Geophysica (1997), 33(2), 17-27

On Waves, Oceanic Turbulence, and Their Interaction

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Abstract

Much progress in understanding the coupling between the atmosphere and the sea has been made during the past years. Here we discuss the implications of the recent discovery of enhanced (above wall layer predictions) dissipation rates in the upper ocean. In particular, we look at estimates of surface roughness and discuss mechanisms of wave-turbulence interaction.

(Dedicated to Professor Sergei Kitaigorodskii on the occasion of his sixtieth birthday.)

Key words: ocean waves, turbulence, air-sea interaction, surface roughness

1. *Introduction*

The starting point for many, if not most, of today's students in air-sea interaction is one of two monographs which first appeared over a quarter century ago: *Phillips* (1966) and *Kitaigorodskii* (1970). Although now somewhat dated, these major works are still relevant, laying down a solid foundation in the field. There has, however, been much progress since then, as summarized in, e.g., *Donelan* (1990). Here, we present recent experimental data on turbulent dissipation rates in the near-surface of the ocean, and discuss the implication of these on energy transfer from the atmosphere to the water column. In particular, we look at the energy transfer to waves from the wind, and then from the waves into oceanic turbulence. Energy may be lost from the waves in several ways, e.g. viscous dissipation, breaking and wave-turbulence interaction. We discuss some of the proposed mechanisms for wave-turbulence interaction.

Published by the Finnish Geophysical Society, Helsinki

2. Waves and surface roughness

The oceanographic community, by and large, has treated the ocean surface as an aerodynamically slightly rough surface in which the roughness increases with wind speed or friction velocity. Typical descriptions for this wind speed dependence are given by *Charnock* (1955), *Smith* (1980), and *Large* and *Pond* (1981). The roughness elements are the very short waves whose propagation speeds are of the order of the friction velocity in the air, u_{*a}. Consistent with this wall layer view of the surface is the idea that the kinetic energy input to the ocean from wind is approximately $\rho_a u_{*a}^3$, where ρ_a is the density of air. In steady state homogeneous conditions this puts a limit on the kinetic energy delivered from atmosphere to ocean.

Kitaigorodskii and *Volkov* (1965) recognized the importance of longer waves in determining the roughness of the surface. Near full development the propagation speed of the largest waves approaches the wind speed and these waves therefore contribute little to the net roughness of the surface. By viewing the interaction of the wind with the surface in a frame of reference propagating with the phase speed of a particular wave component, they produced a simple and elegant means of discovering the contributions of various wave components to the roughness based on the ratio of phase propagation speed and friction velocity:

$$z_o \propto a \exp\left(-\kappa c / u_{*a}\right) \tag{1}$$

where z_o is the contribution to the roughness length from waves of amplitude a and phase speed c. Here, κ is the von Kármán number, $\kappa \approx 0.4$. In terms of the complete wave number spectrum $S(k)$:

$$z_o^2 = \alpha^2 \int_0^\infty S(k) \exp\left[-2\kappa c(k) / u_{*a}\right] dk, \tag{2}$$

where τ in an empirical constant (see *Kitaigorodskii*, 1968). Thus the long, fast waves contribute relatively little to the roughness except when they are young, e.g. on the rising edge of a storm or at short fetch. Various prescriptions of this sort are now in common use in determining the stress of the wind on the water surface (*Donelan*, 1990). Underdeveloped wave fields cause an increase in the stress on the surface by a factor of up to 2-3 for the same wind speed at full development.

Kitaigorodskii and *Volkov* (1965) realized that in fully rough flow the stress is delivered in varying amounts to different components of the wave spectrum, some of which travel at speeds many times greater than the friction velocity. Consequently, the total energy transferred from wind to water can be much larger when the waves are underdeveloped, both because the stress is larger and because the stress receptors travel much faster than the friction velocity. See also *Kitaigorodskii et al.* (1995).

Using the wind input model of *Donelan* and *Pierson* (1987), *Terray et al*. (1996) have calculated the ratio of the average velocity associated with the energy transfer, $\bar{c}$, to the friction velocity in the air, u_{*a}. Standard oceanographic practice takes $\bar{c}$ to be the wind drift velocity at the surface, which was given by *Wu* (1975) to be approximately $u_{*a}/2$, so that the total energy transfer rate per unit area, F_1 is:

$$F_1 = \rho_a u_{*a}^3 / 2 \tag{3}$$

whereas, based on Kitaigorodskii's ideas, *Terray et al*. (1996) find

$$F_2 = \rho_a u_{*a}^2 \bar{c} = 2 F_1 \bar{c} / u_{*a} .$$

Using the empirical relation

$$\bar{c} / c_p \simeq \begin{cases} 0.5 & c_p / u_{*a} \le 13 \\ 6.5 \left(c_p / u_{*a} \right)^{-1} & c_p / u_{*a} > 13 \end{cases} \tag{4}$$

(*Terray et al.* 1996) we have $F_2/F_1 \approx 3$ - 13 over a wide range of wave ages. Hence, the rate of energy transfer to the surface can exceed traditional estimates of F_1 by up to an order of magnitude, depending on the wave age.

Most of the energy supplied to the waves is lost from them locally in breaking. At full development all the energy entering the wave field from the wind is lost locally. Even while the waves are strongly growing, the fraction of energy lost from the wavefield is still in excess of 95 %. The bulk of this energy enters in the water column as turbulence, and is eventually dissipated as heat.

3. Turbulence in the upper ocean

Kitaigorodskii and *Miropolskii* (1968), in their early work on dissipation rates in the upper ocean, recognized two distinct types of fluid motion: potential wave motion, and the residual, referred to as turbulent motion. They asserted that wave breaking was an important means by which energy is transferred to turbulent scales and hypothesized that an interaction between the wave and turbulent motions was responsible for much of the vertical energy exchange in the upper ocean. They did not, however, identify the source of the interaction. The work of Kitaigorodskii and Miropolskii led to an important finding of *Benilov* (1973). Benilov recognized that the standard balance equation for kinetic energy neglected energy supplied by breaking waves. Separating the motion into potential and turbulent components, Benilov wrote down the energy balance equation for turbulent motions. He discovered, in this equation, a new term linking the turbulent and potential wave fields: the divergence of the energy flux from the waves due to the presence of turbulence. Benilov then proposed a model equation for the turbulent kinetic energy dissipation rate, *e*, as a function of depth, the energy of the wave field and the energy flux from breaking waves. The model showed an upper surface layer strongly dependent on the energy flux. Unfortunately, however, the only data existing at that time (*Stewart* and *Grant*, 1962) was recorded at greater depths - and hence could not be used to support (or contradict) the model.

Over the next decade, a series of experiments reported new estimates of *e* in the upper ocean. *Arsenyev et al*. (1975), *Jones* and *Kenney* (1977), *Dillon et al*. (1981), *Oakey* and *Elliott* (1982) and *Jones* (1985) all reported results consistent with those of Stewart and Grant; that is, there was no observed dependence of the dissipation rate on properties of the wave field. Based on these and their own experimental data, *Soloviev et al*. (1988) proposed that the dissipation in the upper ocean could be modelled using wall layer theory with *e* depending only on the water-side friction velocity u_{*w} and *z*, the distance from the surface, viz. $\varepsilon = u_{*w}^3(\kappa z)^{-1}$. In this model, oceanic and atmospheric turbulence are analogous, with the turbulence generated entirely from current shear near the surface. Waves, breaking or not, are assumed to play no role in the process.

The one data set that did not support wall layer scaling was that of *Kitaigorodskii et al*. (1983). Near-surface measurements from a tower in Lake Ontario showed dissipation rates one to two orders of magnitude higher than wall layer theory would predict. *Kitaigorodskii et al*. proposed that the elevated dissipation rates were due to waves - with the excess turbulent energy due either to wave breaking or to wave-turbulence interaction, as proposed by *Benilov* (1973) or *Kitaigorodskii* and *Lumley* (1983). They proposed a two layer structure describing turbulent energy dissipation: an upper layer, with thickness of the order of two significant wave heights, in which the turbulence is generated primarily by an energy flux from waves at the surface, and a lower “wall-layer” in which shear dominates.

Subsequent measurements by *Gregg* (1987), *Gargett* (1989), *Agrawal et al*. (1992), *Anis* and *Moum* (1992) and *Osborn et al.* (1993) confirmed the existence of high near surface dissipation rates. *Agrawal et al*. looked at the intermittency of dissipation events and showed that several intense (order 10 to 100 times the median value) events dominate the total dissipation. It was suggested that these events were related to breaking waves.

Terray et al. (1996) analysed the same data set reported by *Agrawal et al.,* and proposed a model consistent with both the wall layer and “enhanced dissipation” data sets. Their work was based on an extensive tower-based data set collected during the WAVES (Water-Air Vertical Exchange Study) experiments of 1985-1987. The tower, located 1100 m offshore in Lake Ontario was the same tower used by *Kitaigorodskii et al*. (1983). To minimize disturbances created by the tower, only data collected upwind of the tower were analysed; the WAVES dissipation data were thereby restricted to fetch limited waves and strongly forced conditions. The dissipation rate was parameterized with both wind and wave parameters. In addition to the friction velocity and depth, the wave parameters H_s and *F*, the significant wave height and energy input per unit area from wave to sea respectively, were introduced into the problem: $e = f(z, u_*, H_s, F)$. The model proposed a three layer structure of the upper ocean: an upper layer of constant dissipation rate into which energy is injected directly by breaking; an intermediate layer in which both wave breaking and shear are important turbulence sources, $eH_s/F = 0.3\,(z/H_s)^{-2}$; and a lower layer, in which shear dominates (i.e. a classical wall layer with $\varepsilon = u_{*w}^3/\kappa z$) - see Figure 1. The transition from the top to intermediate layers was found, by equating the total energy dissipation to energy input F, to occur at a depth of 0.6 H_s. The transition from intermediate to wall layer was found to occur at $3.6\,\bar{c}/u_{*a}$ or, for wave ages typical of the WAVES experiment, about 10 H_s.

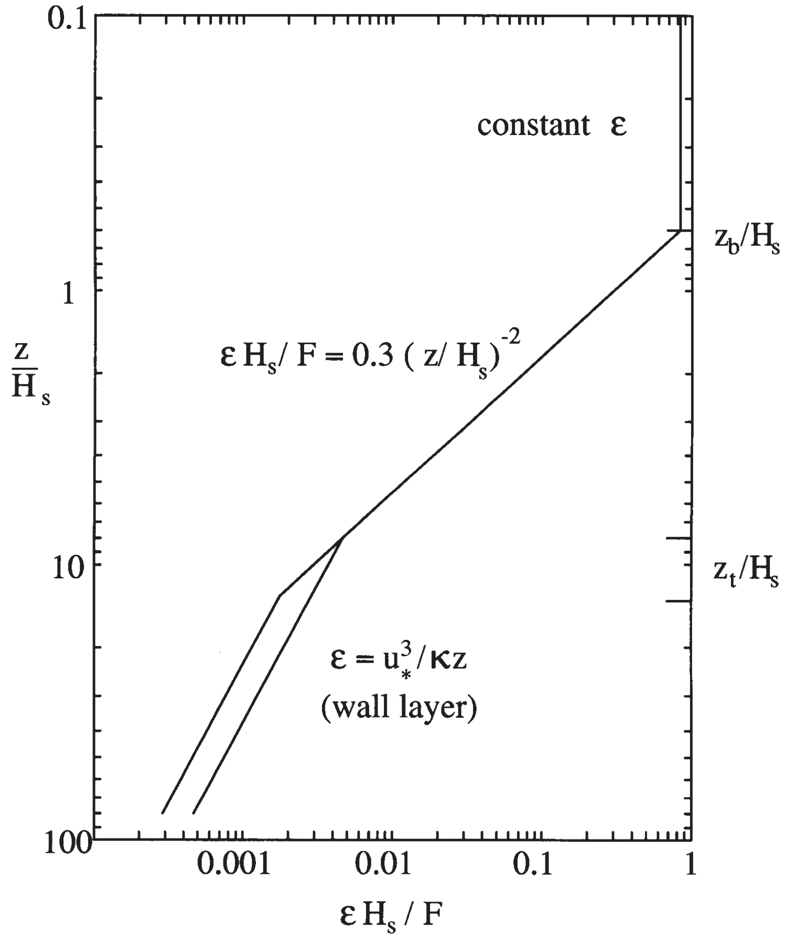

Fig. 1. Schematic showing the variation of dissipation rate, e with depth z using the scaling of *Terray et al.* (1996). H_S in the significant wave height of the wind sea, and F energy input per unit area from the wind into the water column. z_b and z_t mark transition depths between the regimes. The variation in z_t and the wall layer values show that there is an additional wave age dependence in the transition to wall layer behaviour.

The WAVES data, upon which the *Terray et al.* model was based, were strongly forced, short fetch data (with c_p/u_{*a} in the range 4 to 8 and H_s ~ 20 cm) and not representative of more fully developed oceanic data. There thus remained questions regarding the range of validity of the model. Furthermore, there was an additional *possible* wave age dependence in the intermediate layer scaling not apparent in the WAVES data. *Drennan et al*. (1996) addressed these issues, using new experimental data. During the Surface Wave Dynamics Experiment, SWADE, the 20m SWATH (Small Water-plane Area, Twin Hull) ship *Frederick G. Creed* was deployed in the Atlantic ocean, off the coast of Virginia, USA. Amongst the equipment deployed on the *Creed* was an acoustic current meter positioned about 2 m deep, ahead of the bow. During SWADE, data were collected in the wave age range of c_p / u_{*a} = 13 - 25, with $H_s \sim$ 1-2 m. The data were found to support the *Terray et al*. (1996) model, extending its validity to oceanic conditions.

4. *Wave-Turbulence Interaction*

The question of “wave-turbulence” interaction has been one of long-standing interest to Sergei Kitaigorodskii. With Miropolskii he presented the first analysis of the turbulent kinetic energy (TKE) budget in the upper ocean, and identified possible terms that might couple the surface wave field to upper layer turbulence (*Kitaigorodskii* and *Miropolskii*, 1968).

Returning later to this subject, he gave a careful analysis of the possibility of turbulent transport of wave energy, embodied in the flux term $< w'Q^2 >$, where Q^2 denotes the kinetic energy of the waves and w' is the vertical turbulence velocity (*Kitaigorodskii* and *Lumley*, 1983). Such a term had been previously studied by *Boyev* (1971) and *Benilov* (1973) for the cases of strong and weak turbulence, respectively, but the work of Kitaigordskii and Lumley attempted to provide a firm theoretical foundation for it.

Ölmez and *Milgram* (1992) argued that this transport mechanism could explain their laboratory measurements of the attenuation of short waves (having wavelengths less than 10 cm) by turbulence. They identified the ratio $\chi = c_p / w'$, where c_p is the phase speed of the waves, as the principal dimensionless variable controlling the importance of this mechanism. In their experiment χ varied between 25 and 140, with the bulk of their data having values around 50. They further suggested that similar values of χ might occur in wind-forced lakes and oceans. For example, estimating $w' \simeq 3u_{*w}$ we find $\chi \simeq 10 c_p / u_{*a}$, which implies, following Ölmez and Milgram, that this mechanism should be important for waves from early through intermediate stages of development.

In addition to the transport mechanism studied by Kitaigordskii and Lumley, a number of other terms appearing in the TKE budget have been suggested as a means of coupling waves and turbulence. For example, *Anis* and *Moum* (1995) proposed a contribution to the TKE flux of the form $< \tilde{w} q^2 >$ where q^2 denotes the turbulent kinetic energy, and $\tilde{w}$ is the vertical wave velocity. Several mechanisms for the wave-driven production of TKE have also been suggested. Based on field measurements, *Shonting*

(1970) and later *Cavaleri* and *Zecchetto* (1987) proposed that the wave field contains a rotational component. Although these observations are most likely flawed (*Santala*, 1991), recent laboratory measurements under carefully-controlled conditions have found a small wave-related Reynolds stress, t_w (*Chueng* and *Street*, 1988; *Magnaudet* and *Thais*, 1995). The working of this stress on the mean shear is then a source of TKE. *Anis* and *Moum* (1995) parameterize the wave stress as $\tilde{\tau} \simeq \tilde{w}^2 \sin\phi$, where the phase angle ϕ measures the departure of the waves from irrotationality. They point out that for long waves, phase angles as small as a degree or two are sufficient to account for the measured dissipation in the surface layer. Unfortunately, such small departures from irrotationality are unlikely to be observable in field data, and alternate methods of testing this (and other hypotheses) regarding wave-turbulence interaction are required.

The main difficulty with using the TKE budget to identify wave-turbulence interaction mechanisms is that the momentum and TKE equations are not closed, and hence cannot be used (without modelling assumptions) to demonstrate that the necessary correlations are generated by the underlying dynamics. Consequently it may be useful in this regard to return to the governing equations for turbulence forced by surface waves. We know from experiment that enhanced levels of dissipation (*i.e.* relative to a wall layer) exist at depths of order the wavelength of the waves. It is believed that the prompt effects of wave breaking extend to depths of only a few wave heights (*Rapp* and *Melville*, 1990), suggesting that there is a scale separation between the initial region of turbulence generation by breaking, and the subsequent maintenance and deepening of that turbulence by other processes. Since we are interested in the possible role of waves in the latter, we will neglect wave breaking, and look instead at the effect of the fluctuating strain rate of the waves on pre-existing turbulence. *Phillips* (1961) analysed this from the standpoint of the TKE equation, and concluded that the effect is negligible. We begin instead from the Navier-Stokes equations and ask whether wave straining can amplify the turbulence.

Since, apart from advection, the only interaction between waves and turbulence is due to wave straining, we linearize the Navier-Stokes equation to obtain

$$\partial_t u = u \cdot S + (u \cdot \nabla)\mathbf{U}, \tag{5}$$

where u is the velocity of the turbulence, $\mathbf{U}$ the mean current, and $S_{ij} = \partial_i \tilde{u}_j$ the fluctuating strain rate of the waves. We take the waves to be unidirectional and monochromatic (with angular frequency n_o), and further suppose that the current has the form $\mathbf{U} = [U(z),0,0]$. Then equation (5) reduces to the 2-dimensional system

$$\begin{aligned} \dot{u}_x &= \delta\left[u_x \sin\theta + u_z \cos\theta\right] + \gamma u_z \\ \dot{u}_z &= \delta\left[u_x \cos\theta - u_z \sin\theta\right], \end{aligned} \tag{6}$$

where the overdot denotes differentiation with respect to the variable $\theta = n_o t$. The dimensionless parameters $\delta = ka \exp kz$ and $\gamma = \partial_z U/n_o$ characterize the strain rates of the waves and mean shear, respectively.

If the current is not sheared, then it is straightforward to show that the waves can produce only a bounded, periodic fluctuation in the velocity. However, if the mean vertical shear does not vanish, then for $\delta^2/\gamma < 1$ this system of equations has exponentially-growing solutions. The growth rate, τ, is

$$\sigma / n_o = \sqrt{\gamma\delta^2}\sqrt{1-\delta^2/\gamma} \simeq \sqrt{\gamma\delta^2} \tag{7}$$

or in dimensional form

$$\sigma \simeq \sqrt{\partial_z U \partial_z U_s / 2}\,, \tag{8}$$

where $U_s(z)$ denotes the Stokes drift of the waves. The maximum growth rate, $\sigma_{max} = \gamma n_o / 2$ is attained when $\delta^2/\gamma = 1/2$. These results are reminiscent of the Craik-Leibovich theory of Langmuir circulation (*Craik* and *Leibovich*, 1976), where a similar expression for the growth rate is obtained. There the rectified effect of wave straining on the mean current appears as a rotation and stretching of vorticity by the vertically-sheared Stokes drift.

It is of interest to inquire whether the necessary parameter range $\delta^2/\gamma < 1$ is attainable. For shears of wall-layer magnitude, and using $k_p a = 0.32\left(u_{*a}/c_p\right)^{1/2}$ (*Maat et al*. 1991), it is straightforward to show that $\delta^2/\gamma = 1.35 k_p z \exp\left(-2k_p z\right) < 0.25$, where k_p is the wavenumber of the dominant waves. Hence the ratio $\sigma/\sigma_{max} < 0.87$, with the maximum occurring at $z = 1/(2k_p)$. However, the near-surface shear is almost certainly smaller (*Santala*, 1991), which has the effect of moving the depth at which $\sigma \simeq \sigma_{max}$ closer to the surface. We conclude that the required range of δ^2/γ is physically attainable, with growth rates close to the theoretical maximum occurring at depths $z < 1/(2k_p)$.

The growth of the turbulence is, of course, limited by the nonlinear terms in the velocity equation that we have so far neglected. These remove energy from the large scales at a rate $\tau^{-1} \sim q/$, where and q denote the characteristic length and velocity scales of the energy-containing eddies. If these eddies span the "wave layer", then $\sim 1/(2k_p)$. Based on our previous discussion of the dissipation rate given in section [3], we estimate $q = \beta u_{*w}$, with β a number somewhat greater than unity. Then $\tau^{-1} \simeq \beta n_o \gamma$, implying that $\sigma\tau \simeq \beta^{-1}\sqrt{\delta^2/\gamma} < \beta^{-1}$. Hence the "eddy turnover" time is fast enough to limit the growth of the eddies that are most strongly coupled to the waves, but is not so fast as to prevent them from growing appreciably. A more realistic model incorporating the coupling of scales of order ~ $1/2k_p$ to the nonlinear energy cascade should permit a quantitative computation of the saturation TKE level.

We conclude from the analysis discussed above that in addition to breaking, wave straining is likely to play a significant role in maintaining the observed vertical distribution of turbulence in the oceanic surface layer.

Acknowledgements

We gratefully acknowledge support provided by the U.S. Office of Naval Research (Grant N0014-94-1-0629 to the National Water Research Institute and N0014-89-1785 to the University of Washington). E.A.T. acknowledges support from the National Science Foundation (Grants OCE-84-18711 and OCE-93-01440).

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