Effects of Sea Spray on Tropical Cyclone Intensity

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ABSTRACT

The intensity of tropical cyclones is sensitive to the rates at which enthalpy and momentum are transferred between sea and air in the high-wind core of the storm. Present models of the wind dependence of these transfer rates suggest that the effective drag coefficient is more than twice the effective enthalpy transfer coefficient at wind speeds above 25 m s^{-1}. Using this ratio in numerical models, however, makes it impossible to sustain storms of greater than marginal hurricane intensity. Some other physical process must, therefore, enhance enthalpy transfer at very high wind speeds. This paper suggests that *re-entrant* sea spray explains this enhanced transfer. When a spray droplet is ejected from the ocean, it remains airborne long enough to cool to a temperature below the local air temperature but not long enough to evaporate an appreciable fraction of its mass. The spray droplet thus gives up sensible heat and returns to the sea before it has time to extract back from the atmosphere the heat necessary to continue its evaporation. Microphysical modeling, combined with data from the Humidity Exchange over the Sea Experiment (HEXOS), makes it possible to derive an expression for the net enthalpy transfer of re-entrant spray. This spray enthalpy flux is roughly cubic in wind speed. When this relation is used in a numerical simulation of a hurricane, the spray more than compensates for the observed increase in the ratio of drag and enthalpy transfer coefficients with wind speed. The momentum flux associated with sea spray is an important energy sink that moderates the effects of this spray enthalpy flux. Including a parameterization for this momentum sink along with wave drag and spray enthalpy transfer in the hurricane simulation produces results that are similar to ones based on equal transfer coefficients.

1. Introduction

Basic theory (e.g., Emanuel 1986) and numerical experiments (Ooyama 1969; Rosenthal 1971; Emanuel 1995) show that the intensity of tropical cyclones depends strongly on the coefficients for the transfers of momentum (C_D) and enthalpy (C_k) between the ocean and the atmospheric boundary layer. The maximum wind speed, in particular, depends on $(C_k/C_D)^{1/2}$ in the high-wind-speed core of the storm (Emanuel 1986). Unfortunately, there are no simultaneous measurements of the effective values of these coefficients at wind speeds greater than about 25 m s^{-1}, and the theory of air–sea interaction at very high wind speeds is poorly developed. The agitated sea no doubt increases the effective roughness length and, thereby, C_D and the dissipation rate of kinetic energy; while, for wind speeds up to about 20 m s^{-1}, there is little observational evidence to suggest a corresponding increase in C_k (Geernaert et al. 1987; DeCosmo et al. 1996). Emanuel (1995) showed that, if estimated values of the exchange coefficients at 20 m s^{-1} applied at higher wind speeds, maintaining a storm of much greater than marginal hurricane intensity would be impossible. Some mechanism must also serve to enhance air–sea enthalpy exchange at high wind speed.

One candidate for enhancing the sea–air enthalpy flux at high wind speeds is sea spray. Riehl (1954, p. 287) was perhaps the first to suggest that sea spray supplies a significant amount of heat for generating and maintaining tropical storms. Laboratory studies (e.g., Mestayer and Lefauconnier 1988), numerical spray droplet models (e.g., Rouault et al. 1991; Edson et al. 1996; Van Eijk et al. 2001), and open-ocean observations (Korolev et al. 1990) all show that sea spray can redistribute enthalpy between the temperature and humidity fields in the marine boundary layer.

Fairall et al. (1994) were the first to incorporate a reasonable spray parameterization into a larger-scale cyclone model. While drawing no conclusions about whether spray had any effect on the intensity of their modeled storm, they did conclude that, without spray (or another mechanism for exchanging latent heat, such

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as rain), their model did not develop a realistic cyclone boundary layer. Kepert et al. (1999) continued this work and concluded that, although spray had little effect on the net air–sea enthalpy flux, it could increase storm intensity and affect development indirectly, by altering the stratification of the boundary layer, for instance. Lighthill et al. (1994) argued, however, that by reducing the surface layer temperature, evaporating sea spray would actually weaken tropical cyclones.

Because the net enthalpy flux across the air–sea interface, rather than the individual sensible and latent heat fluxes, controls a hurricane's intensity, Emanuel (1995) concluded that spray cannot affect the *net* enthalpy transfer. The marine boundary layer must supply all the latent heat required to evaporate the spray droplets. As we shall show here, however, Emanuel's (1995) conclusions hold only if all the sea spray evaporates. If some of the spray falls back into the sea—that is, is *reentrant*—an appreciable sea–air enthalpy flux results.

Three consecutive presentations at the American Meteorological Society's 23d Conference on Hurricanes and Tropical Meteorology further highlighted the diversity of opinions on spray's role in hurricane thermodynamics. In a preliminary version of this work, Andreas and Emanuel (1999) led off by showing that using an empirically based parameterization for the air–sea spray enthalpy flux in Emanuel's (1995) hurricane model produced a dramatic increase in storm intensity. Wang et al. (1999) followed with a more elaborate numerical model that incorporated spray through a parameterization based on Fairall et al. (1994). Wang et al. concluded that, although the presence of spray affected the structure of the boundary layer in their model and the rate at which their modeled cyclone intensified (cf. Kepert et al. 1999), it did not affect the storm's final intensity. Third in line was Uang (1999), who incorporated spray using the Fairall et al. parameterization in a hurricane model similar to Emanuel's (1995). Like Wang et al., Uang concluded that the spray had little effect on the final intensity of his modeled cyclones. Unlike Andreas and Emanuel (1999), however, these latter two presentations did not consider the effect of re-entrant spray.

At the 24th Hurricanes Conference, we reiterated our conclusion that spray is important for transferring enthalpy *and momentum* between the air and sea in high winds (Andreas and Emanuel 2000). Again, Wang et al. (2000) followed us immediately and showed that their modeled tropical cyclone evolved the same with or without spray when they used the Fairall et al. (1994) spray parameterization. They also showed, however, that when they used a spray parameterization based on Andreas and DeCosmo's (1999) spray model, their modeled cyclone became unrealistically intense. But they had misrepresented the Andreas and DeCosmo model such that their modeled air–sea enthalpy flux was too large. Naturally, their modeled storm became too intense. The irony in all these comparisons is that the Fairall et al. spray parameterization relies heavily on Andreas's (1992) spray model, which is also the precursor of the Andreas and DeCosmo model and our current spray parameterization (Andreas and Emanuel 1999, 2000).

Recently, Bao et al. (2000) reported a numerical simulation of a developing hurricane in the Gulf of Mexico. They included the effects of spray in their simulation using the Fairall et al. (1994) algorithm but also tested our ideas about re-entrant spray (Andreas and Emanuel 1999). They found that, if all the spray evaporates in the marine boundary layer, the spray has a negligible effect on storm intensity, as Emanuel (1995) suggested. But if some of the spray falls back into the sea, there is a marked and realistic increase in the storm intensity, as we contend (Andreas and Emanuel 1999, 2000).

Here we continue our study of the air–sea enthalpy flux when the sea surface is producing copious spray. In section 2, we show theoretically how spray *can* produce a net enthalpy flux from the ocean to the atmosphere. A small part of this net flux arises from the fact that the sea is generally warmer than the overlying air, so spray transfers some sensible heat directly to the atmosphere. A much larger enthalpy flux occurs when some of the spray falls back into the sea; these droplets have cooled below the air temperature but return to the sea before they have had time to extract from the atmosphere the sensible heat necessary to evaporate. This re-entrant spray thus cools the ocean and represents a net enthalpy flux to the atmosphere.

Using the spray model that Andreas (1992) developed and that Andreas and DeCosmo (1999) tuned with eddy-correlation measurements of the sensible and latent heat fluxes during the Humidity Exchange over the Sea experiment (HEXOS), we estimate the magnitude of spray's role in air–sea heat exchange in section 3. From the HEXOS observations, which include flux measurements in 10-m winds up to 18 m s^{-1}, we have shown that the spray contribution can, at times, exceed 100 W m^{-2} out of a combined surface sensible and latent heat flux (i.e., the net enthalpy flux) of 400 W m^{-2} (Andreas and Emanuel 1999). At much higher wind speeds, the spray contribution to enthalpy exchange will be an even larger fraction of the total because the spray enthalpy flux increases as the cube of the friction velocity.

In addition to providing an enhanced enthalpy flux, sea spray also extracts momentum from the wind (section 4). As spray droplets are ripped off the wave crest and injected into the airstream, they accelerate horizontally and, thus, represent a drag on the airflow. To our knowledge, we are the first to include spray momentum in a tropical cyclone simulation.

Finally, in section 5, we put all these ideas together in a series of tropical cyclone simulations using Emanuel's (1995) simple hurricane model. Incorporating realistic wave drag with a Charnock parameterization for surface roughness slows the wind and does not let the storm reach its full potential. Adding the spray enthalpy flux, however, provides the energy for a very intense hurricane. The spray momentum exchange can also be

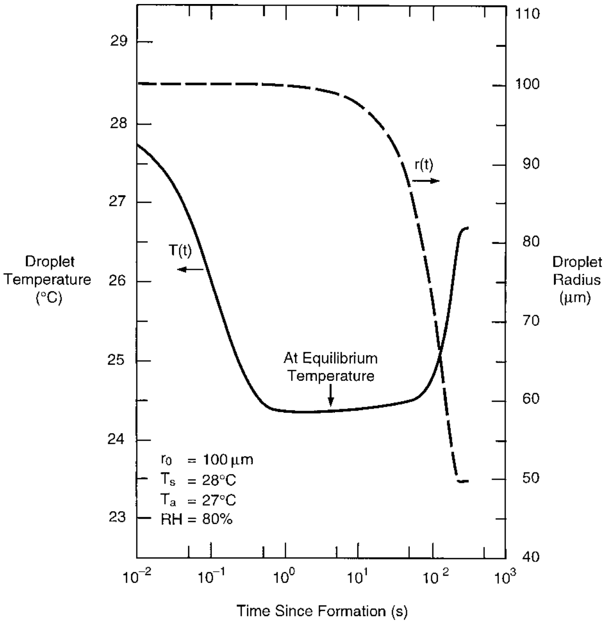

FIG. 1. The temperature (T) and radius (r) evolution of a spray droplet of initial radius 100 μm that is ejected from the sea surface (at temperature $T_s = 28$°C) into air of 27°C and 80% relative humidity. Also, the droplet has initial salinity 34 psu, and the barometric pressure is 1000 mb.

substantial and offsets some of the effects of the enhanced spray enthalpy flux. Combining the spray and wave-drag parameterizations produces a simulation with a cyclone intensity similar to our "control" run, which uses Emanuel's (1995) standard C_D and C_k values that are equal and increase linearly with the wind. Hence, spray-mediated exchange is one possible explanation why this simple parameterization produces realistic hurricanes (e.g., see Emanuel 1999).

2. The thermodynamics of spray

Microphysical modeling (e.g., Andreas 1990, 1992) has provided many insights into the processes that control air–sea heat and moisture transfer over a wind-agitated sea. The sea spray droplets that form from whitecaps and spume (e.g., Monahan et al. 1986) in such conditions, and that are important in transferring heat and moisture, range in radius from 1 to 500 μm. For this entire size range, however, the transfers of sensible and latent heat that these droplets facilitate are decoupled; that is, the sensible heat exchange that they mediate occurs three orders of magnitude more rapidly than the latent heat exchange.

Figure 1 demonstrates this for a 100-μm-radius spray droplet ejected into typical tropical cyclone conditions. In the figure, the droplet falls from 28°C, the sea surface temperature, to its so-called equilibrium temperature within 1 s. Only about 1% of the mass must evaporate for the droplet to reach this equilibrium temperature, as we will demonstrate shortly. Notice, too, that this equilibrium temperature is well below the ambient air temperature T_a of 27°C. Very little evaporation occurs, however, until at least 10 s after formation; and the droplet does not reach its equilibrium radius until at least 100 s after its formation. The final point to notice in Fig. 1 is that this equilibrium radius is not zero. Because the droplet is saline, it retains some liquid as long as the relative humidity is 75% or higher (Twomey 1953; Pruppacher and Klett 1978, Table 4.3). Below that humidity, it will crystallize into a salt particle.

Andreas (1995) showed plots like Fig. 1 for other conditions, and Andreas (1990, 1992) explained that the three-order-of-magnitude difference in timescales for temperature and radius evolution that is apparent for the 100-μm droplet in Fig. 1 also manifests in all spray droplets with radii between 1 and 500 μm. In other words, the sensible and latent heat exchanges from spray droplets are decoupled. The ambient humidity has negligible effect on the temperature timescale; and the sea surface temperature T_s (also the initial droplet temperature) has no effect on the radius timescale because the droplet is at its equilibrium temperature T_{eq} during most of its evaporation.

These essential facts lead us to believe that spray droplets with relatively short atmospheric residence times—that is, re-entrant spray—can accomplish a net sea-to-air enthalpy transfer. Our scenario relies on the rapidity with which droplets reach thermal equilibrium coupled with the much slower rate at which they evaporate. In other words, droplets readily give up heat associated with the falling temperature trace in Fig. 1, but many fall back into the sea before extracting from the atmosphere the latent heat to evaporate as represented by the radius trace in Fig. 1. The recent microphysical modeling by Van Eijk et al. (2001) corroborates this scenario that most spray droplets reach their equilibrium temperature but fail to reach their equilibrium radius before falling back into the sea.

Most others who have considered this problem have discounted the sensible heat carried by the spray and have focused on the latent heat transfer because this is, potentially, a much larger flux (e.g., Fairall et al. 1990; Hasse 1992; Makin 1998). But as Hasse (1992) and Emanuel (1995) pointed out, to evaporate completely a spray droplet must extract as much sensible heat from the near-surface droplet evaporation layer (Andreas et al. 1995) as it gives up in latent heat. That is, if a spray droplet were to evaporate entirely, its evaporation would reflect no net sea–air enthalpy transfer, only a conversion between the latent and sensible heat contributions to enthalpy.

We demonstrate the role of the sensible heat carried by the spray droplets in accomplishing a net sea–air enthalpy exchange in Fig. 2. Consider a system consisting of a near-surface layer of the ocean initially at temperature T_s underneath a near-surface air layer initially at temperature T_a. The near-surface ocean has initial mass m_0; the near-surface air initially contains no

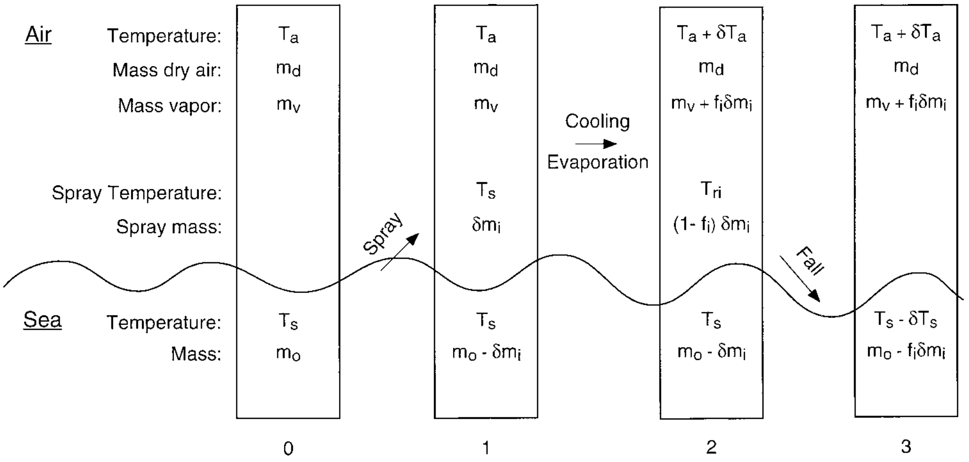

FIG. 2. A schematic diagram of the thermodynamic steps that occur in a control volume enclosing the near-surface ocean and the near-surface atmosphere when the ocean ejects spray, the spray cools and evaporates, and then falls back into the ocean.

spray and has dry air mass m_d and water vapor mass m_v (Fig. 2, panel 0). In panel 1, we let a small amount of ocean mass δm_i be converted to spray droplets of radius r_i. This step adds mass with temperature T_s to the air and diminishes the ocean mass by δm_i. Actually, spray droplets have a wide range of sizes, as we mentioned. Subscript "*i*" here means that we consider only a small, arbitrary radius bin for this demonstration.

In panel 2 of Fig. 2, we let the spray droplets cool to T_{ri} and let a fraction f_i of the original spray mass evaporate. We assume that T_{ri} here is the droplet equilibrium temperature (see Fig. 1) because all droplets up to 500 μm in radius reach their equilibrium temperature within 5 s, which is generally shorter than their residence time when the winds are above 20 m s^{-1} (Andreas et al. 1995). The cooling and evaporation do not change the properties of the ocean, but they do change the air temperature by δT_a. The evaporation also adds vapor in the amount $f_i \delta m_i$ to the air. Since the droplets have given up sensible heat to the air during their cooling but have extracted some sensible heat to evaporate, we cannot be sure of the sign of δT_a. Does the spray warm or cool the air? Most believe that spray cools the air; but the advantage of this thought experiment is that we do not have to know.

Finally, in panel 3, all the spray has fallen back into the sea. The air remains at temperature $T_a + \delta T_a$ and with vapor mass $m_v + f_i \delta m_i$; but this step clearly cools the ocean by δT_s because the re-entrant spray is at T_{ri}, which is less than both T_a and T_s (see Fig. 1). Because enthalpy must be conserved, by cooling the ocean, the spray has obviously accomplished a net enthalpy transfer from the sea to the air. In other words, by comparing the state of the air in panels 0 and 2 in Fig. 2, we cannot tell what net effect the spray has had. Such focus on the near-surface air has been the source of most of the confusion over spray's role in air–sea exchange. But when we compare the state of the ocean in panels 0 and 3, spray's role is clear. The spray has cooled the ocean and, thus, must have accomplished a net enthalpy transfer from the sea to the air. (Note, however, that this cooling of the ocean does not constitute a large sink term in the *ocean's* thermal budget, especially when compared to other effects such as vertical mixing through the base of the oceanic mixed layer.)

One necessary clarification for panel 3 is that the smallest spray droplets do not fall back into the sea. These droplets have very long residence times and thus would contribute little, if any, to the sea–air enthalpy exchange. But Fig. 2 accurately depicts the fate of droplets larger, nominally, than 10 μm, which carry most of the spray sensible and latent heat (e.g., Andreas 1992; Andreas et al. 1995; Van Eijk et al. 2001).

We can formalize the scenario depicted in Fig. 2 by evaluating the total system enthalpy. In panel 0, the system enthalpy is

$$K_0 = (c_{pd} m_d + c_w m_v) T_a + L_v m_v + c_w m_0 T_s, \quad (1)$$

where c_{pd} and c_w are the heat capacities of dry air and liquid water, and L_v is the latent heat of vaporization. The first two terms on the right in (1) represent the total enthalpy of the air, while the third term is the enthalpy of the ocean water.

In panel 2, the system enthalpy is

$$\begin{aligned} K_2 = {} & [c_{pd} m_d + c_w(m_v + f_i \delta m_i)](T_a + \delta T_a) \\ & + (L_v + \delta T_v)(m_v + f_i \delta m_i) + c_w \delta m_i (1 - f_i) T_{ri} \\ & + c_w (m_o - \delta m_i) T_s, \end{aligned} \quad (2)$$

where δL_v is the change in the latent heat of vaporization resulting from the change in air temperature.

Since the total enthalpy of the system must be conserved, K_0 and K_2 must be equal. Equating these gives

$$\begin{aligned} & [c_{pd} m_d + c_{pv}(m_v + f_i \delta m_i)] \delta T_a + L_v f_i \delta m_i \\ & \quad = c_w \delta m_i [(T_s - T_a) + (1 - f_i)(T_a - T_{ri})], \end{aligned} \quad (3)$$

where c_{pv} is the heat capacity at constant pressure of water vapor, and we have used $\delta L_v = (c_{pv} - c_w)\delta T_a$. Our convention throughout this development has been that δT_a, δm_i, and δL_v all have their probable signs: $\delta T_a < 0$, $\delta m_i > 0$, and $\delta L_v < 0$.

The left side of (3) is the change in enthalpy of the air between panels 0 and 2 in Fig. 2; the right side shows the processes that accomplish this change. The enthalpy of the air increases because of the sensible heat that all droplets give up in falling from their initial temperature T_s down to the air temperature T_a. The enthalpy of the air also increases because of the sensible heat lost when the droplet's temperature falls from temperature T_a to T_{ri}.

In the final panel in Fig. 2, the remaining spray falls back into the ocean. This step changes neither the temperature nor the specific humidity of the air and thus does not affect the air's specific enthalpy. The ocean does cool in this process, however, reflecting the net enthalpy transfer to the atmosphere.

We can draw several important conclusions from (3). If the entire droplet evaporates ($f_i = 1$), the net enthalpy gain comes only from bringing the spray from ocean temperature to ambient air temperature. Freshwater droplets may evaporate entirely, but larger seawater droplets rarely do. Therefore, even if the ocean and atmosphere have the same temperature, there can still be a net enthalpy transfer since $f_i < 1$ is the usual case for seawater droplets. In general, if there is no evaporation, the droplet's temperature will be somewhere between T_s and T_a, and there will be a small net enthalpy transfer.

But the major effect comes when evaporation reduces the droplet temperature to T_{ri}. Consider a droplet with initial mass m_i and temperature T_s. The droplet cools from T_s to the air temperature primarily by losing sensible heat. To cool from T_a to T_{ri}, however, it must evaporate some of its water. We ask how much droplet mass must evaporate for the droplet to cool from T_a to T_{ri}. The heat balance of this cooling is simply

$$m_i c_w (T_a - T_{\mathrm{eq}i}) = L_v (m_i - m), \tag{4}$$

where m is the droplet's mass when its temperature reaches $T_{\mathrm{eq}i}$. Clearly,

$$\frac{m}{m_i} = 1 - \frac{c_w}{L_v}(T_a - T_{\mathrm{eq}i}). \tag{5}$$

Therefore, if $T_a - T_{\mathrm{eq}i} \approx 3°\mathrm{C}$, for example (cf. Fig. 1), $m/m_i \approx 0.995$. That is, less than 1% of the droplet's initial mass must evaporate for it to cool well below T_a. Notice how rapid this process is. According to Fig. 1, a droplet with an initial radius of 100 μm reaches $T_{\mathrm{eq}i}$ in less than a second. Larger droplets take somewhat longer; smaller droplets evolve even more quickly (see also Andreas 1990; Andreas and DeCosmo 1999).

The exact temperature that the droplet attains depends on its initial size, salt content, ambient temperature and humidity, and length of time in the air; but the equilibration timescale is typically very short compared to the average time a droplet spends airborne (e.g., Andreas 1992). Thus, to a first approximation, we may consider T_{ri} to be the wet-bulb temperature of a saline, spherical droplet of radius r_i. The largest heat transfer will occur when the fraction of droplet that evaporates, f_i, is just large enough to bring the droplet temperature down close to T_{ri}. Again, this fraction is of order 1%.

3. The net thermodynamic contribution of spray

A tropical cyclone responds to the net enthalpy transfer across the air–sea interface rather than to the individual fluxes of sensible and latent heat. Our hypothesis here is that re-entrant sea spray contributes some of that enthalpy flux. In the context of Andreas's (1992) spray model, Andreas and DeCosmo (1999) write the total air–sea enthalpy flux as

$$Q_{e,T} = Q_{e,\mathrm{int}} + Q_{e,\mathrm{sp}}. \tag{6}$$

Here, $Q_{e,\mathrm{int}}$ is the usual interfacial enthalpy flux parameterized with the transfer coefficient C_k, and $Q_{e,\mathrm{sp}}$ is the spray enthalpy flux.

In light of our discussion in the last section, we formulate $Q_{e,\mathrm{sp}}$ as a sensible heat flux driven by the temperature difference between the sea surface and the spray droplets at their thermal equilibrium. Andreas (1992) and Andreas et al. (1995), however, show that spray droplets with an initial radius of about 100 μm carry most of the spray sensible heat. Therefore, rather than considering droplets of all sizes (each with a unique T_{eq}), to make the parameterization simple, we hypothesize that these 100-μm droplets are the bellwethers of the spray enthalpy transfer and thus parameterize $Q_{e,\mathrm{sp}}$ as

$$Q_{e,\mathrm{sp}} = \rho_w c_w (T_s - T_{\mathrm{eq},100}) V(u_*). \tag{7}$$

Here, ρ_w is the density of seawater, $V(u_*)$ is a function of the friction velocity u_*, and $T_{\mathrm{eq},100}$ is the equilibrium temperature of droplets that are initially 100 μm in radius. In actual practice, though, we approximate this equilibrium temperature with the salinity-modified wet-bulb temperature. (For u_* in m s^{-1} and for the other variables on the right-hand side of (7) in mks units, $Q_{e,\mathrm{sp}}$ is in W m^{-2}.)

Andreas and DeCosmo (1999) extracted spray sensible and latent heat fluxes from DeCosmo's (1991) HEXOS data using Andreas's (1992) spray generation function to predict spray production. Andreas and Emanuel (1999) compared the spray enthalpy flux $Q_{e,\mathrm{sp}}$ implied by this partitioning with the HEXOS measurements of the total enthalpy flux $Q_{e,T}$. Figure 3 likewise shows that spray enthalpy flux but parameterized here as (7) suggests.

The rate at which the wind does work on the sea surface to produce spray droplets has long been presumed to go approximately as the cube of the wind speed or the cube of u_* (e.g., Wu 1979; Monahan et al. 1986; Monahan 1988; Andreas et al. 1995; Andreas 1998a).

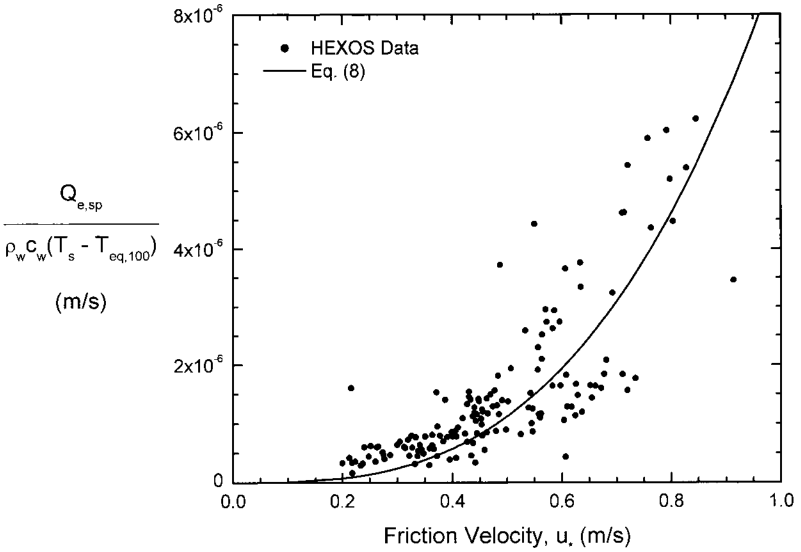

FIG. 3. The spray enthalpy flux that Andreas and DeCosmo (1999) inferred from the HEXOS data is plotted as suggested in (7) to deduce the wind function $V(u_*)$, which is given by (8).

We fitted the line in Fig. 3 to the HEXOS data by doing a least squares fit for an assumed wind dependence of u_*^3. The result is

$$V(u_*) = 9.0 \times 10^{-6} u_*^3 \tag{8}$$

(V in m s^{-1} for u_* in m s^{-1}).

As an example of the magnitude of the spray enthalpy flux, suppose the water temperature is 28°C, the air temperature is 27°C, and the relative humidity is 80% (cf. Fig. 1). The equilibrium temperature of a 100-μm spray droplet with initial salinity 34 psu is then 24.36°C (e.g., Andreas 1995, 1996; Kepert 1996). For a 10-m wind speed of 20 m s^{-1}, we use Large and Pond's (1981) results to estimate that the 10-m drag coefficient is 1.79 $\times$ 10^{-3}; u_* is consequently 0.84 m s^{-1}. Equations (7) and (8) then give $Q_{e,\mathrm{sp}} = 79$ W m^{-2}. This is already a large number. To the extent that we can extrapolate (7) and (8) to cyclone-strength winds, the spray enthalpy flux could be very large indeed. We suggest that these spray effects may be the source of the anomalous surface heat fluxes that tropical cyclones require for their generation and maintenance (Emanuel 1995; Smith 1997).

4. The spray momentum flux

When spray is ejected into the airstream, air drag accelerates it. This exchange extracts momentum from the airflow. Likewise, when spray droplets crash back into the sea, they transfer their momentum to the ocean. For moderate winds, however, Wu (1973) found this spray-mediated momentum exchange to be negligible. Fairall et al. (1994) likewise found the effect negligible for winds up to 50 m s^{-1}. Pielke and Lee (1991), on the other hand, suggest that the spray drag is significant. In their model, it reduces the near-surface wind speed by 15% at a wind speed of 40 m s^{-1} and, consequently, would increase the surface stress by a few percent. We revisit this issue of spray momentum here.

By integrating the equation for the horizontal force balance on a spray droplet, we have calculated how long a droplet of radius r_0 takes to accelerate from zero to within e^{-1} of the local wind speed. We find that, in surface-level winds of 10 m s^{-1} and higher, droplets with radii up to 500 μm reach this speed within 1 s. In turn, using Andreas's (1992) estimate of a droplet's residence time, we further see that all droplets up to 500 μm in radius are airborne long enough to accelerate to the local wind speed U.

We can next use this information to estimate the rate at which the spray extracts momentum from the airflow. Formally, that spray momentum flux is simply

$$\tau_{\mathrm{sp}} = \frac{4\pi}{3} \rho_w u \int_{r_{\mathrm{lo}}}^{r_{\mathrm{hi}}} r_0^3 \frac{dF}{dr_0}\, dr_0, \tag{9}$$

where dF/dr_0 is the so-called spray generation function, the number of droplets of initial radius r_0 produced per square meter of sea surface per second per micrometer increment in r_0. Consequently, in (9), $(4\pi\rho_w r_0^3/3)dF/dr_0$ is the mass flux of spray droplets with radius r_0. The limits of integration, r_{lo} and r_{hi}, are nominally 1 and 500 μm.

For U in (9), we use the wind speed at one significant wave height above mean sea level. That is, (9) assumes that all droplets reach this level. This is a good assumption because the droplets that contribute most to (9) are the large ones torn right off the wave crests.

Equation (9) is also an upper bound on the surface stress that falling spray imparts to the ocean. It is only an upper bound, though, because some of the smaller

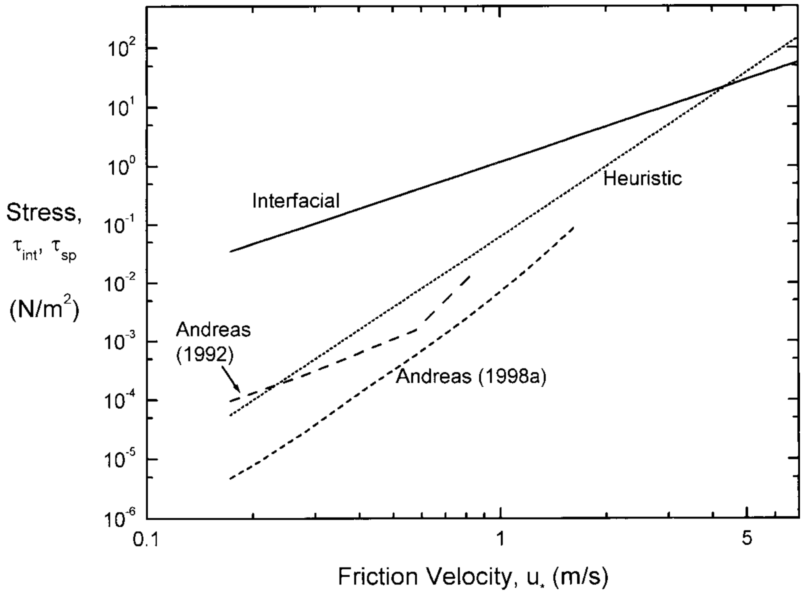

FIG. 4. Estimates of the interfacial and spray stresses. The domain of the spray stresses calculated using (9) and the Andreas (1992) and Andreas (1998a) spray generation functions reflects the wind speed limits over which these functions are defined. The heuristic line is Eq. (11).

spray droplets remain suspended indefinitely and thus do not fall back into the sea. Some of the droplets also leave water vapor behind in the atmosphere and therefore do not return to the sea with radius r_0. Nevertheless, the essential point is that this transfer of momentum from air to sea through the spray is immaterial to the dynamics of the cyclone; the cyclone responds only to the spray-mediated momentum exchange *in the air*, as estimated from (9).

Using (9), we have evaluated τ_{sp} for the Andreas (1992) and Andreas (1998a) spray generation functions. Figure 4 shows the results of these calculations. For comparison, the figure also shows the interfacial surface stress, which is just

$$\tau_{int} = \rho_a u_*^2, \tag{10}$$

where ρ_a is the air density.

Figure 4 suggests that, for moderate winds, the spray stress implied by the Andreas (1992) and Andreas (1998a) spray generation functions is two to three orders of magnitude less than the interfacial stress, as Wu (1973) concluded. But the estimated spray stress increases roughly as the fourth power of u_* because dF/dr_0 is approximately cubic in u_* (Andreas 1998a) and U is linear in u_*. Meanwhile, τ_{int} is quadratic in u_*. We estimate that, when the winds reach hurricane strength, the interfacial and spray stresses become comparable.

The two estimates of spray stress in Fig. 4 set a lower bound on the actual spray stress because they model only droplets with initial radii between 1 and 500 μm. Spume droplets can be much larger than this. And though these very large droplets contribute negligibly to the enthalpy transfer, they should enhance the spray momentum transfer beyond what we predict with the Andreas (1992) and Andreas (1998a) spray generation functions.

Therefore, the line labeled ''Heuristic'' in Fig. 4 depicts our current parameterization for the spray stress,

$$\tau_{sp} = 6.2 \times 10^{-2} u_*^4 \tag{11}$$

(τ_{sp} in N m^{-2} for u_* in m s^{-1}). As the figure shows, the interfacial stress and this spray stress are equal at about $u_* = 4.2$ m s^{-1}; this u_* corresponds to a surface-level wind speed of about 60 m s^{-1}. For higher wind speeds, the spray stress actually exceeds the interfacial stress.

5. Tests using a simple model

As a preliminary test of the new spray formulations, we incorporate them in the simple, balanced axisymmetric model of Emanuel (1995). This model is phrased in potential radius coordinates and thus yields extremely high horizontal resolution in the eyewall region, where the vorticity is large. It also uses a representation of convection based on the subcloud layer entropy equilibrium postulate. The interfacial fluxes of enthalpy and momentum are calculated using standard bulk-aerodynamic formulas but with coefficients that depend linearly on wind speed (but are always equal to each other), as we explain in the appendix. The environmental conditions used in all of these simulations are those of the standard control run in Emanuel (1995); that is, the sea surface temperature is 27°C, the mean outflow temper-

TABLE 1. Summary of the four simulations depicted in Fig. 5.

Simulation	Interfacial stress	Interfacial enthalpy	Spray enthalpy	Spray stress
Control	$C_D = C_k$ and both increase linearly with wind speed			
Wave drag only	Use Eqs. (A2), (A5), and (A6)	Use Eqs. (A8), (A9), and (A10)		
Spray enthalpy and momentum	$C_D = C_k$ and both increase linearly with wind speed		Use Eqs. (7) and (8)	Use Eq. (11)
All effects	Use Eqs. (A2), (A5), and (A6)	Use Eqs. (A8), (A9), and (A10)	Use Eqs. (7) and (8)	Use Eq. (11)

ature is −70°C, and the mean ambient surface-level relative humidity is 80%.

To model the spray enthalpy and momentum fluxes, we add (7), (8), and (11) to the model's standard bulk-flux formulation and assume that $T_{\mathrm{eq},100}$ is the wet-bulb temperature (modified for salinity) of the well-mixed boundary layer air at a height of 10 m. We also incorporate in the model the effects of wave drag on the interfacial momentum and enthalpy fluxes as described in the appendix. In all other respects, the model is identical to that used in Emanuel (1995). Note that Emanuel's model requires only the net enthalpy flux from the sea surface; temperature and moisture are not carried as separate variables. Since moist enthalpy is the sole thermodynamic variable of the model, we need not keep track of the air–sea temperature difference and the moisture gradient separately.

We run four hurricane simulations: the standard (i.e., control) simulation reported in Emanuel (1995), which uses bulk exchange coefficients C_k and C_D that depend linearly on wind speed but are equal to each other at all wind speeds; a simulation using the Charnock relation (A6) and the associated scalar roughness length (A10) to represent wave drag but with no spray effects; a simulation using (7), (8), and (11) for the spray enthalpy and momentum fluxes but with no wave drag; and a simulation using wave drag, (A6) and (A10), together with (7), (8), and (11) for the spray enthalpy and momentum fluxes. Table 1 summarizes the details of these simulations; Fig. 5 shows our results.

It is clear from Fig. 5 that the wave drag effect, uncompensated by any enhanced surface enthalpy flux, leads to a substantial diminution of storm intensity. In fact, the model cannot generate very intense hurricanes even in very favorable environments. While it is unreasonable to suppose that the sea state is in equilibrium in a hurricane—as assumed in the Charnock formula—actual sea states might exert even more drag because waves are steeper and have lower average phase speeds before they reach equilibrium.

When we include the spray enthalpy and momentum effects in the third simulation in Fig. 5, storm intensity increases significantly because of the enhanced enthalpy exchange associated with the re-entrant sea spray. We

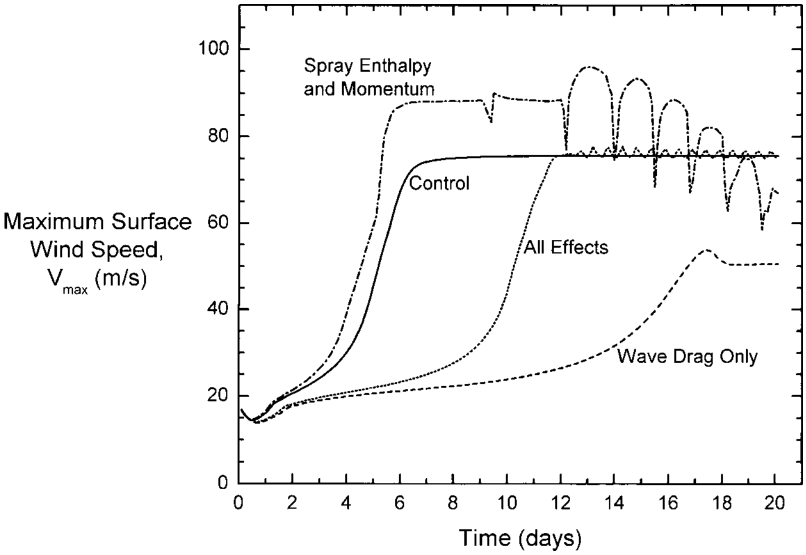

FIG. 5. Evolution with time of the maximum surface wind speed, $V_{\max}$, in four integrations of Emanuel's (1995) tropical cyclone model. The "Control" run uses identical exchange coefficients for enthalpy and momentum. The "Wave Drag Only" run simulates wave effects on the drag coefficient using the Charnock relation, (A6), and a comparable parameterization for the scalar roughness, (A10). The "Spray Enthalpy and Momentum" run models spray enthalpy and momentum fluxes using (7), (8), and (11) but includes no wave drag. The run with "All Effects" includes both the spray and wave parameterizations. See Table 1.

are not really surprised that this spray simulation produces such an intense storm. As Table 1 shows, this simulation simply adds an enthalpy flux mediated by the spray to the control run, which already has enough enthalpy transfer to produce a realistic storm. The spray momentum flux does moderate the effects of the enhanced enthalpy flux in this simulation, though.

The oscillatory pattern in the spray simulation results from eyewall replacement cycles, during which the hurricane eyewall collapses and reforms (e.g., Bister and Emanuel 1998). Eyewall replacement cycles tend to occur in this model when either the ambient middle troposphere is exceptionally moist or the surface enthalpy flux is exceptionally strong (Emanuel 1995). We could eliminate the eyewall cycles here, without affecting the average intensity of the modeled storm, by making the middle troposphere drier.

While our formulations of the spray fluxes, (7), (8), and (11), must be regarded as highly provisional, it is clear that, based on our current understanding, sea spray can have a significant effect on storm intensity. When we include both wave drag and spray enthalpy and momentum exchange in the fourth simulation in Fig. 5 (i.e., ‘‘All Effects’’), the model reaches an intensity similar to that of the control run. That is, in this case, the wave drag and sea spray effects nearly cancel. This perhaps explains why models with very simple formulations of air–sea exchange produce simulated storms of reasonable intensity (Emanuel 1999).

6. Summary

It is clear from both theory and numerical simulations that the observed intensity of tropical cyclones cannot be explained by air–sea interaction physics that accounts only for increased wave drag without an attendant process that enhances the sea–air enthalpy flux. We here propose that the ‘‘missing’’ process is re-entrant sea spray. Spray droplets injected into the air thermally equilibrate after only about 1% of their mass has evaporated. When they fall back into the sea before losing much more of their mass, they give up to the atmosphere sensible heat before they have had time to extract a higher price in latent heat. The HEXOS measurements support a cubic dependence of the spray enthalpy flux on friction velocity. Extrapolating these results to hurricane-force winds in a numerical model shows that the spray-induced enthalpy flux can be very significant—more than enough to offset the increased drag caused by the spray and potentially enough to offset the increased wave drag as well.

We are evidently the first to incorporate this spray drag into a tropical cyclone model. Though previous estimates have suggested that spray droplets have negligible influence on air–sea momentum exchange for moderate winds, our analysis of spray generation rates suggests that the spray drag and the interfacial drag become comparable in hurricane-strength winds. Quite simply, the interfacial drag increases quadratically with the friction velocity, while the spray drag appears to increase as the fourth power. We estimate that the two momentum transfer mechanisms are thus equal for a surface-level wind speed of about 60 m s^{-1}.

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APPENDIX

Gradient-Wind Transfer Laws

The wind speed variable in Emanuel’s (1995) tropical cyclone model is the gradient wind. Consequently, the standard version of that model parameterizes the interfacial fluxes of momentum and enthalpy in terms of transfer coefficients C_D and C_k, respectively, that apply to the gradient wind. In our control run here, C_D and C_k increase linearly with the wind speed but are always equal.

For other runs mentioned in the text that involve spray and wave effects, however, we formulate the spray and interfacial exchanges in terms of the friction velocity u_* since this is, dynamically, a more fundamental variable than the usual 10-m wind speed. We thus need transfer laws that relate the gradient wind and the enthalpy at the gradient-wind level to the interfacial momentum and enthalpy fluxes. For example, in the ‘‘wave drag’’ and ‘‘all effects’’ runs, we model the interfacial momentum flux as

$$\tau_{\text{int}} = \rho_a u_*^2, \tag{A1}$$

where ρ_a is the air density. We therefore need a gradient-wind transfer coefficient, C_{Dg}, that relates u_* to the magnitude of the gradient wind, G:

$$u_* = C_{Dg} G. \tag{A2}$$

Likewise, in these same runs, we model the interfacial enthalpy flux as

$$Q_{e,\text{int}} = -\rho_a u_* k_*, \tag{A3}$$

where k_* is the enthalpy flux scale (comparable to the temperature and moisture flux scales of Monin–Obukhov similarity theory). We thus also need an enthalpy flux transfer coefficient, C_{kg}, that relates k_* to the difference between the enthalpy at the sea surface, K_s, and the enthalpy at the level of the gradient wind, K_a:

$$-k_* = C_{kg}(K_s - K_a). \tag{A4}$$

We derive these transfer coefficients using Rossby similarity and two assumptions: i) in the boundary layer of a tropical cyclone, the gradient wind takes the place of the geostrophic wind in forcing surface exchange; ii) because of the high winds, the boundary layer stratification is near neutral.

Under these assumptions, we can simply write down the drag coefficient that Andreas (1998b) obtained,

$$C_{Dg} = \frac{\kappa}{\{[\ln(h_g/z_0) - A(0)]^2 + B(0)^2\}^{1/2}}. \tag{A5}$$

Here, $\kappa(=0.40)$ is the von Kármán constant; h_g is the reference height for the gradient wind; and $A(0)$ and $B(0)$ are the neutral-stability values of the Rossby similarity functions for longitudinal and transverse wind components, as defined by Yamada (1976), for example. Yamada gives $A(0) = 1.855$ and $B(0) = 3.020$, but here we use the values from Zilitinkevich's (1989a) more recent review, $A(0) = 1.7$ and $B(0) = 4.5$.

To represent the effect of wave drag, we parameterize the roughness length z_0 in (A5) with Charnock's relation,

$$z_0 = \alpha\frac{u_*^2}{g}. \tag{A6}$$

Here, g is the acceleration of gravity; and α is the Charnock constant, which we take as 0.0185 (Wu 1982; Johnson et al. 1998).

The value for C_{kg} in (A4) is likewise derived from Rossby similarity. From Andreas (1998b), we see that

$$-k_* = \frac{\kappa(K_s - K_a)}{\ln(h_k/z_k) - C(0)}. \tag{A7}$$

Here, h_k is the height at which we evaluate K_a, z_k is the roughness length for enthalpy, and $C(0)$ is the neutral-stability value of the Rossby similarity function for temperature (e.g., Yamada 1976). We take $C(0) = 3.7$ (Yamada 1976; Zilitinkevich 1989b). Hence, from (A4) and (A7), the enthalpy transfer coefficient is

$$C_{kg} = \frac{\kappa}{\ln(h_k/z_k) - C(0)}; \tag{A8}$$

then

$$Q_{e,\mathrm{int}} = \rho_a C_{Dg} C_{kg} G(K_s - K_a). \tag{A9}$$

We identify z_k in (A8) with the roughness length for temperature (z_T) and use the COARE algorithm (Fairall et al. 1996) to estimate z_T. The scalar roughness z_T is a decreasing function of the roughness Reynolds number $R_* = u_* z_0/\upsilon$, where υ is the kinematic viscosity of air. We compute z_0 and u_* for R_* from (A6) and (A2), respectively, but find that z_k ($=z_T$) gets quite small rapidly. When u_* is greater than 0.88 m s^{-1}, the Coupled Ocean–Atmosphere Response Experiment (COARE) algorithm predicts that z_k is smaller than the mean free path of an air molecule, about 7×10^{-8} m. It seems unphysical to us for z_k to be any smaller than this. Hence, we simplify the predictions of the COARE algorithm for small u_* and limit z_k to the length of the mean free path for large u_*. Our z_k algorithm is thus

$$z_k = 2.27 \times 10^{-4} \exp(-9.2u_*) \qquad \text{for } 0 \le u_* \le 0.8788 \text{ m s}^{-1}, \tag{A10a}$$

$$z_k = 7.0 \times 10^{-8} \qquad \text{for } 0.8788 \text{ m s}^{-1} < u_*. \tag{A10b}$$

In these, z_k is in meters when u_* is in meters per second. Note that because the z_0 value that went into these estimates acknowledges wave effect, z_k does also.

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