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Key Points:

- A threshold power law relates
 whitecap coverage to wind stress,
 wave steepness, and dissipation
- Whitecap coverage is well correlated with the mean square slope of the equilibrium range waves
- The fit to turbulent dissipation rate shows a strong residual trend with wave age

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Observations of whitecap coverage and the relation to wind stress, wave slope, and turbulent dissipation

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Abstract Shipboard measurements of whitecap coverage are presented from two cruises in the North Pacific, and compared with in situ measurements of wind speed and friction velocity, average wave steepness, and near-surface turbulent dissipation. A threshold power law fit is proposed for all variables, which incorporates the flexibility of a power law with the threshold behavior commonly seen in whitecapping. The fit of whitecap coverage to wind speed, U_{10} , closely matches similar relations from three recent studies, particularly in the range of 6–14 m/s. At higher wind speeds, the whitecap coverage data level off relative to the fits, and an analysis of the residuals shows some evidence of reduced whitecapping in rapidly developing waves. Wave slope variables are examined for potential improvement over wind speed parameterizations. Of these variables, the mean square slope of the equilibrium range waves has the best statistics, which are further improved after normalizing by the directional spread and frequency bandwidth. Finally, the whitecap coverage is compared to measurements of turbulent dissipation. Though still statistically significant, the correlation is worse than the wind or wave relations, and residuals show a strong negative trend with wave age. This may be due to an increased influence of microbreaking in older wind seas.

1. Introduction

Whitecaps are a primary mechanism associated with many of the interactions between the atmosphere and the ocean. They drive mixing near the air-water interface, and facilitate the transfer of energy, momentum, heat, and mass [*Melville*, 1996]. This makes whitecapping a critical component of the global climate system [*Cavaleri et al.*, 2012]. Since whitecaps dissipate surface wave energy, primarily in the form of turbulence and bubble production, they are also fundamentally important in the evolution of the wavefield. Therefore, understanding the energetics of the whitecaps is critical for improving upon the present operational wave forecasts [*Cavaleri et al.*, 2007].

Whitecaps are a visual signature of breaking waves. Wave-breaking is a complex nonlinear phenomenon, which occurs at scales much smaller than most global ocean models can resolve. Therefore, breaking is often implemented using empirical parameterizations based on field observations. Toward that end, we present observations of open ocean whitecapping, accompanied by concurrent wind, wave, and turbulent dissipation measurements.

A common visual measurement of wave-breaking is the whitecap coverage, W, which is the average fraction of sea surface covered by whitecaps. Whitecap coverage is frequently characterized as a function of the 10 m wind speed, U_{10} , since measurements or hindcasts of U_{10} are often readily available. Common parameterizations are power law functions,

$$W = a U_{10}^n, \tag{1}$$

or cubic equations,

$$V = a(U_{10} - b)^3.$$
 (2)

Unlike the power law, the cubic equation contains a threshold, *b*, below which there is no visible whitecapping. Conversely, the power law allows for variation of the exponent of wind speed, *n*.

© 2015. American Geophysical Union. All Rights Reserved. Aggregating the many reported fits to U_{10} , as in Anguelova and Webster [2006], shows variability in predicted whitecap coverage of orders of magnitude across all wind speeds. Such fits date back decades [e.g., *Monahan*, 1969], so it is possible that some of this scatter is due to differences in experimental methods over the years. Most conspicuously, early *W* estimates required manual inspection of a limited collection of photographs, while modern studies tend to use large batches of digital imagery and pixel-wise thresholding techniques [see *Callaghan and White*, 2009; *Kleiss and Melville*, 2011]. In addition, images have been used from a variety of geometries and instrumentation, leading to significant differences in resolution and image quality.

Still, it is likely that some of the variability in the literature fits of W to U_{10} has physical origins. For example, the wind stress, τ , is a better descriptor of the wind energetics, and is less sensitive to atmospheric stability effects. Thus, the wind friction velocity, $u_* = \sqrt{\tau/\rho}$, is often used in place of U_{10} . Furthermore, several recent studies have shown variations in W related to wave age [Sugihara et al., 2007], wave height [Stramska and Petelski, 2003], and wind history [Callaghan et al., 2008a]. This is not surprising, since the process of wave-breaking is almost entirely driven by wave mechanics, rather than direct wind forcing [Babanin, 2011].

The wave steepness, or slope, is fundamentally important to breaking, therefore average wave slope parameters have often been proposed to replace wind variables in parameterizations of whitecapping. Bulk wave steepness variables commonly take the form

$$S = \frac{Hk}{2},$$
(3)

where *H* and *k* are a characteristic wave height and wave number. The significant wave height, H_s , is often used for *H*. Alternatively, *Kleiss and Melville* [2010] found good correlation between *W* and steepness using a peak wave height,

$$H_{p} = 4 \left[\int_{0.7f_{p}}^{1.3f_{p}} E(f) df \right]^{1/2}, \tag{4}$$

based on the work of *Banner et al.* [2000]. Here f_p is the frequency at the peak of the omnidirectional spectrum, E(f). The corresponding peak wave number, k_p , is often used for the characteristic wave number value. Peak variables such as f_p , k_p , and H_p are easily defined in unimodal seas, as in fetch-limited conditions. However, wave spectra from the open ocean routinely exhibit several apparent peaks at different frequencies. Therefore, the energy-weighted mean frequency, f_m , is sometimes used instead:

$$f_m = \frac{\int fE(f)df}{\int E(f)df}.$$
(5)

Similar definitions can be written for mean wave number, k_m , and wave height H_m . Note that the cyclic wave number, rather than radial wave number, is used throughout this paper.

However, measurements of the propagation speeds of whitecaps indicate that most whitecaps are associated with waves shorter than the peak or mean waves [see *Gemmrich et al.*, 2008]. On average, these waves are much steeper than the dominant waves or swell waves. Therefore, other wave slope variables are often used to describe the energy in the short to intermediate wind waves, i.e., the spectral "tail."

To describe the breaking probability at a desired frequency scale, *Banner et al.* [2002] used the azimuthintegrated spectral saturation,

$$B(f) = k^{4}E(k) = \frac{(2\pi)^{4}f^{5}E(f)}{2g^{2}}.$$
(6)

The spectral saturation is related to wave steepness through its association with the mean square slope (*mss*). Given a range in frequency, $f_1 < f < f_2$, the *mss* of the waves in these frequencies can be calculated as

$$mss = \int_{f_1}^{f_2} \frac{2B(f)}{f} df.$$
 (7)

In addition, *Banner et al.* [2002] showed better consistency in their results after normalizing B(f) by the directional spread, $\Delta\theta(f)$. A number of studies have since used the *mss* or mean saturation (\bar{B}), either normalized

or unnormalized, to parameterize breaking [Kleiss and Melville, 2010; Gemmrich et al., 2013; Hwang et al., 2013].

It has frequently been suggested that a better dynamic predictor of *W* is the rate of energy dissipation from wave-breaking, S_{ds} . The idea was proposed early on by *Cardone* [1969] to correct for atmospheric stability effects in the data of *Monahan* [1969]. *Kraan et al.* [1996] derived a theoretical prediction for *W* based on the WAM dissipation source function and a JONSWAP spectrum, which showed reasonable agreement with their measured whitecap coverage. *Hanson and Phillips* [1999] used estimates of dissipation from the *Phillips* [1985] equilibrium range theory, and showed improved correlation over U_{10} fits. Most recently, *Hwang and Sletten* [2008] derived a dissipation parameterization based on a total equilibrium with wind energy input. Using a collection of whitecap coverage measurements, they proposed a linear fit of the form

$$W = a(S_{ds} - b). \tag{8}$$

Much of the energy lost from the waves during breaking is dissipated as turbulence in the ocean surface layer. In measurements, this is seen as a region of "enhanced dissipation," above a standard logarithmic boundary layer [*Craig and Banner*, 1994]. *Gemmrich* [2010] found most of the breaking turbulence to be concentrated in the very near surface, especially in the wave crest. *Thomson* [2012] introduced a drifting measurement platform called the SWIFT (Surface Wave Instrument Float with Tracking), which is designed to measure the nearsurface dissipation profile, $\epsilon(z)$, in a wave-following reference frame. *Schwendeman et al.* [2014] used SWIFTs to measure $\epsilon(z)$ for fetch-limited waves in the Strait of Juan de Fuca. They showed good agreement between the wave dissipation rate based on a fetch-limited energy balance and the integrated turbulent dissipation,

$$S_{ds} \approx \rho_w \int \epsilon(z) dz.$$
 (9)

Sutherland and Melville [2015] measured turbulent dissipation from the R/V FLIP using a combination of subsurface acoustic measurements and stereo infrared (IR) imagery of the sea surface. They showed excellent correlation between integrated turbulent dissipation and estimated wave-breaking dissipation for wave ages below approximately $c_m/u_*=50$.

A different statistic, which is often used to estimate breaking dissipation, is the breaker crest distribution, $\Lambda(c)$. This variable was introduced in *Phillips* [1985], based on laboratory work from *Duncan* [1981], showing a proportionality between the rate of energy loss from a breaking wave and its phase speed to the fifth power. In this formulation, the total breaking dissipation becomes

$$S_{ds} = \rho_w g^{-1} \int b c^5 \Lambda(c) dc, \qquad (10)$$

where $\Lambda(c)$ is the total length of breaking crests per unit area, *c* is the breaker speed, and *b* is a scaling factor sometimes called the "breaking strength." $\Lambda(c)$ is a more complete description of the kinematics of the breaking waves than whitecap coverage. However, the use of $\Lambda(c)$ in practice (i.e., in equation (10)) has shown varying degrees of success. In particular, it is strongly suspected that *b* is not a constant. For example, the laboratory experiments of *Drazen et al.* [2008] demonstrated that *b* increases with the wave steepness, while *Schwendeman et al.* [2014] showed orders of magnitude variations in *b* in both their data and the published literature. Meanwhile, *Banner et al.* [2014] highlighted the large impacts of differing interpretations of the breaking speed, *c*.

Conversely, *Sutherland and Melville* [2013] showed estimates of wave dissipation from $\Lambda(c)$ that were well matched with modeled dissipation values. However, the method of *Sutherland and Melville* [2013] differs from the majority of previous studies in two important ways. First is in the use infrared cameras, leading to measurements of microscale breaking waves, or "microbreakers," which are not visible as whitecaps. *Sutherland and Melville* [2015] noted that anywhere between 20% and 90% of their total estimated dissipation may have come from microbreaking. Additionally, they employed a spectral breaking strength, *b*(*c*), introduced in *Romero et al.* [2012], which uses the spectral saturation (equation (6)) to approximate the dependence on wave slope identified in *Drazen et al.* [2008].

In this paper, the focus is on the relationship of whitecap coverage to measurements of wind, waves, and turbulent dissipation. We choose to avoid $\Lambda(c)$ for several reasons. First, the results of *Sutherland and Melville*



Figure 1. Map showing average ship locations during observations of both 2012 (blue crosses) and 2015 (orange squares) cruises.

[2013] and *Schwendeman et al.* [2014] strongly indicate that estimating S_{ds} from $\Lambda(c)$ is not feasible without measurements of microbreaking, especially in mature wind seas. Furthermore, although $\Lambda(c)$ has the potential to provide more information on the whitecap kinematics, *Kleiss and Melville* [2011], *Schwendeman et al.* [2014], and *Banner et al.* [2014] have shown that it can be very sensitive to changes in processing. Moreover, apart from the total magnitude, *Kleiss and Melville* [2010] showed very little variation in their measured $\Lambda(c)$ from whitecapping. Despite (or perhaps due to) its simplicity, whitecap coverage remains a useful representation of the overall breaking conditions. As such, *W* is frequently used in parameterizations of gas transfer [*Woolf*, 2005] and sea spray aerosol production flux [*de Leeuw et al.*, 2011]. The layout of this paper is as follows: the data collection and processing are described in section 2; the results are shown and discussed in sections 3 and 4; section 5 summarizes the important points.

2. Methods

2.1. Field Experiments

The data used here come from two roughly 3 week cruises in the North Pacific. The first cruise, onboard the R/V *New Horizon*, departed San Diego, CA, on 26 September 2012 and returned on 16 October 2012. The second, onboard the R/V *Thomas G. Thompson*, departed Seattle, WA, on 27 December 2014 and returned on 14 January 2015. The primary objective of both cruises was the replacement of a long-term moored wave buoy (Datawell Waverider MK III) at Ocean Station Papa (50°N, 145°W). Measurements from this buoy can be found in *Thomson et al.* [2013], as well as a manuscript submitted to the Journal of Physical Ocean-ography (J. Thomson et al., Wave breaking turbulence in the ocean surface layer, submitted to *Journal of Physical Oceanography*, 2015). The remaining cruise time was left for further data collection. Figure 1 maps the ships' average location and proximity to Station Papa during days of successful data collection. Drifting instruments were used in both experiments, often deployed from the ship at dawn and retrieved later the same day. The most common combination of drifters was one Waverider buoy (Datawell DWR-G4) supplemented by two SWIFT buoys. Concurrent with the drifter deployments, wind and video measurements were made from the ship. During measurement operations, the ship was directed into the oncoming wind to limit distortion of the wind and waves.

Figure 2 shows histograms of the wind, waves, and dissipation measured during both experiments, averaged over 30 min intervals coinciding with the video data (see below). The amount of data is roughly equal



Figure 2. Histograms of wind, wave, and integrated turbulent dissipation measurements during both 2012 (blue) and 2015 (orange) experiments. between the two cruises. U_{10} ranges from 5.5 to 16.0 m/s and H_s from 1.5 to 5.8 m. Overall, the 2015 experiment resulted in more measurements of high winds, large and long waves, and high dissipation rates than 2012. Further details of the measurements are provided below.

2.2. In Situ Methods

Wind speed and direction were measured at 10 Hz using a triaxis R. M. 8100 Sonic Anemometer Young mounted on the ships' jack staff. The anemometer height above sea level was 11.8 m in 2012, and 15.7 m in 2015. After despiking, the mean speed and direction are calculated over 5 min bursts. The wind friction velocity, u_{*} , is estimated using the inertial dissipation method [Edson et al., 1991; Yelland et al., 1994]. Wind speed is corrected to the 10 m estimate, U_{10} , using a standard turbulent boundary layer assumption with roughness z_0 given by the Charnock relation with constant Charnock parameter, $c_h =$ 0.014 [Garratt, 1977].

Wave spectral measurements come from the Datawell DWR-G4 waveriders, which use GPS to measure the horizontal (east-west and north-south) and vertical wave orbital velocities. Datawell's built-in processing techniques are applied to the velocity measurements to calculate the frequency spectrum, *E*(*f*), mean wave direction, $\bar{\theta}(f)$, and directional spread, $\Delta \theta(f)$ [*de Vries*, 2014]. These were used to calculate the bulk parameters (*H*_s, *f*_m, etc.) described in section 1 and shown in Figure 2.

Figure 3a shows the frequency spectra measured during the experiments, as well as the fourth and fifth moments of the spectra, and directional spread, all colored by U_{10} . After normalizing by the mean frequency, f_{m} , several frequency regimes are apparent. At low frequencies below f_m , there is often significant swell energy, with some swell peaks on the order of the dominant wave energy. There is little relationship between the swell energy and



 $E(f) [m^2s]$ 0.5 1 4 5 0.1 2 3 10⁻² (b) $f^4 E(f) [m^2 s^{-3}]$ 10⁻⁴ 10⁻⁶ 0.1 0.5 2 3 4 5 1 10⁻² (c) $f^5 E(f) \ [{
m m}^2 {
m s}^{-4}]$ 10⁻⁴ 10⁻⁶ 0.1 0.5 2 3 4 5 1 100 (d) $\Delta \theta(f)$ [°] 50 0 0.5 1 0.12 3 4 5 f/f_m

Figure 3. (a) Measured wave spectra, plus their (b) fourth and (c) fifth moments, and (d) directional spread, all colored by U_{10} and plotted against normalized frequency, f/f_m . Dotted lines mark the limits of the clearest f^{-4} decay, $\sqrt{2}f_m \le f \le \sqrt{5}f_m$. At higher frequencies, the decay is more consistent with saturation, f^{-5}

the wind speed, as these waves were not generated by the local winds. The absolute peak in E(f) often occurs in the vicinity of $f_m = 1$, but they do not match exactly. Above f_m , the spectra initially decay as f^{-4} , consistent with the equilibrium range theory of Phillips [1985]. According to Phillips' theory, in this regime, the spectral source terms of wind input, dissipation, and nonlinear energy flux are all in equilibrium, and the total energy is proportional to the local wind stress. Beyond these frequencies there is a transition to an f^{-5} tail, as has been observed in several previous measurements [e.g., Kahma and Calkoen, 1992; Romero and Melville, 2010] and which is often called the saturation range. This transition is more clearly visible in the fourth and fifth moments of the spectra. Based on the spectra, we define the equilibrium range to extend from roughly $\sqrt{2}f_m$ to $\sqrt{5}f_m$ (2 k_m to $5k_m$), as shown in dotted lines on Figure 3.

The fundamental principles of the SWIFT platform are detailed in Thomson [2012]. The SWIFTs are equipped with a Nortek Aquadopp HR current profiler, which measures water velocities from a depth of 0.6 m up to the free surface. The SWIFTs follow the wave orbital motion at the surface, effectively filtering the orbital velocity from their measurement. What remains are only the turbulent velocity fluctuations. The secondorder structure function is calculated from the time series of velocity profiles at 5 min intervals. Using the method of Wiles et al. [2006], based on Kolmogorov's theory, leads to profiles of the turbulent dissipation rate. Throughout this paper, a bulk turbulent dissipation rate is used from integrating the profiles in depth, as in equation (9). The turbulence profiles themselves are analyzed in further detail in a separate paper using the 2015 data, currently in review (Thomson et al., submitted manuscript, 2015).

2.3. Video Methods

The camera configuration differed slightly between the two experiments. In 2012, a single serial "bullet" camera with a wide field-

of-view (FOV) was mounted on a Pan/Tilt system to the foreword rail of the second deck of the R/V New Horizon, at a 10.8 m height, similar to the setup of Schwendeman et al. [2014]. The Pan/Tilt provided active stabilization and the capability to look either starboard or port, depending on lighting conditions. In 2015, the bullet camera was replaced with several Pt. Grey Flea2 and Flea2G cameras, which were mounted directly to the ship's port and starboard rails, 11.0 m above the waterline. One camera on each side had a

Table 1. Camera Specifications and Geometry								
Year	Camera	Focal Length (mm)	Height (m)	Pixels	Rectified Resolution (m ²)			
2012	Mini Bullet, Sony ExView	3.6	10.8	720 imes480	1.2×10 ⁻²			
2015	Pt. Grey Flea2G	2.8	11.0	1288 imes 964	1.2×10 ⁻²			
2015	Pt. Grey Flea2	9	11.0	1032 imes 776	1.5×10 ⁻³			

FOV similar to the bullet camera. To test the effect of better resolution, another camera on each side was fitted with a narrower lens. Video was recorded synchronously from both cameras, and the side was again chosen based on the lighting. Table 1 summarizes the camera configuration and specifications from both experiments. Camera settings such as exposure time and gain were adjusted before every video capture to maximize contrast and minimize saturation.

Figure 4 shows the steps involved in the video processing. First, the resulting video data are georectified from pixel coordinates to meters. *Schwendeman and Thomson* [2015] showed that, in the absence of other data sources, adequate georectification can be performed using an automatic detection of the horizon. This is the method used on the 2012 video data. In 2015, a NovAtel SPAN-IGM-A1 receiver and dual antennas were mounted on the rail next to the port side cameras to aid in the georectification. NovAtel's SPAN products combine an inertial navigation system (INS) with global navigation satellite system (GNSS) for a more robust and accurate position measurement than can be achieved with either method alone. The horizon method is still used, as a means to synchronize the camera frames and INS-GNSS data stream. The rectified images are interpolated onto a rectangular grid for further processing. In each case, the grid was set between 40 and 60 m from the ship, to avoid reflections from the ship while maintaining acceptable camera resolution. The pixel footprint in the rectified image is specific to the camera geometry, the details of which can be found in Table 1.

Next, a brightness threshold is applied to the rectified video to isolate the whitecaps. A method similar to that of *Kleiss and Melville* [2011] and *Schwendeman et al.* [2014] works well for the large quantity of video used here. In this technique, a histogram of pixel brightness is calculated over 5 min lengths of video. The second derivative of the logarithm of the histogram yields a peaked shape, indicating a change of curvature separating the bright foam of the whitecaps from the otherwise dark sea surface. The threshold is set as the brightness at 20% of the peak in the curvature, as recommended in *Kleiss and Melville* [2011].

Finally, the 5 min thresholded bursts are gathered into half-hour segments for calculation of whitecap coverage, *W*. Averaging over 30 min reduces the scatter in the *W* estimate, and matches the duration of the Waverider measurements. Since video was recorded at 15 frames-per-second (fps) in the 2012 experiment, and 5 or 7.5 fps in 2015, this amounts to between 9000 and 27,000 images per 30 min. This is well above the 500 images which *Callaghan and White* [2009] recommended to reduce the variability of *W* to $\pm 3\%$. Still, uncertainty exists due to nonstationarity of the whitecapping and lighting conditions, as well as in the choice of threshold. Variability in the 5 min bursts is used to make error estimates in the 30 min averages. A bootstrapping of the bursts is performed, and error bars are calculated corresponding to one standard deviation around the mean. Some errors were particularly large, on the order of *W*, and were found to correlate with poor thresholding or low data quality in one or more of the bursts. Therefore, these points are removed from the subsequent analysis.

The two cameras from the 2015 experiment are used to determine the effect of the pixel resolution on whitecap coverage. *Schwendeman et al.* [2014] postulated that their data were biased by inadequate resolution of small whitecaps. The narrow FOV camera improved upon the previous resolution by roughly a factor of 10 (Table 1), but Figure 5 shows little difference in *W* from the larger FOV. Therefore, to avoid double counting the 2015 data in curve fitting, only the wide FOV results are used in the remainder of the paper.

Figure 5 indicates that the limiting factor for identifying small breaking events may not be camera resolution. This is encouraging for comparisons of *W* measurements across studies, given that there has been little attempt to standardize equipment and image acquisition. However, there is another factor which may preclude measurement of small whitecaps, namely, lack of contrast due to less production of bubbles and foam. Visual inspection of our images indicates that the quantification of these small events is highly dependent on proper illumination and optimal camera settings (exposure, shutter speed, etc.). Glare, sun glitter, and low



Figure 4. Example of rectification and thresholding for a single video frame. (a) The original frame, undistorted and cropped, showing the sample area (cyan) and detected horizon (red dashed line). (b) The sample area, rectified to world coordinates. (c) The area after thresholding.

light at dawn and dusk can all impede the detection of small breakers. In the limiting case of microbreaking, no foam is produced, meaning that these events cannot be measured from traditional video cameras regardless of camera settings and lighting.

3. Results

3.1. W Versus U₁₀ and u_{*}

As discussed in section 1, the most common parameterization of whitecap coverage is to U_{10} . Figure 6a shows the North Pacific data along with three fits from the recent literature, plus the widely used relation of Monahan and O'Muircheartaigh [1980]. The three more recent studies were chosen because their experimental methods closely match those used here. In each of these studies, whitecap coverage was calculated using digital images from a camera set 10-20 m above sea level with an oblique view 6°-33° below horizontal, and with a pixel-wise brightness threshold. Stramska and Petelski [2003] and Callaghan et al. [2008a] both used shipboard photographs from the North Atlantic Ocean, while the video in Sugihara et al. [2007] came from a tower in Tanabe Bay, Japan, roughly 2 km offshore. In all cases, the local winds were measured in situ and adjusted to U_{10} , without a correction for atmospheric stability. Each fit line is plotted for only the range of the available measurements, with maximum U_{10} of 14 m/s [Stramska and Petelski, 2003], 16.3 [Sugihara et al., 2007], and 23.1 [Callaghan et al., 2008a]. All of these fits are of the cubic variety (equation (2)), although Callaghan et al. [2008a] found that their data were best described by two distinct cubics, with a transition between 9.25 and 11.25 m/s. The literature curves tend to diverge for $U_{10} < 7$ m/s, when whitecapping is infrequent and microbreaking may have a larger impact. In the range of $7 < U_{10} < 14$ m/s, the fits agree to within roughly a factor of 2. This wind speed range also brackets the majority of the North Pacific data.

The *W* data used in *Monahan and O'Muircheartaigh* [1980] were calculated manually, using the method described in *Monahan* [1969]. This may explain why the classical *Monahan and O'Muircheartaigh* [1980] fit is biased high relative to the other curves, which were made using thresholded digital images. In addition, the *Monahan and O'Muircheartaigh* [1980] fit is in the

form of a power law (equation (1)), which diverges from the cubic fits at low wind speeds.

A fit to the North Pacific data is also plotted in Figure 6a, and closely resembles those of the recent literature. A hybrid of equations (1) and (2) is used,

$$W = a(x-b)^n, \tag{11}$$

were x is the explanatory variable, in this case, U_{10} . This style of fit is used throughout the remainder of this paper, and will be called the "threshold power law." It incorporates the threshold behavior which is a



Figure 5. Comparison of whitecap coverage, *W*, for video from wide fieldof-view and narrow field-of-view cameras in the 2015 experiment. The linear fit is barely visible next to the one-to-one line.

common feature of whitecap parameterizations [e.g., *Banner et al.*, 2000], as well as the ability to tune the exponent, *n*. The threshold power law is more versatile than either equations (1) or (2); however, the coefficients of this function cannot be solved for linearly. Instead, the Levenberg-Marquardt algorithm is used to compute the best fit by minimizing the sum of the squares of the log residuals,

$$W_{res} = \log_{10} W - \log_{10} [a(x-b)^n].$$
 (12)

The log residuals are used to give equal weight to the whitecap data across several orders of magnitude. Table 2 shows the best fit values of a, b, and n from equation (11). In addition, 90% confidence intervals are estimated for each coefficient, again using a bootstrap-style technique. The full data set is randomly resampled (with replacement) 100

times, with a best fit calculated after each resampling. The confidence intervals on each coefficient correspond to the 5th lowest and 95th highest value in the set of all fits. It should be noted that because the threshold power law has three coefficients, these confidence intervals are sometimes quite large, particularly when the variables are highly scattered. In the case of U_{10} , the cubic equation likely would have been adequate, as n = 2.8 is the exponent of the best fit, and n = 3 is within the confidence interval $(1.94 \le n \le 3.39)$.

Table 3 lists the error metrics for each fit, which are also calculated from the log residuals. The root-meansquare error (RMSE) is defined as



Figure 6. Whitecap coverage versus (a) U_{10} and (b) u_* , as well as the threshold power law fit for each (black solid line). Included are relations from *Stramska and Petelski* [2003] (blue dotted), *Sugihara et al.* [2007] (red dashed), *Callaghan et al.* [2008aa] (green dash dot), and *Monahan and O'Muircheartaigh* [1980] (purple solid) plotted over the range of conditions from which they were calculated.

Fit Variable, x	а	b	п
U ₁₀	$2.81 imes 10^{-5}$ (4.88 $ imes 10^{-6}$, 2.19 $ imes 10^{-4}$)	$3.87 imes10^\circ$ (2.77 $ imes10^\circ$, 5.21 $ imes10^\circ$)	2.76 (1.94, 3.39)
U*	$6.82 imes 10^{-2}$ ($5.08 imes 10^{-2}$, $1.03 imes 10^{-1}$)	$1.39 imes 10^{-1}$ (9.49 $ imes 10^{-2}$, 1.67 $ imes 10^{-1}$)	2.04 (1.63, 2.56)
U_{10}/c_{m}	$2.71 imes 10^{-2}$ ($1.48 imes 10^{-3}$, $3.31 imes 10^{-2}$)	$3.85 imes 10^{-1}(-2.19 imes 10^{-1},5.41 imes 10^{-1})$	2.66 (1.55, 5.32)
u∗/c _m	9.64 $ imes$ 10 $^{\circ}$ (9.54 $ imes$ 10 $^{-1}$, 6.24 $ imes$ 10 1)	$1.69 imes 10^{-2}$ ($1.34 imes 10^{-2}$, $2.00 imes 10^{-2}$)	1.92 (1.25, 2.50)
u∗/c _p	$4.80 imes 10^{-1}$ ($1.02 imes 10^{-1}$, $6.89 imes 10^{3}$)	$1.14 imes 10^{-2}$ (-6.49 $ imes$ 10 $^{-2}$, 1.32 $ imes$ 10 $^{-2}$)	1.09 (0.67, 5.36)
$H_s k_p/2$	$5.43 imes 10^{3}$ (4.16 $ imes$ 10 $^{-2}$, 3.09 $ imes$ 10 6)	$-3.33 imes 10^{-2}$ ($-3.76 imes 10^{-1}$, $2.51 imes 10^{-3}$)	4.48 (0.43, 9.99)
$H_s k_m/2$	$6.85 imes10^2$ ($3.09 imes10^\circ$, $5.00 imes10^5$)	$5.78 imes 10^{-3}~(-1.79 imes 10^{-3}, 9.10 imes 10^{-3})$	2.66 (1.40, 4.47)
$H_m k_m/2$	$8.26 imes 10^{1}$ (1.47 $ imes$ 10^{-1} , 5.72 $ imes$ 10^{3})	$1.28 imes 10^{-3}~(-6.54 imes 10^{-4}, 5.78 imes 10^{-3})$	2.19 (0.70, 3.27)
$H_p k_p/2$	$1.47 imes 10^{5}(1.60 imes 10^{2},3.35 imes 10^{5})$	$-6.81 imes 10^{-2} (-3.65 imes 10^{-1}, -1.68 imes 10^{-2})$	6.71 (2.81, 17.42)
mss	$7.63 imes 10^7$ (2.61 $ imes$ 10 7 , 1.05 $ imes$ 10 8)	$3.26 imes 10^{-4}$ ($-2.05 imes 10^{-4}$, $5.10 imes 10^{-4}$)	4.01 (3.88, 4.09)
mss/($\Delta \theta$)	$1.46 imes10^{5}$ (2.94 $ imes10^{\circ}$, 1.02 $ imes10^{7}$)	$9.36 imes 10^{-4}$ ($4.66 imes 10^{-4}$, $2.52 imes 10^{-3}$)	3.07 (1.03, 3.87)
$mss/(\Delta f \Delta \theta)$	$2.82 imes10^1$ ($1.90 imes10^\circ$, $1.05 imes10^2$)	$1.44 imes 10^{-2}$ ($1.26 imes 10^{-2}$, $2.12 imes 10^{-2}$)	2.39 (1.45, 2.81)
$\int \epsilon(z) dz$	$1.73 imes 10^4 (1.30 imes 10^3, 6.02 imes 10^7)$	$1.97 imes 10^{-4}$ ($1.40 imes 10^{-4}$, $3.11 imes 10^{-4}$)	1.89 (1.47, 2.92)
$\int \epsilon(z) dz$ (n = 1)	$2.80 \times 10^{1} (2.20 \times 10^{1}, 3.52 \times 10^{1})$	$3.46 imes 10^{-4}$ ($3.26 imes 10^{-4}$, $3.76 imes 10^{-4}$)	1

Table 2 Desults of the Threshold Device Levy Fit $W = c(x - b)^n$ With 000% Confidence Intervals in Deventhesi

$$RMSE = \sqrt{\frac{\sum W_{res}^2}{N}},$$
(13)

where N is the number of observations. This RMSE can be thought of as an average order of magnitude deviation from the fit. The other quantity shown is the coefficient of determination (R^2) , here defined as,

$$R^{2} = 1 - \frac{\sum W_{res}^{2}}{\sum \left(\log_{10} W - \overline{\log_{10} W}\right)^{2}},$$
(14)

where the overbar implies an average. As will be discussed further, the fit to U_{10} has the best statistics (RMSE = 0.28, R^2 = 0.81) of any shown in this paper.

In Figure 6b, W is plotted against u_{*} , with a threshold power law fit and literature relations from Stramska and Petelski [2003] and Sugihara et al. [2007]. u, was not calculated in Callaghan et al. [2008a]. The scatter is slightly more than for U_{10} , as indicated by the fit statistics (RMSE = 0.32, R^2 = 0.74). In addition, the fit from Stramska and Petelski [2003] is qualitatively distinct from the others. This is probably explained by the fact that while u_{*} was estimated here and in Sugihara et al. [2007] using the inertial dissipation method, Stramska and Petelski [2003] use an empirical relation from U_{10} .

One feature that is apparent in this data set is a leveling off of W at the highest winds, $U_{10} > 14$ m/s. In many ways, this is similar to the results of Callaghan et al. [2008a], which also show a muted whitecapping response in the high wind speed regime. Unfortunately, these conditions are not frequently measured, so it is difficult to ascertain whether this is a persistent feature or a product of small sample size. For example, the data above 14 m/s shown here come from a single day (7 January 2015). Similarly, Sugihara et al. [2007] show only two data points above 14 m/s and Stramska and Petelski [2003] do not make any measurements

Table 3. Statistics of the Threshold Power Law Best Fit						
Fit Variable, <i>x</i>	RMSE	R ²				
U ₁₀	0.28	0.81				
<i>u</i> _*	0.32	0.74				
U_{10}/c_{m}	0.34	0.69				
u_*/c_m	0.30	0.76				
u_*/c_p	0.43	0.49				
$H_s k_p/2$	0.56	0.13				
$H_s k_m/2$	0.44	0.48				
$H_m k_m/2$	0.47	0.40				
$H_p k_p / 2$	0.57	0.10				
mss	0.39	0.58				
$mss/(\Delta\theta)$	0.38	0.61				
$mss/(\Delta f \Delta \theta)$	0.30	0.75				
$\int \epsilon(z) dz$	0.50	0.27				
$\int \epsilon(z) dz \ (n=1)$	0.51	0.26				

above 14 m/s.

Stramska and Petelski [2003], Sugihara et al. [2007], and Callaghan et al. [2008a] each identify wave development as a source of scatter in their $W(U_{10})$ plots. Stramska and Petelski [2003] separate their data using the difference between their (visually estimated) H_s and the fully developed significant wave height, H_{Full}, from the Pierson-Moskowitz spectrum [Pierson and Moskowitz, 1964]. They find markedly less whitecapping when the waves are "undeveloped," defined as $H_s < (H_{Full} - 0.5 \text{ m})$, in comparison with the "developed" waves in similar wind speeds. Sugihara et al. [2007] note a similar separation by wave age, c_p/u_* , but only in the case of pure windseas. For $8 < c_p/u_* < 16$, W is noticeably less than the remainder of the data $(16 < c_p/u_* < 29)$. Finally, *Callaghan et al.* [2008a] concentrate on the wind history, or wind acceleration, similar to Hanson and Phillips [1999]. They show that when the wind is increasing (as measured



Wind Speed Residuals



over a 2.5 h period) and U_{10} is above 9 m/s, W is significantly reduced. To summarize, each study found that when the waves were growing or highly forced, the whitecap coverage was less than otherwise expected.

These ideas are tested using the residuals of the U_{10} threshold power law fit. Figure 7a shows the W_{res} data plotted against c_m/u_* . A linear fit of the residuals is also shown to highlight the trend. Interestingly, over the whole data set, there is a decreasing trend with wave age (W reduced in older waves). However, for $c_m/u_* < 29$, as in the *Sugihara et al.* [2007] data set, W_{res} strongly increases with wave age. In other words, W is less than predicted in both particularly young and particularly old waves. This is similar to the results of *Lafon et al.* [2007] in a coastal region, showing a peak in W at around $c_p/U_{10}=0.69$. A fit of W to (inverse) wave age alone was also attempted, however, the overall statistics are not significantly improved (see Table 3) and no peak is apparent. Of the wave age variables, the best fit is to u_*/c_m , rather than c_p or U_{10} .

Next, the residuals are examined as a function of the 2.5 h average wind acceleration, $\Delta U_{10}/\Delta t$, as in *Callaghan et al.* [2008a] (Figure 7b). The linear fit is plotted for all the residuals, as well as for only $U_{10} > 9$ m/s. As predicted, the residuals are negative (*W* reduced) for accelerating winds, and the effect is more pronounced in the higher wind regime. Even for the higher winds, however, the trend is small in comparison with the overall scatter. Finally, the residuals are plotted against the *Stramska and Petelski* [2003] wave development metric, $H_{Full} - H_s$. For this variable, almost no trend can be seen. This is likely due to the influence of swell on H_s . Indeed there are many points for which $H_s > H_{Full}$, which could not occur in the absence of swell energy.

3.2. W Versus Wave Slope

Figure 7 shows that wave development can modulate the general relationship of W to U_{10} . Next, we test how well the whitecap coverage can be predicted from measurements of the waves alone. As discussed in section 1, an increase in the overall wave steepness is expected to lead to more whitecaps. In Figure 8a, W is plotted against wave steepness using the significant wave height, H_{sr} and the mean wave number, k_{m} . There is a clear correlation, but several points look to be extreme outliers. In particular, the points in the lower right of the plot with relatively large steepness but little whitecapping. These are likely points with a significant fraction of swell energy, leading to an unsuitable significant wave height. Again, a threshold power law fit is used to quantify the correlation, as shown in Tables 2 and 3. In addition, the results using several alternative steepnesses of the form Hk/2 are also reported. Despite the issues visible in Figure 8a, H_s $k_m/2$ was found to have the best statistics of these fits (RMSE = 0.44, R^2 = 0.48). The fits using peak variables H_p or k_p are particularly poor for this data. This is important to note, because while breaking parameterizations that have been developed in young wind seas often use these variables [e.g., Hwang and Sletten, 2008; Kleiss and Melville, 2010], in the open ocean the peak waves may not be related to the breaking waves at all. The fit from Kleiss and Melville [2010] to $H_p k_p/2$ is shown for comparison in Figure 8a. Aside from the scatter, the overall behavior is quite similar over the range of their measurements. Note that Kleiss and Melville [2010] use a standard power law fit, which is acceptable since their steepnesses are all relatively



Figure 8. Whitecap coverage versus (a) $H_s k_m/2$ and (b) equilibrium range *mss*, using the frequencies $\sqrt{2}f_m \le f \le \sqrt{5}f_m$, and normalized by the mean directional spread, $\Delta\theta$, and frequency bandwidth, Δf . Black lines are the current fits, and the orange line comes from *Kleiss* and *Melville* [2010], using $H_p k_p$.

large. However, the North Pacific whitecaps show a clear threshold around $H_s k_m/2 \approx 0.01$, which the power law could not reproduce.

Better agreement is found using steepness parameters that focus on the short or intermediate wind waves. In particular, the *mss* of the equilibrium range waves shows good correlation (RMSE = 0.39, $R^2 = 0.58$). Again, the definition of the equilibrium range used here is $\sqrt{2} \le f/f_m \le \sqrt{5}$, as shown in Figure 3. Extending the mean square slope calculation to higher frequencies in the saturation range tends to degrade the correlation. This may have to do with the scale of waves that form whitecaps. The corresponding equilibrium range phase speeds are 0.44–0.71 times c_m . Although the peak in estimates of $\Lambda(c)$ from whitecaps tends to be in the $0.2c_m - 0.5c_m$ range, these equilibrium range waves are well represented in the overall whitecapping. Moreover, the results of *Gemmrich et al.* [2013] indicate that after weighting by c^5 , the equilibrium waves produce the majority of the whitecap energy dissipation. This gives a physical reasoning as to why the slopes of these particular waves best explain the whitecap coverage.

In Figure 8b, *W* is plotted against *mss* after further normalizing by directional spread, $\Delta\theta$, and frequency bandwidth, Δf . The resulting fit nearly matches the performance of the wind variables (RMSE = 0.30, $R^2 = 0.75$). The success of the normalization shows that, for similar *mss*, wavefields that are more unidirectional will exhibit more breaking. Note that Figure 3 shows significant variation in the directional spread with frequency. For consistency with *mss*, the value used in the normalization is the mean value over the equilibrium range frequencies. The *mss* normalized by $\Delta\theta$ is similar to the "weighted, band-averaged saturation" used in *Gemmrich* [2010] and *Gemmrich et al.* [2013],

$$\sigma_b = \int_{f_p}^{4f_p} \frac{B(f)}{f\Delta\theta(f)} df,$$
(15)

where $\Delta\theta(f)$ is the frequency-dependent directional spread and the factor of 1/f relates the saturation, B(f), to *mss* (as in equation (7)). In addition, normalizing by frequency bandwidth, $\Delta f = (\sqrt{5} - \sqrt{2})f_m$, also improves the quality of the fit. Since $f^{-1}B(f) \propto f^4 E(f)$ is approximately constant in the equilibrium range, this normalization removes the erroneous dependence on f_m from the limits of integration.

Compared to Figure 6, the plots of W with wave steepness variables show less evidence of leveling off at the high winds (steep waves). Thus, the unexpectedly low W for these wind conditions is perhaps explained by the wave slope, which is also lower than expected. This effect is also shown and discussed in Thomson et al. (submitted manuscript, 2015). Although the overall fit to *mss* is not quite as good as that of U_{10} , it may be that young, highly forced waves are a case where wave slope parameters outperform the wind variables.



Figure 9. Whitecap coverage versus integrated turbulent dissipation rate, shown on (a) logarithmic and (b) linear axis. Solid lines are the threshold power law fit, while dashed lines are a linear fit.

3.3. W Versus Turbulent Dissipation Rate

Finally, the whitecap coverage is compared with the turbulent dissipation rate, as measured by the SWIFT drifters. Again, the SWIFTs measure profiles of dissipation, $\epsilon(z)$, but since we are only interested in total dissipation, we integrate the profiles to a single value, $\int \epsilon(z)dz$. This bulk turbulent dissipation rate is in good agreement with the wave dissipation rate in many conditions [see *Thomson et al.*, 2013; *Schwendeman et al.*, 2014]. Figure 9a shows *W* plotted against $\int \epsilon(z)dz$. It is clear that the scatter is much larger than both the wind speed and wave steepness parameterizations. This is reflected in the statistics of the threshold power law fit to the dissipation data (RMSE = 0.50, $R^2 = 0.27$). In particular, for integrated dissipation rates below 5 m³/s³, the variability is more than 2 orders of magnitude. Although the confidence intervals on the best fit exponent (1.47 $\leq n \leq 2.92$) do not support the linear model proposed by *Hwang and Sletten* [2008], the statistics are not made significantly worse by constraining the fit to be linear (RMSE = 0.51, $R^2 = 0.26$). In Figure 9b, the same results are plotted using a linear vertical axis for an alternative view of the data.

In Figure 10, the residuals of the fit to dissipation are examined, using c_m/u_* as in Figure 7a. This time, the



Figure 10. Residuals of the threshold power law fit to turbulent dissipation, plotted against wave age. The dashed line is a linear fit to all the residuals, and the dash-dot line is for only $c_m/u_* < 29$.

young waves ($c_m/u_* \le 29$) show no clear pattern. However, for older waves, there is an even stronger negative trend. The lowest residuals come from the periods when whitecap coverage is least, yet there is still significant turbulent dissipation. This could be evidence of the contribution of microbreaking to the dissipation. It is consistent with *Sutherland and Melville* [2015], who suggest that the percentage of dissipation from microbreaking increases with wave age, to as much as 90%.

Alternatively, uncertainty in the in situ measurements of turbulent dissipation may be responsible for both the overall scatter and the bias. There are three broad categories of uncertainties here. First are uncertainties from sources of turbulent dissipation not related to wave breaking. These could be due to direct input from the wind, nonbreaking "swell dissipation," or the SWIFTs' own turbulent wake. The first two have been estimated to be only minor contributions to the total turbulent dissipation [see *Terray et al.*, 1996; *Schwendeman et al.*, 2014], while

the latter is minimized by using acoustic beams directed out and away from the SWIFT hull. Conversely, the turbulent dissipation is only one of several mechanisms for breaking-related dissipation. Most importantly, these measurements cannot quantify the dissipation due to the entrainment of air in the whitecap bubble plume, which *Lamarre and Melville* [1991] estimated to be as much as 50% of the dissipation in laboratory breakers. Finally, there are the uncertainties in the measurement of turbulent dissipation inherent to the SWIFT instrumentation and methodology. *Thomson* [2012] estimated these errors to be on the order of 10%; however, these estimates were made for much smaller wave heights ($H_s < 1$ m). More recently, Thomson et al. (submitted manuscript, 2015) have demonstrated the difficulty in measuring turbulent velocities in the bubble cloud produced by large breaking events. Therefore, it may be that some of the dissipation occurring during active whitecapping is not resolved in the SWIFT measurements.

4. Discussion

Figure 6a stands in stark contrast to Anguelova and Webster [2006], who suggest discrepancies of 2 orders of magnitude in the literature $W(U_{10})$ fits across all wind speeds. By comparing only studies using modern digital image processing and similar camera geometry, many of the main methodological sources of scatter are eliminated. There is a clear distinction from the widely used fit of *Monahan and O'Muircheartaigh* [1980], which was derived from manual tracing of whitecaps in relatively small batches of photographs. The modern fits agree to within roughly a factor of 2 between wind speeds of 7 and 14 m/s. Large intermittence of whitecapping is likely responsible for the deviations at low wind speeds. Meanwhile at high winds, the relative lack of measurements leads to biases associated with the small sample size. Still, despite all the known issues, the fit to U_{10} has the least error of all the variables examined here. This is similar to the conclusions of *Kleiss and Melville* [2010], *Goddijn-Murphy et al.* [2011], and *Salisbury et al.* [2013], who all show little to no improvement using parameterizations that incorporate the wave conditions.

Wave-breaking is related to the local wind speed through the concept of equilibrium. According to the *Phillips* [1985] theory, the spectral energy density of the equilibrium range waves is proportional to u_* . Since *mss* is calculated from the fourth moment of the wave frequency spectrum, and the equilibrium range spectrum decays as f^{-4} , the *mss* of the equilibrium waves is approximately proportional to u_* , as shown in *Thomson et al.* [2013]. Thus, the success of the fit of *W* to equilibrium range *mss* is no coincidence. The Phillips theory also accounts for the directional spread of the waves through the *l(p)* function, although this variable is more challenging to measure.

Thomson et al. [2013] also demonstrated that total wind input and total wave dissipation are often in near equilibrium (i.e., not just in the spectral equilibrium range). This was also shown in *Hwang and Sletten* [2008], based on a scaling of fetch-limited wave growth. Their proposed dissipation scaling (equation (8)) was actually validated using the wind input, which is primarily dependent on u_* , with a small modulation by wave variables. To summarize: the U_{10} and u_* parameterizations are successful because they are variables that are closely linked to both the equilibrium wave steepness and wave-breaking dissipation, yet are measured more precisely than either steepness or dissipation themselves.

An unresolved question is whether to differentiate "active" whitecaps from "residual" foam, as is often done in the literature [e.g., *Scanlon and Ward*, 2013]. It has been hypothesized that measurements of purely active breaking may be more closely related to wave dynamics, since the behavior of residual foam is dependent on water chemistry [*Callaghan et al.*, 2008b]. The experiments of *Callaghan et al.* [2013] show that whitecap foam can indeed be stabilized by high levels of surfactants. Meanwhile, satellite estimates of *W* indicate that U_{10} parameterizations are biased low when the sea surface temperature is high, perhaps because of increased water viscosity or other bubble dynamics [*Salisbury et al.*, 2013]. However, ongoing laboratory work has shown an approximately linear relationship between the volume of a whitecap's submerged bubble plume and its dissipation of wave energy (A. H. Callaghan, personal communication, 2015). A similar conclusion was presented in *Lamarre and Melville* [1991]. The bubble plume volume is in turn correlated with the surface decay of the whitecap foam patch, although water chemistry (e.g., large surfactant concentrations) may modify this relationship. In this way, the appearance and decay of residual foam may actually be a more useful expression of the breaking energetics than previously thought.

As discussed in *Scanlon and Ward* [2013], it is difficult to separate active and residual breaking on the basis of pixel brightness alone. Instead, we use an implicit calculation to determine the influence of residual foam in *W*.



Figure 11. The characteristic residual foam time, τ , calculated using $T_{min} = 5$ s (blue) and $T_{min} = 10$ s (orange), plotted against (a) the turbulent dissipation rate, and (b) normalized mean square slope. The dashed lines are linear best fits.

This strategy is based on the observation that as a whitecap propagates through the camera field-of-view, it "flips" pixels from zero to one in the thresholded images (Figure 4c). For an idealized whitecap, these pixels remain "flipped" until the residual foam fully decays. In this case, only the newly flipped pixels are associated with active whitecapping, while the total whitecap coverage also incorporates residual foam. The ratio of these quantities, multiplied by the time between images, *dt*, produces a characteristic foam time scale,

$$\tau = \frac{\sum l_{total}}{\sum l_{flipped}} dt,$$
(16)

where *dt* is needed to account for differences in camera frame rate. It should be noted that, though related, this τ is not directly comparable to the average whitecap duration as calculated from $\Lambda(c)$ (as in *Kleiss and Melville* [2010]), or from tracking individual breakers (as in *Callaghan* [2013]). Rather, τ represents the average residence time of whitecap foam over an individual image pixel.

Unfortunately, visual inspection of the thresholded images shows that the actual whitecaps do not necessarily behave like the idealized whitecap described above. Instead, real whitecaps leave patchy foam which may be further advected by the subsequent wave orbitals. This means that pixels may flip several times over the course of a single breaking event. Therefore, a further condition is implemented to better isolate the foam time scales. Specifically, pixels are only counted in $I_{flipped}$ the first time they flip within a defined time frame, T_{min} . This assumes that all subsequent flips within T_{min} are assumed to be due to the same wave, therefore T_{min} should be on the order of the mean wave period, $T_{m'}$ or between 5 and 10 s (see Figure 2).

In Figure 11, τ is plotted against the turbulent dissipation rate and normalized *mss*, for T_{min} equal 5 and 10 s. Increasing T_{min} leads to slightly larger values of τ , but in either case a similar trend is apparent, namely, that steeper waves and increased dissipation are associated with an increase in residual foam. As mentioned above, this is likely due to bubbles being injected deeper into the water column for breaking under these conditions. This also shows that the increase in whitecap coverage with these quantities (as in Figures 6 and 8, and 9) is not due to an increase in the rate of breaking alone, but is also the result of a transition to larger, longer-lasting, whitecap events.

Finally, in relating whitecap measurements to wave dissipation, it is tempting to switch from W to $\Lambda(c)$. But would it improve the results? The potential gain is from the additional dependence on the breaker speeds. However, as discussed in section 1, not only is the measurement of breaking speed problematic, most studies have shown the overall shape of $\Lambda(c)$ to be remarkably consistent. All together, *Gemmrich et al.* [2008], *Thomson et al.* [2009], *Kleiss and Melville* [2010], *Gemmrich et al.* [2013], and *Schwendeman et al.* [2014] produce dozens of $\Lambda(c)$ with similar unimodal distributions, peaking somewhere in the range of 0.2–0.5 times the dominant phase speed, and decaying at a rate of around c^{-6} . In other words, the main dynamic parameter in $\Lambda(c)$ is its amplitude, or the total length of breaking crests per area, which is similar to the whitecap coverage. It is not surprising that the large scatter in W over the years [see *Anguelova and Webster*, 2006] is mirrored by similar variability in the breaking strength, *b* [see *Schwendeman et al.*, 2014], since these

approaches share many similar sources of uncertainty (e.g., image processing, microbreaking and residual foam, noisy ancillary measurements). For example, the negative residual trend of W with wave age shown here is a similar result to the increase in b with wave age shown in *Gemmrich et al.* [2008]. One advantage of $\Lambda(c)$ is that it is a spectral measurement, and could potentially be used to calculate spectral wave dissipation. In practice, however, small errors in $\Lambda(c)$ are exaggerated by the calculation of the fifth moment, particularly for large c. Furthermore, if the breaking strength is indeed spectral (as in *Romero et al.* [2012]), that value has not yet been well validated.

5. Conclusion

Whitecap coverage was measured during two experiments in the North Pacific Ocean, and was found to correlate in varying degrees with wind speed and wind stress, wave steepness, and turbulent dissipation. Interestingly, no significant differences in *W* were found when decreasing the camera pixel footprint by an order of magnitude in the 2015 experiment. The coefficients and errors for a threshold power law fit to *W* and each variable are compiled in Tables 2 and 3. The residuals of these fits were used to investigate potential biases in the parameterizations.

The fit to U_{10} has the best overall statistics, and shows good agreement with U_{10} relations from similar recent studies [*Stramska and Petelski*, 2003; *Sugihara et al.*, 2007; *Callaghan et al.*, 2008a], particularly where the density of data is high (roughly $7 \le U_{10} \le 14$ m/s). There is a general leveling off of whitecap coverage at high winds, as was also shown in *Callaghan et al.* [2008a]. There is some minor evidence of bias in the $W(U_{10})$ relation due to wave development, as inferred from measurements of wind acceleration and wave age. However, the variability in the residuals is large relative to these secondary trends.

Of the wave slope parameterizations, the best correlation is with the mean square slope of the equilibrium range waves, defined as $\sqrt{2}f_m \le f \le \sqrt{5}f_m$. This is probably because the majority of whitecaps form in this range. Meanwhile, other bulk steepness values are based on the dominant waves or are influenced by swell. The worst parameterizations involve the "peak" variables, which may be completely associated with swell. The fit to equilibrium range *mss* is further improved by normalizing by the directional spread and frequency bandwidth.

Whitecap coverage does correlate with turbulent dissipation, but the variability is larger than for the wind or wave parameters. The best fit is not linear, as has frequently been proposed, but the large scatter means a linear trend cannot be completely dismissed. The residuals show a strong trend with wave age, possibly related to microbreaking. There are clear parallels to the $\Lambda(c)$ literature, due to many similar sources of uncertainty.

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