

Relationship Between Oceanic Energy Fluxes and Surface Winds During Tropical Cyclone Passage

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Abstract

Based on velocity profiles acquired from the deployment of airborne expendable current profilers (AXCPs) during the passage of hurricane Gilbert, WKBJ-scaled vertical wavenumber spectra were estimated and compared to internal wave spectra. For the stratification, observed kinetic energies were typically up to 10-times larger than the background spectral values. ClockWise (CW) rotating spectral energy densities exceeded the CounterClockWise (CCW) rotating densities in 83% of the profiles by a factor of four. The downward vertical energy fluxes ranged from 0.5 to 10 $\text{ergs cm}^{-2} \text{s}^{-1}$, and the average value was $\sim 2 \text{ ergs cm}^{-2} \text{s}^{-1}$, equaling those found in oceanic fronts. These near-inertial energy fluxes followed a *tanh* dependence where the argument involves the ratio of the CW and CCW energies.

Amplitudes of the wave groups determined from scaled-velocity profiles across the top of the thermocline ranged between 15 to 50 cm s^{-1} . Equating 10-m winds to near-inertial energy fluxes suggested a nonlinear relationship for the drag coefficient, valid for winds up to $\sim 38 \text{ m s}^{-1}$. Comparisons of this function to various accepted drag coefficient formulations were favorable between wind speeds of 12 to 20 m s^{-1} . For wind speeds beyond 28 m s^{-1} , this function leveled off, compared to monotonically increasing wind-speed dependent drag coefficient formulations. Mixed layer current simulations from a high-resolution ocean model, forced with observed winds with varying surface drag coefficients, suggest reasonable agreement to those simulated from this analytical function. Simulated currents were within $\sim 20\%$ to those observed from AXCPs. Understanding the relationships associated with the surface momentum flux is important for air-sea coupling physics during strong wind conditions.

1. Introduction

Tropical cyclone passage over an ocean represents one of the more extreme upper ocean response cases, as the ocean loses heat to the atmosphere. The most apparent effect of tropical cyclone (TC) passage is the marked cooling of the sea surface temperatures (SSTs) of 1 to 5°C, which occurs to the right of the storm track, as shown for hurricane Juliette in the eastern Pacific Ocean (Figure 1). Although warm SSTs ($> 26^{\circ}\text{C}$) are required to maintain the TC, the *rightward bias* of the maximum SST cooling and mixed layer depth deepening by 20 to 40 m are primarily due to entrainment mixing of the cooler thermocline water with the warmer ocean mixed layer (OML) water, which accounts for about 65 to 80% of the observed SST decreases. These SST decreases are due to either vertical shear of the horizontal currents or surface-generated turbulence within the OML. Both mechanisms play important roles in the upper ocean thermodynamics and air-sea fluxes.

The ocean's momentum response is classified into two regimes: the directly-forced or near-field; and the evolving 3-dimensional wake or far-field (Shay [1]). In the near-field, the cyclonically-rotating wind field of a TC causes OML currents of about 1 to 1.5 m s⁻¹ to diverge from the storm track, starting within one-quarter of an inertial wavelength behind the eye, where the inertial wavelength (Λ) is defined as the product of the storm translation speed (U_h) and the local inertial period (IP). This divergence causes the upwelling of cooler water underneath the storm track, thereby decreasing the OML depth. Over the next one-half inertial cycle, OML currents converge towards the storm track, causing an increase in the OML depth as warmer water is downwelled into the thermocline. This alternating cycle of upwelling and downwelling of the isotherms (and isopycnals) occurs over distances of Λ and establishes horizontal pressure gradients that couple the OML to the thermocline as part of a spreading 3-dimensional wake.

The near-inertial response represents one of the more dominant components that have been observed and modeled during TC passage. Within this framework, near-inertial motions are characterized by large horizontal and vertical wavelengths (Geisler [2]) due to oscillations in the vertical current structure by low-order baroclinic modes (Gill [3], Kundu and Thompson [4], Shay et al. [5]). An important aspect of these forced motions has been the frequency shift above the local Coriolis frequency by about 5 to 10%. This shift is necessary for the near-inertial energy to vertically propagate from the surface-intensified OML into the thermocline. However, during strong forcing events, both the rate of vertical energy flux and the level where energy accumulates have remained unresolved due to the lack of observational evidence.

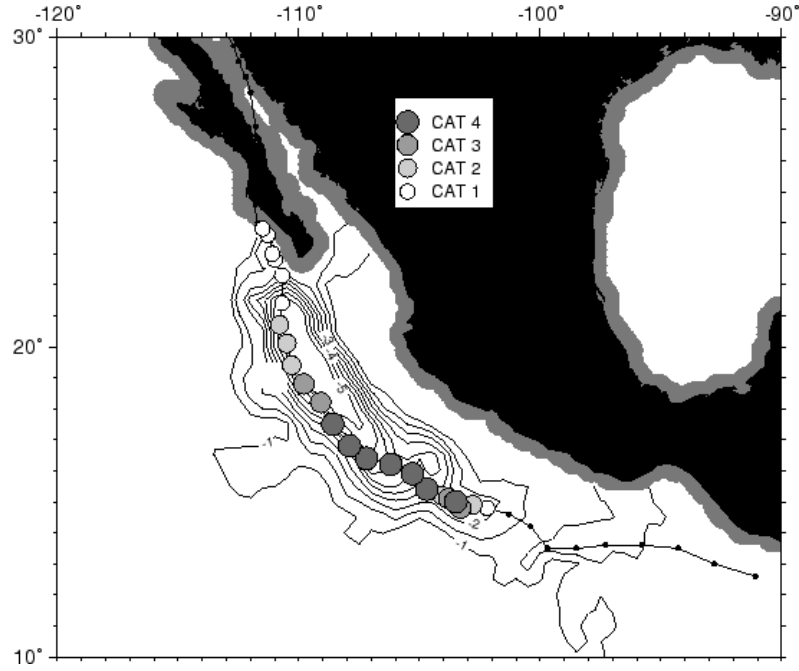


Figure 1: SST response to the passage of hurricane Juliette in Sept 2001 relative to the track (black curve) and storm intensity (see Legend). SST is derived from TRMM Microwave Imager (TMI), using a 75 km coastal ocean blanking.

From a series of Airborne eXpendable Current Profiles (AXCPs) successfully deployed in TCs Josephine (1984), Norbert (1984), and Gloria (1985), as part of the Ocean Response to A Hurricane Project (ORTAH) sponsored by a consortium of oil companies, high quality current profiles were obtained in a hurricane environment for the first time (Sanford et al. [6]). While these baroclinic current and temperature profiles were studied observationally (Sanford et al. [6]), analytically (Shay et al. [5]), and numerically (Price et al. [7]), the time-evolution of the spreading 3-dimensional near-inertial wave wake was not resolved with only one snapshot of measurements. Subsequently, Shay et al. [8,9] isolated the near-inertial response induced by hurricane Gilbert from AXCPs deployed before, during and after hurricane passage. Upon removal of the geostrophically balanced currents associated with a Loop Current Warm-Core Ring (LCWCR), velocity profiles were placed in a storm-based coordinate system where time represented along-track distance for a steadily-moving storm (Geisler [2]). Using least squares, these profiles were modeled to isolate the

near-inertial content and compared to forced, analytical simulations. A summation of the first four baroclinic modes described up to 77% of this near-inertial current variability over the first 1.7 Inertial Periods (IP $\sim$ 30 h) following Gilbert.

There is a characteristic time scale for the phase of each baroclinic mode to separate from the wind-forced OML currents when the wind stress scale ($\sim 2 R_{\max}$ defined as the radius of maximum winds) is greater than the deformation radius associated with the first baroclinic mode of (O(40 km) (Gill [3])). This time scale, which is required for a phase difference of $\pi/2$ to develop between the upper thermocline and the OML currents, increases with mode number due to decreasing phase speeds. These baroclinic time scales represent the onset of rapid energy propagation from the OML into the thermocline (Zervakis and Levine [10]). In a theoretical framework, an average energy flux from moving TCs to the internal wave field is found to have a global value of 10^{10} W by summing forced normal modes (Nilsson [11]). However, estimating the energy fluxes has been more problematic in strong TCs due to the lack of three-dimensional current profiles in the near-inertial wave wake. Moreover, the relationship of these fluxes to the near-surface winds has been even more difficult due to the lack of synopticity in concurrent oceanic current and atmospheric wind measurements over storm scales.

An important question emerging from studies of a strongly forced OML is the relationship between the internal wave flux and surface wind stress. Under storm conditions, considerable uncertainty remains with respect to the surface drag coefficient used to estimate the wind stress. For example, Large et al. [12] found a monotonically increasing drag coefficient, valid in winds to 25 m s^{-1} . For the response to hurricanes Norbert and Josephine, simulated currents agreed well with observed mixed layer currents from Price et al. [7] based on Large et al. [12] drag coefficient extrapolated to higher winds. Buoy measurements in hurricanes Anita, Camille and Frederic suggest a slightly higher bias in the drag coefficient, but with the same slope up to 45 m s^{-1} (WAMDI Group [13]). However, the wave-age dependent drag coefficient formulation of Donelan [14] indicates considerable variability in the surface drag between 10 to 20 m s^{-1} compared to these wind-speed dependent drag coefficients. While direct methods have been used to estimate the surface drag coefficient, Smith et al. [15] suggested using inverse methods and constraints to determine the exchange coefficients associated with heat and momentum fluxes. This approach may be particularly important at very high winds and sea states where direct measurements at the air-sea interface are difficult to acquire in TCs. Thus, one possibility is to use the ocean's momentum response as a tracer of surface wind stress, and hence, determine the drag coefficient from observations during TCs.

Accordingly, a functional dependence between the observed surface winds and wind-forced internal wave fluxes is examined here using the current profiles for hurricane Gilbert, estimated by Shay et al. [9] in a first attempt to use this approach.

The resulting functional relationship as well as several accepted drag coefficient methods (Large and Pond [12], WAMDI Group [13], Smith [16]) are then used to simulate OML currents from the Hybrid Coordinate Ocean Model (HYCOM) to assess differences in the observed and simulated OML profiles (Bleck [17]).

As shown in Figure 2, these measurements included seventy-six AXCPs deployed prior, during, and subsequent to the passage of hurricane Gilbert from 14 to 19 Sept 1988 (Shay et al. [8]). Given the predominance of the near-inertial wave excitation by the wind field, vertical wavenumber spectra are estimated for WKBJ-scaled profiles and compared to estimates from the internal wave spectra (Garrett and Munk [18]). Kinetic energy spectral densities are decomposed into ClockWise (CW)- and CounterClockWise (CCW)-rotating components to estimate the rates of downward energy flux. That is, excess spectral energy densities (i.e. CW-CCW rotating components) are multiplied by the group velocity and integrated over vertical wavenumber space following Leaman [19] to determine the internal wave flux in the 3-dimensional wake structure. These internal wave fluxes are equated to the observed surface (10-m) winds to derive a functional relationship. These results are placed into a framework of the surface drag coefficients found from previous studies (Large and Pond [12], WAMDI Group [13], Donelan [14], Smith [16], Alamaro et al. [20], Powell et al. [21]). The implications of this approach are summarized and similarities to characteristics of well-established drag coefficients are noted.

2. Observations

This section focuses on analysis of measurements made during hurricane Gilbert (1988) as it propagated across the Gulf of Mexico. Thus, we provide surface winds, air-sea variables and velocity profiler data in estimating the kinetic energy fluxes into the thermocline from the wind-forced OML. Measurements and profiler analyses have been described elsewhere (Shay et al. [8,9]).

2.1 Wind Structure

Flight-level winds were reduced to 10-m using a boundary layer model of Powell [22] assuming a translating wind field and objectively analyzed at 8 km intervals (Figure 3). Maximum winds were ~ 38 and 34 m s^{-1} at radii of 60 km (R_{max}) and 90 km from the storm track, respectively. The wind field was broad and complex where winds exceeded 15 m s^{-1} as far as $5 R_{\text{max}}$ to the right of the track over a LCWCR in the Gulf of Mexico.

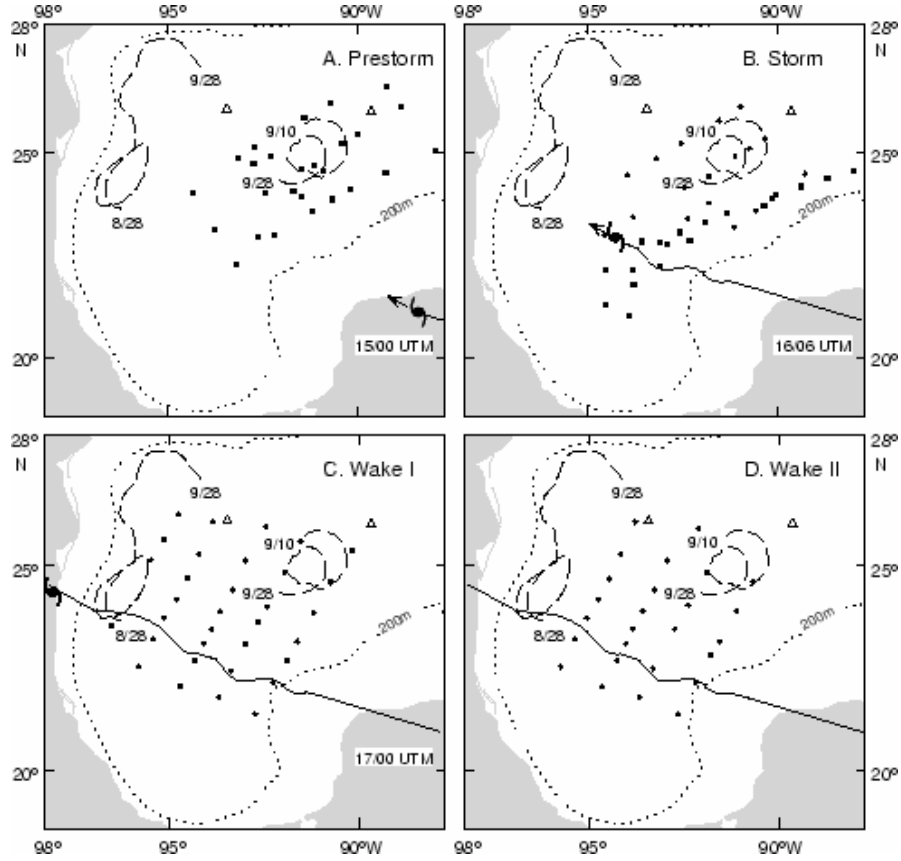


Figure 2: Locations of AXCPs (circles) and AXBTs (boxes), tracks of drifting buoys (dashed), National Data Buoy Office buoys (triangles) and Minerals Management Service moorings EE, FF, and GG (open circles in panel a) relative to the track of hurricane Gilbert (solid) in the Gulf of Mexico from (a) 14 Sept, (b) 16 Sept (c) 17 Sept and (d) 19 Sept 1988. The 200 m bottom contour is dotted (from Shay et al. [8]).

This strong wind forcing field shifts the frequency of forced near-inertial motions by a few percent above the local Coriolis frequency (Price [23]). This frequency shift is proportional to the OML Burger number, which is a function of the storm speed (U_h), R_{max} , the local Coriolis parameter (f) and stratification beneath the OML. In the present case, the Burger number is 0.08. Based on scaling arguments, the predicted

blue-shift in the near-inertial frequency is $1.04f$, which allows energy to propagate vertically from the OML into the thermocline (Shay et al. [9]). The predicted wind-driven ocean velocity (V_{is}) scale, defined as

$$\tau R_{\max} (\rho_o h U_h)^{-1},$$

is $\sim 1.1 \text{ m s}^{-1}$. This is based upon a wind stress (τ) using an extrapolated drag coefficient from Large et al. [12], where ρ_o is the reference ocean density, and h is the initial OML depth (40 m). The corresponding thermocline velocity ($V_{th}=44 \text{ cm s}^{-1}$) is approximately twice the scaled velocity of 22 cm s^{-1} , as estimated from $V_{is} h b^{-1}$ where b is a seasonal thermocline thickness of about 200 m suggested by observed velocity and temperature profiles.

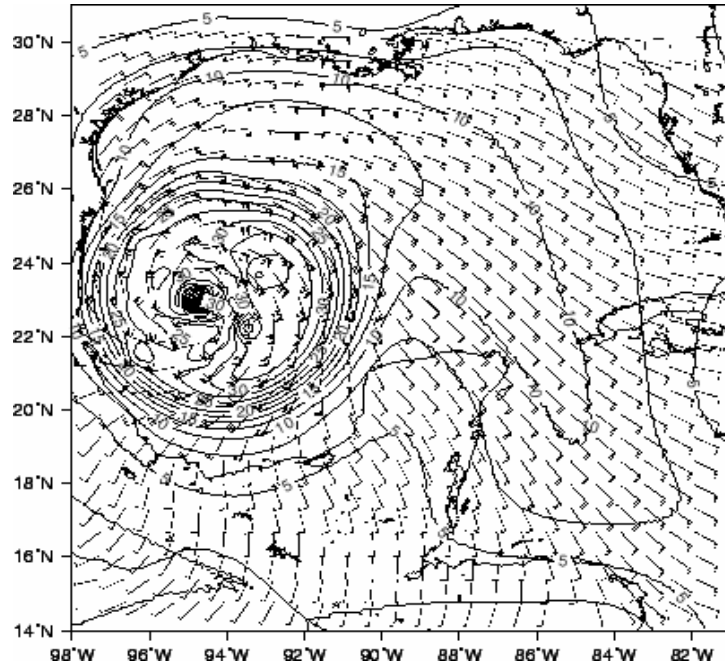


Figure 3: Surface winds (10-m) derived by combining winds reduced from flight-level, ECMWF surface analyses and buoy observations for 06UTC 16 Sep 1988 observed during hurricane Gilbert. Analyzed wind data are subsampled every eighth data point as wind barbs (in knots) with isotachs representing wind magnitudes (m s^{-1}) (from Jacob et al. [24]).

2.2 Velocity Profiles

Since TC Gilbert moved steadily at 290°T and at a nearly constant speed, along-track distance can be converted into time (Geisler [2]), and velocity profiles deployed at different times have an along-track position in storm coordinates relative to Gilbert's eye ($22^\circ 57'\text{N}$, $94^\circ 44'\text{W}$). In this storm-based coordinate system, drop sites are a function of the along-track coordinate (Λ), or inertial period (IP), and the cross-track coordinate R_{max} .

Along-track velocity sections indicate a CW-rotation of the current vector relative to the storm profile, particularly in the OML (Shay et al. [9]). The maximum current response (1.3 m s^{-1}) normalized to the predicted wind-driven current was about 1.1 located at about $2 R_{\text{max}}$ which agrees with the rightward bias in the maximum response. This behavior is consistent with strong near-inertial currents that develop within the thermocline following Gilbert's passage. That is, a large fraction of the variability is associated with forced near-inertial motions and the first few baroclinic modes (i.e. low-mode response). After removal of the geostrophic current and surface wave-induced orbital velocities (Sanford et al. [6]), a large fraction of the current variability was associated with the near-inertial current response, and the CW-rotation of the velocity vector with depth.

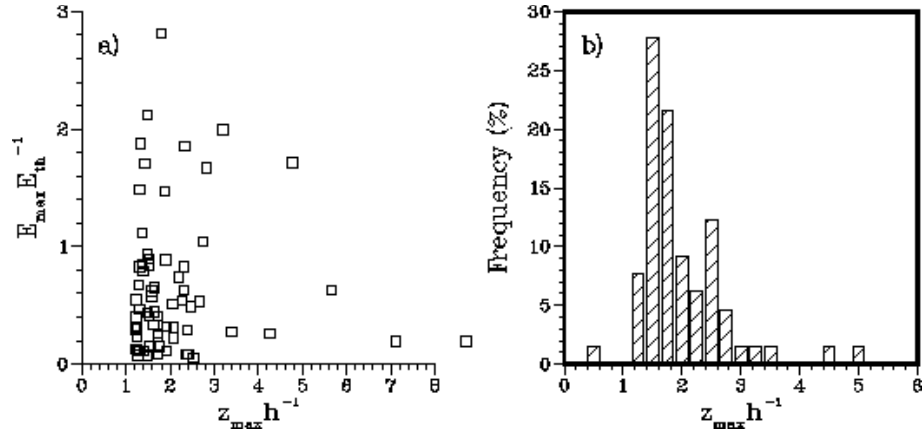


Figure 4: a) Scatterplot and b) histogram of levels (m) of maximum observed WKB-scaled kinetic energy scaled by the predicted thermocline kinetic energy (E_t) relative to the respective depth, normalized by the observed mixed layer depths (h) at each of the current profiles for Gilbert. Note that V_{th} is twice the scaled thermocline velocity, using scaling arguments from Price [23].

2.3 Maximum Energy Levels

Maximum energy levels for the WKBJ-scaled current profiles, for Gilbert, normalized by the mixed layer depths (h) at each AXCP drop site, are shown in Figure 4. Levels of energy maxima were located from the OML base (35 to 55 m) to about 150 m. Approximately 95% of the velocity profiles indicated maximum energy levels ($E_{\max} \propto E_{th}^{-1}$, where E_{th} is estimated from the scaled thermocline velocity of 44 cm s^{-1}) from 1 to 4 h where the average depth was 2 h , and the standard deviation was 0.8 h . The distribution tends to follow a χ^2 distribution starting at about one layer depth (h). Since OML temperatures showed regions of upwelling (Storm and Wake 1- see Figure 2) and downwelling (Wake 2), these depths of maximum WKB-scaled energy were representative of an energetic thermocline response to the applied atmospheric forcing. Maximum current shears coincided with the region of the largest buoyancy frequencies to 12 cph [cycles per hour], suggestive of strongly forced, near-inertial motions. These near-inertial motions reduced the bulk Richardson number to below criticality, which induced vertical mixing events across the entrainment zone (Shay et al. [8]). Using a diagnostic approach, Jacob et al. [24] found that shear-induced mixing events dominated the OML heat balance in the wake of Gilbert between 1 to 3 $R_{\max}$ where the maximum observed OML temperature response was $\sim 4^\circ\text{C}$.

3. Vertical Wavenumber Spectra:

Velocity profiles were stretched, normalized by the buoyancy frequency and interpolated to 3-m vertical resolution following Leaman and Sanford [25]. While the vertical wavelengths of the forced motions were large, the WKBJ-scaling allows a direct comparison to the internal wave spectra (Garrett and Munk [18]) as well as estimation of the vertical energy fluxes from the profiler data. That is, OML velocity components and energies are precluded from this calculation since the buoyancy frequency (N) is zero in a well-mixed surface layer. This approach to scaling the velocity profiles provides a more realistic estimate of the energies and fluxes penetrating into the thermocline, then would be obtained retaining OML currents.

3.1 Garrett-Munk Comparisons

Following Leaman [19], the form of spectra is

$$KE(m) = N_o^2 M_o^{-3} 3 E_o (2j^* \pi)^{-1} A(\lambda_v), \quad (1)$$

where $A(\lambda_v) = (t-1)(1 + \lambda_v)^{-1}$, $\lambda_v = m (j^* \pi M_o)^{-1}$, $M_o = 0.122 \text{ cycles km}^{-1}$, $j^* = 6, 15$, and 20 , m is vertical wavenumber, $E_o = 2 \pi \times 10^{-5}$ and $t = 2.5$. Within a factor of 2 to 3, this nondimensional energy density is a universal factor for a background internal wave continuum in the ocean, based on the thermal structure (Munk [26]). However, average values of the normalized spectral energies observed during Gilbert were 23, 11 and 9% for $j^* = 6, 15$ and 20 , respectively. Thus, the energies excited by Gilbert were significantly above the spectral levels of eqn (1), for the common values of j^* which represent larger wavelengths supported by the stratification and the low-baroclinic mode response.

j^*	$E_{iw}E_o^{-1}(\%)$	$u_{iw} \text{ (cm s}^{-1}\text{)}$
6	23	8
10	11	4
20	9	3

Table 1: Summary of comparisons between the theoretical values from the internal wave spectra (E_{iw}), normalized by the WKBJ-scaled observed energies (E_o), for a thermocline thickness of 1300 m in the western Gulf of Mexico as a function of j^* . The amplitude of the wave group from all of the profilers was $\sim 34 \text{ cm s}^{-1}$ with a standard deviation of 12 cm s^{-1} .

3.2 Rotary Spectra

Rotary vertical wavenumber spectra (Gonella [27]) of the WKBJ-scaled kinetic energy spectral densities were estimated by smoothing the 3-m velocity profiles using a Tukey window, transforming the data, and spectrally-averaging to decrease leakage to adjacent wavenumber bands (Otnes and Enochson [28]). After 1.5 IP following passage of Gilbert, the cross-track spectral structure encompassed the Gulf common water and the LCWCR (Figure 5). Between $-R_{\max}$ to $2 R_{\max}$, there was a predominance of CW-rotating motions at the low wavenumbers including the transition zone in the LCWCR regime. The transition between vertical wavenumbers became more apparent at vertical wavelengths of about 100 m. In the cross-track direction, the CW-rotating spectral peak at 500 m wavelength ($7 \times 10^{-3} \text{ cycles m}^{-1}$) was shifted to about 300 m wavelength ($10^{-2} \text{ cycles m}^{-1}$) at $1 R_{\max}$, and to 100 to 200 m wavelength ($1.5 \text{ to } 3 \times 10^{-2} \text{ cycles m}^{-1}$) at $2 R_{\max}$, but was not as clearly evident at $3 R_{\max}$. Presumably, the distinct change, even in the higher vertical wavenumbers, may have been due to this transition zone between Gulf common water and LCWCR F. Between 4 and $5 R_{\max}$, a distinguishable peak in the spectral energy densities was shifted to a vertical wavelength of 100 m. At low vertical wavenumbers (wavelengths

of 500 to 2000 m), the WKB-scaled, CW-rotating energy dominated the CCW-rotating components. The presence of shorter wavelengths from 50 to 500 m indicated that energy input from the wind field excited a spectrum of wave modes. This is not a surprising result given the predominance of the energies in the larger vertical wavelength oscillations (low wavenumbers). This is known as the low-mode dominance of the oceanic current response excited by TCs (Price et al. [7], Shay et al. [9]). Similarly, spectral energy densities at $2 R_{\max}$ in the along-track direction indicate a cascade of energy from low to high vertical wavenumbers (not shown).

3.3 Energy Partitioning

WKB-scaled rotary spectral energies were integrated over vertical wavenumber space to estimate the partitioning of the CW and CCW energy densities during Gilbert. Of all the velocity profiles, sixty-three (83%) revealed a predominance of the CW-rotating energy (Figure 6). Energy ratios from 0 to 1 indicated that 28% of the profiles had energy ratios from 0 to 1, including the 17% where CCW-rotating energy was dominant plus the profiles where this ratio was close to unity. The average ratio of the CW to CCW-rotating components was 3.6, indicative of a preference for downward energy propagation from the wind-forced OML into the thermocline. Correlation levels between the CW and CCW components exceeded 0.74 for the scaled profiles, where the regression slope was about 1.5 and the bias was 15 ergs cm^{-3} . A weighted rate of turning of the velocity vector (direction) with the scaled energies was also used to determine the directionality of the downward energy flux from strong forcing events (D'Asaro [29]). Conceptually, both approaches are analogous to a double-helix structure in the vertical with the stronger CW-rotating strand dominating the CCW-rotating strand, corresponding to downward energy propagation and upward phase propagation (Leaman [19]).

3.4 Energy Fluxes

Excess spectral energy densities were multiplied by the vertical group velocity for each wavenumber contained within the spectra to provide an estimate of the vertical energy flux from the expression

$$F_{iw} = \int_{2\pi/D}^{2\pi/6} \Delta E(m) c_{gv}(m) dm, \quad (2)$$

where D represents the profile depth less the start depth of the profiles (nominally 10 m), $\Delta E(m)$ represents the difference between the CW and CCW energies (CW(m)-CCW(m)), and c_{gv} is the vertical group velocity. Based on the vertical structure of

internal waves with an e^{imz} dependence, Rossby and Sanford [30] determined the vertical group velocity to be

$$c_{gv} \cong -2f \varepsilon \text{ m}^{-1}, \quad (3)$$

where ε represents departures from the local Coriolis parameter (f) and m is the vertical wavenumber. The departure of the observed frequency from f is ~ 0.04 , as described above. For the largest resolvable vertical wavelength of 2000 m from the profiles, c_{gv} ranged from 0.1 to 0.2 cm s^{-1} , which agreed with previous estimates from moored observations (Brooks [31], Shay and Elsberry [32]) and simulations (Price [23]).

The product of the excess CW energy and vertical group velocity was integrated across the wavenumber spectrum, following eqn (2). For the sixty-three CW energy-dominated profiles (Figure 7), downward energy fluxes ranged from 1 to 8 $\text{ergs cm}^{-2} \text{ s}^{-1}$, with an average value of 1.6 $\text{ergs cm}^{-2} \text{ s}^{-1}$. This flux estimate was 5 to 10 times larger than the flux found by Leaman [19], but was similar to the values of 2 to 4 $\text{ergs cm}^{-2} \text{ s}^{-1}$ obtained in the North Pacific Subtropical Front experiment (Kunze and Sanford [33]). This result suggests that the TC wind field excited as much internal wave energy in the thermocline as that found in the vicinity of oceanic fronts.

From a global perspective, Nilsson [11] estimated the near-inertial wave power for TC's translating at 7 m s^{-1} to be $\sim 2.6 \times 10^{10} \text{ W}$. If a rate of $2 \text{ ergs cm}^{-2} \text{ s}^{-1}$ is used as found here, the near-inertial wave power is estimated to be about $7.4 \times 10^{11} \text{ W}$, or roughly 30 times this theoretical estimate. Similar estimates of wave power have been found in the conversion of barotropic to internal tides of $3 \times 10^{11} \text{ W}$ (Bell [34]), and 10^{12} W (Sjoberg [35]). The estimated energy flux from TC force winds found here represents a fairly realistic overall value based on these comparisons. Regression analyses between the energy fluxes and the ratio of the scaled energies indicated a hyperbolic tangent relation between these two regimes, of the form:

$$F_{iw} = 0.2 + 1.4 \tanh \{ (CW \times CCW^{-1}) - 1.2 \}, \quad (4)$$

where a linear relationship exists between energy flux and the CW/CCW Energy Ratio (ER). The first regime occurs between ER values of 0 to 4, and corresponds to an increase in the energy flux for the larger ER values. Beyond ER values of 4.5, a second regime consists of a nearly constant flux region on average, where the hyperbolic tangent function approaches a value of $1.6 \text{ ergs cm}^{-2} \text{ s}^{-1}$. The correlation coefficient between energy flux and ER is about 0.7. Note that the ER range from 3 to 4 agrees well with studies on the internal wave spectrum, indicating a preference for CW rotation of the velocity vector and downward energy propagation for surface-intensified flows.

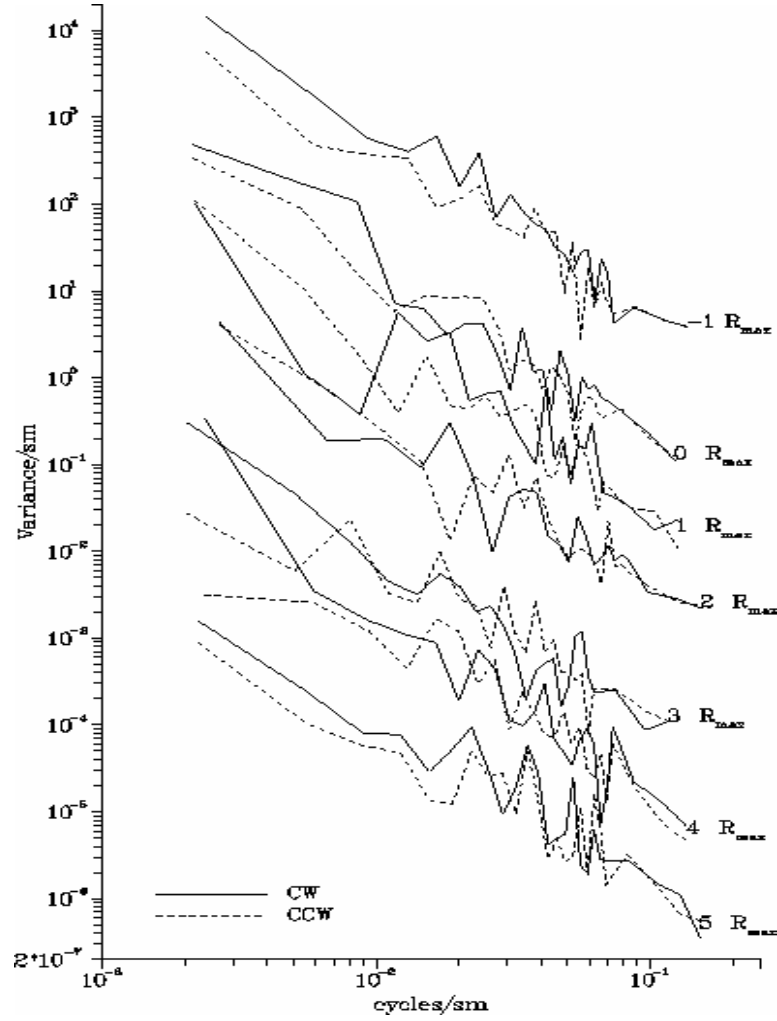


Figure 5: WKB-scaled rotary vertical wavenumber spectra ($\text{cm}^2 \text{s}^{-2} \text{sm}^{-1}$) in the cross-track direction at 1.5 IP following Gilbert's passage beginning at $-R_{\text{max}}$ and extending to $5R_{\text{max}}$ for the anticyclonically (CW: solid) and cyclonically rotating (CCW: dashed) energies. Spectral energy levels from each consecutive profile are scaled downward by a decade.

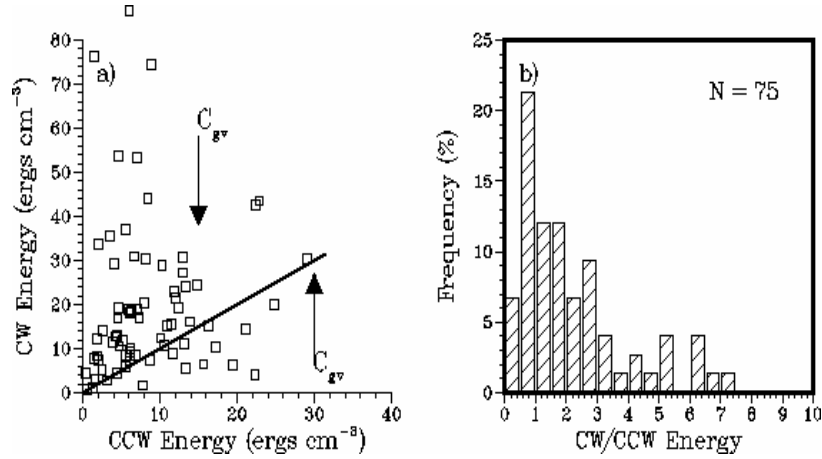


Figure 6: a) Scatterplot (left panel) of the integrated CW and CCW rotating spectral energy densities (ergs cm^{-3}), and b) histogram (right panel) of the ratio of CW (m) to CCW (m) spectral energy densities derived from the scaled wavenumber spectra from 75 AXCPs, deployed for Gilbert. The line $\text{CW}=\text{CCW}$ separates the two counter-rotating regimes. Arrows depict downward (upward) propagating wave group speeds for the CW (CCW) rotating domains, respectively.

4. Relationship to Forcing Structure

4.1 Approach

Following Kunze and Sanford [33], we assume that the near-inertial energy flux based on eqn (2) is due to surface wind in the following form:

$$F_{\text{iw}} \cong V_o \rho_a c_d U_{10}^2, \quad (5)$$

where V_o is the scaled wave group amplitude, ρ_a is the air density, c_d is the surface drag coefficient, and U_{10} is the 10-m wind. Amplitudes of the wave groups were estimated from the near-inertial profiles at the top of the thermocline where the largest energies were found. Scaled wind-driven current amplitudes ranged from 15 to 50 cm s^{-1} , using only the profiles for CW-rotating energies associated with downward energy propagation (Figure 8). The average amplitude of the wave group was 34 cm s^{-1} with a standard deviation of 12 cm s^{-1} . By contrast, the background internal wave amplitudes ranged between 3 to 8 cm s^{-1} for the range of j_* values

shown in Table 1. Moreover, similar wave group amplitudes of 30 cm s^{-1} were found in moored thermocline current data observed during the passage of TC's Allen (Brooks [31]) and Frederic (Shay and Elsberry. [32]) where the response persisted for at least 4 weeks following storm passage. Notice that the scatter between the internal wave energy fluxes and these amplitudes in Figure 8 indicates a direct relationship. That is, larger energy fluxes were generally associated with larger wave amplitudes, consistent with theoretical considerations. These fluxes and the wave amplitudes were determined from the velocity profiles.

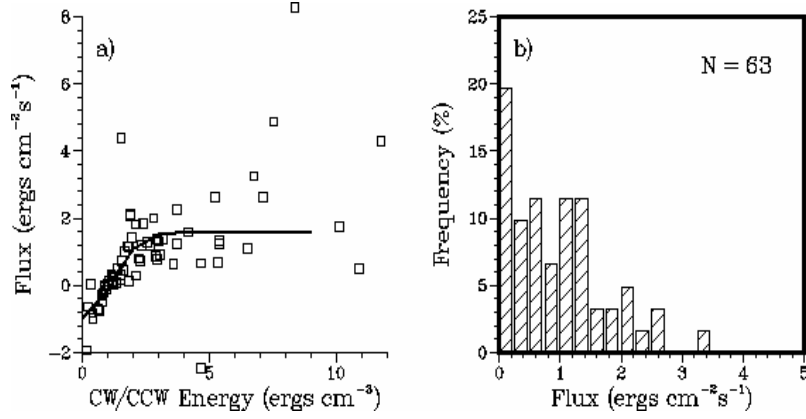


Figure 7: a) Scatterplot (left panel) of the energy ratio (ER) of the anticyclonically (CW) and cyclonically (CCW) rotating spectral energy densities and internal wave energy fluxes ($\text{ergs cm}^{-2} \text{s}^{-1}$), and b) histogram (right panel) for the scaled current profiles from 63 AXCPS. The hyperbolic tangent dependence is shown as a solid curve. Notice two distinct regimes in the energy flux scatter, between 0-5 ER and beyond 5 ER, which are accentuated for scaled profiles.

As per eqn (5), these quantities are related to the surface wind stress. Since the winds were observed (as shown in Figure 3), the air density is taken as $1.16 \times 10^{-3} \text{ gm cm}^{-3}$. The amplitudes of the wave groups used the average value of 34 cm s^{-1} for fluxes between 0.5 and $2 \text{ ergs cm}^{-2} \text{s}^{-1}$ and for V_o below and above these fluxes, values of 22 and 46 cm s^{-1} , respectively (ie one standard deviation) were used in this analysis. The remaining unknown is the drag coefficient. The scatter in wind speed squared and the fluxes normalized by the amplitude of the wave group is shown

below in Figure 9. At wind speeds from 10 to 20 m s⁻¹, there is a significant increase in the fluxes. The inflection point, or change in the slope occurs at about 28 m s⁻¹. This is particularly interesting since recent tank experiments by Alamaro et al. [20] suggest a change in the slope at about this wind speed. Moreover, as winds increased beyond 28 m s⁻¹, normalized internal wave fluxes approach a relatively constant value, in an average sense. At very high winds, the scatter becomes much larger in these normalized fluxes. Thus, caution is needed, given the atmospheric variability in the density structure as well as the oceanic variability associated with the LCWCR, where significant exchanges of heat occur because of deep, warm layers (Shay et al. [36]).

The fit between the internal wave flux normalized by the air density and the wave group amplitudes from eqn (5) gives a hyperbolic tangent dependence in Figure 7a

$$\mathfrak{Z}(U_{10}) = 3.755 \times 10^{-3} + \tanh(-0.295 \bullet U_{10}^{-2}), \quad (6)$$

valid for observed winds from 10 to 38 m s⁻¹. This nonlinear functional dependence suggests that beyond a wind speed of 28 m s⁻¹, internal wave fluxes remain relatively constant, in contrast to the large variations occurring for winds between 10 to 20 m s⁻¹. This result also suggests that there may be a limit to the amount of energy input into the thermocline ocean current spectrum during TC passage.

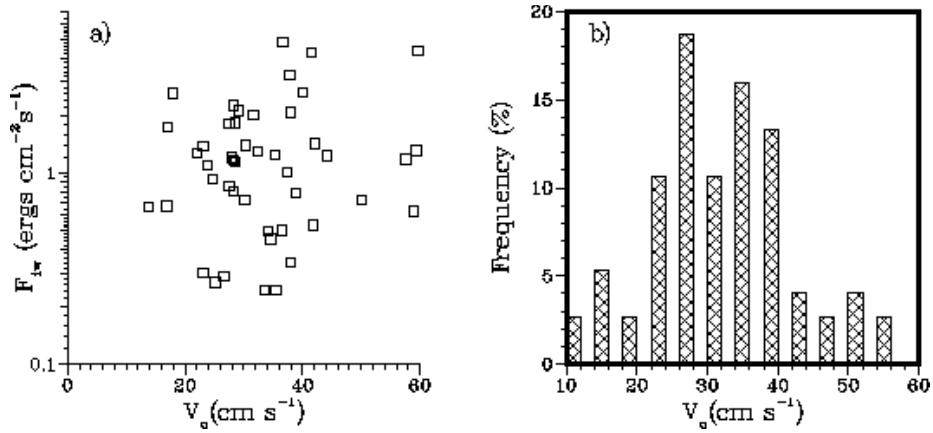


Figure 8: a) Internal wave flux (F_{iw}) versus amplitude of the WKB-scaled wave group (V_o), b) histogram of V_o , for hurricane Gilbert. F_{iw} is plotted on log scale for clearer depiction of the process and analyses are confined to the region of 0~5 ER in Figure 7 for the scaled velocity profiles.

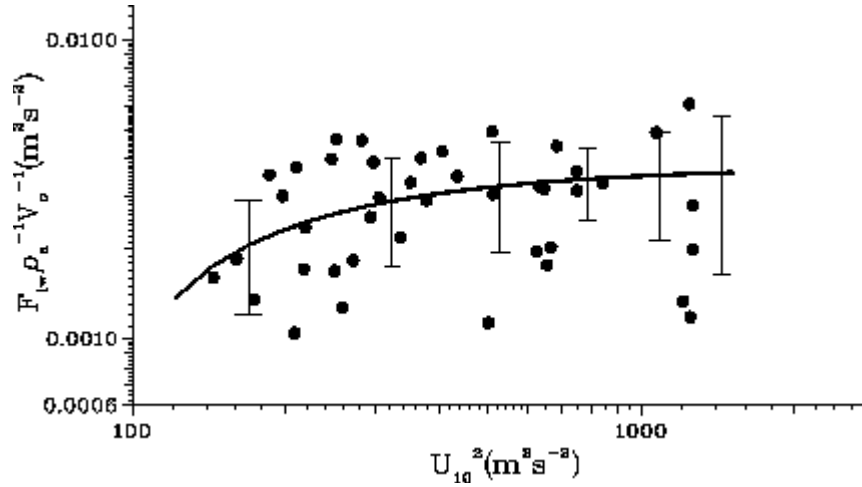


Figure 9: Scatterplot of WKBJ-scaled internal wave flux $F_{iw} (V_o \rho_a)^{-1}$ versus observed U_{10}^2 including the best line fit (solid) and binned standard deviations relative to the regression curve. Analyses are confined for values between 1×10^{-3} and 8×10^{-3} based on wave fluxes and scaled velocity profiles.

4.2 Comparisons

From a bulk aerodynamic perspective, Large and Pond [12], the WAMDI group [13], and Smith [16] determined various wind-speed dependent drag coefficients for moderate to strong winds. These drag coefficients ($\times 10^{-3}$) have the following form:

$$c_d = a + b \times U_{10}, \quad (7)$$

where a is the bias and b is the slope of the regression curve, times the U_{10} . For Large and Pond [12], $a = 0.49$ and $b = 0.065$, for WAMDI [13] $a = 0.80$, and $b = 0.065$, and for Smith [16], $a = 0.61$, and $b = 0.063$. The Large and Pond [12] formulation is valid over the range from 11 to 25 m s^{-1} whereas the WAMDI [13] formulation has been applied to TCs based on observed and simulated significant wave heights from hurricanes Camille, Frederic and Anita. The Smith [16] formulation is valid up to 22 m s^{-1} . For a 40 m s^{-1} near-surface wind, drag coefficients range from $3.1 - 3.5 \times 10^{-3}$ resulting in a 10 to 15% uncertainty in the stress, *plus* errors in U_{10} based on a reduction of the flight-level winds to 10-m (Powell [22]). In many modeling studies,

the underlying assumption is that one can extrapolate the drag coefficients to increase *ad-infinitum* - well beyond the range of data used in their formulation. Based on observed wind-driven currents from several TCs, there certainly appears to be a limit in the forced OML currents. Recently Powell et al. [21] have suggested a leveling off of c_d at a particular threshold wind speed, whereas Alamaro et al. [20] have suggested that c_d decreases with increased wind speeds beyond a threshold U_{10} .

To examine the functional dependence of eqn (6), a comparison is made to the drag coefficient formulations based on wind speed (Large and Pond [12], the WAMDI Group [13], and Smith [16]) and those based on wave-age (Donelan [14]). For example, Alamaro et al. [20] constructed an annular wind-wave flume and used it to estimate surface drag coefficients and the air-sea enthalpy fluxes for equivalent 10-m wind

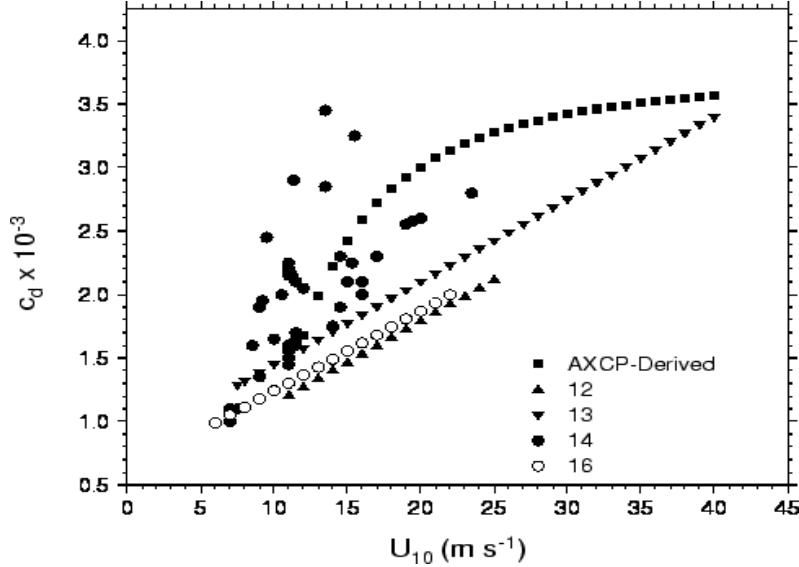


Figure 10: Relationship between the AXCP-derived analytical function (squares) in eqn (6) compared to the wind speed-dependent (Large and Pond [12], WAMDI [13], Smith [16]) and wave age-dependent (Donelan [14]) drag coefficients as per the legend.

speeds up to 55 m s^{-1} . They found that the drag coefficient increases with wind speed but levels off at wind speeds between 26 to 30 m s^{-1} , at $c_d \sim 3.8 \times 10^{-3}$. As shown in Figure 10, this analytical function is essentially the same as that of Large and Pond [12] at low-wind speed. One of the interesting aspects of hyperbolic tangent

dependence of eqn (6) is its rate of increase from 0.002 to 0.003 compared to c_d estimates by Large and Pond [12], in the 10 to 20 m s^{-1} range. Moreover, there is marked agreement with the wave age-based c_d estimates of Donelan [14], for this U_{10} range. In the present formulation, orbital velocities associated with surface waves have been removed from the profiles following Sanford et al. [6], and the residual surface wave-induced currents in the OML have not been included in the analysis as per WKBJ-scaling arguments. In the 20 to 30 m s^{-1} range, the function approaches a threshold where the slope of the curve approaches zero at about 28 m s^{-1} , and the limit approaches ≈ 0.0035 at higher wind speeds. Up to 25 m s^{-1} in this leveling off region, the Large and Pond [12] c_d monotonically increases with wind speed as does the WAMDI [13] c_d , although at a slightly faster rate of increase. At a wind speed of 40 m s^{-1} , the value of this analytical function would be nearly equal to that of the WAMDI formulation.

In recent experiments, Global Positioning System (GPS) dropsondes were deployed from research aircraft to measure the quasi-Lagrangian wind profiles in the atmospheric boundary layer in TCs. A logarithmic variation of mean wind speed was found in the lowest 200 m, to a maximum of 500 m, with a gradual weakening with height to 3 km (Powell et al. [21]). In these estimates, surface stress, roughness length and neutral stability drag coefficient determined by the profile method also suggest a leveling of the surface momentum flux, and as winds increase above hurricane force, there is with a slight decrease of the drag coefficient with increasing winds.

There are parallels that can be drawn from the analytical function (eqn 6) based on internal wave fluxes and drag coefficients derived from more conventional approaches. For instance, similar drag coefficient trends were inferred from a model study of storm surges in the North Sea (Mastenbroek et al. [37]) as found by a direct evaluation (Amorocho and DeVries [38]) over a range of wind speeds from 0 to 40 m s^{-1} . The nonlinear functional dependence suggests that wind stress increase is primarily due to winds beyond 30 m s^{-1} . Yet, considerable uncertainty remains at these very high-wind speeds, and caution has to be applied in estimating surface wind stress from any drag coefficients and the analytical function found above (eqn 6). Perhaps other factors such as the directionality of the wind, stress and wave fields need to be accounted for in a hurricane environment as suggested by Geernaert [38]. In a TC environment, however, as the surface wave field is not fully developed, it may lead to more uncertainties in the momentum transfer between the two fluids.

5. OML Current Simulations

Variability in the upper ocean momentum response due to stresses estimated from different drag coefficient formulations is investigated using the primitive equation Hybrid Coordinate Ocean Model (HYCOM, Bleck [17]). The simulated current

structure is compared to the AXCP profiles for the four wind stress fields to quantify OML current differences due to these differing formulations.

5.1 Model Configuration

HYCOM uses a hybrid coordinate system in the vertical that is designed to transition from z-levels and isopycnic coordinates in the ocean interior, and to sigma coordinates in shallow water regions. The model used in this study is implemented for a domain extending from 14° to 31° N and from 80° to 98° W with a horizontal resolution of 0.07° . The model has 250×242 horizontal grid points and 50 levels in the vertical. The model

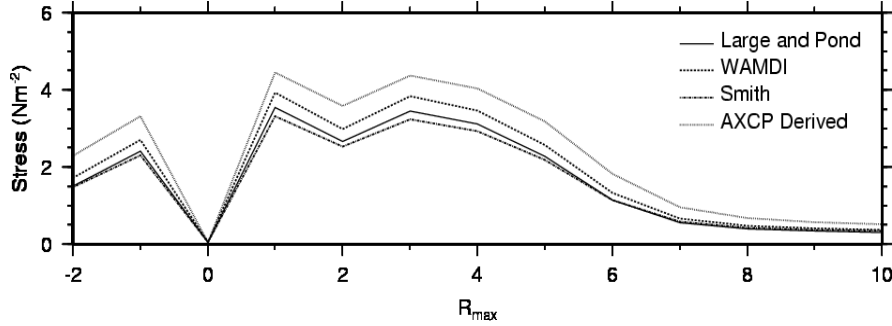


Figure 11: Cross-section ($R_{\max}$) of estimated surface wind stresses (N m^{-2}) from Large and Pond [12], WAMDI [13], Smith [16] c_d drag coefficient parameterizations and that inferred from the AXCP-derived analytical function (eqn 6) estimated using the wind field in Figure 3.

is initialized with uniform stratification throughout the domain and therefore the pre-storm fields remain quiescent. This contrasts the modeling results of Jacob and Shay [41], where the initial ocean state was not at rest. With a minimum vertical spacing of 3 m, the upper 30 model levels remain in the z-coordinate system. The model is configured with the K-profile vertical mixing parameterization of Large et al. [40]. The boundaries along the Florida Straits and the Caribbean Sea are closed by vertical sidewalls. The model is integrated for six days using the stresses derived from realistic wind fields covering Storm, Wake 1 and Wake 2 observational snapshots (Figure 2).

Simulated OML currents are defined as a vertical integral from the surface to the depth of the layer. The OML depth is defined as the level where the temperature changes by more than 0.2°C , which is consistent with the thermistors on the AXCP

profilers (Shay et al. [5]). While Jacob and Shay [41] showed that determining accurate OML depths from an isopycnic model is difficult, the HYCOM model circumvents this issue with z-coordinates in the OML. This provides a closer depiction of the observed currents and densities within the layer.

5.2 Surface Stresses

Bulk drag coefficient formulations from Large and Pond [12], WAMDI [13], Smith [16] and the AXCP-derived function (eqn 6) are used to estimate wind stresses for forcing the ocean model. With a maximum wind speed of $\sim 38 \text{ m s}^{-1}$ associated with Gilbert, the drag coefficients are extrapolated as necessary. For the AXCP-derived function (eqn 6), a constant coefficient of 1.14×10^{-3} is used for wind speeds less than 10 m s^{-1} . Estimated stresses for winds in Figure 3 suggest a maximum difference of 20% for the four drag coefficient formulations, as shown in Figure 11, with Smith [16] and the AXCP-derived function giving the minimum and maximum stresses, respectively.

5.3 Comparisons

	Large and Pond [12]		WAMDI Group [13]		Smith [16]		<i>tanh</i>	
U: Slope/ Bias	0.78	7	0.82	7	0.76	8	0.83	9
V: Slope/Bias	1.13	17	1.17	19	0.73	17	1.14	23
Mean U	0		-1		1		-3	
Mean V	-19		-22		-18		-25	
σ_u	48		50		48		54	
σ_v	31		33		30		35	
<i>RMS U</i>	47		49		47		53	
<i>RMS V</i>	36		39		34		43	

Table 2: Linear regression statistics quantifying the differences between observed storm currents and simulated currents for 4 different drag coefficients. All values are expressed as cm s^{-1} except the regression curve slope.

As the internal wave fluxes were derived for scaled current profiles, observed OML currents were not included in the analysis ($N=0$ in the OML). Thus, to examine the sensitivity of the derived analytical function (eqn 6), simulated OML currents are compared to those observed for the Storm snapshot (Figure 2) of hurricane Gilbert as

well as the other snapshots. As shown in Figure 12, the horizontal structure of the simulated current response in the OML during the Storm snapshot suggests good

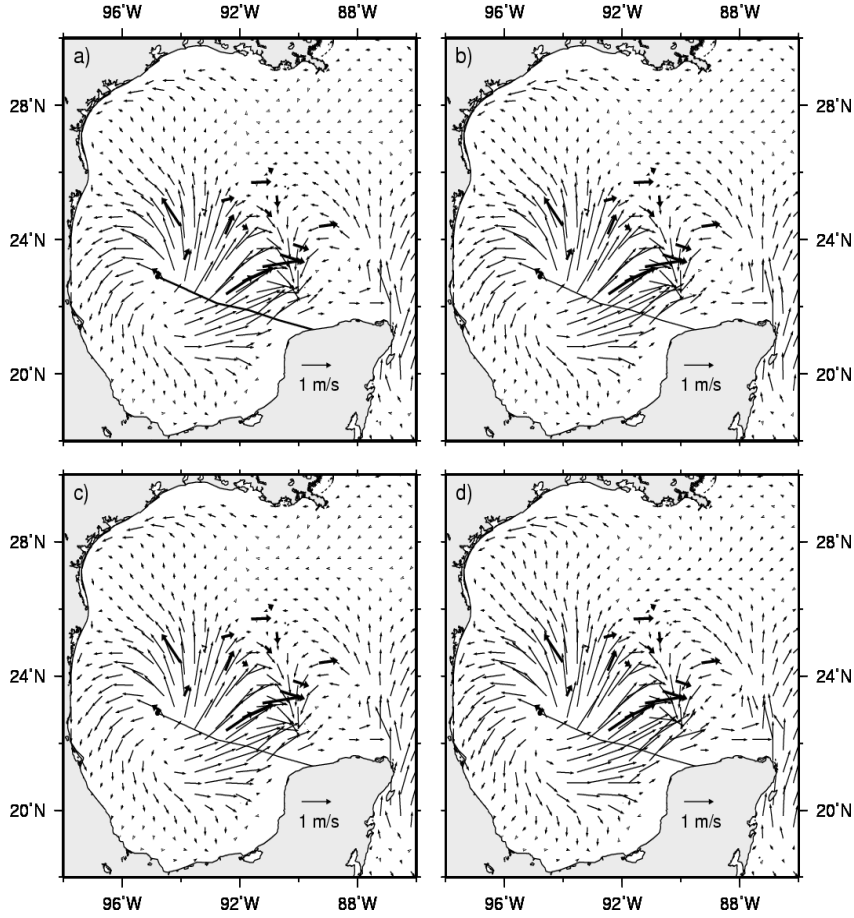


Figure 12: Simulated OML currents for the four stress distributions in Figure 11 for a quiescent ocean for a) Large and Pond [12], b) WAMDI [13], c) Smith [16] and d) AXCP-derived function (eqn 6) where thicker arrows are observed storm-induced currents. For the AXCP-derived function (d), currents are stronger extending farther away from the storm track (solid black line).

agreement among the four cases. Maximum simulated mixed layer currents occur for the AXCP-derived function (eqn 6), corresponding to maximum stress values. In this

case, stronger OML currents exist away from the track due to higher stress values, compared to the other three formulations. However, the overall OML currents are slightly weaker in the AXCP-derived model compared to observations. While the near-surface currents in this model are stronger, due to the rapid decay of currents with depth, averaging within the OML reduces this current magnitude. Maximum values are within 10% ~ 15% of the observed currents for the AXCP-derived formulation.

Simulated mixed layer currents for the four drag coefficients are compared to the observed currents from AXCP data using linear regression statistics for hurricane Gilbert, including the total data set of all three snapshots in Tables 2 and 3. While RMS differences between the drag coefficient formulations remain quantitatively similar, the slope of the regression line suggests a marginally better fit for the AXCP-derived function (eqn 6) to the current data, as shown in Figure 13. Compared to previous layered-model simulations of currents by Jacob and Shay [41], the magnitude of simulated OML currents from HYCOM are closer to observed values. For the Smith [16] drag coefficient (not shown) similar agreements were found. Using data from all three snapshots, did not significantly reduce the RMS differences.

	Large and Pond [12]		WAMDI Group [13]		Smith [16]		<i>Tanh</i>	
U: Slope/Bias	0.88	7	0.91	8	0.85	8	0.89	8
V: Slope/ Bias	1.05	10	1.08	13	1.00	0.11	1.10	14
Mean U	-5		-7		-5		8	
Mean V	-11		-13		-11		-14	
σ_u	44		46		42		46	
σ_v	34		35		33		38	
<i>RMS U</i>	44		45		43		47	
<i>RMS V</i>	36		38		34		41	

Table 3: As in Table 2, except all three snapshots (defined in Figure 2) of mixed layer currents for four different drag coefficient parameterizations.

A key question emerging from our analysis concerns the qualitative similarities between the simulated currents implicit in an accepted drag coefficient such as WAMDI [13] compared to those based on the AXCP-derived analytical function (eqn 6). As shown in Figure 14, comparisons suggest a direct relationship for both current components. Both simulated current components are well predicted by the analytical function (eqn 6) compared to the WAMDI [13] drag formulation, with a spread of $\approx 10 \text{ cm s}^{-1}$, relative to the ideal line with a slope of unity. This is the noise level of AXCP observations in a hurricane environment (Sanford et al [6]). These results suggest a tight relationship. Given this level of agreement between WAMDI [13] and

the $\tanh$ function (eqn 6), the results suggest that perhaps this latter function may reasonably represent the drag coefficient: simulated currents are certainly as good as those resulting from extrapolated Large and Pond [12] and Smith [16] values, and are consistent with results from the WAMDI [13] drag coefficient.

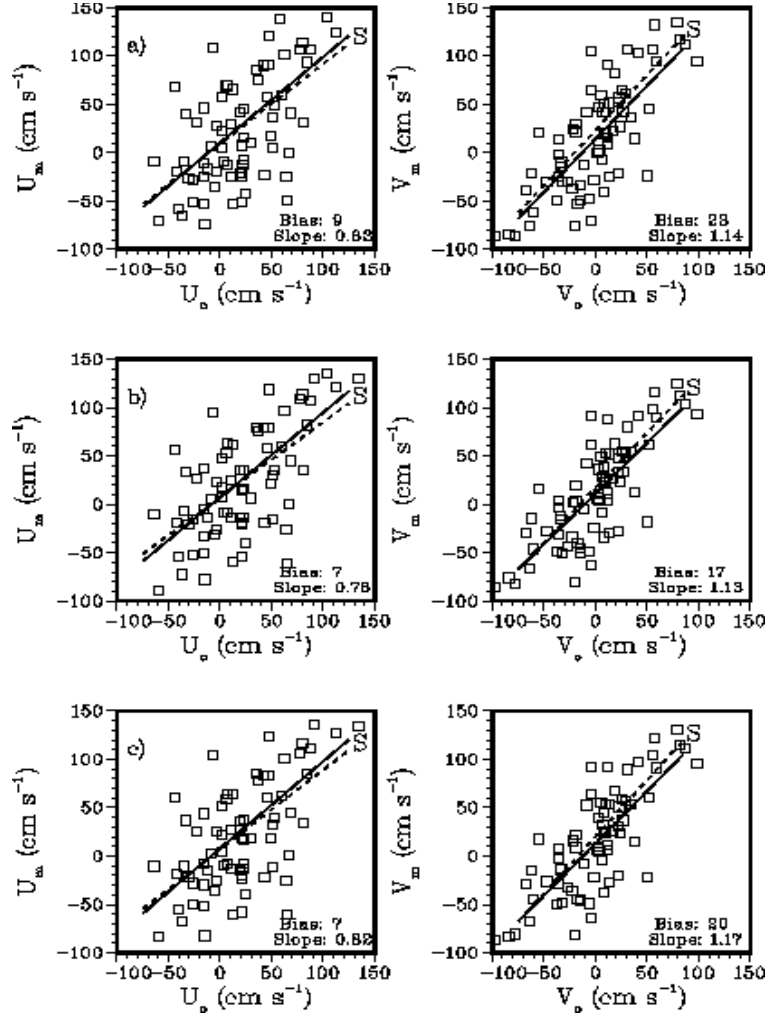


Figure 13: Comparison of simulated and observed OML currents for the U-component (left panels) and V-component (right panels) for a) AXCP-derived analytical function, b) Large and Pond [12], and c) WAMDI [13]

drag coefficients. Dashed line is for least squares regression fit, for hurricane Gilbert, snapshot only.

While the accepted drag coefficients continue to increase the surface wind stress, the analytical function (eqn 6) and drag coefficients suggested by Alamaro et al. [20] and Powell et al. [21] tend to level off at a particular threshold wind speed as mentioned in Section 4.2. The black line is for all profiles from three snapshots (see Figure 3). Bias and slopes of the regression curves for the storm snapshot (only) are given in the lower left of each panel.

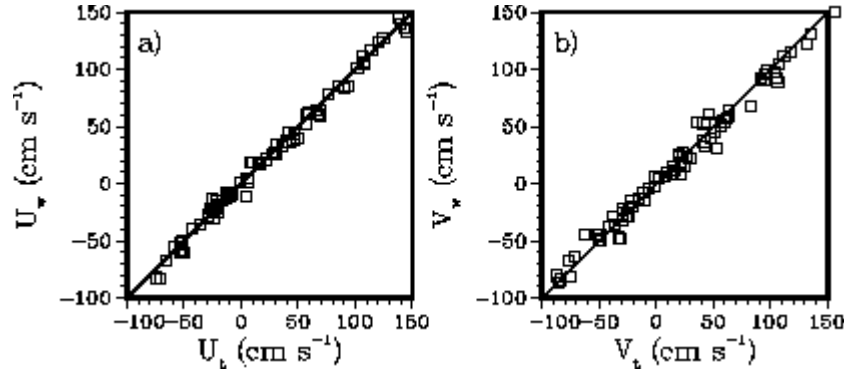


Figure 14: Scatterplots between simulated currents from AXCP-derived $\tanh$ (eqn 6) function (abscissa) and the WAMDI drag coefficient (ordinate) for a) U-component and b) V-component of current (cm s^{-1}). The solid black line represents the unity slope line.

However, the major differences between drag coefficients of Large and Pond [12], WAMDI [13] and this AXCP-derived function (eqn 6) may only become more apparent for winds exceeding 40 m s^{-1} , in terms of dynamical processes required to keep the stress nearly constant or to decrease for winds above this threshold. This issue will be further investigated using upper ocean current measurements from AXCPs acquired during higher surface winds such as those recently observed in TC's Isidore and Lili as they deepened to category 3 and 4 storms. Notwithstanding, the importance of acquiring high-resolution current profiles during strong wind events is very useful in understanding the amount of momentum being input to the OML by the wind stress (Sanford et al. [6]). Our results also point to the fact that the ocean can be used as a tracer of the momentum flux, when high-resolution air-sea flux measurements may not be possible under very strong wind and wave conditions.

6. Concluding Remarks

There has been significant progress in improving our understanding of the basic oceanic and atmospheric processes that occur during the passage of TCs. Central to this progress is the need to isolate fundamental physical processes involved in the air-sea interactions through detailed process studies using experimental, empirical, theoretical and numerical approaches. As has been found from previous sets of measurements (hurricane Gilbert, Spectrum 90, Pudov and Petrichenko [42]), these approaches are needed to eventually improve predictions of storm intensity and structure (Shay [43]).

The goal of the Gilbert experiment in Sept. 1988 was to measure the evolution of the 3-dimensional current and temperature structure excited by the passage of a TC over the scales of the atmospheric forcing. For the first time, the evolving 3-dimensional oceanic response to the passage of a TC was accurately determined from the successful deployment of AXCPs in the western Gulf of Mexico (Shay [1]). The sampling strategy resulted in a reliable time-space series to address the 3-dimensional evolution of the near-inertial wave wake (Shay et al. [9]) and the heat and mass balance (Jacob et al. [24], Jacob and Shay [41]). This approach has recently been duplicated and enhanced during the passages of TCs Isidore and Lili during deepening cycles in the Gulf of Mexico and Caribbean Sea in 2002.

Current profiles revealed a CW-rotation of the current vector with depth, indicating vertical propagation of near-inertial wave groups from the OML into the thermocline. Clockwise-rotating spectral energies dominated the scaled energy spectra in the profiles by an average factor of four, and followed a *tanh* dependence in the ratio of CW and CCW energies. Downward energy fluxes ranged from 0.5 to 10 $\text{ergs cm}^{-2} \text{s}^{-1}$ whereas an average value was $\sim 2 \text{ ergs cm}^{-2} \text{s}^{-1}$ for scaled profiles. These energy fluxes were 4 to 10 times larger than in unforced cases (Leaman and Sanford [25]), but were similar to those found in frontal regimes (Kunze and Sanford [33]).

By comparing the internal wave fluxes to the square of the wind speed, an analytical function was found that suggests a nonlinear dependence on the wind speed. This approach differs considerably to previous treatments of the wind stress by using the oceanic response as a tracer of the amount of momentum flux transferred to the oceanic thermocline from the forced ocean mixed layer, and is consistent with the propagation of energy arguments. As shown in previous studies, we found that the response is dominated by near-inertial wave processes where the wave group amplitudes are observed at the top of the thermocline (Shay et al. [5], Brooks [31],

Shay and Elsberry [32]). Moreover, as the vertical wavenumber spectral energy densities are between 4 to 10 times larger than the background internal wave spectrum (Smith [16], Garrett and Munk [18]), therefore the wave power was found to be at least 30 times larger than theoretical estimates from a global perspective (Nilsson [11]). It is no surprise that hurricane forcing of the thermocline current response, which may persist for at least a month following storm passage, is as effective as oceanic fronts in focusing internal wave energy close to the near-inertial frequency ($\sim f$).

The derived analytical function, based upon three sets of point measurements, agrees well with the Large and Pond [12] formulation at a wind speed of 11 m s^{-1} , and at about 40 m s^{-1} for the WAMDI [13] drag coefficient. Between these two wind regimes, there is fairly good agreement with the wave-age dependent drag coefficient of Donelan [14]. The latter suggests that a considerably higher drag coefficient between 10 to 20 m s^{-1} , as reflected by the analytical function. At a threshold wind speed of $\sim 28 \text{ m s}^{-1}$, the slope of the function approaches zero suggesting that beyond this wind speed, the internal wave spectrum tends to saturate as in the case of surface waves. Comparison of this analytical function to accepted other drag coefficients indicates that there may also be a limit to the wind-speed dependence of the drag coefficient. Beyond about 28 m s^{-1} , near-surface wind stress may be dependent solely on wind speed increases, rather than monotonically increasing functions.

Detailed comparisons using these formulations in the HYCOM ocean model were completed. Since the AXCP-derived function is based on energy entering the thermocline (profiles were WKBJ-scaled), simulated OML currents compared to observed currents represent an independent test. While *RMS* differences in simulated and observed currents were quite large for all cases, the analytical AXCP-derived function suggested currents did as well as those implicit in extrapolated (Large and Pond [12], WAMDI [13], Smith [16]) drag coefficients. Results from the WAMDI [13] formulation were based on buoy data from three hurricanes. The regression slope for the function is also slightly larger compared to the other methods relative to the observed OML currents. The results suggest that this issue needs to be looked at from both indirect approaches (as described here) and direct measurements.

Finally, there is a relationship between surface winds and the internal wave flux, which is dominated by near-inertial processes. While the analytical function is not a conventional drag coefficient in the strictest sense (as determined by the more direct and higher resolution measurements), using the oceanic current response as a tracer of input by the wind stress not only opens new questions concerning the momentum flux at high wind speeds, but raises the possibility of oceanic current measurements as a tracer in understanding the momentum transfer between the two fluids (Smith et al. [15]). Clearly, this issue needs further inquiry using high-resolution measurements of the surface wind and wave processes with concurrent oceanic velocity and atmospheric wind profiles. Given uncertainties in ocean surface variables during

strong winds, the air-sea exchange coefficients need to be better understood and parameterized for oceanic and coupled models to realistically simulate the oceanic response and feedback to strong forcing events.

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