9C.4 DIRECT AIRBORNE MEASURMENTS OF MOMENTUM FLUX IN HURRICANES

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1. Introduction

For wind speeds from 5 to 20 m s^{-1} over the ocean, it has been repeatedly shown that the drag coefficient increases nearly monotonically (Smith 1980; Large and Pond 1981; Geernaert et al. 1986, Smith et al. 1992) with wind speed. To explicitly model conditions at wind speeds greater than 20 m s^{-1}, including tropical cyclones, requires some sort of extrapolation of existing flux measurements. The simplest approach leads to a monotonically increasing drag coefficient with wind speed. However, evidence is mounting that the drag coefficient does not continue to increase at higher wind speeds. Modeling studies (ie Moon et al. 2004) support a 'roll-off' of the drag coefficient at wind speeds of approximately 35 m s^{-1} due to wave effects. Donelan et al. (2004) provided the first direct measurements from laboratory wave-tank studies. Their results indicate a 'saturation' or roll-off of the drag coefficient to a near-constant value at wind speeds greater than 33 m s^{-1}. Powell et al. (2003) found similar results from inferred drag coefficients from 331 GPS sonde wind profiles in 15 tropical cyclones. Assuming a mean log-profile, Powell et al.'s results indicate a leveling off and even a decrease of the drag coefficient somewhere between 30 and 35 m s^{-1}.

A primary objective of the Coupled Boundary Layer Air-Sea Transfer (CBLAST) Hurricane program is to obtain measurements suitable for the computation of near-surface fluxes of heat and momentum in hurricanes. The work presented herein focuses on measurements of the near-surface fluxes of momentum obtained from an instrumented aircraft.

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2. Measurements

A specially instrumented National Oceanic and Atmospheric Administration (NOAA) WP-3D Orion aircraft (hereafter referred to by its call sign, N43RF) was used to obtain boundary layer measurements suitable for the computation of heat and momentum flux. Because the hurricane environment makes it difficult to obtain measurements with an aircraft very near the surface a flight strategy was developed to utilize a series of straight and level flux runs beginning near the top of the boundary layer with successively lower-altitude legs until a final leg of roughly 60 to 100 m altitude was completed. Such a 'stepped descent' series of runs may include as many as five flux legs of roughly 15 to 55 km in length. Ideally, a full stepped descent pattern consists of a series of legs oriented parallel to the mean flight-level wind vector and a second series of legs oriented perpendicular to the mean flight-level wind vector.

Safety requirements dictate all boundary layer legs are to be flown in rain-free conditions. A typical flight pattern then consists of a survey pattern (i.e. a figure four through the storm typically between 1.5 and 3 km altitude) followed by one or two full stepped descents in a suitably identified region of the storm. Primary considerations for suitability include space available between rain-bands for completing flux legs and location within storm with regard both to distance from the center and storm quadrant.

For the CBLAST-Hurricane experiment, N43RF was instrumented with three independent systems for measuring air-relative aircraft velocity: a 5-hole nose radome system (Brown et al. 1983; Khelif et al. 1999), a system consisting of two Rosemount 858Y probes and a pitot tube, and an NOAA Air Resources Laboratory (ARL) designed 9-hole gust probe system (BAT; Crawford and Dobosy 1992; Hacker and Crawford 1999). From data obtained during calibration maneuvers performed in advance of the 2003 measurement campaign, it is noted that the one of the pressure ports from the nose radome system leaked. Further, examination of data from flux runs conducted after penetrations of regions of heavy rain indicate water intrusion into one or more the radome ports was common. Thus, for the remainder of the analysis, data will be presented only from the Rosemount system and the ARL BAT probe.

3. Momentum Flux

The flux data presented herein are from measurements made during six flights in two storms in 2003. The flights occurred on September 2, 3, and 4 into Hurricane Fabian and September 12, 13, and 14 into Hurricane Isabel. During all six flights the hurricanes were either category 4 or 5.

Data presented from Hurricane Fabian are derived from the Rosemount system (the BAT was inoperable during these flights). Data from Hurricane Isabel are derived from the BAT system. To assure continuity of measurements between probes and hence storms, comparisons are shown in Figure 1 for the covariance from Rosemount measurements and from BAT measurements for flux runs in Hurricane Isabel. The correlation between the two data sets is 0.75 and is reasonable given the noisy nature of the calculation.

For both storms, fifty-nine flux runs from altitudes below 400 m were completed. Measurements from drop-sondes in both storms suggest that the top of the boundary layer at the locations of the flux runs was normally between 300 and 500 m. Wind, temperature and humidity data from these flux runs are used to compute wind stress, friction velocity, and ultimately estimate the 10 m neutral drag coefficient. Beginning and ending times for a given run are chosen initially based on markers set in-flight and are modified in post-flight analysis to remove sections where the aircraft is not straight and level or the plane passes through rain at the beginning or end of the leg leading to spikes in the wind data.

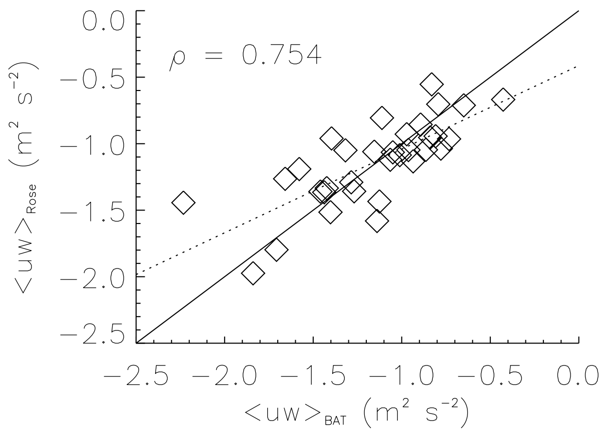

Figure 1: Scatter diagram of the along wind covariance from the Rosemount probe and the BAT probe from all flux runs during which both probes were operable.

Data quality assurance for individual flux legs include inspection of the linear cumulative summation of the covariance, the power spectra for individual wind components, and co-spectra and ogives (Friehe et al. 1991) for along wind and cross wind stress. Eleven runs are discarded based on this analysis. Figure 2 shows six panels: two each containing graphs from two flux runs for (a and b) linear cumulative summation of along wind covariance, (c and d) along wind co-spectra, and (e and f) ogives for along wind covariance. Data from two discarded runs are shown on the left panels. The right panels show data from two good runs. For the discarded runs, disturbances over small portions of the run affect the total wind stress by as much as 20%-30%. This is reflected in the cumulative sum as sharp changes over small distances. Note that the co-spectra and ogives do not appear ‘well-behaved’ for these runs. It is possible that organized mesoscale features such as boundary layer rolls are responsible (Foster et al. 2005), but that is only conjecture at this point. For the analysis presented herein, data from discarded runs are excluded but do warrant further investigation.

For the accepted runs, the linear cumulative summation remains with a near constant slope over the entire run. Both the co-spectra and the ogives reveal consistent behavior between runs and suggest that the majority of the energy in the momentum flux is from eddies ranging from 100 m to 2 to 3 km in size.

The majority of the runs (34 out of 48) are oriented along the mean wind vector, reflecting the difficulty in finding a long enough path of rain-free space in the cross wind direction. The average air-relative leg length is roughly 28 km, with a minimum of 13 km from one of the lower altitude legs, and a maximum of more than 55 km. The majority of legs are between 20 and 30 km in length. The lowest altitude leg is 70 m. Only six legs are at altitudes less than 100 m.

Leg-averaged mean flight level winds speeds vary from 21 m s^{-1} to just over 40 m s^{-1}, with most between 30 and 35 m s^{-1}. The near-surface neutral-stability wind speed, U_{10N}, is taken from the nadir-pointing stepped frequency microwave radiometer (SFMR; Uhlhorn and Black 2003). Several methods were tested to estimate U_{10}, however SFMR estimates were found to be the most consistent with other measurements. Values of U_{10N} vary from a minimum of roughly 17 m s^{-1} to a maximum of 31 m s^{-1}.

The covariance is computed by rotating the wind vector into its along wind and cross wind components followed by removing the mean over the entire leg. The magnitude of the momentum flux may then be calculated:

$$\tau = -\rho\left[\overline{u'w'} + \overline{v'w'}\right], \qquad (7)$$

where ρ is the air density and $\overline{u'w'}$ and $\overline{v'w'}$ are the along wind and cross wind components of the covariance for a given flux leg. In this analysis we do not assume a constant wind direction throughout an entire leg. Thus, the reader is cautioned that the magnitude of the along and cross wind components of the covariance is highly dependant on the amount of turning of the wind and how changes in wind direction and magnitude over the course of a single run are interpreted in the analysis (*i.e. how variations in the horizontal wind vector are broken into variations in the along wind component and the cross wind component)*. However, the total covariance and hence the stress is unaffected.

The friction velocity is estimated by assuming constant stress throughout the depth of the hurricane boundary layer. Following Donelan (1990) and Banner et al. (1999) we compute the friction velocity correcting for the influence of the Coriolis force and the horizontal pressure gradient.

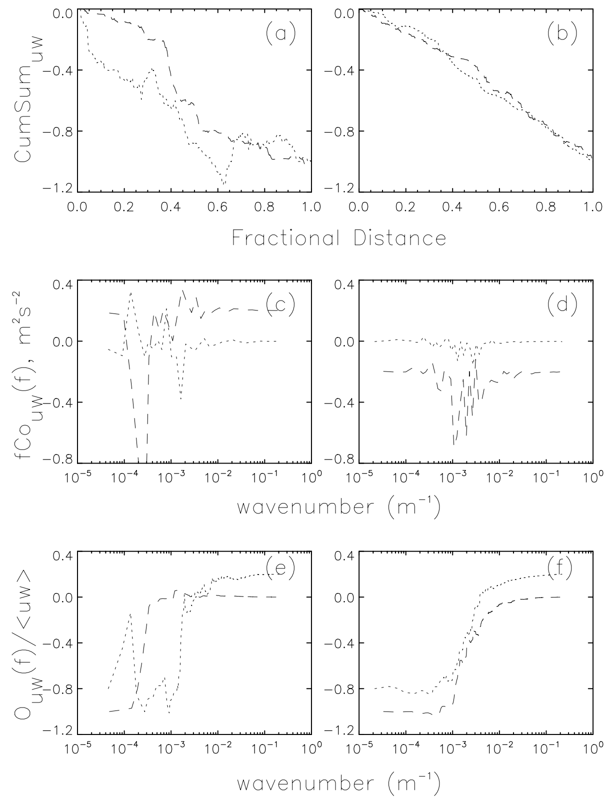

Figure 2: Data are shown from two discarded runs in the left panel and two good runs in the right panel. The top graphs (a & b) show the linear cumulative summation, the middle two (c & d) show the freequency-weighted co-spectra, and the bottom two (e & f) show ogives normalized by the total covariance. In each of the middle and lower graphs, one of the lines has been offset to aid in comparison.

4. Results and Discussion

The drag coefficient provides a means to parameterize surface fluxes based on 'bulk' measurements. Here we compute the 10 meter neutral drag coefficient from our measurements of friction velocity and 10 m wind speed such that: $C_{D,10\mathrm{N}} = u_*^{\,2} / U_{10\mathrm{N}}^{\,2}$. Figure 3 shows $C_{D,10N}$ computed from the 48 flux runs plotted as a function of near surface neutral wind speed, U_{10N}. Virtually all of the data fall below extrapolated results from earlier studies (Large and Pond 1981, Smith 1980). The data are delineated both by storm and by orientation of the flux leg to the environmental wind. Allowing for the small sample size, there is no significant difference between results when separated by leg orientation or by storm. The bold stars and line in the figure represent the bin averaged drag coefficient for 2.5

m s^{-1} wide bins centered at 20, 22.5, 25 m s^{-1}, etc. For the conditions under which these measurements were made there appears no dependence of $C_{D,10N}$ on U_{10N}.

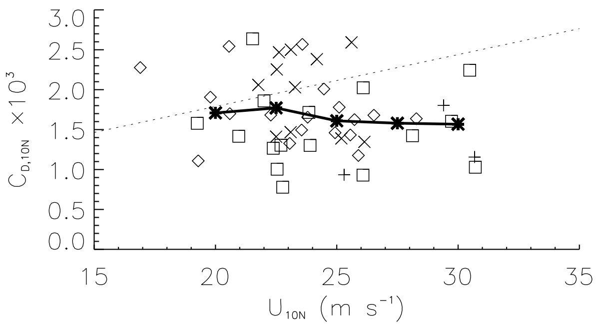

Figure 3: Computed $C_{D,10N}$ as a function of U_{10N} for the 48 flux runs from this study. Data are delineated by storm (Fabian: squares and plusses; Isabel: diamonds and X's) and by leg-wind orientation (cross: plusses and X's; along: squares and diamonds). Thick line represents bin-averaged results.

Three of the stepped descents were completed in the right rear quadrant. One stepped descent was completed in the left front quadrant and two were completed in the right front quadrant Wright et al. (2001) presented wave spectra from a Category 3 hurricane over the open ocean in 1998. Their results included individual spectra for several locations in the four storm quadrants within roughly 200 km from the storm center. Directly behind the eye, they found tri-modal wave spectra that, as one moved through the right rear quadrant towards the right front, merge into one broad dominant swell propagating roughly along the wind direction. Continuing into the right front quadrant the wave spectra remain dominated by one primary swell, but oriented 30 to 60 degrees to the right of the wind. Assuming such a pattern is typical; one may expect the stress in the right front and right rear quadrants to be diminished by the presence of following swell (c.f. Drennan et al. 1999). On the other hand, stronger winds in the right front quadrant might be expected to increase the drag coefficient through wave age enhancement (i.e. younger, more strongly forced wind-sea waves). As indicated above, our data show no dependence when delineated by storm quadrant. However, further investigation, including acquisition of additional measurements and combining wave spectral measurements with flux measurements would allow for a test of this hypothesis.

Katsaros et al. (2002) reported on roll-type features in hurricanes observed from satellites. Such features could carry a significant portion of the momentum. Under-sampling these features would lead to an under estimation of the flux and hence the drag coefficient. However, for measurements in this study there is little evidence in the aircraft data of the widespread existence of such features. Further, one might expect that coherent structures, depending on their alignment with the environmental wind, might lead to preferential sampling depending on the orientation of the flight leg (along wind or cross wind). Given that all of the runs used in this study passed the data quality assurance tests described in the preceding section and since we found no evidence for systematic differences between calculations from cross wind and along wind flight legs, it is unlikely that under-sampled coherent boundary layer structures contributed significantly to momentum fluxes. However, it should be pointed out, that if indeed coherent structures do exist and these structures are indeed under-sampled, it is likely that they would be difficult to detect in our data. Thus the further investigation into the possible existence coherent structures is crucial for future studies.

Figure 4 shows the binned results from this study superimposed on results from earlier studies. The Large and Pond (1981) line is from eddy-correlation measurements up to 20 m s^{-1} and eddy dissipation measurements to 25 m s^{-1}. The Smith (1980) line is from eddy-correlation measurements to 22 m s^{-1}. Both data sets are from open-ocean long fetch conditions. In addition to our calculations at wind speeds from 20 to 32 m s^{-1}, we included the results of Powell et al. (2003) for log-profile fits for a 10-150 m surface layer and surface winds to 42 m s^{-1}. Results from Donelan et al. (2004) taken from wave-tank studies are also shown on the figure.

Results from this study are in general agreement with results from the earlier studies for wind speeds of 20 and 22.5 m s^{-1}. However, results from the more recent studies all begin to diverge at higher wind speeds. While Donelan et al. note a roll-off that begins around 32 m s^{-1}, it is not near as pronounced as the decrease noted in the Powell et al. results (at least not for wind speeds less than 50 m s^{-1}). Our measurements suggest a roll-off at even much lower wind speeds and at a smaller value of the drag coefficient. Unfortunately, data from this study were not collected at wind speeds much greater than 30 m s^{-1}, and thus we cannot speculate on the behavior of the drag coefficient at even greater wind speeds.

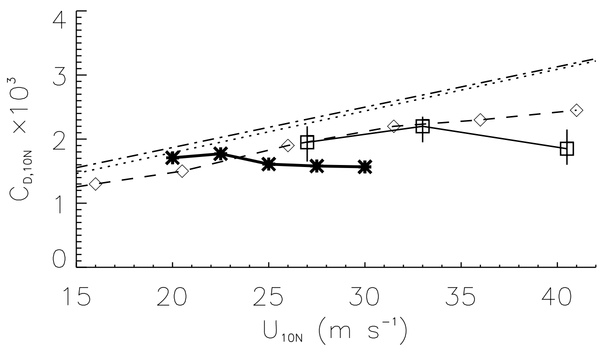

Figure 4: As in Figure 4 except only showing the binned values from this study. Also shown are extrapolations of results from Large and Pond (1981) and Smith (1980). The Diamonds represent results from Donelan et al. (2004) and the squares are taken from Powell et al. (2003).

5. Concluding Remarks

In this study the first-ever direct measurements of momentum flux within a hurricane boundary layer were presented. The measurements were made from an instrumented aircraft in rain-free regions of two hurricanes. Surface values for momentum flux and 10-meter drag coefficient were extrapolated from measurements made at altitudes between 70 and 383 m with near-surface wind speeds from 17 to 31 m s^{-1}.

For the lowest wind speeds in this study, up to 22.5 m s^{-1}, the results agree well with data from several earlier studies. For wind speeds greater than 22.5 m s^{-1}, calculations of drag coefficients are less than values inferred from studies conducted in wave tanks (Donelan et al. 2004) and using GPS drop-sonde winds assuming log profiles (Powell et al. 2003). It is clear that additional measurements are needed. Further studies should attempt to obtain measurements at wind speeds ranging from 20 to 45 m s^{-1}. One of CBLAST's primary objectives was to obtain measurements in regions of surface winds approaching 50 m s^{-1}. While we were not able to accomplish that goal, it is the opinion of the authors that obtaining measurements in winds at least up to 40 m s^{-1}, and perhaps higher, is possible and a concerted effort should be placed on obtaining such measurements.

Finally, future investigations should also focus on acquiring data in all storm quadrants and coupling the results to remotely sensed sea surface conditions. Again, this was also part of the overall objectives of CBLAST, but with the limited number of storms sampled, it was not fully achieved. There exists compelling evidence, at least in lighter winds, that suggest one may expect significant differences depending on storm quadrant and hence wind/wave/swell directional relationships.

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