# ETL Sea Spray parameterizations

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## 1 Background

Present parameterizations of air-sea fluxes are reasonably valid up to wind speeds of about 25 m/s. Extrapolation of these parameterizations to higher wind speeds are inconsistent with theoretical analyses of the strength of tropical cyclones by Emmanuel. One major issue is the relative balance of momentum and scalar (heat/moisture) transfers. Emmanuel's model considerations require the ratio of enthalpy to momentum exchange to be near 1.0 while algorithms based on the data presently available extrapolate to much lower ratios. It is speculated that this balance is affected by evaporation of sea spray droplets at high wind speeds (u>25 m/s). This is illustrated in the Fig. 1, which shows the known behavior of the turbulent transfer coefficients and that which is required to explain observed cyclone intensities.

At high wind speeds, the ocean is a major source of droplets produced by bursting bubbles and spume (i.e., the shearing off of wave tops) to the lower troposphere. Droplets may play a large role in latent heat transfer between the ocean and, under extremely high winds such as found in hurricanes, may also have a large effect on the air-sea exchange of momentum. However, the relative importance of droplets in air-sea interaction at high wind speeds is largely unknown, due in large part to the difficulty in measuring droplet concentrations at high wind speeds. If droplet concentrations were available, existing models of the atmospheric boundary layer (ABL) incorporating droplet dynamics could be employed to understand droplet-mediated fluxes.

The fundamental parameter required for representing the effect of sea spray on air-sea exchange processes is the size dependent *source function* for droplets, or number of droplets of a given size produced at the sea surface per unit surface area per unit time as a function of wind speed. Because the source function cannot be measured directly at present, it must be estimated from the height-dependent number-size distribution of droplets, $n(r, z)$ (i.e., the number of droplets of given radius per unit volume of air per increment of radius at height $z$) and a model for the atmospheric boundary layer that incorporates droplet dynamics. However, progress in determining the source function has been frustrated due to the difficulty of measuring $n(r, z)$. The present droplet parameterizations are based on droplet concentrations determined on a beach, 10 hours of data at a wind speed of 21 m/s from the HEXOS program, and inferences from various laboratory studies. The data from Smith et al. go to 30 m/s but, besides not being representative of the open ocean, contain no measurements of the larger size (r>20 μm) droplets believed to be important to this problem. .

Once the important properties of the source function are characterized, then it is still quite complicated to determine the effect of the sea spray on the fluxes of heat and moisture. This becomes a balance between the rate at which droplets of various sizes are thrown into the atmosphere, how quickly they respond to the thermal/moisture environment, and how they modify that environment by cooling and evaporating. There have been two approaches to analyzing/modeling these processes. One is a scaling model approach that considers the thermal and evaporation time response of the spectrum of droplets versus their suspension lifetime (essentially the ejection height divided by mean fall velocity). The second approach is to use explicit numerical into which the spectrum of droplets is ejected and spectral budget equations are used to compute their vertical diffusion, evaporation, and modification of the local temperature/humidity profiles. Explicit models can be either Lagrangian or Eulerian in nature. Scaling models lead directly to parameterizations while explicit models can, in principle, be imbedded in the host model of interest.

From a parameterization point of view we again draw a distinction between 'resolvable' and 'subgridscale' modifications to the environment by the evaporating droplets. Subgridscale modification by the droplets of their own evaporation environment is called the *feedback* problem. It originated in early scaling models that used MO similarity to relate the surface turbulent fluxes and flux-profile relationships to describe fraction of droplet mass lost to evaporation before the droplet re-impacted the ocean. Because the profiles are cooled and moistened in the droplet evaporation layer, these early formulae overestimated the thermal effect of the droplets on the PBL.

To summarize, parameterization of sea spray effects involves 1) droplet source strength as a function of wind speed (or wave processes), 2) the characteristic height of the droplet sources, 3) vertical diffusion of droplets, 4) droplet evaporation microphysics, 4) feedback effects.

## 2 Scaling Approach to Thermodynamic Effects

The ETL sea spray parameterization has been developed to account for all of these processes. To begin, we write the four principal thermodynamic flux components of interest (Fairall et al., 1994)

$$H_s' = \rho_a c_{pa} C_H U (T_o - T_a)$$

$$H_l' = \rho_a L_e C_E U (q_s(T_o) - q_a)$$

$$Q_s' = \rho_w c_{pw} F_v (T_o - T_w)$$

$$Q_l' = \rho_w L_e F_E$$

Here $H_s$ and $H_l$ are the direct turbulent sensible and latent heat fluxes and the Q terms are the equivalent fluxes associated with the heat carried by the droplets into the atmosphere and the water vapor deposited by evaporation of the droplets in the air. The primes denote fluxes without feedback effects; $T_o$ is the ocean temperature, $T_a$ the air temperature, $q_s$ the saturation specific humidity of seawater, $q_a$ the atmospheric specific humidity, and $T_w$ the wet bulb temperature for a seawater droplet. The Q terms are computed by specifying the droplet source strength, $S_n(r)$, as a function of droplet size, r, and making arguments about the relative sizes of the evaporation response time ($\tau_r$) and the suspension time ($\tau_f$=h/$v_f$) where h is the effective droplet source height (one-half the significant wave height) and $v_f$ is the size-dependent mean gravitational fall velocity of the droplet.

$$F_v = 4\pi/3\int r^3 S_n(r)dr$$

$$F_E = 4\pi/3\int \frac{3\tau_f}{\tau_r} r^3 S_n(r)dr$$

The time constants are specified analytically following Fairall et al., (1990) and Andreas (1992).

## 3 Source Function

These integrals require a specification of $S_n$. Following Fairall et al. (1994), we used a form characteristic of spume droplets (sea spray blown off the tops of breaking waves at height h) that has a fixed shape as a function of droplet size, $S_{no}(r)$, plus a wind speed dependence

$$S_n = f(U)S_{no}(r)$$

The shape of $S_{no}$ is shown in Fig. 2. The wind speed dependence was originally specified as that for the fractional area of whitecap coverage

$$f(U) = W_b = 3.8\times 10^{-6}U^{3.4}$$

The latest version of the ETL sea spray parameterization has modified this wind speed dependence and changed the form of $S_n$ slightly based on parameterization of a physically-based model that was developed in terms of energy lost to the wave breaking process. The actual shape of the source is not independent of wind speed in the physically-based model, but we have incorporated that into an additional windspeed dependence of the integrals.

$$F_v = 2.5\times 10^{-6}u_*^2 W_b$$

$$F_E = \frac{0.08W_b}{1 + 7\exp(-0.04U^{4/3})} h\beta(T_o)[q_s(T_a) - q_a] = G(U)h\beta(T_o)[q_s(T_a) - q_a]$$

Note we have changed from $T_w$ to $T_a$ in this relation by using the wet bulb formulae

$$\beta = (1 + \frac{L_e \gamma q_s}{c_{pa}})^{-1}$$

$$\gamma = \frac{1}{q_s}\frac{\partial q_s}{\partial T}$$

$$T_a - T_w = \frac{1-\beta}{\gamma}(1 - s)$$

where s is the saturation ratio (relative humidity in % divided by 100).

## 4 Feedback Effects

The flux equations above describe the fluxes in the presence of droplets but in the absence of feedback effects (i.e., the droplets are not distorting the profiles of temperature and humidity below h). However, if droplet fluxes are significant compared to the direct turbulent fluxes, then it is likely that they are affecting the profiles. Thus, we need a method to account for feedback. To do this, we define the feedback as a perturbation in the temperature and humidity below h; $T_a$ is decreased and $q_a$ is increased. Note that $T_a$ and $q_a$ are the *resolved* reference values of temperature and humidity - the changes occur nearer the surface within the droplet production layer. Rather than use $q_a$, we express the change an increase in dew point temperature, $T_d$, where $q_a = q_s(T_d) = s\, q_s(T_a)$. Thus, in the presence of feedback, the fluxes are

$$H_s = \rho_a c_{pa} C_H U(T_o - T_a - \delta T_a)$$

$$H_l = \rho_a L_e C_E U[q_s(T_o) - q_s(T_d + \delta T_d)]$$

$$Q_s = \rho_w c_{pw} F_v (T_o - T_w)$$

$$Q_l = \rho_w L_e G(U) h \beta(T_o)[q_s(T_a - \delta T_a) - q_s(T_d + \delta T_d)]$$

Note that evaporation of droplets changes $T_a$ and $T_d$ but does not change $T_w$.

We parameterize the feedback effects by relating $\delta T_a$ to a feedback coefficient that is computed as the ratio of droplet evaporation without feedback to the heat available to evaporate droplets. The idea is that it takes heat to evaporate the droplets so they will consume heat by cooling and moistening the droplet layer; this will reduce the amount of evaporation until an energy balance is reached. To compute this coefficient, we must account for all sources of heat available:

$$feed = \frac{Q_l'}{H_{sm} + Q_s' + Q_l' + H_{s\varepsilon}}$$

where $H_{sm}$ is the sensible heat flux with $T_a$ set to $T_w$ and $H_{s\epsilon}$ is the heat generated in the droplet evaporation layer by the dissipation of turbulent kinetic energy, $\epsilon$,

$$H_{s\varepsilon} = 0.5 \rho_a u_*^3 / \kappa[\ln(h/10) + \kappa U / u_*]$$

The feedback coefficient is applied by computing the changes in $T_a$ and $T_d$

$$\delta T_a = \frac{1-\beta}{\beta} \delta T_d = feed * (T_a - T_w) = feed * \frac{1-\beta}{\gamma}(1-s)$$

This definition leads to $T_a = T_d = T_w$ (i.e., no evaporation) when feed=1.

## 5 Example

The parameterization is illustrated in the next two figures for a typical tropical cyclone boundary layer with water temperature of 29 C and a relative humidity of 80%. The feedback coefficient as a function of wind speed is shown in Fig. 3. At 50 m/s only about 30% of the evaporation computed without feedback is realized. The fluxes for this case are shown in Fig. 4. The direct turbulent fluxes increase roughly linearly with wind speed ( the x and O symbols). The dots are the dissipation heating. The blue line is the heat carried directly by the droplets ($Q_s$) and the redline the evaporation without feedback. The magenta triangles are the droplet evaporation with feedback. For this situation, the droplets enhanced the total enthalpy flux by about 50% at U=50 m/s, which is probably not enough to satisfy the Emmanuel constraint. However, we consider the representation of droplet mass flux to be uncertain by at least a factor of 3.

## 6 Matlab Programs

A matlab version of the physical model and the parameterized version of the physical model can be found at

ftp://ftp.etl.noaa.gov/et7/users/cfairall/onr_droplet/parameterization/

The programs on this site include:

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| drop_source_2 | Physically-based model |
| spray_param | Parameterized scaling version of physical model |
| test_spray | A driver that runs spray-param for specified conditons |
| spry_mass | A program that runs drop_source_2 |
| qsat | Saturation specific humidity function |
| wf2 | Droplet gravitational fall velocity function |
| drop2.pdf | Description of the physically-based model |

## 7 References

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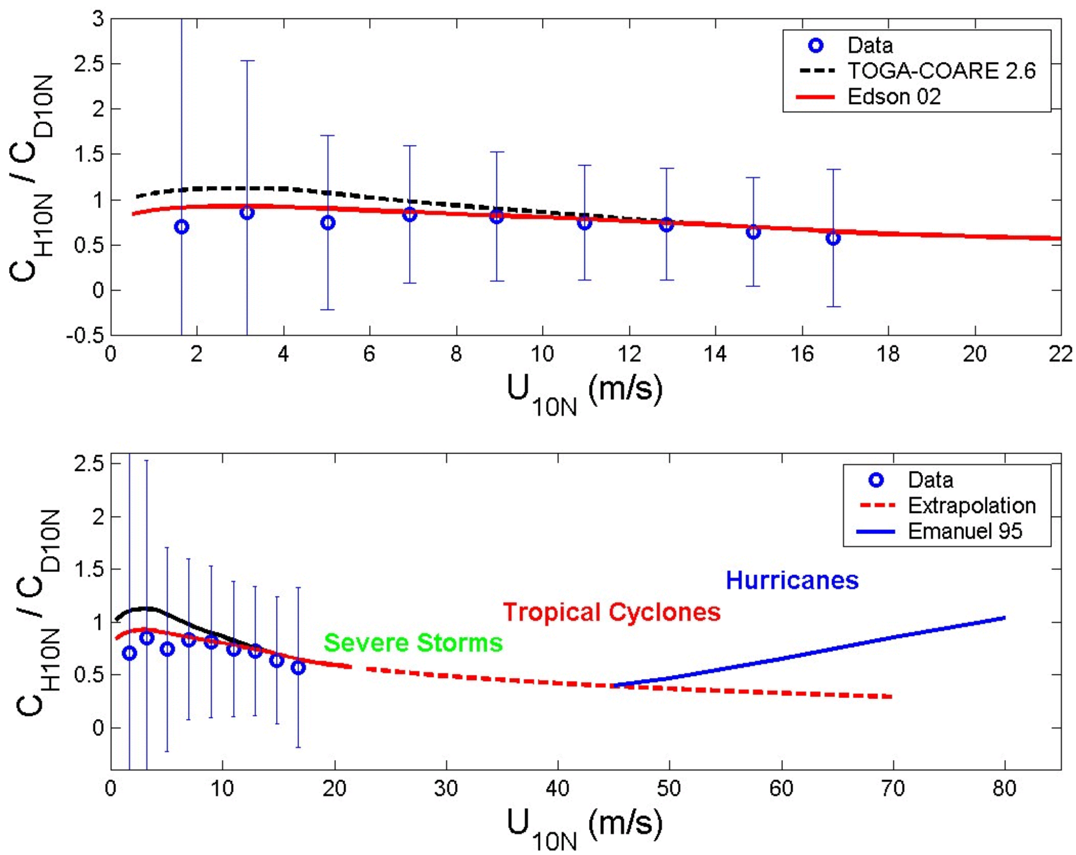


Figure 1. Ratio of enthalpy to momentum flux transfer coefficients. Upper panel: examples of data with comparisons to the COARE algorithm. Lower panel: as above, except showing extrapolations to wind speeds relevant to hurricanes.

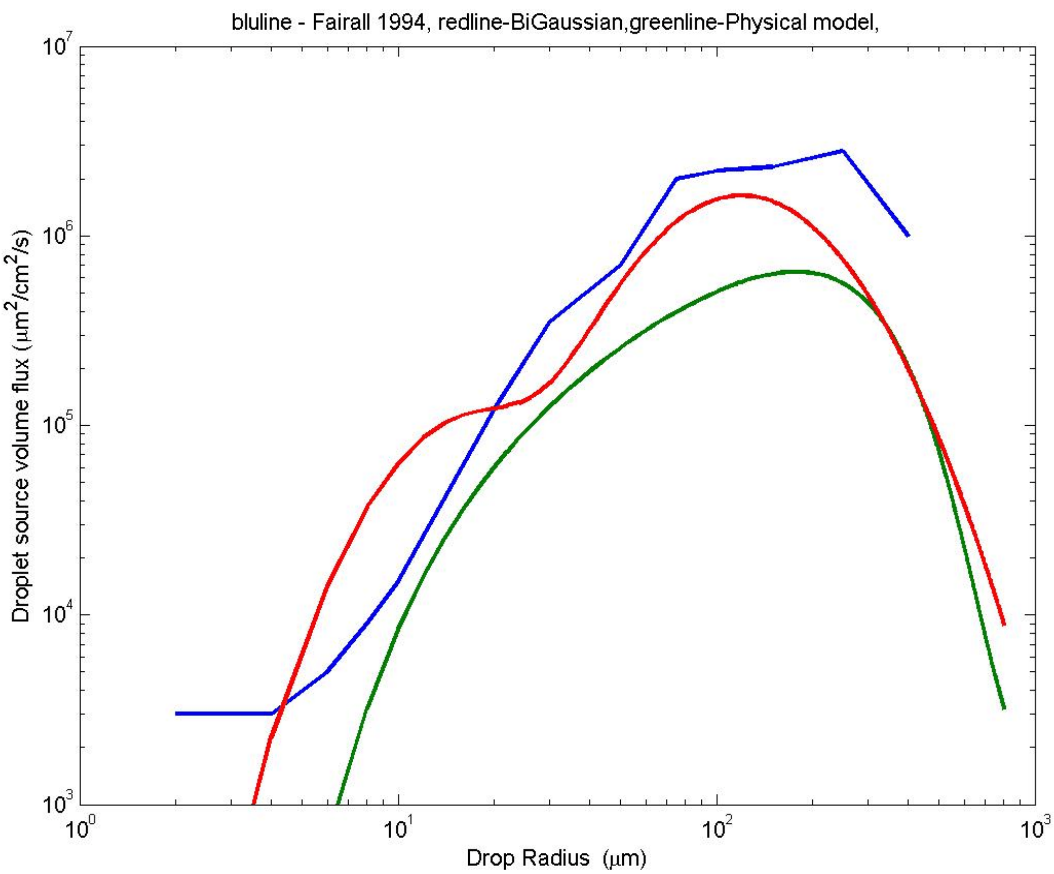


Figure 2. Normalized droplet source functions from different estimates. The Fairall et al. (1994) model - blue line; a two-mode log-normal estimate, red line; the physical model at U=30 m/s divided by the whitecap fraction at that wind speed.

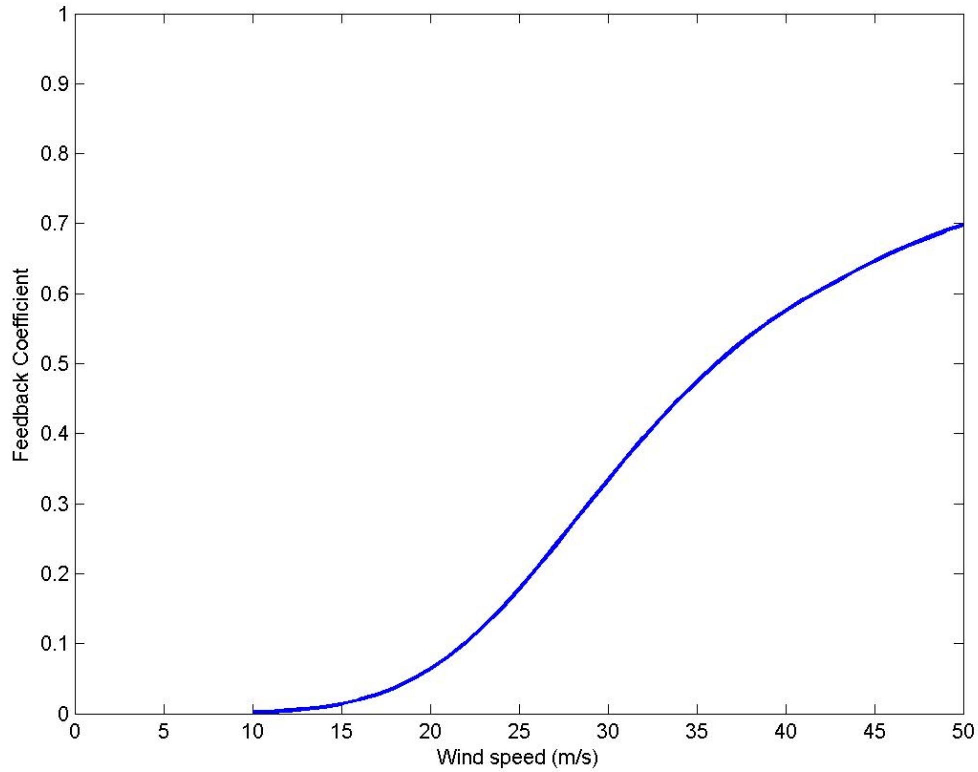


Figure 3. Feedback coefficient as a function of wind speed for a boundary layer with 80% RH.

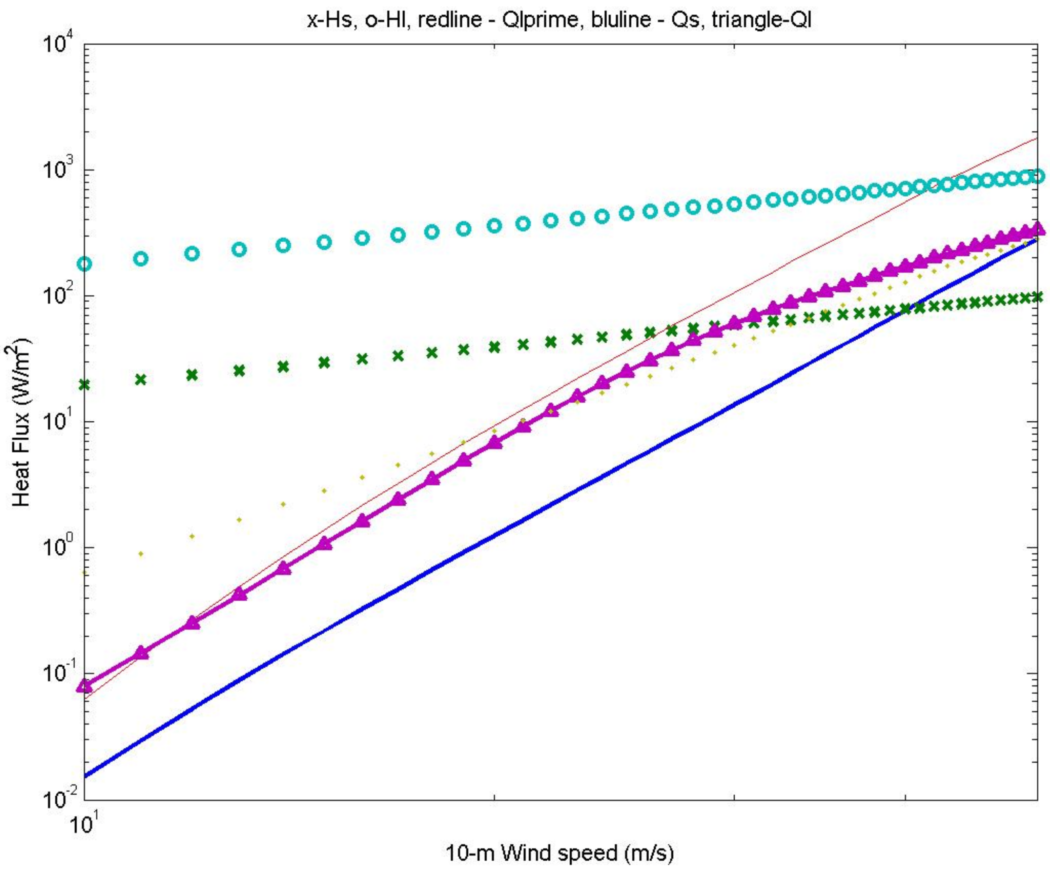


Figure 4. Heat fluxes as a function of wind speed for a boundary layer with RH=80%: $H_s$, x's; $H_l$, circles; $Q_s$, blue line; $Q_l$', weird color line; $Q_l$, magenta triangles; $H_{se}$, dots.