Predicting the Turbulent Air-Sea Surface Fluxes, Including Spray Effects, from Weak to Strong Winds

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LONG-TERM GOALS

The goal is to investigate, through theory and by analyzing existing data, sea surface physics and air-sea exchange in winds that range from weak to hurricane-strength. Ultimately, we want to develop unified parameterizations for the fluxes of momentum, sensible and latent heat, and enthalpy across the air-sea interface. These flux parameterizations will provide improved model coupling between the ocean and the atmosphere and, in essence, set the lower flux boundary conditions on atmospheric models and the upper flux boundary conditions on ocean models.

OBJECTIVES

1. Develop a theoretical framework for predicting air-sea fluxes from mean meteorological conditions and apply uniform analyses, based on this framework, to datasets that we will assemble.
2. Assemble a large collection of quality air-sea flux data that represents a wide variety of conditions.
3. Compute fluxes from these datasets using an improved analysis that better accommodates measurements made over heterogeneous surfaces, such as coastal zones. Focus the analyses on common problems where existing bulk formulations perform poorly—such as over surface heterogeneity, in weak winds, and in very strong winds.
4. Develop a unified algorithm for predicting the turbulent air-sea surface fluxes that spans the environmental range in our datasets, obeys theoretical principles and constraints, and substantially exceeds the correlation due to fictitious correlation.

APPROACH

This project is a collaboration between Ed Andreas and Larry Mahrt. In addition, Dean Vickers of Oregon State University is a contractor on this project. Mahrt has been focusing on boundary-layer processes in weak winds, when stratification and surface heterogeneity are important issues and when Monin-Obukhov similarity theory breaks down. He has recently started a study of near-surface wind maxima in the coastal zone. Andreas, in contrast, has been concentrating on high winds, when sea spray becomes an important agent for modifying the usual interfacial fluxes of momentum, heat, and moisture. Vickers brings expertise in processing large datasets—especially, aircraft data—and in parameterizing air-sea exchange. Together, we will develop flux parameterizations that span wind speeds from near zero to hurricane-strength.

In light of a new parameterization for the air-sea drag that we will describe soon, we have modified the usual equations for parameterizing the *interfacial* air-sea fluxes of momentum (τ, also called the surface stress), sensible heat (H_s), latent heat (H_L), and enthalpy (Q_{en}). In our formulation, these four equations now read (Andreas et al. 2012)

$$\tau \equiv \rho u_*^2 = \rho\left[f\left(U_{N10}\right)\right]^2, \tag{1a}$$

$$H_s = -\rho c_p u_* \theta_*, \tag{1b}$$

$$H_L = -\rho L_v u_* q_*, \tag{1c}$$

$$Q_{en} = -\rho u_*\left[c_p \theta_* + L_v q_*\right]. \tag{1d}$$

Here, ρ is the density of moist air; c_p, the specific heat of air at constant pressure; L_v, the latent heat of vaporization; and f, a function that predicts the friction velocity, u_*, from the 10-m, neutral-stability wind speed, U_{N10}. Equation (1a) defines the friction velocity.

The θ_* and q_* in (1b), (1c), and (1d) are flux scales that we compute through Monin-Obukhov similarity:

$$-\theta_* = \frac{k\left(\Theta_s - \Theta_r\right)}{\ln\left(r/z_T\right) - \psi_h\left(r/L\right)}, \tag{2a}$$

$$-q_* = \frac{k\left(Q_s - Q_r\right)}{\ln\left(r/z_Q\right) - \psi_h\left(r/L\right)}. \tag{2b}$$

Here, Θ_s and Q_s are the potential temperature and specific humidity at the surface of the ocean; Θ_r and Q_r, the temperature and humidity at reference height r; k, the von Kármán constant (= 0.40); z_T and z_Q, the roughness lengths for temperature and humidity, which we compute from the algorithm in Liu et al. (1979; also in Fairall et al. 1996); and ψ_h, an empirical stratification correction that depends on r/L, where L is the Obukhov length.

To account for spray effects, we formulate the *total* scalar fluxes as

$$H_{s,T} = H_s + Q_{s,sp}\ , \tag{3a}$$

$$H_{L,T} = H_L + Q_{L,sp}\ , \tag{3b}$$

$$Q_{en,T} = H_s + H_L + Q_{en,sp}\ . \tag{3c}$$

In these, subscript T denotes the total flux across the air-sea interface. The first term on the right [first two terms in (3c)] in each of these is the *interfacial* flux, parameterized as in (1) and (2); and the right-most term is the *spray-mediated* flux. Because the spray-mediated fluxes do not scale the same way that the interfacial fluxes do, this separation into spray and interfacial components is crucial in high winds (above about 12 m/s; Andreas 2011). In fact, Andreas (2011) recently demonstrated that, when spray-mediated transfer is in play, the total sensible heat flux can at times be countergradient, contrary to the down-gradient assumption that is common in modeling air-sea heat transfer.

To address these various ideas, we have assembled—and are still assembling—a large set of air-sea flux data. We currently have in hand 20 datasets comprising about 7000 air-sea flux measurements. In this set, surface-level wind speeds range from near zero to 72 m/s; and sea surface temperatures range from –1° to 32°C. This dataset thus covers almost all oceanic conditions. Flux parameterizations developed from these data should, indeed, be "unified."

WORK COMPLETED

Traditionally, the surface stress, (1a), is formulated in terms of a drag coefficient as

$$\tau = \rho C_{Dr} U_r^2 . \tag{4}$$

Here, U_r is the wind speed at reference height r, and C_{Dr} is the drag coefficient. From (1a) and (4), we see that C_{Dr} is evaluated as $(u_*/U_r)^2$. As such, C_{Dr} is prone to large random scatter because u_* is typically uncertain by at least ±10% and U_r is also uncertain. C_{Dr} is thus especially ill-posed in light winds and low stress, when uncertainties are generally even larger. Consequently, a consensus formulation for the drag coefficient is yet to emerge (e.g., Jones and Toba 2001).

Recently, Foreman and Emeis (2010) suggested an alternative drag relation that overcomes many of the shortcomings in a relation formulated in terms of C_{Dr}. Their idea was to parameterize u_* directly; and from a dataset of about 1000 points measured in aerodynamically rough flow, they obtained

$$u_* = a U_{N10} + b , \tag{5}$$

where a = 0.051, b = –0.14, and both u_* and U_{N10} are in m/s.

With the large dataset that we have assembled, we investigated Foreman and Emeis's (2010) suggestion (i.e., Andreas et al. 2012). Figure 1 shows our results. Actually, Andreas started this project with data in hand—our so-called "original" data. Mahrt brought to the project the "aircraft" data. We analyzed these original and aircraft sets separately and found them to yield statistically identical results. That is, we corroborated the results from the original dataset with the aircraft dataset or vice versa.

Our result, formulated as (5), for aerodynamically rough flow—10-wind speed above 9 m/s—is

$$u_* = 0.0583 U_{N10} - 0.243 , \tag{6}$$

where both u_* and U_{N10} are in m/s. Meanwhile, in aerodynamically smooth flow, the data in Figure 1 tend to follow aerodynamically smooth scaling for which the roughness length z_0 obeys

$$z_{0s} = 0.135 \frac{\nu}{u_*} , \tag{7}$$

where ν is the kinematic viscosity of air.

In a plot of u_* versus U_{N10}, (7) produces essentially a straight line. Hence, we joined the straight line in the aerodynamically rough regime, (6), and the straight line in smooth flow implied by (7) with a hyperbola. This hyperbola,

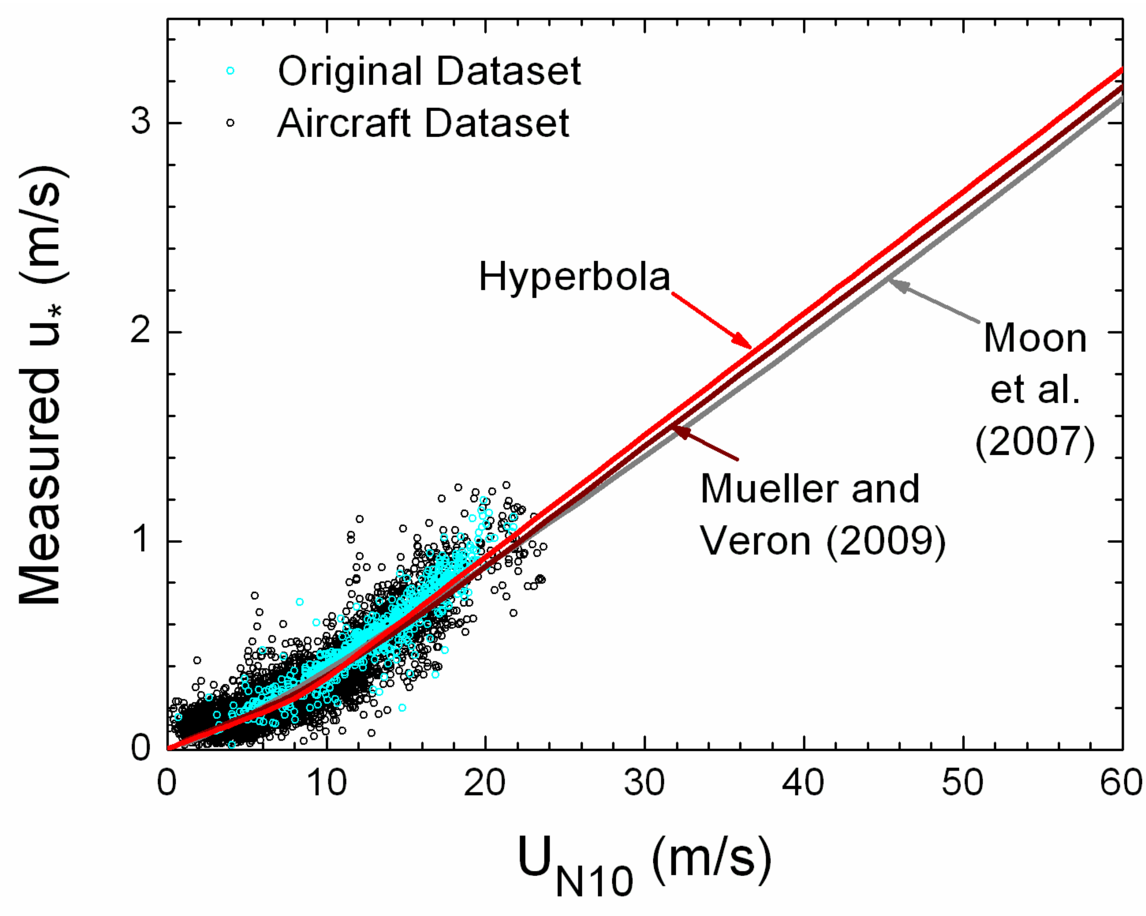

Figure 1. The "original" and "aircraft" datasets [see Andreas et al. (2012) for descriptions] are plotted as u_* versus U_{N10} for U_{N10} up to 60 m/s. In aerodynamically smooth flow—low U_{N10}—the data tend to follow the straight line implied by (7). In aerodynamically rough flow—U_{N10} above 9 m/s—the data follow (6). We thus smoothly joined these two lines with the hyperbola (8). The theoretical results from Moon et al. (2007) and Mueller and Veron (2009) are not much different from our main straight-line result (6) for U_{N10} above 20 m/s.

$$u_* = 0.239 + 0.0433\left\{\left(U_{N10} - 8.271\right) + \left[0.120\left(U_{N10} - 8.271\right)^2 + 0.181\right]^{1/2}\right\}, \qquad (8)$$

is thus a continuous, differentiable function that parameterizes the air-sea momentum flux (1a)—through u_*—from U_{N10} of 0 m/s up as far as we want to extrapolate.

In Figure 1 are two theoretical results that model the air-sea drag as a consequence of just wind-wave coupling. That is, Moon et al. (2007) and Mueller and Veron (2009) modeled the surface stress as a combination of the tangential stress or skin friction, form drag on the waves, and flow sheltering by the waves. Yet both models predict essentially the same straight-line behavior for winds above 20 m/s that our data-based extrapolation predicts. (Neither Moon et al. nor Mueller and Veron evidently realized that they were predicting u_* to be a linear function of U_{N10} in high winds.) In other words, exotic processes like sea spray loading (e.g., Barenblatt et al. 2005; Ingel 2011; Bianco et al. 2011) or the disintegration of the air-sea interface (Emanuel 2003; Soloviev and Lukas 2010) are not necessary to explain our observations.

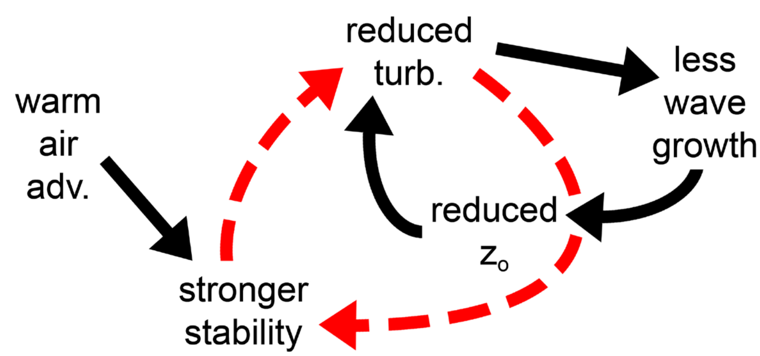

Figure 2. Warm air advection off a coast over cooler water leads to stable stratification in the near-surface air. This stratification inhibits the downward momentum flux from aloft such that the turbulence is reduced in the off-shore flow. In turn, wave growth and the associated surface roughness (z_0) are limited. Consequently, the stability increases further in a feedback loop that can ultimately decouple the atmosphere and the sea surface.

As a counterpoint to this work on high winds, we have also been looking at weaker winds and the resulting collapsed turbulence just above the air-sea interface. Warm air advecting from a rougher land surface over colder water leads to buoyancy destruction of turbulence and restricts the downward mixing of momentum from aloft. Consequently, the surface wave field and the associated surface roughness are reduced (Figure 2). The decreasing turbulence and decreasing surface roughness evolve together and can cause almost complete decoupling of the wind field from the sea surface.

As a result of the weak surface stress, the flow becomes almost free from the surface and forms a low-level wind maximum (a jet). This feedback mechanism is one of several processes that lead to a low-level wind maximum in the coastal zone.

Such wind maximum can be substantially sharper than nocturnal jets over land. A Long-EZ flight during the CBLAST weak wind experiment collected a large number of soundings. These wind speed profiles suggests that such wind maxima can be quite variable over a period of several hours (Figure 3). Here the horizontal domain was about 20 × 40 km. The profiles include very sharp wind maxima below 50 m and a second group of profiles with relatively sharp maxima above 150 m. A third group of profiles corresponds to more diffuse maxima.

We are investigating the role of spatial variation and possible instability of the sharp wind maxima. The profiles of potential temperature (not shown) indicate strong stratification in the lowest 50 m and weaker stratification above. Weak turbulence in the lowest 20–30 m defines a surface boundary layer.

In the layer containing the sharp wind maximum, the turbulence is extremely weak. At higher levels, the turbulence is still generally quite weak. This flight day (Figure 3) is the first of several case studies that we will conduct for compiling statistics of low-level wind maxima in the coastal zone.

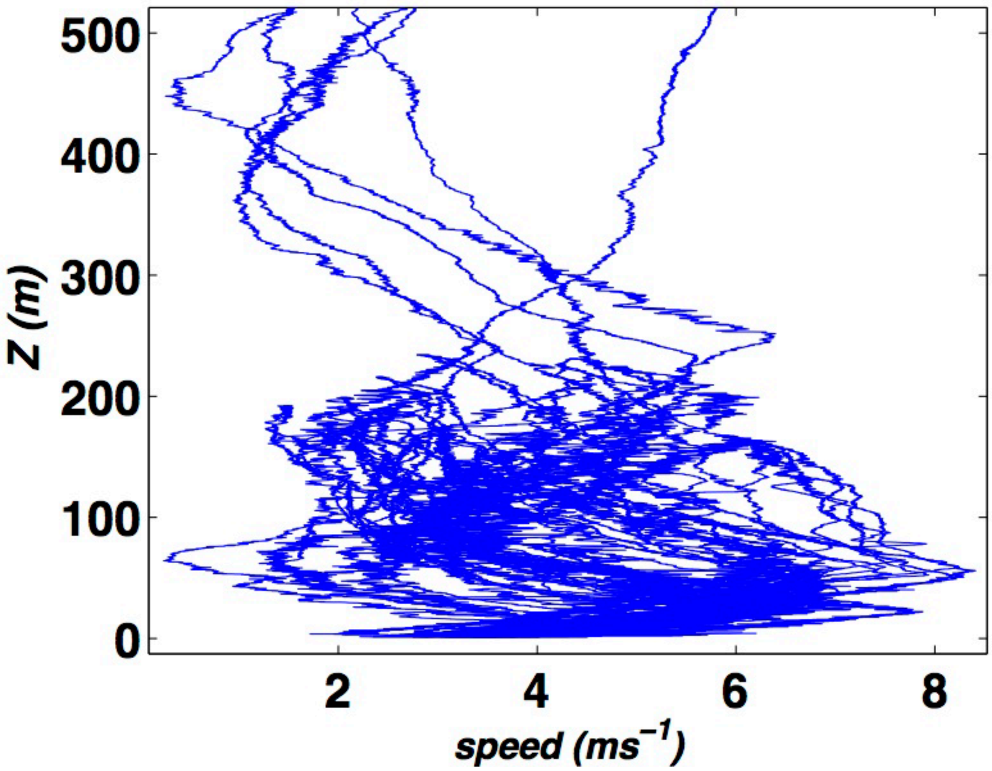

Figure 3. Fifty vertical profiles of wind speed that the Long-EZ aircraft measured between the sea surface and 500 m on 8 August 2001 during the CBLAST weak wind pilot experiment near Martha's Vineyard. The vast majority of the soundings include a wind maximum (jet) below 100 m. Sometimes the jet is quite sharp.

RESULTS

From (6), we can calculate the neutral-stability, 10-m drag coefficient, $C_{DN10} \equiv \left(u_* / U_{N10}\right)^2$, that is usually converted to C_{Dr} in (4). That is, from (6),

$$C_{DN10} = 3.40\times10^{-3}\left(1 - \frac{4.17}{U_{N10}}\right)^2, \tag{9}$$

where U_{N10} is still in m/s. Because of the negative intercept in (6), (9) predicts that C_{DN10} rises with increasing wind speed, rolls off, and asymptotes to 3.40×10^{-3}. Hurricane modelers have been trying to justify this behavior in C_{DN10} ever since Emanuel (1995) explained that models could not produce realistic hurricanes if the air-sea drag were parameterized as an extrapolation of, for example, the Garratt (1977) or Large and Pond (1981) result, which had C_{DN10} increasing with U_{N10} without bound.

Figure 4 shows our hyperbola, (8), recast as C_{DN10}. The figure also shows the Charnock-plus-smooth equation this is currently in vogue (e.g., Fairall et al. 1996, 2003; Andreas et al. 2008). It also shows the Moon et al. (2007) and Mueller and Veron (2009) model predictions for C_{DN10}. Lastly, the figure shows continuous functions that Sanford et al. (2007) and Chiang et al. (2011) formulated from the discrete results in Powell et al. (2003) and used in models of the ocean mixed layer under hurricanes.

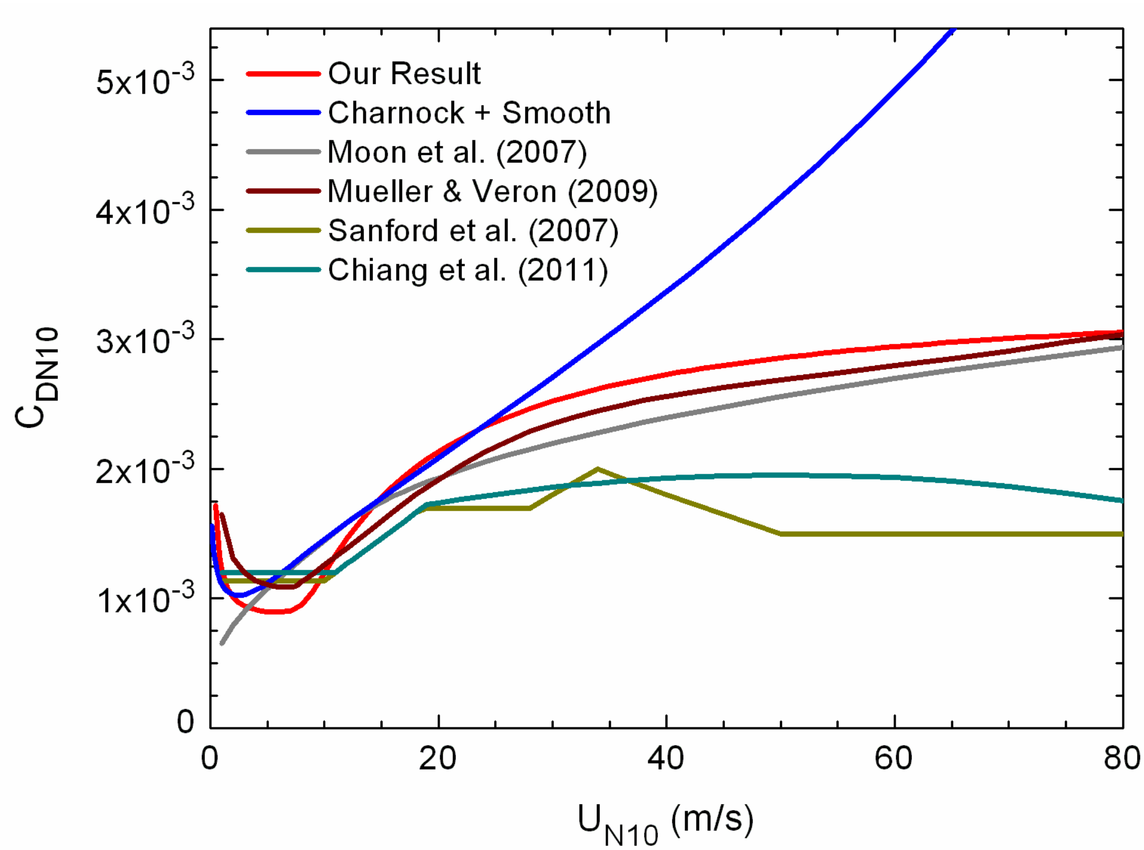

Figure 4. The neutral-stability, 10-m drag coefficient, C_{DN10}, is plotted against U_{N10} for U_{N10} between 0 and 80 m/s. Our hyperbola, (8), and the theoretical results of Moon et al. (2007) and Mueller and Veron (2009) agree well; all predict that C_{DN10} rises rapidly for wind speeds up to about 20 m/s, then rolls off to values of about 3.0×10^{-3}, and becomes nearly constant for U_{N10} above 100 m/s. The commonly used "Charnock + Smooth" relation, on the other hand, predicts that C_{DN10} rises without bound. The Sanford et al. (2007) and Chiang et al. (2011) curves, which are based on Powell et al. (2004), roll off to much lower values, of order 1.5×10^{-3} to 1.9×10^{-3}, and are probably too low in hurricane-strength winds.

Because we do not have much confidence in the Powell et al. (2003) results, the Sanford et al. (2007) and Chiang et al. (2011) lines in Figure 4 probably represent a smallest lower bound on C_{DN10} in hurricane-strength winds. The "Charnock + Smooth" curve, on the other hand, is what Emanuel (1995) was evaluating: its extrapolation would produce so much momentum loss to the sea surface that modeled hurricanes would not intensify. Our result and the Moon et al. (2007) and Mueller and Veron (2009) results, however, produce moderately valued drag coefficients that are in the range found appropriate for hurricane modeling (e.g., Tang and Emanuel 2012).

IMPACT/APPLICATIONS

Our inspection of the aircraft datasets has shown widespread problems: The measurements depend on aircraft heading with respect to the wind vector. We are continuing to work on this problem. Our analysis also indicates that current formulations of air-sea fluxes based on Monin-Obukhov similarity are not always well posed in weak winds, an effect that can mask important wave effects (Mahrt and Khelif 2010; Mahrt et al. 2012).

The behavior of the drag coefficient in tropical cyclones has been a crucial knowledge gap at least since Emanuel (1995) reported that hurricane models could not produce realistic storms if their drag parameterization was simply an extrapolation of drag relations obtained at lower wind speeds—like the Charnock relation in Figure 4. Equations (6) and (8) now provide a rational, data-based estimate for C_{DN10} in hurricane-strength winds. Moreover, our analysis explains why C_{DN10} must roll off with increasing wind speed: Known processes for wind-wave coupling seem to explain the roll off.

TRANSITIONS

Journal articles and conference presentations document our work on air-sea exchange. Andreas has also developed a software "kit" that contains instructions and the Fortran programs necessary to implement a bulk air-sea flux algorithm that includes the spray effects, as described by Andreas et al. (2008) and Andreas (2010). The current version of that code is 3.4, and the kit is posted at http://www.nwra.com/resumes/andreas/software.php, where it can be freely downloaded.

One of the goals of this project is to update that bulk flux algorithm using the newly available data that we have assembled. But at the Naval Research Laboratory in Monterey during the July 2012 workshop on the ONR DRI on Unified Parameterization for Extended Range Prediction, of which this project is a part, NRL modelers seemed interested in testing the new drag parameterization that Andreas described there. Hence, he quickly developed a revised bulk flux algorithm based on the drag relation (8) and equations (1) and (2) and provided it to several NRL modelers. That algorithm has no spray processes in it yet but will let modelers try out our new drag relation.

RELATED PROJECTS

Andreas is in the third and final year of a project funded under the National Ocean Partnership Program. This project is on "Advanced Coupled Atmosphere-Wave-Ocean Modeling for Improving Tropical Cyclone Prediction Models," with Isaac Ginis at the University of Rhode Island and Shuyi Chen at the University of Miami as lead PIs. Andreas is a subcontractor to the University of Rhode Island and has been supplying code and expertise to help the project understand surface momentum and heat exchange in hurricane-strength winds—especially spray-mediated exchange.

Andreas and Kathy Jones of the Army Cold Regions Research and Engineering Laboratory just started in FY12 a project to study spray icing that is funded under the new ONR Arctic program. We plan two field experiments during the three-year project and will measure sea spray size distributions and the associated meteorological conditions from off-shore platforms or at other well-exposed marine sites where we can expect high winds. In particular, we will make eddy-covariance measurements of the momentum and sensible and latent heat fluxes and thus will add to the inventory of flux datasets that we have already assembled under the current project.

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