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On the Relatonship between Wave Breaking and Marine Aerosol Concentration in Deep Sea Areas

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Abstract

Aerosol fluxes from the sea surface are one of the important factors determining the dynamics of the air-sea interaction. Not numerous available data showed that the intensity of aerosol fluxes strongly depends on the intensity of wave breaking. In the paper theoretical formulas to determine the probability of breaking crests and percentage of whitecaps coverage are discussed. These formulas are a starting basis for the set-by-step procedure to determine the aerosol fluxes in deep water under the steady sea state conditions.

1. Introduction

The heights of wind-induced irregular waves do not increase infinitely, but are limited by breaking phenomenon or energy dissipation due to bottom friction. Wave breaking occurs whenever a momentarily high crest reaches an unstable condition. It is an intermittent process and its frequency depends on the severity of the sea. Breaking wind waves play an important role in upper ocean dynamics, including surface layer currents, turbulent mixing and dynamic loading on offshore and coastal structures which may cause serious safety problems and structural damage to these structures. It is therefore highly desirable to estimate the frequency of wave breaking occurrence and the limiting height of breaking waves in evaluation of the hydrodynamic forces.

The dynamics of the wave breaking have a direct influence on the generation of currents and turbulence in the water body, hence on the phenomena depending on the mixing of upper layers of the ocean. Breaking waves disrupt the chemical and organic surface films and produce fluxes of the sea-salt aerosols. Most of the aerosol generated from natural waters is in the form of jet and film drops from the bursting of air bubbles (Monahan and Van Patten 1989). The aerosol droplets may transfer water vapour, heat, pollutants and bacteria through the air-water interface. They can be very easily transported by wind over large distances. In this way, marine aerosols influence the optical features of the atmosphere, which are of fundamental importance for the remote sensing of sea surface, as well as playing an important role in climate changes.

In this paper the relationship between aerosol concentration and wave breaking is discussed. The amount of marine aerosol rising from the sea surface depends on the coverage of the sea by breaking waves or whitecaps, and on the rate of intensity of breaking. The wind speed, commonly used in prediction of the whitecaps coverage, is only one of the factors determining the wave energy and probability of breaking occurring. It is more appropriate to find the linkage between the percentage of sea surface covered by white caps and sea state characteristics, such as the significant wave height H_s and peak frequency ω_p.

The central point in such an approach is the determination of the probability of wave breaking. In this paper, the vertical acceleration threshold concept is used to predict the limiting wave height in wind-induced wave trains in the deep ocean. The threshold acceleration criterion is applied at the wave crest for one-dimensional case and everywhere on the sea surface for two-dimensional case.

This paper is organised as follows. First, in Section Two, the breaking criteria for the irregular waves are determined. Section Three contains analysis of the available experimental data. In Section Four an estimation of energy dissipation in the white-capping is discussed in brief. Section Five summarises the methods of estimating aerosol fluxes in deep waters. Finally in Section Six the main conclusions are formulated.

2. Breaking Criteria and Whitecaps Coverage for Irregular Waves

Estimation of the limiting wave height for wind-induced waves is complicated as our present understanding of most aspects of wave breaking remains fragmentary. Most of the available statistics of breaking waves are based on limiting steepness criterion. Such a criterion is used to identify the part of the joint probability density function of wave height and wave period where the waves are assumed to be breaking (Nath and Ramsey 1974, Ochi and Tsai 1983). The integration over the joint probability function yields the fraction of breaking waves (or probability distribution of breaking wave heights).

For steep regular waves, the crest attains a sharp point with an angle of 120° and the particle acceleration at the crest equals $\frac{1}{2}g$. However, waves near limiting height have rounded crests with a very small radius of curvature. It was shown by Longuet-Higgins and Fox (1977) that the local crest profile of such waves approaches a self-similar form. Furthermore the maximum surface slope can exceed the 30° slope of the limiting wave, ultimately reaching a value of about 30.38°. The vertical acceleration at the crest is not $\frac{1}{2}g$ but approximately $0.39g$ (Longuet-Higgins 1986).

To extend the above results to the case of breaking of irregular waves it is assumed that the downward acceleration at the crest of wave has to be greater than αg for breaking to occur, i.e.:

$$\left|\frac{d^2\zeta}{dt^2}\right| > \alpha g, \tag{1}$$

in which α is a constant. Snyder et al. (1983) have found that α varies from 0.4 to 0.52. The laboratory experiments of Ochi and Tsai (1983) provide the value $\alpha \approx 0.4$.

The starting point for calculation of the probability of wave crests breaking in a given wave train is the probability density function of positive maxima (crests) with a downward acceleration greater than αg which can be expressed as follows (Cartwright and Longuet-Higgins 1956, Massel 1996):

$$f_{\max}\left(\zeta_{\max}\right) = \frac{\int_{-\infty}^{-\alpha g} f_3\left(\zeta_{\max}, 0, \ddot{\zeta}\right)\ddot{\zeta}d\ddot{\zeta}}{\int_0^{\infty}\int_{-\infty}^{-\alpha g} f_3\left(\zeta_{\max}, 0, \ddot{\zeta}\right)\ddot{\zeta}d\ddot{\zeta}d\zeta}, \quad 0 \le \zeta \le \infty, \tag{2}$$

in which:

$$f_3\left(\zeta_{\max}, \dot{\zeta}, \ddot{\zeta}\right) = \frac{1}{(2\pi)^{3/2}\sqrt{m_2\widetilde{\Delta}}}\exp\left\{-\frac{1}{2\widetilde{\Delta}}\left[m_4\,\zeta_{\max}^2 + \right.\right.$$

$$\left.\left. + \; 2m_2\,\zeta_{\max}\,\ddot{\zeta} + m_0\,\ddot{\zeta}\right] - \frac{1}{2m_2}\dot{\zeta}^2\right\}, \tag{3}$$

where $\widetilde{\Delta} = m_0 m_4 - m_2^2$ and m_n are spectral moments defined as:

$$m_n = \int_0^{\infty} \omega^n S(\omega)\, d\omega. \tag{4}$$

The dots over variables denote differentiation in time. After some algebra (details are given by Srokosz (1986) and Massel (1998)), the probability that a crest of any height will break becomes:

$$F_{cr} = \exp\left(-\frac{\alpha^2 g^2}{2m_4}\right). \tag{5}$$

F_{cr} represents the probability that breaking will occur at a crest at a given point on the sea surface. It can be shown that as $m_4 \to \infty$, $F_{cr} \to 1$ (see Fig. 1). The formula (5) is of practical importance for the two-dimensional case of waves propagating in the wave flume as it determines the relative number of breaking wave crests in the wave train.

The probability F_{cr} should be distinguished from that of Snyder and Kennedy (1983) which deals with the fraction of the sea surface covered by breaking waves.

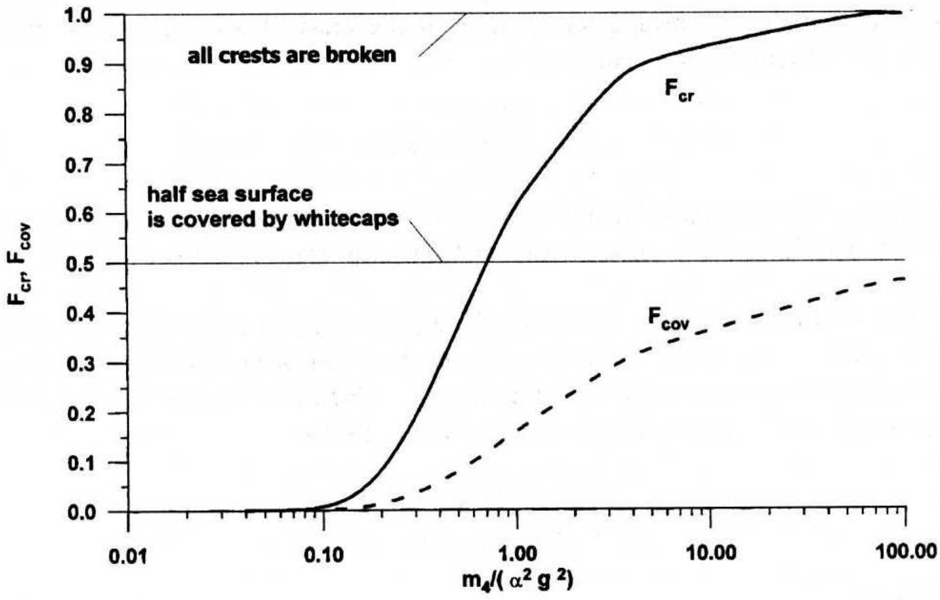

Fig. 1. Probability of breaking crests F_{cr} and whitecap coverage of sea surface F_{cov} as a function of $m_4/\alpha^2 g^2$

Let us assume that wave breaking occurs in the regions of fluid where the surface motion (not only in the wave crest vicinity) requires downward acceleration to exceed the dynamic threshold αg. Snyder and Kennedy (1983) found that this definition of breaking yields the percentage of sea surface covered by whitecaps in the form:

$$F_{cov} = 1 - \Phi\left(\frac{\alpha g}{\sqrt{m_4}}\right), \tag{6}$$

in which $\Phi(x)$ is the Laplace integral (Abramowitz and Stegun 1975)

$$\Phi(x) = \frac{1}{\sqrt{2\pi}} \int_{-\infty}^{x} \exp\left(-\frac{t^2}{2}\right) dt. \tag{7}$$

For very small waves, $m_4 \to 0$ and $F_{cov} \to 0$. During a heavy storm, $m \to \infty$ and $F_{cov} \to 1/2$. This means that half of the sea surface, where acceleration of the water elements is directed downwards and greater than critical, is covered by whitecaps. This fact is illustrated in Fig. 1.

The probabilities F_{cr} and F_{cov} are independent of any assumption on the spectral width, assuming that moment m_4 exists. Moreover, this remarkably simple result holds also for finite water depth assuming that probability density function (3) is approximately true for finite water depth.

Spectral properties of wind-induced waves in deep water are usually modelled by the Pierson-Moskowitz spectrum (Pierson and Moskowitz 1964) or by the JONSWAP spectrum (Hasselmann et al. 1973). It is well known that the Pierson-Moskowitz spectrum can be obtained as a special case of the JONSWAP spectrum when the peak enhancement factor γ equals 1. Therefore, in this paper we will concentrate mainly on the JONSWAP spectrum which can be represented as:

$$S(\hat{\omega}) = \frac{\beta g^2}{\omega_p^4} \hat{\omega}^{-5} \exp\left(-\frac{5}{4}\hat{\omega}^{-4}\right) \gamma^r, \tag{8}$$

in which β is the Phillips constant, ω_p the peak frequency, $\gamma = 3.3$ is the standard peak enhancement factor, and function r takes the form:

$$r = \exp\left[-\frac{1}{2}\frac{(\hat{\omega}-1)^2}{\sigma_0^2}\right], \tag{9}$$

in which:

$$\sigma_0 = \begin{cases} 0.07 & \text{when} \quad \omega < \omega_p \\ 0.09 & \text{when} \quad \omega \geq \omega_p \end{cases} \tag{10}$$

The Phillips constant β and peak frequency ω_p can be determined from the JONSWAP experiment as (Hasselmann et al. 1973):

$$\beta = 0.076\left(\frac{gX}{U^2}\right)^{-0.22}. \tag{11}$$

and

$$\omega_p = 7\pi\left(\frac{g}{U}\right)\left(\frac{gX}{U^2}\right)^{-0.33}. \tag{12}$$

We define the spectral moments of order n as follows:

$$m_n = \int_{\omega_l}^{\omega_h} \omega^n S(\omega)\, d\omega = \omega_p^n \int_{\hat{\omega}_l}^{\hat{\omega}_h} \hat{\omega}^n S(\hat{\omega})\, d\hat{\omega}, \qquad \hat{\omega} = \frac{\omega}{\omega_p}. \tag{13}$$

Theoretically, the lower and upper limits of integration in Eq. (13) should be equal to 0 and ∞, respectively. However, the form of JONSWAP spectrum which is based on experimental data, indicates that negligible energy is contained in the frequency band $0 < \hat{\omega} < 0.5$. We assume therefore that $\hat{\omega}_l = 0.5$.

The upper limit required more attention as its influence on the spectral moments (especially for higher moments) is substantial. After substitution of Eq. (8) into Eq. (13) we obtain the m_n moment as:

$$m_n = \beta g^2 \omega_p^{n-4} \int_{\hat{\omega}_l}^{\hat{\omega}_h} \hat{\omega}^{n-5} \exp\left(-\frac{5}{4}\hat{\omega}^{-4}\right) \gamma^r\, d\hat{\omega}. \tag{14}$$

Let us assume that $\hat{\omega}_l = 0$, $\hat{\omega}_h = \infty$, and $\gamma = 1$ (the Pierson-Moskowitz spectrum). Hence, the moment m_n becomes (Massel 1998):

$$m_n = \beta g^2 \omega_p^{n-4} \cdot \int_0^\infty \hat{\omega}^{n-5} \exp\left(-\frac{5}{4}\hat{\omega}^{-4}\right) d\hat{\omega}$$
$$= \frac{\beta g^2 \omega_p^{n-4}}{4} \cdot \left(\frac{5}{4}\right)^{\frac{n-4}{4}} \cdot \Gamma\left(\frac{4-n}{4}\right), \qquad (15)$$

in which $\Gamma(x)$ is a gamma function (Abramowitz and Stegun 1975). Equation (15) indicates that fourth moment m_4 becomes infinite as $\Gamma(0) = \infty$. The only way to calculate this moment for practical applications is to impose some threshold frequency $\hat{\omega}_h$. Taking into account the peak frequencies ω_p observed in practice, it has been assumed that $\hat{\omega}_h = 6$. Usually, the waves with frequency $\omega < 6\omega_p$ can still be considered as gravity waves, when the viscous effects are negligible.

Using the limits $\hat{\omega}_l = 0.5$ and $\hat{\omega}_h = 6$ in Eq. (14) we obtain for the fourth moment:

$$m_4 = \beta g^2 \int_{0.5}^{6.0} \hat{\omega}^{-1} \cdot \exp\left(-\frac{5}{4}\hat{\omega}^{-4}\right) \gamma^r d\hat{\omega}. \qquad (16)$$

Assuming now that $\gamma = 3.3$ for the JONSWAP spectrum and $\gamma = 1$ for the Pierson-Moskowitz spectrum, Eq. (16) yields:

$$m_4 = \begin{cases} 1.7057\beta g^2 & \text{for the JONSWAP spectrum} \\ 1.6344\beta g^2 & \text{for the Pierson – Moskowitz spectrum,} \end{cases} \qquad (17)$$

It should be noted that the fourth moment depends only on the non-dimensional fetch through the Phillips constant β, and does not depend on the peak frequency ω_p. However, other moments m_n ($n = 0, 1, 2, 3$) still depend on ω_p (see Eq. 14).

3. Available Experimental Data on Whitecaps Coverage

The experimental data on the breaking of irregular waves and whitecapping coverage are very limited. Ochi and Tsai (1983) obtained the breaking criterion from measurements of irregular waves generated in tank. They found that the probability of occurrence of wave crests breaking depends to a great extent, on the shape of the wave spectrum and the probability increases significantly with the fourth moment of the spectrum, exactly as suggested by Srokosz's formula (see Eq. 5).

However, for the purpose of this paper, the data on whitecaps coverage of the sea surface are of special interest. Whitecaps coverage has been investigated by Monahan (1971), Toba and Chaen (1973), Wu (1979), Monahan and O'Muircheartaigh (1981), Koepke (1984), Marks (1987) and others. In particular, Monahan (1971) collected 71 observations of the whitecapping at locations on the Atlantic Ocean and adjacent salt water basins. The basic motivation for this study was to obtain observations suitable for direct comparison with the existing several contradictory descriptions of the wind dependence of salt water whitecaps. The optimal power-law expression for the dependence of oceanic whitecaps coverage fraction F_{cov} on 10 m elevation wind speed V is usually given in the form:

$$F_{cov} = a\, V^{\lambda}. \tag{18}$$

The least squares fitting method based on (Monahan 1971) data suggests that $a = 1.35 \times 10^{-5}$, $\lambda = 3.4$ for 4 ms^{-1} $< V <$ 10 ms^{-1}. As shown in Fig. 2, this curve coincides with the highest whitecaps coverage observations at various wind speeds. In fact, it forms an envelope over all the data points in this range.

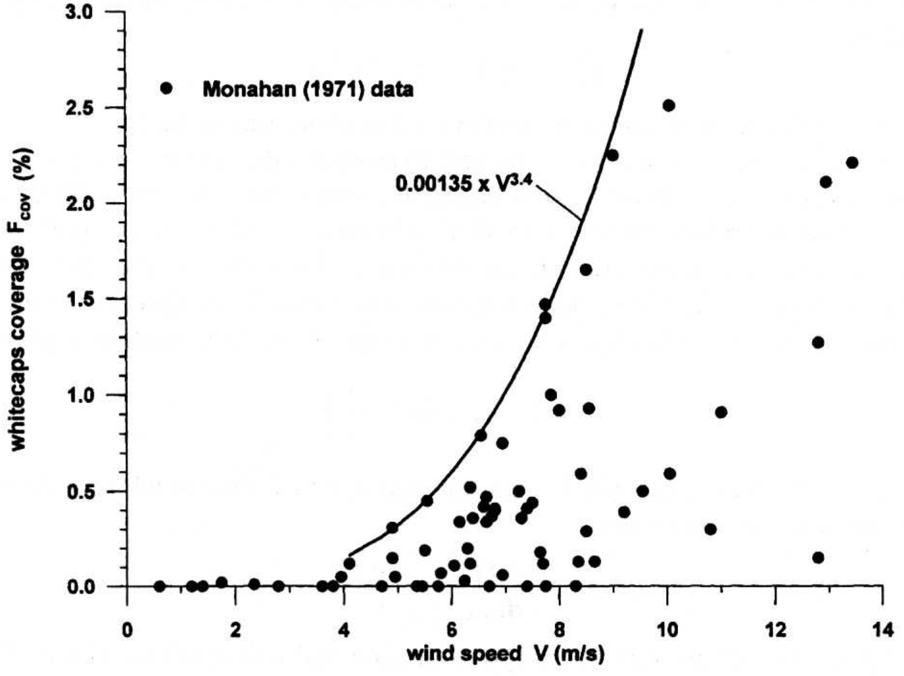

Fig. 2. Monahan (1971) data and their envelope for wind speed 4 m/s $< V <$ 10 m/s

Cardone (1969) using the results from fresh-water whitecap observations and assuming that the fraction of the water surface covered by whitecaps is directly related to the rate of energy transfer from the air flow to the fully developed sea, obtained $a = 1.2 \times 10^{-5}$, $\lambda = 3.3$ for 4 ms^{-1} $< V <$ 10 ms^{-1}.

Toba and Chaen (1973) observations of the whitecaps in the East China Sea and coastal waters of Japan yield the following parameters: $a = 1.55 \times 10^{-6}$, $\lambda = 3.75$. Combining the Atlantic Ocean data collected by Monahan (1971) and the Pacific Ocean data of Toba and Chaen (1973), Wu (1979) found that: $a = 1.7 \times 10^{-6}$, $\lambda = 3.75$.

Monahan and O'Muircheartaigh (1981) reanalysed the previous data using the ordinary least squares fitting (OLS) and robust biweight fitting (RBF) methods and they found:

$$F_{cov} = 2.95 \times 10^{-6}\, V^{3.52} \qquad \text{for OLS method} \tag{19}$$

and

$$F_{cov} = 3.84 \times 10^{-6}\, V^{3.41} \qquad \text{for RBF method.} \tag{20}$$

Marks (1984) analysed the data collected during a cruise the research vessel "Polarstern" in the North Atlantic and Greenland Sea (polar expedition Arkis–III). The whitecaps coverage was recorded by modern video-camera-system mounted about 15 m above the sea surface. The final result is given by the following relationship:

$$F_{cov} = 2.54 \times 10^{-6}\, V^{3.58}, \tag{21}$$

in which V is the wind speed at standard 10 m above sea surface.

For further work, it will be of interest to compare the oceanic and fresh water whitecaps coverage. Following Monahan (1971), let us assume that under identical meteorological conditions, the rate R of whitecaps production per unit area of water surface and initial area A_0 of individual whitecaps are the same for salt and fresh water. Therefore, the whitecaps area formed per time and unit areas becomes A_0R. The whitecaps area $A(t)$, at time t from its formation is given by:

$$A(t) = A_0 \exp\left(-\frac{t}{\tau}\right), \tag{22}$$

where τ is the time constant. Hence the rate at which individual area decay per unit and unit area becomes:

$$\frac{dA(t)}{dt} = -\frac{A(t)}{\tau}, \tag{23}$$

and for the whitecaps area decay (per unit time and unit area) we obtain F_{cov}/τ, where F_{cov} is the total area of whitecaps per unit area of sea surface.

In steady state conditions, the rate of whitecaps area formation is equal to the rate of whitecaps area decay, i.e.:

$$A_0R = \frac{F_{cov}}{\tau} \tag{24}$$

and

$$F_{cov} = A_0R\tau. \tag{25}$$

Monahan and Zieltow (1969) found that the time constant τ for salt water is about 1.5 times that for fresh water. This means that given the same meteorological conditions, the fraction of a sea surface covered by whitecaps should be 1.5 times that of a fresh water surface covered by whitecaps.

It is generally recognised that whitecaps coverage is negligibly small for wind speeds of less than 3 ms^{-1}. Also the mechanical tearing away of wave crests, which results in the formation of spume lines, is an additional mechanism of white water formation. It become of special importance for wind speeds of above 9 ms^{-1}. All these facts strongly suggest that the use of a more complex form for $F_{cov}(V)$ than a simple power-law is required to describe precisely the dependence of F_{cov} upon V.

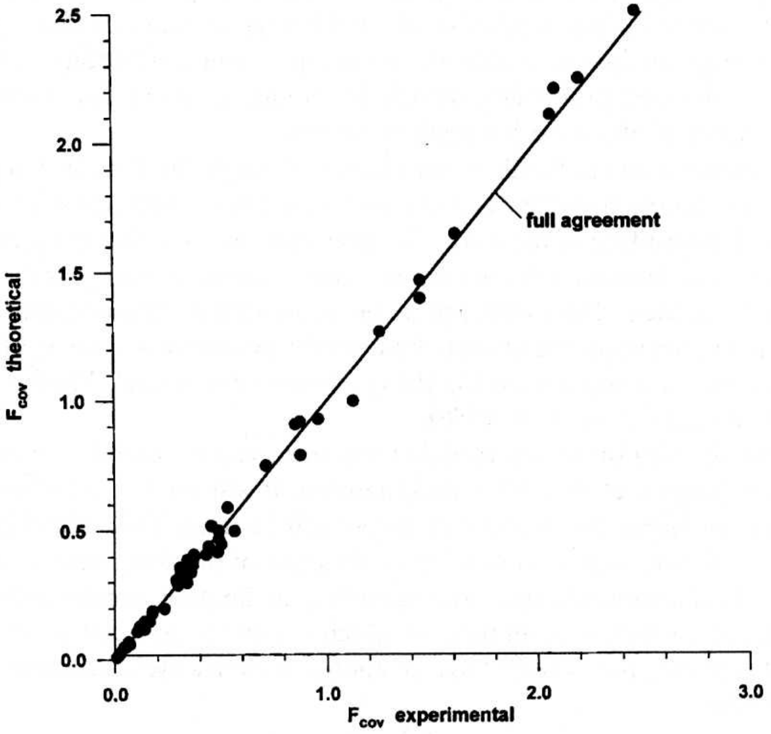

Fig. 3. Comparison of theoretical and experimental whitecaps coverage with the optimal coefficient α

The whitecaps coverage given by Eq. (6) depends on the α value which should be determined a priori. To estimate the range of α value variation, we use the Monahan (1971) data in which the tests with negligible whitecaps coverage (data noted as $F_{cov} = 0.0$) have been neglected. For each remaining test the α values, giving the optimal agreement of the theoretical F_{cov} value with the experimental one, were determined – see Fig. 3. It was found that these α values are in the range of $0.21 < \alpha < 0.44$, with the mean value of 0.27. Fig. 3 demonstrates that Eq. (6) is capable of representing the whitecaps coverage on condition that the α value is known. However, the coefficient α is not constant. In general, it depends on the intensity and type of breaking. It should therefore be considered as a function of local hydrodynamical and meteorological conditions as well as a function of wave breaking intensity. In the following Section we discuss shortly the available models of wave breaking in deep water.

4. Energy Dissipation in the Whitecapping

Estimation of rate of energy dissipated in the whitecaps in deep water is highly complex due to lack of full theory and incomplete description of the physics of the process of wave breaking. Komen et al. (1994) suggest three distinct approaches to model energy dissipation within the whitecaps, namely whitecap model, quasi-saturated model and probability model. It should be noted that none of these models includes all the wave breaking processes.

Once a wave starts to break it loses energy through the fluid injected into the whitecaps to become turbulent and the turbulence itself interacts with the orbital flow on the forward face of the wave. The presence of this whitecap on the forward face of the wave induces a downward pressure extracting energy from the wave. This is the basic idea of the whitecap model suggested by Hasselmann (1974). On the other hand, the imposing of some limit on the permitted steepness of the waves is a fundamental assumption used in the quasi-saturated model (Phillips 1985) and probability model (Yuan et al. 1986).

A detail description of the models mentioned can be found in source papers as well as in Komen et al. (1994), and therefore it will not be given here. For the purpose of this paper the probability model will be used. This model is based on the limiting of downward acceleration at the crest of breaking waves (see Section 2). The critical amplitude a_{br}, corresponding to limiting acceleration $g/2$, can be considered to be the amplitude to which a wave is reduced in the breaking process. Therefore, the expected loss of energy per wave cycle becomes (Longuet-Higgins 1969):

$$\Delta E = \frac{1}{2}\rho_w g \int_{a_b}^{\infty} \left(a^2 - a_{br}^2\right) f(a) da, \tag{26}$$

in which $f(a)$ is the probability density function of wave heights and limiting amplitude a_b, is:

$$a_b = a \left\{ \frac{1}{2} \frac{g}{a\omega^2} \right\}. \tag{27}$$

In order to evaluate the a_b value we need an information on the joint probability density function of wave height and frequency. However assuming the narrow banded wave field, the probability density function of frequency can simply be represented as follows:

$$f(\omega) = \delta\left(\frac{\omega}{\bar{\omega}} - 1\right), \tag{28}$$

in which $\delta(x)$ is the Dirac's delta and $\bar{\omega}$ is the mean frequency.

In addition we assume that wave heights are Rayleigh distributed. Using these assumptions, the energy loss per wave cycle becomes:

$$\Delta E = \exp\left(-\frac{g^2}{8\bar{\omega}^4 m_0}\right)E, \tag{29}$$

where m_0 is the zero spectral moment, i. e.:

$$m_0 = \int_0^\infty S(\omega)^{d\omega}. \tag{30}$$

Representing m_0 in terms of significant wave height we obtain for the non-dimensional energy loss:

$$\frac{\Delta E}{E} = \exp\left\{-\frac{g^2 T_p^4}{8\pi^4 H_s^2}\right\}. \tag{31}$$

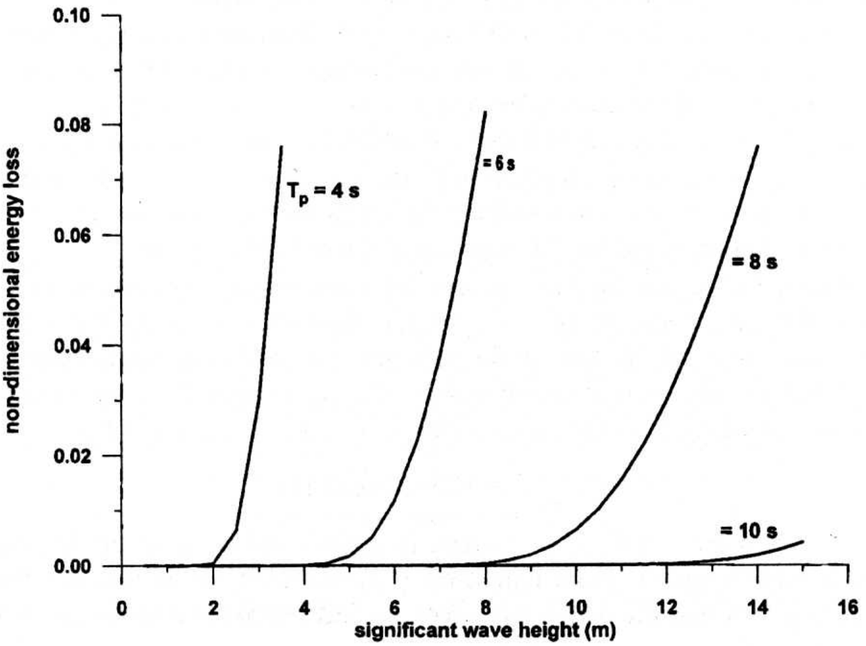

Fig. 4. Non-dimensional energy loss according the probability approach

The $\Delta E/E$ is illustrated in Fig. 4 as a function of selected significant wave heights H_s and peak periods T_p. It should be noted that in this figure, the combination of significant wave height and wave period providing a wave steepness of less than 1/7 is shown only. It means from physical point of view that the highest significant wave heights, for the particular wave periods, correspond to the onset of wave breaking.

5. Intensity of Wave Breaking and Aerosol Concentration in a Deep Sea

As mentioned in Section 3, the most common approach to determine the marine aerosol fluxes in deep water is to establish the relationship between wind speed and whitecap coverage. Using the sea salt aerosol concentration measured by the impactors, Marks (1987) showed that the relation between the aerosol concentration and whitecaps coverage can be described by the following relationship:

$$C_{aerosol} = 10.23 + 15.72 \times F_{cov}, \tag{32}$$

where $C_{aerosol}$ is the aerosol concentration in μg/m^3 and the whitecaps coverage is in percentages.

As mentioned above, the concentration of the aerosol is related to the dissipation of wave energy. In particular, Petelski and Chomka (1996) and Chomka and Petelski (1997) reported the results of measurements and modelling of the mean aerosol emission fluxes in the coastal zone during the BAEX Experiment at the Lubiatowo Station on the southern coast of the Baltic. They found that the aerosol fluxes varied from 3.2 to 384 μgm^{-2}s^{-1}. Assuming that the length of the Polish Baltic coastline is about 500 km and the mean width of the coastal zone 50 m, the mean annual emission from the coastal zone amounts to about 150 tons per year. Moreover, it was found that the aerosol emission fluxes are proportional to the average wave energy dissipation to the power of 3/4 (Chomka and Petelski 1997). This relationship was based on the estimation of wave energy dissipation in the coastal zone using the Thornton and Guza (1983) model.

Contrary to the coastal zone, where the wave energy dissipation area is well defined, the determination of wave energy dissipation in deep water and subsequent estimation of the aerosol fluxes is much more complicated. Marks (1987) showed that the sea aerosol concentration $C_{aerosol}$ of particles larger than 0.1μm for a wind speed of V in the range 1 m/s $< V <$ 12 m/s, is given by:

$$C_{aerosol} = 6.08 \times \exp(0.13V), \tag{33}$$

in which concentration $C_{aerosol}$ is given in μg/m^3 and wind speed in m/s. If we eliminate wind speed V from equations (21) and (33), we obtain the following relationship between the whitecaps coverage and sea aerosol concentration:

$$C_{aerosol} = 6.08 \exp\left[4.7516\, F_{cov}^{0.27933}\right]. \tag{34}$$

The relationship (34) of the sea aerosol concentration on the whitecaps coverage is given in Fig. 5.

6. Conclusions

Aerosol fluxes from the sea surface are one of the important factors determining the dynamics of air-sea interaction. However, they are very difficult to measure

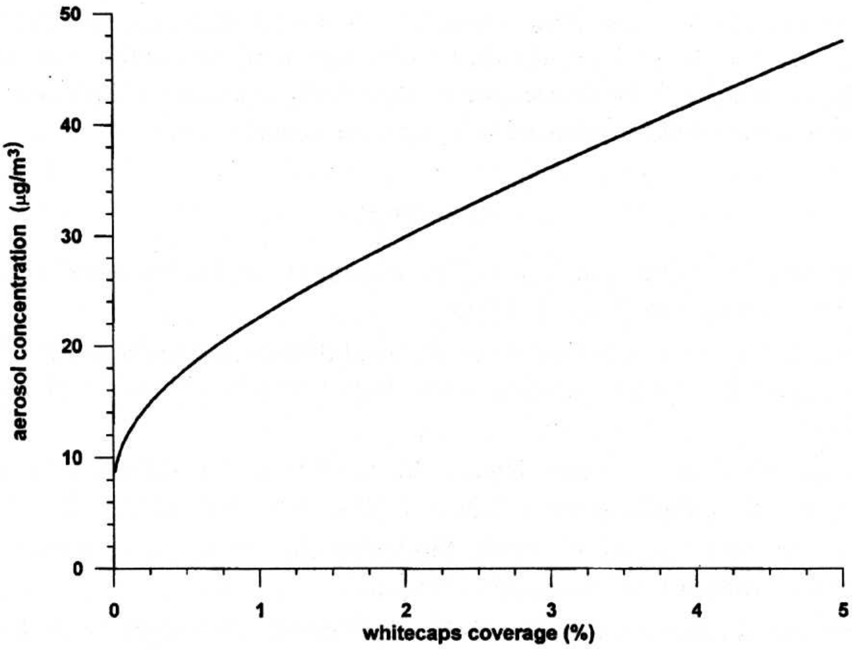

Fig. 5. Sea aerosol concentration C as a function of whitecaps coverage F_{cov}

and to model theoretically. The rarely available data showed that the intensity of aerosol fluxes strongly depends on the intensity of wave breaking.

In the paper, two theoretical formulas to determine the probability of breaking crests and percentage of whitecaps coverage have been discussed. These formulas can be a good starting basis for further studies on the relationship between the sea surface state and aerosol concentration. Results discussed in the previous Sections suggest the following step-by-step procedure for the determination of the aerosol fluxes in deep water under steady sea state conditions:

1. determination of whitecaps coverage of the sea surface F_{cov}. However, control laboratory experiments will be needed to calibrate the theoretical models
2. determination of the wave energy dissipated during whitecapping per unit of sea surface E_{diss}
3. determination of the aerosol emission flux per unit of sea surface using the following:
$$F_{aerosol} = F_{cov} \times E^{n}_{diss}, \tag{35}$$
in which power n for deep water should be determined in the control experiments which involve the simultaneous measurements of wave energy dissipation and aerosol fluxes in natural conditions. The value of coefficient n in

deep waters will probably be different from that suggested by Chomka and Petelski (1997) for shallow water. The proposed methodology depends on the careful laboratory and field measurements of the relationship between the sea state and the whitecaps coverage. Such experiments are planned and their results will be reported in a separate paper.

References

Abramowitz M. and Stegun I. A. (1975), *Handbook of Mathematical Functions*, Dover Publ., New York, 1045 pp.

Cardone V. J. (1969), Specification of the wind distribution in the marine boundary layer for wave forecasting, *Tech. Rep. GSL-69-1*. New York University, 1–131.

Cartwright D. E. and Longuet-Higgins M. S. (1956), The statistical distribution of the maxima of a random function, *Proc. Roy. Soc.* **A237**, 212–232.

Chomka M. and Petelski T. (1997), Modelling the sea aerosol emission in the coastal zone, *Oceanologia*, **39**, 211–225.

Hasselmann K., Barnett T. P., Bouws E., Carlson, H., Cartwright D. E., Enke K., Ewing J. A., Gienapp H., Hasselmann D. E., Kruseman P., Meerburg A., Müller P., Olbers D. J., Richter K., Sell W. and Walden H. (1973), Measurements of wind-wave growth and swell decay during the Joint North Sea Wave Project (JONSWAP), *Deutsches Hydr. Zeit.* **A12**, 1–95.

Hasselmann K. (1974), On the spectral dissipation of ocean waves due to whitecapping, *Boundary Layer Meteo.*, **6**, 107–127.

Koepke P. (1984), Effective reflectance of oceanic whitecaps, *Appl. Optics*, **23**, 1816–1824.

Komen G. J., Cavaleri L., Donelan M., Hasselmann K., Hasselmann S. and Janssen P. A. E. M. (1994), *Dynamics and modelling of ocean waves*, Cambridge Univ. Press, Cambridge, 532 pp.

Longuet-Higgins M. S. (1969), On wave breaking and the equilibrium spectrum of wind-generated waves, *Proc. Royal Soc.*, **310**, 151–159.

Longuet-Higgins M. S. (1986), Acceleration in steep gravity waves, *Jour. Phys. Ocean.*, **15**, 1570–1579.

Longuet-Higgins M. S. and Fox M. J. H. (1977), Theory of the almost-highest wave: the inner solution, *Jour. Fluid Mech.* **80**, 721–741.

Marks R. (1984), Marine aerosols and whitecaps in the North Atlantic and Greenland Sea regions, *Dt. Hydrogr*, **40**, 71–79.

Massel S. R. (1996), *Ocean Surface Waves: their Physics and Prediction,* World Scientific Publ., Singapore, 491 pp.

Massel S. R. (1998), The limiting wave height in wind-induced wave trains, *Ocean Engineering*, **25**, 735–752.

Monahan E. C. (1971), Oceanic whitecaps, *Jour. Physical Oceanogr.*, **1**, 139–144.

Monahan E. C. and Zietlow C. R. (1969), Laboratory comparisons of freshwater and saltwater whitecaps, *Jour. Geoph. Res.*, **74**, 6961–6966.

Monahan E. C. and O'Muircheartaigh I. (1980), Optimal power-law description of oceanic whitecap coverage dependence on wind speed, *Jour. Physical Oceanogr.*, **10**, 2094–2099.

Monahan E. C. and Van Patten M. A. (Eds.) (1989), *The climate and health implications of bubble-mediated sea-air exchange*, The Connecticut Sea Grant College Program, 170 pp.

Nath J. H. and Ramsey F. L. (1976), Probability distribution of breaking wave heights emphasising the utilisation of the JONSWAP spectrum, *Jour. Phys. Ocean.* **6**, 316–323.

Ochi M. K. and Tsai C. H. (1983), Prediction of occurrence of breaking waves in deep water, *Jour. Phys. Ocean.* **13**, 2008–2019.

Petelski T. and Chomka M. (1996), Marine aerosol fluxes in the coastal zone - BAEX experiment, *Oceanologia*, **38**, 469–484.

Phillips O. M. (1985), Spectral and statistical properties of the equilibrium range in wind-generated gravity waves, *Jour. Fluid Mech.*, **156**, 505–531.

Pierson W. J. and Moskowitz L. (1964), A proposed spectral form for fully developed wind seas based on the similarity theory of S.A. Kitaigorodskii, *Jour. Geoph. Res.* **69**, 5181–5190.

Snyder R. L. and Kennedy R. M. (1983), On the formation of whitecaps by a threshold mechanism. Part I: Basic formalism, *Jour. Phys. Ocean.* **13**, 1482–1492.

Snyder R. L., Smith L. and Kennedy R. M. (1983), On the formation of whitecaps by a threshold mechanism. Part III: Field experiment and comparison with theory, *Jour. Phys. Oceanogr.* **13**, 1505–1518.

Srokosz M. A. (1986) On the probability of wave breaking in deep water, *Jour. Phys. Ocean.* **15**, 382–385.

Toba Y. and Chaen M. (1973), Quantitative expression of the breaking of wind waves on the sea surface, *Rec. Oceanog. Works Japan*, **12**, 1–11.

Yuan Y., Tung C. C. and Huang N. E. (1986), Statistical characteristics of breaking waves, [in:] Phillips O. M. and Hasselmann K. (Eds.) *Wave dynamics and radio probing of the ocean surface*, Plenum, 265–272.

Wu J. (1979), Oceanic whitecaps and sea state, *Jour. Phys. Oceanogr.*, **9**, 1064–1068.