Energy Loss by Breaking Waves

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ABSTRACT

Observations of the frequency of wind wave breaking in deep water are combined with laboratory estimates of the rate of energy loss a from single breaking wave to infer the net rate of energy transfer to the mixed layer from breaking waves, as a function of wind speed. Breaking waves with wavelengths much shorter than the dominant waves can contribute energy at a rate that is a significant fraction of the total turbulent kinetic energy dissipation rate in the ocean surface mixed layer.

1. Introduction

Wind waves breaking in deep water are an important component in the process of air–sea interaction. Melville and Rapp (1985), for example, conclude that much of the momentum flux from the wind field may pass through the waves before being transferred by breaking to the mean current in the water column. Kitaigorodskii (1984) argues that turbulence generated by breaking waves reduces the local thickness of the gas diffusion sublayer and hence promotes air–sea gas transfer. Woolf and Thorpe (1991) have shown that bubbles produced by breaking wind waves provide an important pathway in the transfer of the less soluble gases from the atmosphere to the ocean. Resch et al. (1986) and Bortkovskii (1983) have described the role of bursting bubbles in the production of airborne drops and aerosols.

Waves break on a variety of scales. Small-scale ripples or capillary–gravity waves formed on the crest or leading face of steep gravity waves (Longuet-Higgins 1963) may cause small-scale breaking, trapping bubbles near their troughs or producing flow separation (Longuet-Higgins 1992). The more familiar breaking occurs at a larger scale, with plunging or spilling waves observed as whitecaps (e.g., see Cokelet 1977; Longuet-Higgins 1988). The relative contributions of the different scales of breaking to air–sea transfer are presently unknown, and may well depend on the transferred component (e.g., momentum, heat, or gas) under investigation. Recent studies have, for example, pointed to the importance of subsurface turbulence enhancement by very short waves in the air–sea transfer of gases (Jahne et al. 1987; Wanninkhof 1992) and heat (Katsaros et al. 1978). The relation between capillary ripples, flow separation, and fluid circulating below the crest of short surface gravity waves, described by Longuet-Higgins (1992), may play a fundamental role in this process. We shall here, however, be principally concerned with the breaking of larger waves, where breaking is clearly visible and relatively easy to detect.

2. The observations

Measurements have been made of the frequency of wave breaking at a fixed position. Thorpe and Humphries (1980) made observations of the frequency of breaking wind waves in a loch at a fetch of about 20 km, using a capacitance wire probe and a technique designed and used by Longuet-Higgins and Smith (1983) in their measurements in the southern North Sea at a fetch of about 100 km. (Thorpe and Humphries also considered the relation between measurements at a fixed point and spatial variations.) Holthuijsen and Herbers (1986) reported visual observations of whitecap formation in 8719 waves, also in the southern North Sea with fetch varying from 20 to several hundred kilometers, finding much higher rates of breaking than Longuet-Higgins and Smith. Weissman et al. (1984) made measurements using a fine resistance wire probe in a lake with a fetch of 7 km, and more recently Katsaros and Atakturk (1992) have substantially extended these observations. They report, in particular, an increase in the relative frequency of breaking with inverse wave age in the wind speed range 3.5 to 7.5 m s^{-1}. Here the frequency of wave breaking, f, is measured by the number of waves breaking at a fixed position in a given time divided by the number of waves of the dominant wave frequency that pass in the same period of time, that is, by the number of breaking events per period of the dominant waves, and the wave age is the friction velocity in the air divided by the phase

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speed of the dominant waves, c_0. We have made additional measurements in winds from 3 to 28 m s^{-1}, again in the 20-km fetch loch, with high-frequency side-scan sonar and using the technique described by Thorpe (1992).

The data are shown in Fig. 1. Here the breaking frequency is plotted as a function of the 10-m wind speed, W_{10}, divided by the dominant phase speed. We have resorted to the use of this parameter rather than the more appropriate wave age, since the friction velocities were generally not measured or reported. Katsaros and Atakturk's data points include both plunging and spilling breakers, but omit data, typically of low wind speed, in which f was less than 0.01. Holthuijsen and Herber's data lie well above the others, possibly being affected by wave refraction in shallow water and sand banks; the dominant wavelength was about twice the water depth. Longuet-Higgins and Smith's data also lie above the other points, perhaps because they include some steep, but not breaking, waves or are affected by the influence of swell or longer fetch. There is, however,

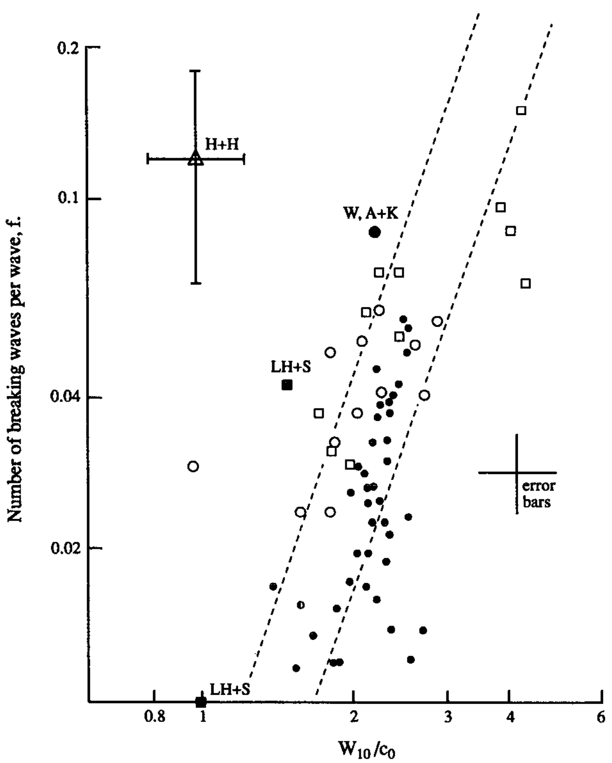

FIG. 1. The number of breaking waves per wave, f, versus wind speed divided by the wave speed of the dominant waves, W_{10}/c_0, both plotted on log scales. The symbols denote: H+H, Holthuijsen and Herbers (1986), with bars showing the range of the measurements and conditions; LH+S, Longuet-Higgins and Smith (1983); WA+K, Weissman, Atakturk, and Katsaros (1984); circles, Thorpe and Humphries (1980); points, Katsaros and Atakturk (1992); and squares, data derived using 250-kHz side-scan as described by Thorpe (1992). The error bars apply to the latter dataset. The LH+S point shown in Thorpe (1992, Fig. 5b) was plotted incorrectly.

some general consistency in the remaining data points and in their general trend. The dashed lines, drawn to represent the points, are given by

$$f = (4.0 \pm 2.0) \times 10^{-3} \times (W_{10}/c_0)^3. \qquad (1)$$

This equation is consistent with the prediction by Phillips (1985) that the number of breaking waves passing a fixed point per unit time is proportional to the cube of the friction velocity in the air, and hence to the cube of the wind speed over a range in which the drag coefficient is uniform. It will tend to underestimate f in the conditions represented by Holthuijsen and Herbers' and Longuet-Higgins and Smith's datasets.

3. Discussion

The probability that a wave passing a fixed point is breaking is equal to the fraction of its alongcrest dimension that is breaking. If, for simplicity, we suppose that the wave field is uniform, consisting of a two-dimensional array of waves with the period of the dominant waves, and that breaking occurs in association with their parallel wave crests, then we can regard f as an estimate of the fraction of the breaking crest length of any one wave.

Duncan (1981) made laboratory measurements of the rate of loss of energy per unit crest length, E, from a quasi-steady breaking wave produced by a subsurface hydrofoil. The estimated energy loss increases rapidly with the phase speed of the breaking wave, c_b,

$$E = (0.044 \pm 0.008)\rho c_b^5/g, \qquad (2)$$

where ρ is the density of the water and g is the acceleration due to gravity. [The value of the coefficient is less than that, 0.06, quoted by Phillips (1985). We have used Duncan's Eq. (17), together with the range of wave slopes of his experiments, to determine the breaking wave drag. The uncertainty in the estimates of Eq. (2) reflects this range of slopes.] Consider now a section of length L of the crest of one wave in the simplified wave field. The breaking length is Lf, and the rate of energy loss is ELf. This is now repeated by the next wave and, to find the spatial average of the rate of dissipation, we must divide the rate by the area over which the dissipation is applied, $\lambda_0 L$, where λ_0 is the wavelength of the dominant waves. The rate of energy loss from the waves per unit surface area is E_w, $= Ef/\lambda_0$, or

$$E_w = (1.9 \pm 1.1) \times 10^{-4} \times \rho(W_{10}/c_0)^3 c_b^5/g\lambda_0, \qquad (3)$$

using (1). For deep-water surface gravity waves, $c_0^2 = g\lambda_0/2\pi$, and so we find that

$$E_w = (3.0 \pm 1.8) \times 10^{-5} \times \rho W_{10}^3 (c_b/c_0)^5. \qquad (4)$$

Oakey and Elliott (1982) have estimated the vertically integrated rate of dissipation of turbulent kinetic energy per unit surface area in the ocean mixed layer as a function of wind speed. Their measurements were made using a free-fall profiling instrument, OCTOPROBE. Accurate measurements of dissipation were possible only at depths below 5.5 m. The mixed layer was typically 20 m thick. The dissipation from the surface to 5.5 m was assumed to be equal to the mean dissipation measured from 5.5-m depth to 10 m above the bottom of the mixed layer. The total rate of dissipation per unit volume integrated from the surface to the base of the mixed layer, E_{ml}, is shown in Fig. 2 (adapted from Oakey and Elliott's Fig. 16) plotted as a function of W_{10}^3. If c_b, the phase speed of the breaking waves measured relative to the underlying flow, is chosen to be equal to the speed of the dominant waves, c_0, the estimate (4) exceeds Oakey and Elliott's estimates by about 10^3. We have superimposed, as dashed lines in Fig. 2, the range of estimates of E_w [Eq. (4)] with $\rho = 1000$ kg m^{-1} and $c_b/c_0 = 0.25$, chosen so as to include most of Oakey and Elliott's mean and maximum vertically integrated dissipation estimates. We conclude that if only the loss of energy from the wave field by breaking supports the turbulence in the mixed layer (and we ignore turbulence production by shear stress or buoyancy flux), the required loss of wave energy is not derived from the breaking of the dominant waves near the peak of the wave spectrum (which would provide far more energy than is found to be dissipated), but from higher-frequency, shorter waves traveling through the water at speeds typically one-quarter of that of the dominant waves, and therefore with wavelengths typically one-sixteenth of the dominant waves.

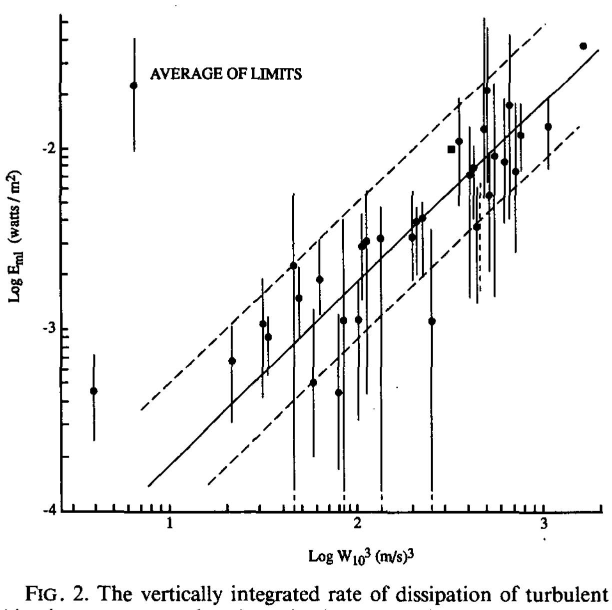

FIG. 2. The vertically integrated rate of dissipation of turbulent kinetic energy per unit volume in the ocean mixed layer estimated by Oakey and Elliott (1982), E_{ml}, versus the cube of the wind speed, W_{10}^3, both plotted on log scales. Each dot represents the average obtained by several measurements in one half-hour period; the lines join the maximum and minimum values observed in the period. The "average of limits" indicates the mean deviation of the maximum and minimum values from the means. The solid square is from Stewart and Grant (1962) and the vertical dashed line is from Gargett et al. (1979). The range of energy flux per unit surface area from breaking surface waves with a length of 1/16 of the dominant waves, given by Eq. (4) with $c_b/c_0 = 0.25$, is shown by the dashed lines. (The figure is adapted from Oakey and Elliott's Fig. 16.)

Oakey and Elliott's estimates may, however, underestimate the total dissipation. In temperate latitudes, the dissipation rate in the upper ocean is subject to considerable temporal variations in response to changes in seasonal and diurnal heat flux (e.g., see Shay and Gregg 1986). Recent measurements by Agrawal et al. (1992) and Osborn et al. (1992) show that the average dissipation may be large at depths $z_m < 10^5 u_*^2/g$, where u_* is the friction velocity in the water, exceeding that estimated by the "Law of the Wall" by factors as great as 70, local enhancement being found in association with bubble clouds and so linked to the breaking wind waves (see also Thorpe 1992). It is not known to what extent this enhances the mean vertically integrated dissipation. If we assume, however, that the stress τ at the sea surface is continuous (perhaps thereby overestimating u_*) so that $\tau = C_D \rho_a W_{10}^2 = \rho u_*^2$, where ρ_a is the density of air, and if we select an extreme value for the drag coefficient, $C_D = 2 \times 10^{-3}$ (Large and Pond 1981), then we find that z_m is about 2.4 m in winds of 10 m s^{-1}, close to the upper limit shown in Fig. 2. Smaller values would be estimated in lower winds. This estimate is much less than the mixed-layer depths, but also less than the depth ranges sampled by Oakey and Elliott. Although integration from the depth range of Oakey and Elliott's observations to the surface is uncertain, the enhancement of the mean vertically integrated dissipation by the superactive near-surface region seems unlikely to be greater than a factor of 10. The total flux of energy from the wind to the water surface may be alternatively estimated as $E_{\text{wind}} = \tau U_s$, where U_s is the surface drift speed. Taking again extreme values for the drag coefficient, $C_D = 2 \times 10^{-3}$ and for $U_s = 0.04\, W_{10}$ (Wu 1975), we can derive an upper bound to E_{wind}, $8 \times 10^{-5} \rho_a W_{10}^3$. Since $\rho_a \ll \rho$ and, in a state of local equilibrium, $E_w \leqslant E_{\text{wind}}$, this, together with Eq. (4), again implies that c_b must be less than c_0.

If the *position* of wave breaking is associated with the dominant wave [and an association of breaking events with the crests of waves in groups of long waves has been noticed; Donelan et al. (1972); Thorpe and Humphries (1980)], but the *scale* of breaking is that of much shorter waves, then there are implications for scaling in the mixed layer. Is it appropriate, for example, to seek scales for the turbulence or bubble clouds produced by breaking waves, which are related to the dominant waves?

Rapp and Melville (1990) have made measurements in which progressive waves are caused to break in a laboratory tank in otherwise quiescent water. Over 90% of the turbulence generated by breaking is dissipated within four wave periods. If no reduction or enhancement of this dissipation time occurs in the sea, for example, as a result of the interaction between wave and shear generated turbulence, we may use (1) to estimate the fraction of time, F, for which turbulence near the sea surface may be enhanced as a result of wave breaking. Enhanced turbulence persists for a time $< 4T_b$, where T_b is the period of the breaking waves, and the time interval between breaking is T_0/f, where T_0 is the period of the dominant waves, so that $F < 4T_b/(T_0/f)$ is approximately equal to f, if $T_b/T_0 = c_b/c_0$ is about 0.25, as found above. The fraction F is therefore generally quite small as shown by Fig. 1. The estimates relate to the problems of adequately sampling turbulence in the near-surface waters; very many vertical profiles taken at random times are required to sample the extreme values and to measure the mean dissipation. The value of F may also relate to the fraction of time a surfactant film may be broken by intense subsurface eddies (Tryggvason et al. 1992) created by breaking wave turbulence, and hence to the rate of air–sea transfer of gases (Goldman et al. 1988).

Bubble clouds are known to persist for 2–5 min (Thorpe and Hall 1983), many times the period of the breaking or dominant waves and, at least when the dominant waves are short, a time that may exceed several times the time interval between breaking at a fixed point. If the depth of a bubble cloud produced by a breaking wave is proportional to its scale (e.g., the wave amplitude or wavelength), then occasional bubble clouds produced by the dominant or larger waves breaking may persist, and obscure, at least to upward-pointing sonars, the subsequent bubble injection of the shorter waves. The mean bubble cloud depths may therefore be determined by relatively rare breaking of large-scale waves (see Thorpe 1992).

4. Conclusions

We have offered simple arguments to show that the input of energy from breaking wind waves at a scale much less than the dominant waves provides a source of energy flux equal to a significant fraction of the total energy dissipated in the mixed layer of the ocean. It is salutary to note that attempts have been made to measure and explain the rate of dissipation of turbulent kinetic energy near the sea surface for over 30 years (e.g., see Stewart and Grant 1962). Further study is required of the energy dissipation in breaking waves, and of the variation in breaking wave frequency, the effects of swell and fetch and nonuniform winds, the role of short waves and other processes such as mean shear, surface buoyancy flux, and Langmuir circulation. The existing description and quantification of the processes of air–sea transfer are not adequate to provide accurate parameterizations of air–sea fluxes for coupled ocean–atmosphere models designed to predict the present and future motions and compositions of the air and seas.

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