1} \) follows it. Such consistent behavior is never found in plots of \( C_{DN10} \).

The significant part of our analysis is that this new relation has a negative intercept. Consequently, the 10-m, neutral-stability drag coefficient rolls off and asymptotes to a constant in high winds:

\[
    C_{DN10} = \left( \frac{u_*}{U_{N10}} \right)^2 = 3.40 \times 10^{-3} \left( 1 - \frac{4.17}{U_{N10}} \right)^2. \tag{5.2}
\]

This behavior is exactly what hurricane modelers have been trying to justify and theorists have been trying to explain.

We suggest that wind–wave coupling explains (5.1). Theoretical models for the surface stress by Moon et al. (2007) and Mueller and Veron (2009) include terms for only the skin friction, form drag on the waves, and flow sheltering. Yet, these produce nearly straight-line relations between \( u_* \) and \( U_{N10} \) up to winds of major hurricane strength. (Neither group evidently realized this behavior.) Furthermore, both model predictions are very close to our (5.1). As a result, we conclude that known processes involving wind–wave coupling may be enough to explain the behavior of the air–sea drag for all wind speeds. These theoretical results also motivate our extrapolating (5.1) to hurricane-strength winds.

The literature contains data-based estimates that suggest \( C_{DN10} \) can be as low as \( 1.5 \times 10^{-3} \) in 50 m s\(^{-1}\) winds (e.g., Powell et al. 2003). We believe that this estimate is the smallest lower bound on the drag coefficient in hurricane-strength winds. On the other hand, one way to view our result (5.2) is as the greatest upper bound on the drag coefficient.

Processes that the models of Moon et al. (2007) and Mueller and Veron (2009) did not include—such as spray loading—may reduce the drag coefficient from what (5.2) predicts. We hypothesize, however, that any such effects will be second order, reducing \( C_{DN10} \) from the level in (5.2) by, perhaps, 10%. Because the sea surface is so strongly forced in high winds, we also hypothesize that swell will negligibly affect air–sea drag for wind speeds above \( \sim 15 \text{ m s}^{-1} \).

From the behavior of the roughness Reynolds number in our two datasets, we also estimated the wind speed and the friction velocity at which the sea surface becomes aerodynamically rough. Although the roughness Reynolds numbers are scattered, we have enough data to reliably determine mean behavior. We conclude that the sea surface becomes aerodynamically rough for \( U_{N10} \) between 8 and 10 m s\(^{-1}\); as an operational estimate, we use \( U_{N10} = 9 \text{ m s}^{-1} \) as the wind speed at transition. From (5.1), this wind speed gives \( u_* = 0.28 \text{ m s}^{-1} \) as the friction velocity when the sea surface becomes aerodynamically rough.

Although flux measurements at sea in light winds have larger uncertainty, our data suggest that \( u_* \) follows...
aerodynamically smooth scaling at low \( U_{N10} \), where the roughness length is presumed to obey \( z_0 = 0.135(v/u_*) \).

For \( U_{N10} < 9 \text{ m s}^{-1} \), this expression produces nearly straight-line behavior in plots of \( u_* \) versus \( U_{N10} \). We thus fitted this smooth region with a straight line, (4.2), to complement our straight-line result in rough flow, (5.1).

It was then natural to smoothly join these two straight lines with a hyperbola that constitutes a unified drag parameterization that encompasses weak-to-strong winds:

\[
 u_* = 0.239 + 0.0433\{(U_{N10} - 8.271) + [0.120(U_{N10} - 8.271)^2 + 0.181]^{1/2}\}. \tag{5.3}
\]

Here, \( u_* \) and \( U_{N10} \) are in meters per second.

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APPENDIX

Bulk Flux Algorithm

A motivation for our developing the \( u_* \) parameterization (4.3) was to use it in a bulk flux algorithm for high-wind, spray conditions. Sea spray enhances the air–sea exchange of sensible and latent heat; consequently, in such an algorithm, the total sensible \( (H_{s,T}) \) and latent \( (H_{L,T}) \) heat fluxes each include two terms, an interfacial part \( (H_s \) and \( H_L \)) and a spray-mediated part \( (Q_{s,sp} \) and \( Q_{L,sp} \)) (Andreas et al. 2008; Andreas 2011a). With the surface stress or momentum flux, \( \tau \), the bulk flux algorithm thus comprises three main equations:

\[
\tau = \rho_a c_p u_*^2, \tag{A1a}
\]

\[
H_{s,T} = H_s + Q_{s,sp}, \tag{A1b}
\]

\[
H_{L,T} = H_L + Q_{L,sp}. \tag{A1c}
\]

Equation (4.3) goes directly into (A1a) and provides \( \tau \) once \( U_{N10} \) is calculated from (2.1). Andreas et al. (2008) give the equations for computing \( Q_{s,sp} \) and \( Q_{L,sp} \) which also depend on \( u_* \). Our specifying \( u_* \) directly from \( U_{N10} \), however, allows us to deviate from standard methods for calculating the interfacial fluxes \( H_s \) and \( H_L \).

By definition, these are

\[
H_s = -\rho_a c_p u_* \theta_s, \tag{A2a}
\]

\[
H_L = -\rho_a L_v u_* q_s. \tag{A2b}
\]

Here \( c_p \) is the specific heat of air at constant pressure, \( L_v \) is the latent heat of vaporization, and \( \theta_s \) and \( q_s \) are temperature and specific humidity flux scales analogous to \( u_* \). These \( \theta_s \) and \( q_s \), in turn, appear in the Monin–Obukhov similarity expressions for the profiles of average potential temperature \( \Theta \) and specific humidity \( Q \) in the atmospheric surface layer (Garratt 1992, 52–54; Andreas et al. 2008):

\[
\Theta_r = \Theta_s + \frac{\theta_s}{k} \ln(r/z_T) - \psi_h(r/L), \tag{A3a}
\]

\[
Q_r = Q_s + \frac{q_s}{k} \ln(r/z_Q) - \psi_h(r/L). \tag{A3b}
\]

Here \( r \) is a reference height; \( \Theta_r \) and \( Q_r \) are thus the average potential temperature and specific humidity at height \( r \), \( \Theta_s \) and \( Q_s \) are the corresponding values at the sea surface. Also, \( L \) is the Obukhov length—a function of \( u_* \), \( \theta_s \), and \( q_s \). For the stability correction \( \psi_h \), we use the same sources that we did for the \( \psi_m \) function in (2.1): Paulson (1970) for unstable stratification, and Grachev et al. (2007a) for stable stratification. Andreas et al. (2008) explain how we estimate the roughness lengths \( z_T \) and \( z_Q \).

Rearranging (A3) and inserting them, respectively, into (A2) yields our algorithm for the interfacial heat fluxes:

\[
H_s = \frac{\rho_a c_p k u_* (\Theta_s - \Theta_r)}{\ln(r/z_T) - \psi_h(r/L)}, \tag{A4a}
\]

\[
H_L = \frac{\rho_a L_v k u_* (Q_s - Q_r)}{\ln(r/z_Q) - \psi_h(r/L)}. \tag{A4b}
\]

As usual, (A1a) and (A4) must be solved iteratively because these and \( U_{N10}, (2.1) \), are coupled through the Obukhov length.

Notice that this new algorithm requires no parameterization for \( z_0 \), which is one of the most variable and uncertain quantities in surface layer similarity theory.


