INT. J. REMOTE SENSING, 1986, VOL. 7, NO. 5, 627–642

Review Article

Whitecaps and the passive remote sensing of the ocean surface

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(Received 10 October 1985)

Abstract. Whitecap coverage (W), which influences the apparent microwave brightness temperature and short-wave albedo of the sea surface, is not only a strongly non-linear function of the 10 m-elevation wind speed (U), but also varies with changes in the stability of the lower atmosphere (i.e. with alterations in the water–air temperature difference ΔT), and with changes in the surface-sea water temperature (T_w). Thus wind retrieval algorithms to be applied to open ocean data from whitecap-detecting satellite instruments should ideally be of the form, $U(W, \Delta T, T_w, d)$, where d is a measure of the effective wind duration. The wind speed associated with the onset of whitecapping, while also varying with ΔT and T_w, is typically 3 to $3{\cdot}5\,\mathrm{m\,s^{-1}}$, not the often quoted $7\,\mathrm{m\,s^{-1}}$.

1. *Introduction*

Anyone who has stood on a clifftop and looked seaward on a windy day is aware that whitecaps are brighter than those regions of the sea surface which are devoid of bubble rafts. Indeed, the albedo of a recently formed whitecap is about ten times the albedo of the adjacent, roughened, sea surface. By contrast, in the microwave portion of the electromagnetic spectrum, a whitecap is a very poor reflector of incident radiation, with an emissivity almost as great as that of a blackbody, while the bubble-free regions are characterized by relatively high microwave reflectivities, often exceeding 60 per cent. Since at microwave wavelengths the sea has a potential to emit more energy per unit area than it could be expected to reflect, a whitecap with its high emissivity represents a local 'hot spot' on the sea surface, and as such causes a high resolution radiometer to record an elevated apparent brightness temperature when such a bubble patch fills a significant fraction of the main lobe of the radiometer antenna, as has been demonstrated by Ross *et al.* (1970).

While it is clear that a satellite-borne instrument whose antenna footprint is several orders of magnitude wider than an individual whitecap will not resolve, or indeed detect, a lone foam patch, it is equally obvious that when the sea state is such that upwards of 1 per cent of the ocean surface is covered with bubble rafts the radiance measured by a satellite sensor will be significantly affected by the aggregate influence of the many whitecaps which lie within its field of view at any instant. The contribution of oceanic whitecaps to the short-wave albedo of the sea surface has been described by Gordon and Jacobs (1977) and the role of whitecaps in elevating the sea surface microwave brightness temperature has been pointed out by Gloersen and Barath (1977).

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An obvious approach to modelling the influence of whitecaps on passive satellite measurements of the visible and microwave character of the sea surface is to combine laboratory, shipboard and low-elevation aircraft measurements of the relevant electromagnetic properties of individual whitecaps with expressions, based primarily on shipboard observations, describing the dependence of open-ocean fractional whitecap coverage upon the wind speed, the thermal stability of the lower atmosphere, etc. One can then predict the integrated effect of the many individual whitecaps that will be present, under the given meteorological conditions, on the satellite signature from a large sea-surface area.

A more comprehensive, and thus perhaps more satisfactory, approach would involve first calculating from physical theory the pertinent electromagnetic properties of sea surface foam, and then combining these results with a model of fractional foam coverage based on a hydrodynamic model which would predict the frequency of wave breaking as a function of the various governing meteorological and oceanographic parameters. Some such model, based on either the empirical or theoretical approach, or resulting from a combination of these approaches, might logically be used in the initial engineering design studies for the passive microwave and visible sensors of any satellite dedicated to marine applications. Certainly, such a model should prove useful in the interpretation of the data from such satellite instruments and, ideally, could provide the basis of an algorithm for the extraction of such quantities as near-surface wind speeds from such satellite measurements. Algorithms based on such physical models would be amenable to extrapolation to conditions not encountered during an instrument's post-launch calibration, which is not the case for any algorithm obtained by simply correlating initial satellite readings with limited sea-truth measurements.

The microwave properties of whitecap foam have been deduced from theory by Droppleman (1970), Rosenkranz and Staelin (1972) and Wentz (1974, 1983) and has been measured by Williams (1969, 1971). Descriptions of the variation with wind speed of the microwave signature of oceanic whitecaps, based on the results of aircraft measurements, have been reported by Ross *et al.* (1970), Nordberg *et al.* (1971) and Webster *et al.* (1976), while a similar description based on data from the satellite-mounted Scanning Multichannel Microwave Radiometer (SMMR) has been provided by Wentz (1983).

The albedo of whitecap foam has been measured by Whitlock *et al.* (1982) and derived theoretically from a foam model of simplified geometry by Stabeno and Monahan (1985). The effective albedo of whitecaps has been investigated by Koepke (1985) and the variation with wind speed of the whitecap contribution to the albedo of the sea surface has been discussed by Gordon and Jacobs (1977) as well as by Stabeno and Monahan (1985).

In spite of all this modelling activity none of the aforementioned models includes a sufficiently adequate and explicit description of the variation of oceanic whitecap coverage with the relevant meteorological and oceanographic parameters to make the model amenable to effective extrapolation over the range of global sea surface conditions. While acknowledging that no entirely adequate description of the dependence of oceanic whitecap coverage, W, upon 10 m elevation wind speed, U, and upon all the other relevant environmental parameters is as yet available, it should be noted that the current models dealing with the effect of whitecaps on satellite measurements do not adequately reflect the present state of understanding of the factors that control whitecapping.

What follows will be devoted to a description of what is presently known about the dependence of oceanic whitecap coverage upon the various marine environmental factors in the hope that this information will prove to be of use to the remote-sensing community.

2. The onset of whitecapping

Some four decades ago Munk (1947) postulated the existence of a critical wind speed associated with the conjectured transition of the sea surface from hydrodynamically smooth to hydrodynamically rough, a transition identified with the onset of a Kelvin–Helmholtz instability. In support of this contention, observations suggesting 'discontinuities' at Beaufort wind force 4 in the sea-surface drag and evaporation coefficients, and in the soaring patterns of seagulls, were summarized. In this same context, Munk concluded from the analysis of the number of whitecaps found in a limited number of photographs of the sea surface that there was a 'striking' transition from smooth to whitecap-covered sea, and assigned a wind speed of $7\,\mathrm{m\,s^{-1}}$ to this step-function in foam patch concentration. While Munk's remark that the typical size of individual whitecaps increases markedly with strengthening winds has been qualitatively verified (Monahan *et al.* 1984), subsequent observations of open-ocean whitecap coverage do not support the suggestion of a $7\,\mathrm{m\,s^{-1}}$ critical velocity associated with the onset of whitecapping. From the analysis of 432 sea-surface photographs taken during the Barbados Oceanographic and Meteorological Experiment (BOMEX), and during other warm-sea (17·4–30·6°C) cruises, Monahan (1971) identified $4\,\mathrm{m\,s^{-1}}$ as the 10 m elevation wind speed below which whitecap coverage is less than 0·1 per cent. Further consideration of these data, combined with the results from comparably warm seas reported by Toba and Chaen (1973) and with the whitecap observations (599 photographs) obtained during the Joint Air–Sea Interaction Study (JASIN), where the sea surface temperatures fell in the 12·5–14·0°C range (Monahan *et al.* 1981), confirmed that whitecap coverage is only negligible when the wind speed is less than 3 or $4\,\mathrm{m\,s^{-1}}$. Specifically, O'Muircheartaigh and Monahan (1983 a) arrived at $3{\cdot}98\,\mathrm{m\,s^{-1}}$ as the speed for the onset of whitecapping from a detailed consideration of these three data sets, and at 3·80 and $4{\cdot}23\,\mathrm{m\,s^{-1}}$ by the application of two differing statistical approaches to the JASIN data set considered in isolation. In spite of these findings many of the authors working in the field of remote sensing have continued to repeat the claim that whitecaps first appear at a wind speed of $7\,\mathrm{m\,s^{-1}}$ (see, for example, Gloersen and Barath 1977, Zheng *et al.* 1983).

When a further detailed statistical analysis was carried out, involving the three above-mentioned whitecap data sets, supplemented by the Storm Transfer Response Experiment (STREX) whitecap observations (779 photographs) (Doyle 1984) obtained in the Gulf of Alaska (surface sea-water temperatures between 5·1 and 11·1°C) and the results from the 1983 Marginal Ice Zone Experiment (MIZEX 83) (322 photographs) (Monahan *et al.* 1984) conducted in Fram Strait (surface water temperatures between −1·4 and 14·4°C), it became apparent that there was no unique threshold velocity for whitecapping, but rather that the wind speed at which whitecaps first appear, henceforth referred to as the Beaufort velocity, U_B, varies in response to changes in various meteorological and oceanographic parameters.

The strong dependence of the Beaufort velocity on the thermal stability of the lower atmosphere, as indicated by ΔT, the sea surface water temperature (T_w) minus the deck height air temperature (T_a), is apparent from the character of the surface depicted in figure 1. The quantity plotted along the ordinate of this figure is the probability that no whitecap will be found in any of the set of shipboard photographs (typical number, 10)

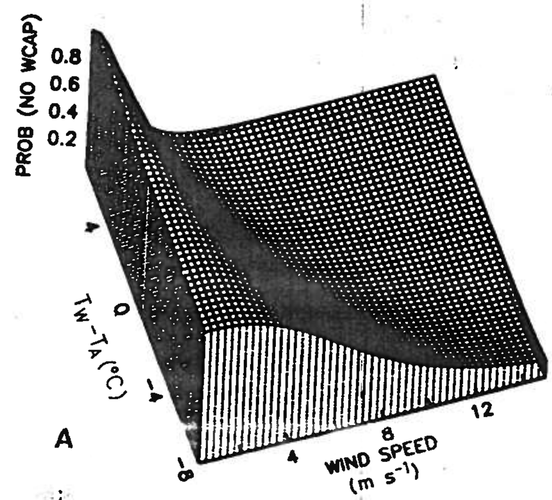

Figure 1. The probability that no whitecaps will be found in any of the set of (typically ten) photographs analysed from a particular whitecap observation interval, plotted against 10 m elevation wind speed and $T_w - T_a$, the surface sea-water temperature minus the deck height air temperature.

analysed from a particular whitecap observation interval with the associated ΔT, U values. It is manifest from figure 1 that there is a good chance of finding no whitecaps even at wind speeds as high as 5 or 6 m s^{-1} when the lower atmosphere is very stable (i.e. for large negative ΔT), but, conversely, when the lower atmosphere is markedly unstable (i.e. ΔT is large and positive) whitecaps will probably be present at wind speeds as low as 2 or 3 m s^{-1}. If the wind speed at which there is a 50 per cent probability of detecting one or more whitecaps in the set of photographs corresponding to an observation interval is adopted as the arbitrary but explicit operational definition of U_B, then the dependence of U_B on ΔT is as shown in figure 2 and is given by

$$U_B = 3{\cdot}27 \times 10^{-0{\cdot}0458\Delta T}\ \mathrm{m\,s^{-1}} \tag{1}$$

It is to be noted that in the case where the lower atmosphere is neutrally stable, i.e. where ΔT is about 0·1 deg C, the U_B obtained from equation (1) is 3·24 m s^{-1}, which is consistent with the non-stability-specific values previously reported (see, for example, Monahan *et al.* 1981). This is understandable, since, well away from coasts and remote from regions of strong horizontal gradients of surface water temperature (such as would be encountered in the vicinity of the Gulf Stream), the marine atmospheric boundary layer is usually close to a condition of neutral stability. The actual distribution, in terms of ΔT, of the 305 whitecap observation intervals that make up the five data sets described above is shown in figure 3.

While the influence of ΔT, i.e. atmospheric stability, on U_B may in many instances be ignored because of the prevalence of near-neutral conditions over much of the world's ocean, the influence on U_B of sea-water surface temperature must always be considered. Using the explicit definition of U_B provided previously, the dependence of this Beaufort velocity on sea surface water temperature, as deduced from the aforementioned five data sets, is given by

$$U_B = 3{\cdot}36 \times 10^{-0{\cdot}00309 T_w}\ \mathrm{m\,s^{-1}} \tag{2}$$

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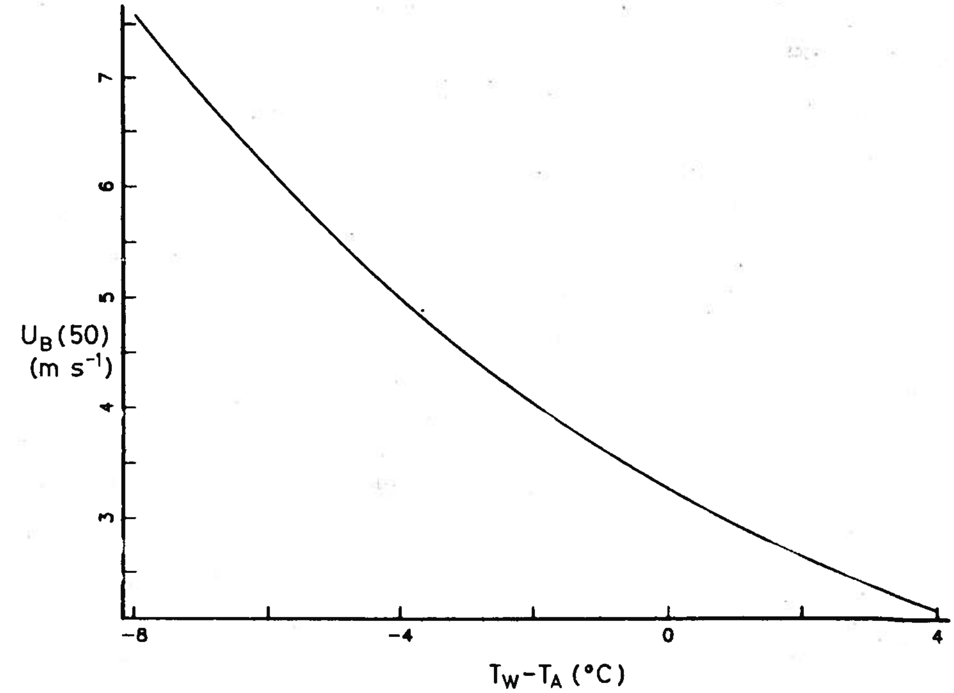


Figure 2. $U_B(50)$, the Beaufort velocity, defined as the 10 m elevation wind speed at which there is a 50 per cent probability of encountering one or more whitecaps in the set of (typically ten) sea-surface photographs from a particular whitecap observation interval, versus $T_w - T_a$. This curve corresponds to the intersection of the surface of figure 1 by a plane parallel to the U and $T_w - T_a$ axes which intersects the Prob (No Wcaps) axis at 0·5.

and depicted in figure 4. A histogram showing the distribution in T_w of all the whitecap observation intervals from these BOMEX, East China Sea, JASIN, STREX and MIZEX 83 data sets is included as figure 5. The decrease of U_B with increasing T_w may be an indication that air is entrained at lower wind speeds when the water is less viscous. The extent by which kinematic viscosity decreases as water temperature increases can be seen in figure 6. The analysis of the numerous whitecap photographs recently acquired during MIZEX 84 (Monahan *et al* 1984), when completed, will result in the addition of many more data points associated with T_w's less than 5°C to the combined data set. It will be most interesting to see how the inclusion of these additional cold-water data will alter the strong increase in U_B as T_w decreases towards 0°C apparent in the statistical analysis of the already available cool and cold sea-water whitecap data sets (JASIN, STREX, MIZEX 83), when taken in isolation. Until the MIZEX 84 whitecap data are included in the general analysis, Equation (2) must be considered to be provisional.

It should now be apparent that the threshold wind velocity above which whitecaps can be expected to cause changes in the sea surface emissivity and albedo which may be detectable by various extant and proposed satellite systems, being tantamount to U_B, is closer to 3 m s^{-1} than to the value of 7 m s^{-1} which is still usually quoted in the remote-sensing literature. Furthermore, this threshold velocity for whitecap influence, varying as does U_B with ΔT and T_w, can be expected to vary geographically and seasonally in a predictable fashion.

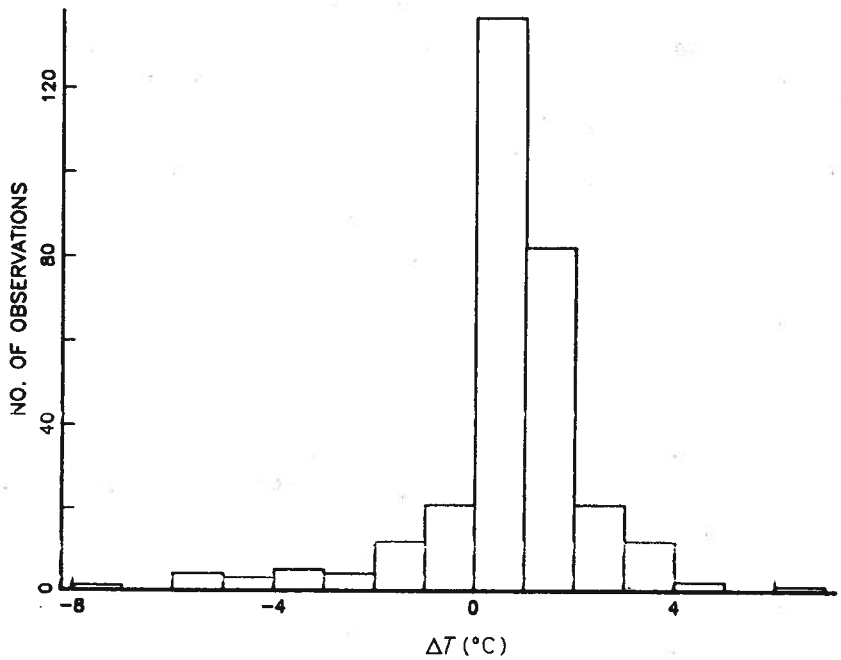


Figure 3. Distribution in ΔT, i.e. $T_w - T_a$, of the 305 whitecap observation intervals that comprise the BOMEX, East China Sea, JASIN, STREX and MIZEX 83 data sets.

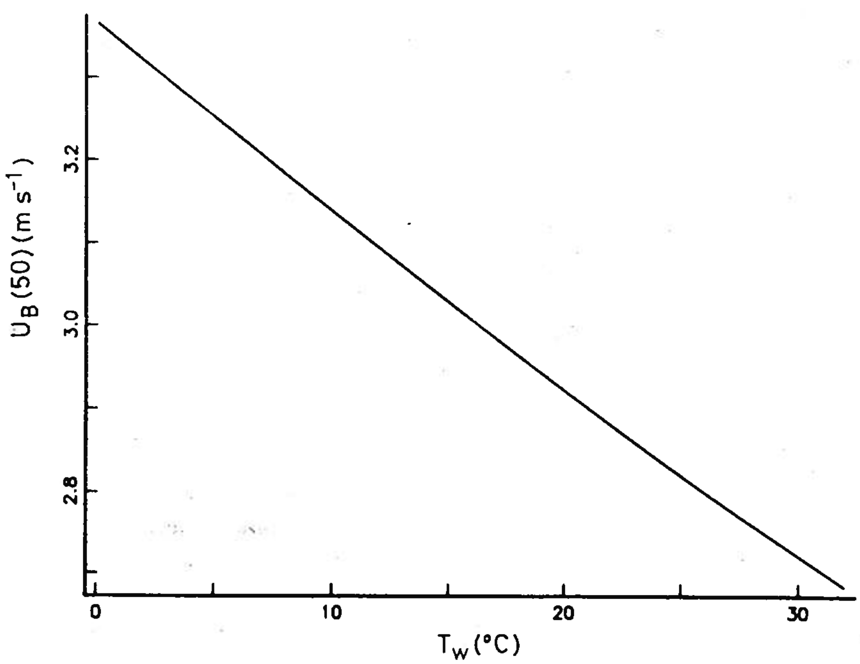
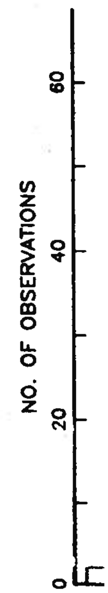


Figure 4. The Beaufort velocity, $U_B(50)$, versus T_w, the sea surface temperature.



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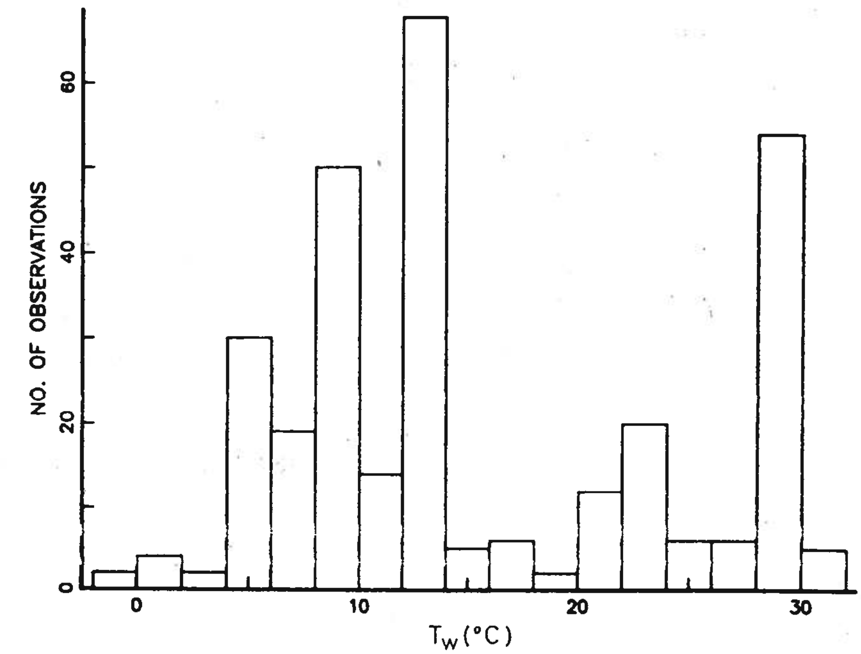

Figure 5. Distribution in T_w of the 305 whitecap observation intervals that make up the five data sets.

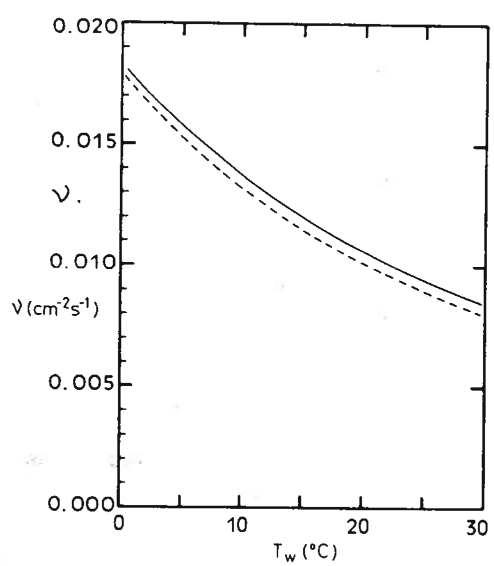

Figure 6. The kinematic viscosity, ν, of sea water of 35‰ (solid line), and of freshwater (dashed line), versus surface water temperature, T_w.

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3. Variations in whitecap coverage

Having described the conditions for the onset of whitecapping, it is now appropriate to describe how, at wind speeds above U_B, whitecap coverage varies with 10 m elevation wind speed U, and with the other relevant environmental parameters. A simple power law will be used to describe $W(U)$, as this type of expression has been used, with few exceptions, for this purpose since Blanchard (1963) first quantitatively described the dependence of whitecap coverage on wind speed using a simple quadratic equation. Monahan and O'Muircheartaigh (1980), applying ordinary least squares (OLS) and robust biweight fit (RBF) analyses to the two warm-water whitecap data sets (BOMEX, $T_w = 26{\cdot}9°C$, Monahan (1971), East China Sea, $T_w = 24{\cdot}7°C$, Toba and Chaen (1973)) obtained the power-law

$$W_{OLS} = 2{\cdot}95 \times 10^{-6} U^{3{\cdot}52}, \qquad U > U_B \tag{3 a}$$

$$W_{RBF} = 3{\cdot}84 \times 10^{-6} U^{3{\cdot}41}, \qquad U > U_B \tag{3 b}$$

where U is measured in metres per second, and W is expressed as a simple fraction. These expressions are consistent with the conclusion of Wu (1979), based primarily on theoretical grounds, that on the open ocean with a fully developed sea, W should be proportional to the cube of the friction velocity u^*, u^{*3} being a measure of the rate at which the wind is doing work on the sea. It must be noted that equations (3 *a*) and (3 *b*) share with all other simple power-law $W(U)$ expressions the character of predicting a jump from a W value of zero to a finite $W(U_B)$ at the Beaufort velocity. A more consistent formulation would be

$$W = A(U^{\lambda} - U_B^{\lambda}), \qquad U > U_B \tag{4}$$

where the quantity AU_B^{λ} is a measure of the maximum rate-of-work by the wind on the sea that can be spent in wave breaking without the entrainment of air and hence without the formation of whitecaps.

The dependence of oceanic whitecap coverage on wind speed and low-elevation atmospheric stability (ΔT), as deduced from a consideration of all five available data sets, is shown in figure 7. This result is in accord with the known dependence of $u^*(U)$ on stability. For a fixed value of U, u^* will increase as the atmosphere becomes less stable, i.e. as ΔT shifts towards positive values. These same findings are summarized by

$$W = 1{\cdot}95 \times 10^{-5} U^{2{\cdot}55} \exp(0{\cdot}0861 \Delta T) \tag{5}$$

It is to be noted that the exponent of U here is smaller than the exponents obtained from a consideration of the warm-water data sets alone (equations (3 *a*) and (3 *b*)), a matter that will be returned to when the influence of T_w on U is discussed. Likewise, for a U of 10 m s^{-1} and a ΔT of 0 deg C, the W value predicted by equation (5) is some 30 per cent below the W values obtained using equations (3 *a*) and (3 *b*).

Since the reduction in W predicted by equation (5), when the air temperature goes from 1 deg C cooler to 1 deg C warmer than the water temperature while the wind speed remains fixed, is in excess of 15 per cent, it is clear that the influence of stability must always be considered when attempting to estimate U from satellite measurements of quantities proportional to W.

The dependence of whitecap coverage upon surface water temperature is not as clearly defined as is its dependence on ΔT. An analysis utilizing the BOMEX and East China Sea whitecap data sets suggests only a modest increase in W with increasing T_w and constant U, as can be seen from figure 8. In contrast to these findings, the results of Bortkovskii (1983), as summarized in figure 9, indicate a strong positive dependence of

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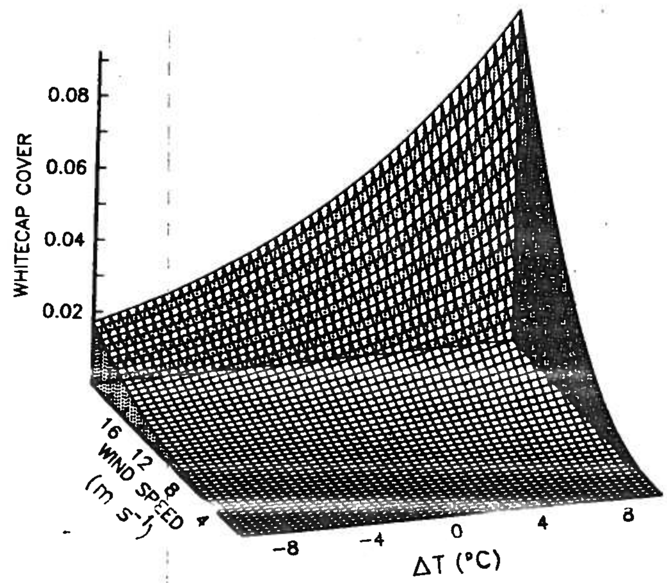

Figure 7. Oceanic whitecap coverage, expressed as a fraction, plotted against ΔT and 10 m elevation wind speed. Surface determined from a consideration of all five data sets.

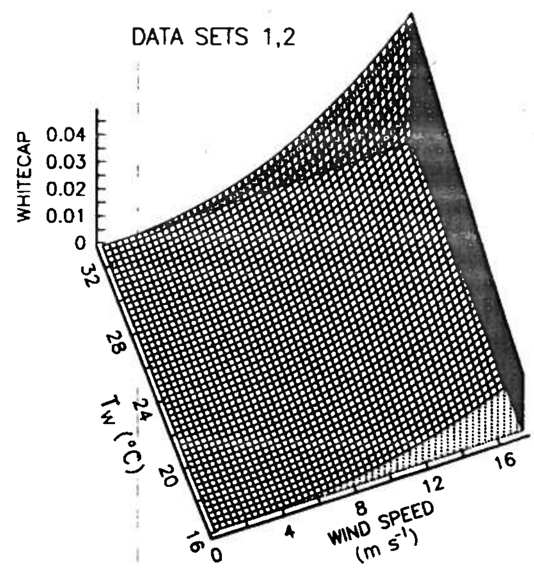

Figure 8. The fraction of the sea surface covered by whitecaps, as a function of 10 m elevation wind speed and T_w. Figure derived from the two warm-water data sets, i.e. the BOMEX and East China Sea sets.

W on T_w. It is to be hoped that when the MIZEX 84 cold-water whitecap data set is added to the other five data sets a clearer picture of the dependence of *W* on T_w will emerge.

It was pointed out earlier that the power-law exponent λ determined for the composite data set, as given in equation (5), was smaller than the λ values obtained

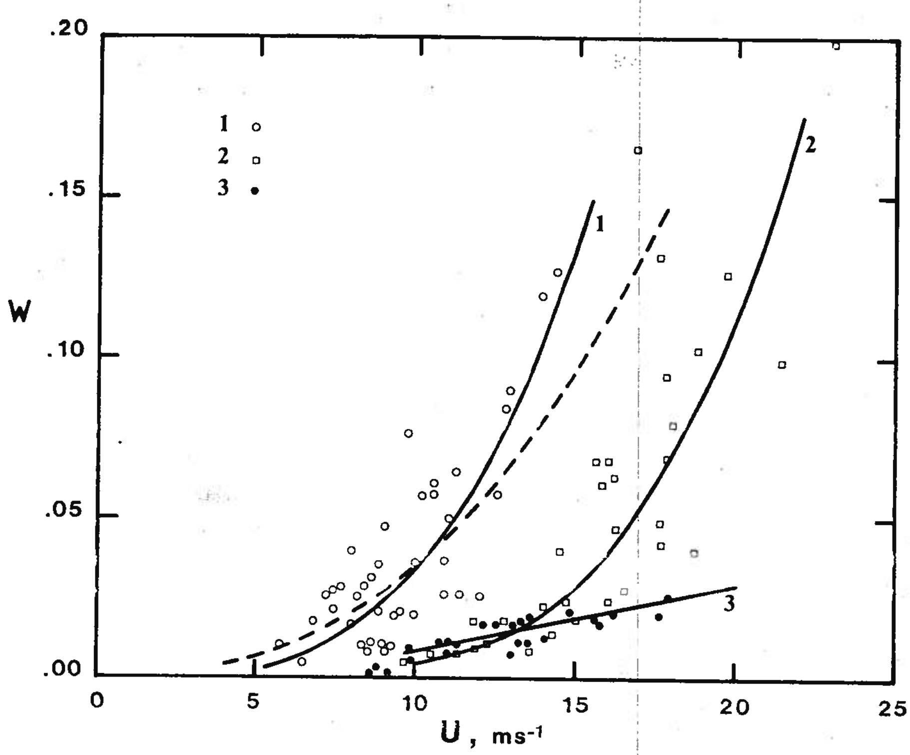

Figure 9. Oceanic foam coverage (active whitecaps plus foam streaks), expressed as a fraction of the sea surface, versus wind speed. This figure is based on figure 2.3 in Brotkovskii (1983). Curve 1: description of Brotkovskii's warm-water (T_w of 27°C) whitecap measurements (open circles); curve 2: description of whitecap data (open squares) collected in waters of intermediate temperatures ($3 < T_w < 15$°C); and curve 3: description of coldwater ($T_w < 3$°C) whitecap observations (solid circles). Curve 1 also corresponds with equation (1) of Monahan (1971), where it was used to describe an envelope over all the BOMEX ($\bar{T}_w$ of 26·9°C) whitecap observations in the wind speed range from 4 to 10 m s^{-1}. The dashed curve is an alternate, OLS, power-law description of Bortkovskii's warm-water observations.

from the analyses of the two warm-water data sets taken in isolation, as reflected in equations (3 *a*) and (3 *b*). This observation led to the suggestion that λ might vary with T_w. Each of the five whitecap data sets was analysed in three ways: (i) a linear regression of ln W on ln U was carried out, excluding, of necessity, the null Wpoints; (ii) a similar analysis was carried out, but in this second treatment all W values less than 1×10^{-4} were taken to be zero in accord with the approach of Wu (1979); and (iii) an OLS fit was obtained, taking into account all non-zero W data points. These results were then combined with the λ values incorporated in equations (3 *a*) and (3 *b*), i.e. the values obtained by applying the OLS and RBF techniques to the BOMEX and East China Sea data sets excluding the low W points as per Wu (Monahan and O'Muircheartaigh 1980); the λ values determined in the same manner from the JASIN data (Monahan *et al.* 1983); and the λ value obtained by the use of a Box–Cox transformation on the same JASIN data set (O'Muircheartaigh and Monahan 1983 b). Finally, the average exponent for each data set, $\bar{\lambda}$, was calculated, as was the mean sea-water temperature associated with each set, $\bar{T}_w$, and the resulting $\bar{\lambda}$, $\bar{T}_w$ points were plotted (figure 10). These results support the hypothesis that λ increases with T_w.

An explanation for this variation of λ with T_w rests on the observation that both typical wind duration and mean sea-water temperature vary latitudinally. Many of the warm-water observations were made in trade-wind regions, where even the fairly high winds persist long enough to produce a fully developed sea, one commensurate with those winds. On the other hand, the whitecap observations made over colder waters were taken in regions such as the Atlantic north-west of Ireland, the Gulf of Alaska and Fram Strait, where the typical duration associated with high wind events was often quite short, and thus the wave spectrum had usually not become fully saturated. Now Cardone (1969) concluded from his analysis of a freshwater whitecap data set (Monahan 1969) that the rate at which energy was being transferred from the wind to the fully developed portion of the wave spectrum, which he took to be proportional to the rate at which it was being dissipated through wave breaking, was proportional to whitecap coverage, and Monahan (1971, 1985) subsequently refined this statement by substituting the rate of whitecap area formation in place of instantaneous whitecap coverage. It follows that in the colder regions such as where the JASIN, STREX and MIZEX 83 observations were made, the rate of whitecap formation and the whitecap coverage would frequently be less than they would have been in the case of a fully developed sea, and it would be expected that this reduction in whitecap coverage associated with the lack of spectral saturation, would become more pronounced, the

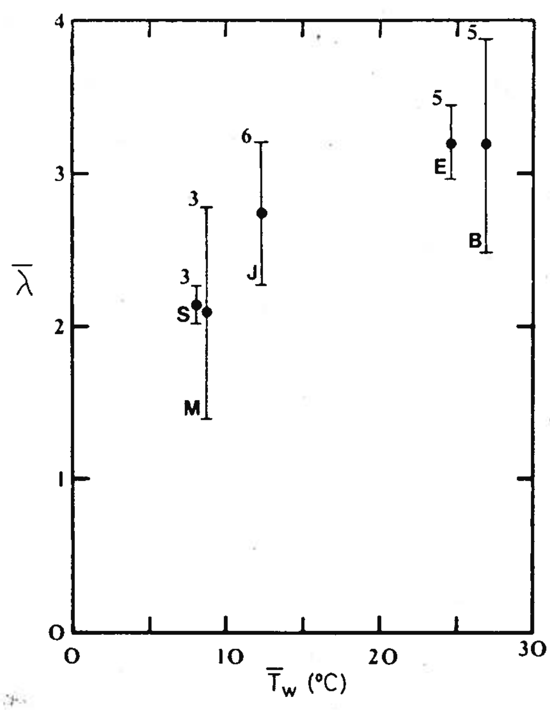

Figure 10. The average power-law exponent, $\bar{\lambda}$, determined using three or more statistical models, for each of the five whitecap data sets, versus $\bar{T}_w$, the average sea surface temperature associated with each data set. S, STREX; M, MIZEX 83; J, JASIN; E, East China Sea; and B, BOMEX plus. The numbers indicate the numbers of λ's determined for each set. The extent of the vertical bars, representing $\sigma(\lambda)$, reflects the sensitivity of the λ determinations to the statistical approaches chosen.

higher the windspeed category, and the shorter the wind duration. Thus, while W_{sat} may vary as $U^{3\cdot 2}$ (figure 10), W_{unsat} would be expected to vary as U^{λ}, where λ here will be less than 3·2 by an amount that will increase as the typical wind duration associated with a data set decreases.

Two additional factors will contribute to an increase in W with increasing T_w, for fixed values of U. As has been pointed out by Donelan and Pierson (1985), the rate at which wind energy is being transferred to the waves will, in the case of an equilibrium wave spectrum be balanced by the rate at which energy is being dissipated due to breaking plus the rate of viscous energy dissipation. Now the rate of viscous dissipation will decrease markedly as T_w increases, because the kinematic viscosity of sea-water decreases significantly as T_w increases over the range from 0 to 30°C, as is illustrated in figure 6. Thus, as T_w increases, more and more of the energy coming from the wind will be dissipated in relatively large-scale wave breaking, i.e. in whitecap formation.

As mentioned previously, it is $\dot{W}$, the rate of whitecap area formation, that is proportional to the rate at which energy is being dissipated through wave breaking. Now $\dot{W}$, W and τ, the exponential time constant characterizing the decay of individual whitecaps, are interrelated by (Monahan 1971)

$$\dot{W} = W\tau^{-1} \tag{6}$$

It has been shown that τ varies inversely with the terminal rise velocity of the small bubble fraction that contributes to the whitecap subsurface bubble cloud (Monahan 1985), and Pounder (1985) has demonstrated that the bubble size distribution shifts towards smaller radii as the water temperature is increased. Since smaller bubbles have much slower terminal rise velocities than do larger bubbles, τ increases as T_w increases. This effect far outweighs the influence on τ of the increase in terminal rise velocity of bubbles of a fixed radius, due to a decrease in viscosity, that is observed when T_w is increased. It then follows from equation (6) that for a fixed $\dot{W}$, W will increase with increasing T_w. This consideration, when coupled with the increase in $\dot{W}$, for fixed U, with increasing T_w associated with a continuing shift of spectral dissipation towards wave breaking and away from viscous dissipation due to the decreasing kinematic viscosity, leads to the conclusion that for fixed U, ΔT and duration, W should increase with increasing T_w. If the apparent microwave brightness temperature, T_B, measured by the SMMR 10·7 GHz channel, horizontal polarization, is taken as a measure of whitecap coverage, then the SMMR wind retrieval algorithm, as described by Cardone *et al.* (1983), is qualitatively in agreement with these conclusions, in that for a fixed T_B, i.e. for a constant W, the wind speed U decreases as the sea surface temperature T_w increases.

A factor influencing W, in addition to U, ΔT, T_w and duration, is the fetch of the wind. Cardone's (1969) theory for whitecap coverage, based on the wave spectrum approach of Pierson and Moskowitz (1964), predicts less whitecapping in cases of limited fetch (e.g. figure 3 for Ross and Cardone (1974)). This model successfully fitted the whitecap observations made on the Laurentian Great Lakes (Monahan 1969). The effect of fetch on W predicted by Cardone's model also seems to be borne out by the preliminary findings from the study of some 1500 Alte Weser lighthouse observations (Monahan and Monahan 1985).

While the relatively modest variations of surface sea-water salinity, S, encountered in the open oceans of the world may not lead to significant variations in whitecap coverage, the same contention certainly cannot be made when comparing oceanic whitecap coverage with the whitecap coverage observed on freshwater lakes. A variety

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of laboratory studies (Miyake and Abe 1948, Monahan and Zietlow 1969, Scott 1974) have demonstrated that the bubble spectra produced in fresh- and salt-water are markedly different. The freshwater bubble spectrum does not extend down to the small bubble sizes encountered in sea-water, perhaps, as suggested by Scott, because the smallest bubbles in the freshwater case immediately coalesce. Since the smallest bubbles in the freshwater bubble cloud are much larger than the smallest bubbles in the sea-water bubble cloud, the freshwater whitecap τ is only about 66 per cent as large as the sea-water τ (Monahan and Zietlow 1969). It follows from equation (6) that for the same U, ΔT, T_w, duration, d, and fetch, f, there should be some 51 per cent greater whitecap coverage on a sea- than on a freshwater body. This conclusion is supported by the comparison of Great Lakes and oceanic whitecapping reported by Monahan (1971).

In addition to $U, \Delta T, T_w, d, f$ and S it has been suggested that W should also vary with the concentration of dissolved organic material in the water and with the presence or absence of an organic monolayer on the sea surface. Certainly, an organic coating can stabilize a bubble and allow it to persist indefinitely on the sea surface (Abe 1962). But during MIZEX 84 it was observed that the only persistent bubbles were those on the outer rim of a whitecap. These bubbles are the first to have reached the surface, and thus are the ones that, having scavenged some of the organic material from the bulk water as they rose towards the sea surface, came into contact with the relatively intact organic film when they reached that surface. The presence of the many bubbles in the whitecap cloud make the effective density of the associated volume of water less than that of the surrounding, relatively bubble-free, water. As a consequence, there is a rising buoyant plume associated with the typical whitecap which gives rise to a horizontally divergent flow at the surface. This flow carries the first, stabilized, bubbles and the surface film outwards from the centre of the plume and thus leaves the water surface above the centre of the bubble cloud quite free of organic contamination. Therefore, the later, typically smaller, bubbles rise to strike a clean surface and burst almost immediately. The small halo of stabilized bubbles in the MIZEX examples did not represent a significant fraction of the initial whitecap area, and as a consequence could not be expected to alter estimates of either τ or W markedly. In regions of the ocean characterized by extreme organic richness it is possible that the influence of this organic material on W may be profound. Certainly bubble streaks or rafts are often encountered along lines of surface convergence, where as a consequence of the convergent flow there is a high concentration of buoyant organic material.

The observational evidence clearly suggests that the early $W(U)$ expressions need to be replaced by more comprehensive $W(U, \Delta T, T_w, d, f, S)$ formulations. At present the $W(U, \Delta T)$ expression that appears as equation (5) is the most general statement of the dependence of whitecap coverage on environmental parameters. It is to be hoped that a further analysis of all the extant whitecap data sets, supplemented by the MIZEX 84 set, will soon shed light on the influence of T_w on W due to the effect of T_w on the bubble spectrum and on the rate of viscous dissipation, and on the independent effect of d on W, so that a quantitative expression for $W(U, \Delta T, T_w, d)$ can be made available.

4. Conclusions

While the descriptions, in terms of the basic meteorological and oceanographic parameters, of the Beaufort velocity, U_B, and of whitecap coverage, W, provided in the foregoing sections are by no means complete, these formulations should be more useful to those designing satellite sensors capable of detecting sea surface foam than those now quoted in the remote-sensing literature. It follows from the description of bubble spectra that the practice of using a foam patch, produced in freshwater and stabilized with detergent, to calibrate microwave sensors that are to be employed to detect salt-water whitecaps (see, for example, Williams 1969, Whitlock *et al.* 1982) is one that might well be revised.

It is worth noting that the optimal $U(W, \Delta T)$ expression for use in a wind retrieval algorithm cannot be obtained by simply inverting a $W(U, \Delta T)$ expression such as equation (5). The optimal $U(W)$ expressions determined from the same two whitecap data sets as were used to obtain equations (3 *a*) and (3 *b*) are given by (Monahan and O'Muircheartaigh 1981)

$$U_{\mathrm{OLS}} = 22{\cdot}64 W^{0{\cdot}19} \tag{7 a}$$

$$U_{\mathrm{RBF}} = 24{\cdot}63 W^{0{\cdot}23} \tag{7b}$$

A further study, incorporating the JASIN data and considering the dependence of U on T_{a} and T_{w}, as well as on W, is described in Monahan *et al.* (1981).

Acknowledgments

This paper was written during a summer spent by E.C.M. as a Visiting Investigator at the Woods Hole Oceanographic Institution. His visit was supported via the U.S. Office of Naval Research Contract NOOO14-85-M-0065, NR 211-229. During this period I.G.O'M. was a Senior Research Associate at the Naval Postgraduate School, Monterey, where his stay was sponsored by the U.S. National Research Council. Many of the observations discussed in this paper (JASIN, STREX, MIZEX 83 and MIZEX 84) were collected as part of the University College, Galway, Whitecap Project. This ongoing effort, supervised by E.C.M., has been sponsored throughout by ONR via Grant N00014-78-G-0052. The assistance in the field of E. M. Murphy, D. M. Doyle, P. A. Bowyer and of Professor K. L. Davidson and his NPS colleagues is gratefully acknowledged. The careful work of the numerous photographic analysts in Galway is recognized with thanks, and the authors are grateful to Prof. D. Gaver of NPS for his statistical insights. This paper is Contribution No. 6110 from the Woods Hole Oceanographic Institution.

References

ABE, T., 1962, On the stable foam formation of sea water in seas (preliminary report). *J. Oceanog. Soc. Japan,* 20th Anniversary Volume, 242.

BLANCHARD, D. C., 1963, The electrification of the atmosphere by particles from bubbles in the sea. *Progress in Oceanography,* Vol. 1 (Oxford: Pergamon Press), pp. 71–202.

BORTKOVSKII, R. S., 1983, *Heat and Moisture Exchange between Atmosphere and Ocean under Storm Conditions* (Leningrad: Hydrometeorological Publishing House) (in Russian).

CARDONE, V. J., 1969, Specification of the wind field distribution in the marine boundary layer for wave forecasting. Tech. Report GSL-69-1, New York University.

CARDONE, V., CHESTER, T., and LIPES, R., 1983, Evaluation of SEASAT SMMR wind speed measurements. *J. geophys. Res.*, **88,** 1709.

DONELAN, M. A., and PIERSON, W. J., JR., 1985. Bragg-scattering and equilibrium ranges in wind-generated waves, with application to scatterometry. *J. geophys. Res.* (in the press).

DOYLE, D. M., 1984, Marine aerosol research in the Gulf of Alaska and on the Irish West Coast (Inishmore). Whitecaps and the Marine Atmosphere, Report No. 6, to the Office of Naval Research from University College, Galway.

DROPPLEMAN, J. D., 1970, Apparent microwave emissivity of sea foam. *J. geophys. Res.*, **75**, 696.

GLOERSEN, P., and BARATH, F. T., 1977, A scanning multichannel microwave radiometer for Nimbus-G and Seasat-A, *I.E.E.E. Jl Oceanic Engng*, **2**, 172.

GORDON, H. R., and JACOBS, M. M., 1977, Albedo of the ocean–atmosphere system: influence of sea foam. *Appl. Optics*, **16**, 2257.

KOEPKE, P., 1985, Remote sensing signatures of whitecaps. In *Oceanic Whitecaps and Their Role in Air–Sea Exchange Processes*, edited by E. C. Monahan and G. MacNiocaill (Galway: Galway University Press).

MIYAKE, Y., and ABE, T., 1948. A study of the foaming of sea-water. Part 1. *J. Mar. Res.*, **7**, 67.

MONAHAN, E. C., 1969, Fresh water whitecaps. *J. atmos. Sci.*, **26**, 1026.

MONAHAN, E. C., 1971, Oceanic whitecaps. *J. Phys. Oceanogr.*, **1**, 139.

MONAHAN, E. C., 1985. The ocean as a source for atmospheric particles. *Proceedings of the NATO Advanced Study Institute Conference on the Role of Air–Sea Exchange in Geochemical Cycling* (Dordrecht: Reidel).

MONAHAN, E. C., FAIRALL, C. W., DAVIDSON, K. L., and JONES BOYLE, P., 1983, Observed inter-relations between 10 m winds, ocean whitecaps and marine aerosols. *Q. Jl. R. met. Soc.*, **109**, 379.

MONAHAN, E. C., and MACNIOCAILL, G., 1985, *Oceanic Whitecaps and Their Role in Air–Sea Exchange Processes* (Galway: Galway University Press).

MONAHAN, E. C., and MONAHAN, C. F., 1985, The influence of fetch on whitecap coverage as deduced from the Alte Weser lightstation observer's log. In *Oceanic Whitecaps and Their Role in Air–Sea Exchange Processes*, edited by E. C. Monahan and G. MacNiocaill (Galway: Galway University Press).

MONAHAN, E. C., and O'MUIRCHEARTAIGH, I. G., 1980, Optimal power-law description of oceanic whitecap coverage dependence on wind speed. *J. Phys. Oceanogr.*, **10**, 2094.

MONAHAN, E. C., and O'MUIRCHEARTAIGH, I. G., 1981, Improved statement of the relationship between surface wind speed and oceanic whitecap coverage as required for the interpretation of satellite data. In *Oceanography from Space*, edited by J. F. R. Gower (New York: Plenum), pp. 751–755.

MONAHAN, E. C., O'MUIRCHEARTAIGH, I. G., and FITZGERALD, M. P., 1981, Determination of surface wind speed from remotely measured whitecap coverage, a feasibility assessment. *Proceedings of the EARSeL-ESA Symposium on Application of Remote Sensing Data on the Continental Shelf*, Voss, 19–20 May 1981, pp. 103–109.

MONAHAN, E. C., SPILLANE, M. C., BOWYER, P. A., HIGGINS, M. R., and STABENO, P. J., 1984. Whitecaps and the Marine Atmosphere, Report No. 7, to the Office of Naval Research from University College, Galway.

MONAHAN, E. C., and ZIETLOW, C. R., 1969, Laboratory comparisons of fresh-water and salt-water whitecaps. *J. geophys. Res.*, **74**, 6961.

MUNK, W. H., 1947, A critical wind speed for air–sea boundary processes. *J. Mar. Res.*, **6**, 203.

NORDBERG, W., CONAWAY, J., ROSS, D. B., and WILHEIT, T., 1971, Measurements of microwave emission from a foam-covered, wind-driven sea. *J. atmos. Sci.*, **28**, 429.

O'MUIRCHEARTAIGH, I. G., and MONAHAN, E. C., 1983 a, Aspects of oceanic whitecap coverage dependence on wind speed; heteroscedasticity in the data, and the estimation of the Beaufort velocity. *Proceedings of the Second International Meeting on Statistical Climatology*, 26–30 September 1983, Lisbon, pp. 2.7.1–4.

O'MUIRCHEARTAIGH, I. G., and MONAHAN, 1983 b, Use of the Box–Cox transformation in determining the functional form of the dependence of oceanic whitecap coverage on several environmental factors, *Proceedings of the Eighth Conference on Probability and Statistics in Atmospheric Sciences*, 16–18 November 1983, Hot Springs, Arkansas, pp. 55–58.

PIERSON, W. J., JR., and MOSKOWITZ, L., 1964, A proposed spectral form for fully developed wind seas based on the similarity theory of S. A. Kitaigorodskii. *J. geophys. Res.*, **69**, 5181.

POUNDER, C., 1985, Sodium chloride and water temperature effects on bubbles. In *Oceanic Whitecaps and Their Role in Air–Sea Exchange Processes*, edited by E. C. Monahan and G. MacNiocaill (Galway: Galway University Press).

ROSENKRANZ, P. W., and STAELIN, D. H., 1972, Microwave emissivity of ocean foam and its effect on nadiral radiometric measurements. *J. geophys. Res.* **77**, 6528.

Ross, D. B., and Cardone, V., 1974, Observations of oceanic whitecaps and their relation to remote measurements of surface wind speed. *J. geophys. Res.*, **79**, 444.

Ross, D. B., Cardone, V. J., and Conaway, J. W., 1970, Laser and microwave observations of sea-surface condition for fetch-limited 17 to 25 m/s winds. *I.E.E.E. Trans. Geosci. Electron.*, **8**, 326.

Scott, J. C., 1975, The preparation of water for surface-clean fluid mechanics. *J. Fluid Mech.*, **69**, 339.

Stabeno, P. J., and Monahan, E. C., 1985, The influence of whitecaps on the albedo of the sea surface. In *Oceanic Whitecaps and Their Role in Air–Sea Exchange Processes*, edited by E. C. Monahan and G. MacNiocaill (Galway: Galway University Press).

Toba, Y., and Chaen, M., 1973, Quantitative expression of the breaking of wind waves on the sea surface. *Rec. Oceanogr. Works Japan*, **12**, 1.

Webster, W. J., Wilheit, T. T., Ross, D. B., and Gloersen, 1976, Spectral characteristics of the microwave emission from a wind-driven foam-covered sea. *J. geophys. Res.*, **81**, 3095.

Wentz, F. J., 1974, The effect of surface roughness on microwave sea brightness temperatures. Contract Report 3–35345, National Environmental Satellite Service, N.O.A.A.

Wentz, F. J., 1983, A model function for ocean microwave brightness temperatures. *J. geophys. Res.*, **88**, 1892.

Whitlock, C. H., Bartlett, D. S., and Gurganus, E. A., 1982, Sea foam reflectance and influence on optimum wavelength for remote sensing of ocean aerosols. *Geophysical Res. Lett.*, **9**, 719.

Williams, G. F., Jr., 1969, Microwave radiometry of the ocean and the possibility of marine wind velocity determination from satellite observations. *J. geophys. Res.*, **74**, 4591.

Williams, G. F., Jr., 1971, Microwave measurements of bubbles and foam. *I.E.E.E. Trans. Geosci. Electron.*, **9**, 221.

Wu, J., 1979, Oceanic whitecaps and sea state. *J. Phys. Oceanogr.*, **9**, 1064.

Zheng, Q. A., Klemas, V., Hayne, G. S., and Huang, N. E., 1983, The effect of oceanic whitecaps and foams on pulse-limited radar altimeters. *J. geophys. Res.*, **88**, 2571.