



RESEARCH LETTER

10.1002/2014GL059746

Key Points:

- Spray dominates enthalpy flux at the hurricane air-sea interface
- Enthalpy fluxes can be estimated from dropsonde thermodynamic data
- Trends of enthalpy flux coefficient can be misleading due to uncertainty

Correspondence to:

D. H. Richter,
David.Richter.26@nd.edu

Citation:

Richter, D. H., and D. P. Stern (2014), Evidence of spray-mediated air-sea enthalpy flux within tropical cyclones, *Geophys. Res. Lett.*, *41*, 2997–3003, doi:10.1002/2014GL059746.

Received 26 FEB 2014

Accepted 1 APR 2014

Accepted article online 4 APR 2014

Published online 23 APR 2014

Evidence of spray-mediated air-sea enthalpy flux within tropical cyclones

David H. Richter¹ and Daniel P. Stern²

¹Department of Civil and Environmental Engineering and Earth Sciences, University of Notre Dame, Notre Dame, Indiana, USA, ²National Center for Atmospheric Research, Boulder, Colorado, USA

Abstract It has long been conjectured that spray ejected from the high-wind ocean surface enhances air/sea enthalpy fluxes, but a lack of observational data, particularly at wind speeds exceeding 20 m s^{-1} , has prevented either confirmation or refutation of this hypothesis. The current study has two aims: first, to provide an estimate of surface enthalpy fluxes obtained from dropsonde data and second, to provide evidence of spray-mediated enthalpy transfer. These are accomplished first by assuming that Monin-Obukhov similarity is satisfied throughout the bottom 100 m of the high-wind boundary layer, then by focusing on the enthalpy flux H_K rather than its transfer coefficient C_K . The scaling of H_K with wind speed in observational data sets reveals similarities to spray-mediated fluxes predicted by a newly developed surface flux model, in contrast with measurements made in a laboratory setting. This behavior supports the claim that surface enthalpy fluxes are dominated by spray within tropical cyclones.

1. Introduction

The widespread use of bulk parameterizations for computing surface fluxes of momentum, heat, and moisture within mesoscale and global-scale meteorological models emphasizes the need for providing accurate estimates of the surface flux coefficients as well as a full understanding of their dependencies. This is particularly true in the field of tropical cyclone modeling, where the development and potential strength of a hurricane is highly sensitive to the surface drag coefficient C_D [Davis *et al.*, 2008; Bryan, 2013] and the relative strength of C_D to the surface enthalpy flux coefficient C_K [Emanuel, 1995; Bao *et al.*, 2011]. This sensitivity demonstrates the need for characterizing how these coefficients change with wind speed as well as understanding what physical processes they are required to represent. Many models, for instance, predict changes to both C_D and C_K due to the presence of sea spray at high winds [Bao *et al.*, 2011; Andreas, 2011; Kudryavtsev, 2006].

Currently, relatively few measurements of C_K exist at hurricane strength wind speeds due to the practical difficulties of instrumenting the ocean surface in such extreme conditions. Unfortunately, the lack of reliable observational measurements leads to a limited understanding of high-wind air-sea transfer, where large uncertainties and discrepancies between measurements and models continue to plague predictive efforts.

Only a few direct observations of heat, moisture, and enthalpy fluxes at high winds have been made using eddy covariance measurements, including those determined from the Humidity Exchange Over the Sea (HEXOS) and Coupled Boundary Layer Air-Sea Transfer Experiment (CBLAST) [DeCosmo *et al.*, 1996; Black *et al.*, 2007; French *et al.*, 2007; Zhang *et al.*, 2008]. These indicate no statistical dependence of C_K on wind speed up to roughly 30 m s^{-1} . Other indirect measurements have also been made, both in a laboratory setting [Haus *et al.*, 2010; Jeong *et al.*, 2012] and using energy budgets constructed from observed meteorological data [Bell *et al.*, 2012]. Although with a large uncertainty, C_K again appears to be insensitive to wind speed up to roughly 70 m s^{-1} .

The present goal is to add to the body of surface enthalpy flux measurements by using dropsonde data obtained during numerous hurricane flights and to combine these estimates with surface flux models to assess the impact of sea spray. The joint use of dropsonde profiles and Monin-Obukhov similarity theory is conceptually similar to the procedure used by Powell *et al.* [2003] to compute C_D , although in the present case this requires approximations of the sea surface temperature to arrive at estimates of C_K . By focusing instead on the surface enthalpy flux itself (as opposed to its transfer coefficient), we demonstrate similarities with previous studies and provide evidence of the effects of spray.

2. Methodology

For this study, we analyze a total of 2425 dropsonde profiles from 35 different tropical cyclones and 2 tropical depressions. The GPS dropsondes measure wind speed and direction, pressure, temperature, and humidity [Hock and Franklin, 1999], transmitting data either 2 or 4 times per second as they fall at a rate of roughly 10–12 m s⁻¹. All sondes in the data set were dropped either by the Air Force C130 or the NOAA P3 aircraft, generally from heights of 1.5–4 km. The data are acquired from NOAA's Hurricane Research Division (NOAA/Atlantic Oceanographic and Meteorological Laboratory (AOML)/Hurricane Research Division (HRD)), who have quality controlled and postprocessed the soundings using HRD's Editsonde software. Based on the typical fall speed, data are sampled on average every 5–6 m vertically. The measurement accuracy is 0.5 mb, 0.2°C, 2%, and 0.5 m s⁻¹ for pressure, temperature, relative humidity, and wind speed, respectively [Hock and Franklin, 1999].

To compute surface fluxes from the dropsonde data, one can construct mean vertical profiles of wind speed and other thermodynamic quantities and subsequently invoke Monin-Obukhov (MO) similarity theory. For example, Powell *et al.* [2003] use mean velocity profiles retrieved from dropsondes to obtain the surface momentum flux and, assuming no ocean current, the drag coefficient C_D as a function of wind speed. By utilizing MO theory, one makes an inherent assumption that turbulent transport and the resulting mean profiles depend only on fluxes at the surface (i.e., the outer part of the boundary layer does not influence the atmospheric surface layer, nor does large-scale advection) and that the predominant length scale is the elevation z taken substantially above the characteristic roughness height z_0 : $z \gg z_0$.

Under these assumptions, as well as assuming neutral stability (it is assumed that in the high-wind hurricane boundary layer, shear turbulence production dominates buoyancy production), the mean velocity profile takes the form

$$\langle u \rangle = \frac{u_*}{\kappa} \ln \left(\frac{z}{z_0} \right), \quad (1)$$

where u_* is the friction velocity and κ is the von Kármán constant taken to be $\kappa = 0.4$ throughout. The current study is concerned with surface flux of specific enthalpy k , defined as follows:

$$k = [(1 - q)c_{p,a} + qc_{p,l}] \theta + L_v q. \quad (2)$$

Here q is the specific humidity, $c_{p,a}$ is the specific heat of dry air, $c_{p,l}$ is the specific heat of liquid water, and L_v is the latent heat of vaporization. We use the potential temperature θ , where the temperature profile recorded by each individual sonde is referenced to the temperature associated with the surface pressure measured by that sonde. Hence, the values of T and θ are identical at the surface for each profile, and any deviations with height are due to adiabatic expansion. Since we are only interested in enthalpy changes due to fluxes at the surface, using θ effectively removes enthalpy changes due to adiabatic expansion (θ defined in this study differs from the typical convention of referencing to an assumed surface pressure of 1000 mb. Likewise, the enthalpy of equation (2) differs from its conventional definition since it is based on θ and not T). We then invoke MO theory for relating mean profiles and surface fluxes of k by assuming that k is conserved throughout the boundary layer (i.e., neglecting external/elevated sources such as that resulting from rainfall):

$$\langle k \rangle - k_s = \frac{k_*}{\kappa} \ln \left(\frac{z}{z_k} \right), \quad (3)$$

where k_s is the enthalpy at the ocean surface, z_k is the roughness height for enthalpy, and k_* is related to the surface flux of enthalpy through the friction velocity u_* :

$$\tau = \rho_a u_*^2, \quad (4)$$

$$H_K = -\rho_a u_* k_*. \quad (5)$$

Here, ρ_a refers to a constant reference dry air density, τ represents the total momentum flux at the surface, and H_K is the total enthalpy flux at the surface. At the same time, one can define bulk flux parameterizations of momentum and enthalpy in the standard way:

$$\tau = \rho_a C_D U_{10}^2, \quad (6)$$

$$H_K = -\rho_a C_K U_{10} (k_{10} - k_s). \quad (7)$$

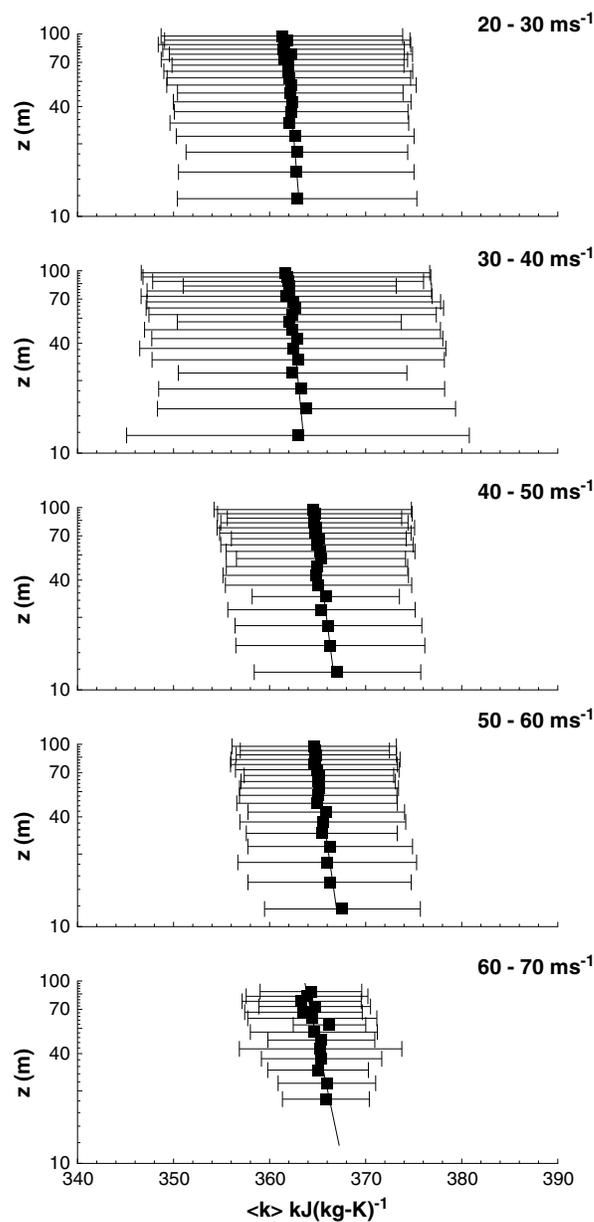


Figure 1. Vertical profiles of mean enthalpy $\langle k \rangle$ for wind speeds up to the 60–70 m s^{-1} bin for a single representative SST bin of 300–301 K. Error bars represent the 95% confidence interval based on the standard deviations of the mean profiles. The solid line represents the fit of $\langle k \rangle$ versus $\ln(z)$.

When plotting $\langle u \rangle$ versus $\ln(z)$, equation (1) indicates that in the region where MO theory holds, the mean profile should lie on a straight line with slope of u_* / κ . This fit is performed over the lower 100 m of the mean profiles (only profiles with 15 or more points in the vertical are fit), and the computed values of u_* and C_D compare very well with the results of Powell et al. [2003] and Holthuijsen et al. [2012] (not shown).

To obtain the surface enthalpy fluxes, each individual profile is again binned according to its mean velocity. Within each wind speed range, however, each profile is additionally binned according to SST, ranging between 295 K and 305 K with 1 K intervals. This procedure places profiles with values of SST within 1 K into the same bin within a certain wind speed range. Other criteria, such as binning based on the difference between SST and a low-level temperature or only on the mean boundary layer temperature, produce similar results to those presented.

In equations (6) and (7), the subscript $[\]_{10}$ refers to mean values at 10 m, and both expressions neglect surface currents.

In principle, the mean profiles of velocity and enthalpy can be used to obtain u_* and k_* by fitting equations (1) and (3). C_D can then be readily obtained through equation (6). To obtain C_K , however, the sea surface temperature (SST) is required in order to compute the surface enthalpy k_s (sea salinity is assumed to be 34 psu). Since the dropsonde does not measure SST, it must be retrieved in some other way. We choose to use the quarter-degree Reynolds daily SST analyses (www.ncdc.noaa.gov/cdr/operationalcdrs.html), interpolated to the dropsonde's latitude-longitude position, as an estimate of SST beneath the recorded profile. By doing this, it is important to emphasize that the computed values of C_K will be much more uncertain than the values of k_{*r} , since C_K depends on SST while k_* does not.

The procedures for obtaining fluxes of momentum and enthalpy differ slightly. To obtain momentum fluxes, each individual profile is binned according to its mean velocity throughout the lower 500 m, similar to what was done by Powell et al. [2003] and Holthuijsen et al. [2012]. Each wind speed bin is chosen to have a width of 10 m s^{-1} , ranging between 20 and 70 m s^{-1} . Furthermore, each individual dropsonde measurement within a single profile is placed into a corresponding 5 m bin in the vertical direction. Averages are then taken over the samples in each vertical bin (only averages containing five or more samples are retained), resulting in a mean velocity profile for each wind speed range, spaced every 5 m in the vertical.

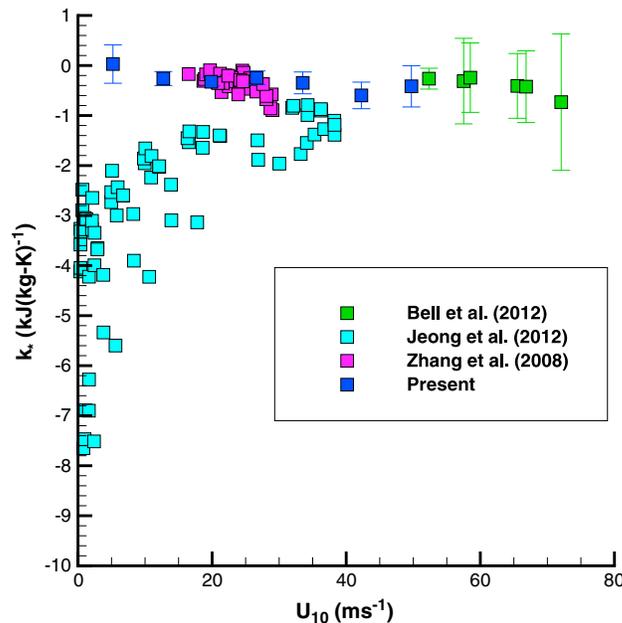


Figure 2. Values of k_* , averaged over all SST bins, versus U_{10} . Error bars represent the 95% confidence interval. Current data shown with blue squares. Other colors are referenced in the figure legend.

the hurricane inner core may be questionable. Other factors, such as the appropriateness of constructing mean velocity and thermodynamic profiles from individual sonde measurements, may compromise the inferred surface enthalpy flux measurements as well. Despite these potential violations, we deem our flux computation strategy as adequate for providing quantitative estimates given both the dearth of available data and similar magnitudes of uncertainty in other existing measurements.

3. Results and Discussion

For each 1 K wide range of SST and each 10 m s^{-1} wide range of wind speed, we compute a value of k_* . These values are then averaged across SST, resulting in a single value of k_* for each wind speed bin. Figure 2 plots the values of k_* as a function of 10 m wind speed. Here U_{10} is the mean velocity computed from the fit of equation (1) within each wind speed bin, evaluated at $z = 10 \text{ m}$.

Figure 2 shows that k_* , as determined from the dropsonde data, remains relatively insensitive to U_{10} in the range of roughly $20\text{--}50 \text{ m s}^{-1}$. Furthermore, the range of values is in quantitative agreement with the covariance measurements of Zhang et al. [2008] and the indirect estimates of Bell et al. [2012]. However, the agreement between the present case, Zhang et al. [2008], and Bell et al. [2012] is in stark contrast to the data of Jeong et al. [2012], who show k_* becoming much less negative with increasing wind speed between 0 and 30 m s^{-1} , suggesting an increasingly vertical profile of $\langle k \rangle$ when plotted in the same way as Figure 1.

The values of the enthalpy flux coefficient C_K are then computed using the values of k_* , u_{*r} , and k_s obtained from the values of SST. These are shown in Figure 3 as a function of U_{10} . The current estimates are found to lie in the same range as the previous high-wind values, albeit with a high degree of uncertainty and a slight underprediction. It should be noted that when combining equations (5) and (7) to obtain C_K from k_* and u_{*r} , the difference $(k_{10} - k_s)$ appears in the denominator, and as a result the present estimates of C_K are highly sensitive to the value of sea surface temperature when SST is close to θ_{10} . Therefore, Figure 3 excludes profiles where the difference between the lowest recorded temperature reading from the dropsonde (usually near 10 m elevation), and SST is found to be less than 1.5 K, leaving a total of 1479 sonde profiles. While a slightly increasing trend appears in C_K versus U_{10} , the large uncertainty of SST (and thus C_K) precludes one from drawing conclusions based solely on C_K .

We therefore turn our attention instead to H_{Kr} , the dimensional value of enthalpy flux determined from the mean profiles via equation (5), with the goal of understanding the discrepancy in Figure 2 between the

Figure 1 shows mean enthalpy profiles over all wind speed ranges for one representative bin ranging between 300 and 301 K (the other 10 SST bins are qualitatively similar). Profiles of $\langle k \rangle$ appear to satisfy MO theory, in that each profile (taken below 100 m) follows a straight line when plotted on a semilogarithmic plot. As before, the slope of this line is used to obtain k_* from equation (3), and the linear fits of $\langle k \rangle$ versus $\ln(z)$ are included as straight lines in Figure 1.

At this point we again stress that by using MO theory to obtain k_{*r} , we make certain assumptions about the hurricane boundary layer which may or may not be fully satisfied. For example, despite the evidence offered by Zhang et al. [2008], it is not firmly established that fluxes (and k_* and u_{*r}) remain constant to heights approaching 100 m within tropical cyclones. Moreover, Smith and Montgomery [2014] claim that even the existence of a logarithmic layer within

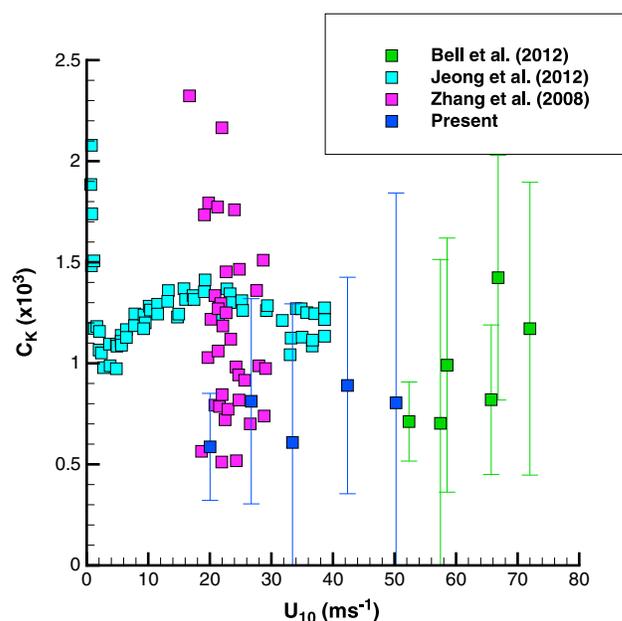


Figure 3. Estimates of C_K based on k_* and the interpolated value of SST as a function of U_{10} . Current data shown with blue squares. Other colors are referenced in the figure legend.

values of k_* obtained using in situ data (i.e., the present case, Zhang et al. [2008], and Bell et al. [2012]) and those made in a laboratory setting [Jeong et al., 2012]. Figure 4 therefore plots H_K versus U_{10} on a log-log plot, in order to investigate the scaling of the enthalpy flux with wind speed.

Several key features should be noted in this figure. First, the solid color lines indicate fits of the various experimental data sets, and their slope β would suggest a power law scaling of $H_K \sim U_{10}^\beta$. The fitted slopes of the cyan [Jeong et al., 2012], magenta [Zhang et al., 2008], green [Bell et al., 2012], and blue (present) lines are $\beta = [0.55, 3.2, 3.7, 1.8]$, respectively. Therefore, the sublinear value of β obtained using the reported data of Jeong et al. [2012] shows that their measurements of H_K scale much less strongly with U_{10} compared to all other measurements. Since u_* scales nearly linearly with U_{10} for all data sets (not shown), this is

consistent with the trends of k_* in Figure 2—i.e., the decrease in magnitude of k_* with wind speed of Jeong et al. [2012] offsets the increase of u_* and thus, equation (4) shows that H_K will scale less strongly with U_{10} .

We argue that the sublinear scaling of H_K versus U_{10} in the laboratory data is due to a dominance of interfacial (i.e., aerodynamic) exchange between the water and the air above, while the strong scaling of the observational data gives evidence of spray-induced enthalpy fluxes. Standard boundary layer theory suggests a sublinear increase of scalar transfer (in this case enthalpy) with wind speed [see, for example, Schlichting, 1968, equation 23.19]. The production of sea spray, meanwhile, is thought to scale as u_*^3 based on the wind energy required to tear droplets from the surface [Andreas et al., 1995; Andreas and Emanuel, 2001]. This increasing loading of water droplets with wind speed is consistent with a much stronger increase of surface enthalpy flux due to partial droplet evaporation and reentry back into the ocean [Andreas and Decosmo, 2002; Andreas, 2011].

This fundamental difference in scaling between interfacial and spray-mediated fluxes is the basis for the flux model of Andreas et al. [2008], which uses the Coupled Ocean-Atmosphere Response Experiment 2.6 (COARE 2.6) algorithm [Fairall et al., 1996] to distinguish between the two routes of enthalpy transfer. To therefore illustrate the dominance of interfacial transfer within the laboratory data, we utilize an updated version of the spray-induced flux model of Andreas et al. [2008] (version 4 of this algorithm is available from <http://www.nwra.com/resumes/andreas/software.php>). The interfacial, spray-mediated, and total enthalpy fluxes are computed based on the ambient experimental conditions of Jeong et al. [2012] and are presented in Figure 4. Note that we do not apply the flux algorithm to the field data sets due to incompatibilities of the reported measurement data with the required model inputs.

Figure 4 shows that over the entire range of wind speeds tested, the total enthalpy flux of Jeong et al. [2012] increases with U_{10} in the same way that the COARE 2.6 model predicts for the interfacial fluxes only. Given their experimental conditions, the flux model predicts that spray-induced enthalpy fluxes should approach the magnitude of interfacial fluxes only near the maximum wind speeds obtained in their wind tunnel. Therefore, while it may appear that no transition in β exists for their laboratory data, measurements at higher wind speeds would be necessary to confirm a continued interfacial flux scaling.

The agreement between the laboratory measurements and the model-predicted interfacial enthalpy flux scaling is in contrast with the three in situ data sets, each of which exhibits a much stronger scaling of H_K with increasing wind speed—closer to that predicted by the model for spray-mediated fluxes. This suggests

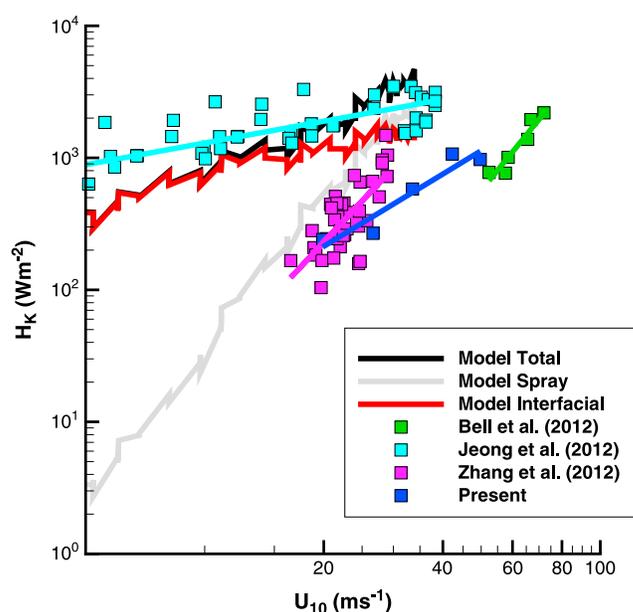


Figure 4. Enthalpy flux H_K (W m^{-2}) versus U_{10} with logarithmic axes. Square symbols indicate observational data (defined in the legend), and lines indicate fits through the symbols with the corresponding colors. The black, light gray, and red lines show the predictions of the surface flux model of *Andreas et al.* [2008] for the total, spray-mediated, and interfacial flux components, respectively, based on the experimental conditions of *Jeong et al.* [2012].

second is the relatively dry ambient air which is present in the laboratory (mean of 56% relative humidity) compared to values found in the hurricane boundary layer. These factors would lead to higher magnitudes of H_K due to increased interfacial fluxes of sensible and latent heat at low wind speeds and push the model-predicted transition to spray-mediated fluxes to higher wind speeds.

Third, the finite length of the experimental setup forces spray to be blown out of the test section, effectively preventing it from contributing to the spray-mediated flux. *Jeong et al.* [2012] estimate that this effect is not large by comparing droplet suspension time scales with sensible and latent heat time scales, but only measurements at higher winds (or lower air-water temperature differences) could confirm this based on the flux scalings presented here. In either case, the differences in β between the observational and laboratory data strongly suggest a spray-mediated enthalpy flux within tropical cyclones.

Finally, it is important to note that although there are differences in the scaling of H_K with U_{10} between the laboratory and field observations, one could not make an argument for the effects of spray based solely on the behavior of C_K . As presented in Figure 3, C_K determined from all available sources above 20 m s^{-1} does not exhibit a strong dependence on wind speed, despite the differences in β . The flux model, however, predicts a significant increase in C_K once the spray-mediated fluxes begin to dominate over the interfacial fluxes [*Andreas*, 2011]. For the values of C_K computed via the dropsonde profiles, this discrepancy emphasizes the much stronger uncertainty of C_K versus H_K which results from making interpolated estimates of SST (similar limitations in estimating SST affect the predictions of *Zhang et al.* [2008] and *Bell et al.* [2012] as well). We therefore suggest that the scaling of H_K with wind speed is an additional, possibly more robust, method for assessing the impacts of spray.

4. Conclusions

We have invoked Monin-Obukhov similarity theory to compute surface fluxes of enthalpy within tropical cyclones at 10 m wind speeds up to 50 m s^{-1} . This is done using over 2000 dropsonde profiles taken from 37 different tropical cyclones, where individual profiles are binned based on their mean boundary layer velocity and a sea surface temperature interpolated to the dropsonde position using a quarter-degree sea surface

a fundamentally different mechanism governing enthalpy transfer between the in situ and laboratory measurements. As mentioned above, this stronger scaling with wind speed is due in large part to the predicted u_*^3 scaling of spray droplet production by wind work at the surface [*Andreas et al.*, 1995; *Andreas and Emanuel*, 2001] (*Andreas* [2010] demonstrates a scaling of $\beta \approx 2.73$ based on fits with the HEXOS [*DeCosmo et al.*, 1996] and Fronts and Atlantic Storm-Track Experiment [*Joly et al.*, 1997] data sets). Physically, this implies that the surface enthalpy flux is closely connected with the predicted increase of spray droplet production with wind speed, rather than resulting from purely aerodynamic interactions with the surface.

In the laboratory wind tunnel, a few factors could lead to the dominance of the interfacial flux scaling. The first is air-water temperature difference of *Jeong et al.* [2012], which, ranging between 1.2 and 9.3° C , reaches higher values than typically found within tropical cyclones [*Cione et al.*, 2000]. The

temperature analysis product. Assuming that the requirements of MO theory are satisfied, this process produces evaluations of the enthalpy flux-scale k_{*} , and this value, combined with the SST interpolants, provides estimates of C_K which lie in the same range as the few existing measurements at high winds.

By focusing on the enthalpy flux H_K instead of C_K , it is found that the few existing measurements taken from within hurricanes produce a power law scaling which closely resembles the spray-mediated enthalpy flux predicted by an updated version of the model of *Andreas et al.* [2008], while measurements taken in a laboratory context [*Jeong et al.*, 2012] exhibit a power law scaling closely resembling that predicted for the interfacial enthalpy transfer. We suggest that this discrepancy is due to conditions within the laboratory setting which favor interfacial fluxes, including large air-water temperature differences and spray blown out of the tank, and indicates that enthalpy fluxes within hurricanes are likely dominated by spray at high winds.

Acknowledgments

The authors would like to sincerely thank Jun Zhang for sharing his CBLAST data, as well as Edgar Andreas for generously providing access to his interfacial/spray flux algorithm code and providing many helpful comments and critiques of the manuscript. We would also like to acknowledge the NOAA/AOML Hurricane Research Division for making the dropsonde data available for this analysis. D. Stern was supported by an NSF-AGS Postdoctoral Research Fellowship (AGS-1231193). NCAR is supported by the National Science Foundation. All sonde profiles used here are publicly available on the NOAA HRD website: <http://www.aoml.noaa.gov/hrd/>.

The Editor thanks two anonymous reviewers for their assistance in evaluating this paper.

References

- Andreas, E. L. (2010), Spray-mediated enthalpy flux to the atmosphere and salt flux to the ocean in high winds, *J. Phys. Oceanogr.*, *40*, 608–619, doi:10.1175/2009JPO4232.1.
- Andreas, E. L. (2011), Fallacies of the enthalpy transfer coefficient over the ocean in high winds, *J. Atmos. Sci.*, *68*, 1435–1445, doi:10.1175/2011JAS3714.1.
- Andreas, E. L., and J. Decosmo (2002), The signature of sea spray in the HEXOS turbulent heat flux data, *Boundary Layer Meteorol.*, *103*, 303–333, doi:10.1023/A:1014564513650.
- Andreas, E. L., and K. A. Emanuel (2001), Effects of sea spray on tropical cyclone intensity, *J. Atmos. Sci.*, *58*, 3741–3751, doi:10.1175/1520-0469(2001)058<3741:EOSSOT>2.0.CO;2.
- Andreas, E. L., J. B. Edson, E. C. Monahan, M. P. Rouault, and S. D. Smith (1995), The spray contribution to net evaporation from the sea: A review of recent progress, *Boundary Layer Meteorol.*, *72*, 3–52, doi:10.1007/BF00712389.
- Andreas, E. L., P. O. G. Persson, and J. E. Hare (2008), A bulk turbulent air-sea flux algorithm for high-wind, spray conditions, *J. Phys. Oceanogr.*, *38*, 1581–1596, doi:10.1175/2007JPO3813.1.
- Bao, J.-W., C. W. Fairall, S. A. Michelson, and L. Bianco (2011), Parameterizations of sea-spray impact on the air-sea momentum and heat fluxes, *Mon. Weather Rev.*, *139*, 3781–3797, doi:10.1175/MWR-D-11-00007.1.
- Bell, M. M., M. T. Montgomery, and K. A. Emanuel (2012), Air-sea enthalpy and momentum exchange at major hurricane wind speeds observed during CBLAST, *J. Atmos. Sci.*, *69*, 3197–3222, doi:10.1175/JAS-D-11-0276.1.
- Black, P. G., E. A. D'Asaro, W. M. Drennan, J. R. French, P. P. Niiler, T. B. Sanford, E. J. Terrill, E. J. Walsh, and J. A. Zhang (2007), Air-sea exchange in hurricanes: Synthesis of observations from the Coupled Boundary Layer Air-Sea Transfer experiment, *Bull. Am. Meteorol. Soc.*, *88*, 357–374, doi:10.1175/BAMS-88-3-357.
- Bryan, G. H. (2013), Comments on Sensitivity of tropical-cyclone models to the surface drag coefficient, *Q. J. R. Meteorol. Soc.*, *139*, 1957–1960, doi:10.1002/qj.2066.
- Cione, J. J., P. G. Black, and S. H. Houston (2000), Surface observations in the hurricane environment, *Mon. Weather Rev.*, *128*, 1550–1561, doi:10.1175/1520-0493(2000)128<1550:SOITHE>2.0.CO;2.
- Davis, C., et al. (2008), Prediction of landfalling hurricanes with the advanced hurricane WRF model, *Mon. Weather Rev.*, *136*, 1990–2005, doi:10.1175/2007MWR2085.1.
- DeCosmo, J., K. B. Katsaros, S. D. Smith, R. J. Anderson, W. A. Oost, K. Bumke, and H. Chadwick (1996), Air-sea exchange of water vapor and sensible heat: The Humidity Exchange Over the Sea (HEXOS) results, *J. Geophys. Res.*, *101*(C5), 12,001–12,016, doi:10.1029/95JC03796.
- Emanuel, K. A. (1995), Sensitivity of tropical cyclones to surface exchange coefficients and a revised steady-state model incorporating eye dynamics, *J. Atmos. Sci.*, *52*(22), 3969–3976, doi:10.1175/1520-0469(1995)052<3969:SOTCTS>2.0.CO;2.
- Fairall, C. W., E. F. Bradley, D. P. Rogers, J. B. Edson, and G. S. Young (1996), Bulk parameterization of air-sea fluxes for Tropical Ocean–Global Atmosphere Coupled–Ocean Atmosphere Response Experiment, *J. Geophys. Res.*, *101*(C2), 3747–3764, doi:10.1029/95JC03205.
- French, J. R., W. M. Drennan, J. A. Zhang, and P. G. Black (2007), Turbulent fluxes in the hurricane boundary layer. Part I: Momentum flux, *J. Atmos. Sci.*, *64*, 1089–1102, doi:10.1175/JAS3887.1.
- Haus, B. K., D. Jeong, M. A. Donelan, J. A. Zhang, and I. Savelyev (2010), Relative rates of sea-air heat transfer and frictional drag in very high winds, *Geophys. Res. Lett.*, *37*, L07802, doi:10.1029/2009GL042206.
- Hock, T. F., and J. L. Franklin (1999), The NCAR GPS dropwindsonde, *Bull. Am. Meteorol. Soc.*, *80*, 407–420, doi:10.1175/1520-0477(1999)080<0407:TNGD>2.0.CO;2.
- Holthuijsen, L. H., M. D. Powell, and J. D. Pietrzak (2012), Wind and waves in extreme hurricanes, *J. Geophys. Res.*, *117*, C09003, doi:10.1029/2012JC007983.
- Jeong, D., B. K. Haus, and M. A. Donelan (2012), Enthalpy transfer across the air-water interface in high winds including spray, *J. Atmos. Sci.*, *69*, 2733–2748, doi:10.1175/JAS-D-11-0260.1.
- Joly, A., et al. (1997), The Fronts and Atlantic Storm-Track Experiment (FASTEX): Scientific objectives and experimental design, *Bull. Am. Meteorol. Soc.*, *78*, 1917–1940, doi:10.1175/1520-0477(1997)078<1917:TFAAST>2.0.CO;2.
- Kudryavtsev, V. N. (2006), On the effect of sea drops on the atmospheric boundary layer, *J. Geophys. Res.*, *111*, C07020, doi:10.1029/2005JC002970.
- Powell, M. D., P. J. Vickery, and T. A. Reinhold (2003), Reduced drag coefficient for high wind speeds in tropical cyclones, *Nature*, *422*, 279–283, doi:10.1038/nature01481.
- Schlichting, H. (1968), *Boundary-Layer Theory*, 6th ed., McGraw-Hill, New York.
- Smith, R. K., and M. T. Montgomery (2014), On the existence of the logarithmic surface layer in the inner core of hurricanes, *Q. J. R. Meteorol. Soc.*, *140*(678), 72–81, doi:10.1002/qj.2121.
- Zhang, J. A., P. G. Black, J. R. French, and W. M. Drennan (2008), First direct measurements of enthalpy flux in the hurricane boundary layer: The CBLAST results, *Geophys. Res. Lett.*, *35*, L14813, doi:10.1029/2008GL034374.