THE AIR-SEA INTERFACE
Radio and Acoustic Sensing, Turbulence and Wave Dynamics

Edited by

M. A. Donelan
Rosenstiel School of Marine and Atmospheric Science
University of Miami
Miami, Florida, USA

W. H. Hui
Department of Mathematics
Hong Kong University of Science and Technology
Kowloon, Hong Kong

W. J. Plant
Applied Physics Laboratory
University of Washington
Seattle, Washington, USA

Rosenstiel School of Marine and Atmospheric Science
University of Miami, Miami, Florida

ON THE COUPLING BETWEEN A SURFACE WAVE MODEL AND A MODEL OF THE MIXED LAYER IN THE OCEAN

P. Lionello
Department of Physics, University of Padua, Padua, Italy

K. Hasselmann
Max Planck Institut für Meteorologie, Hamburg, Germany

and

G. L. Mellor
Princeton University, Princeton, New Jersey, USA

Abstract. We investigated the effect of surface waves on the turbulence near the air-sea interface and on the structure of the upper ocean. The study is organized as a series of numerical experiments. We carried out a one-way coupling between a wave model (namely the third generation wave model WAM) and a second order closure model of the upper ocean (namely the Mellor and Yamada 2.5 level model). We investigated the effect of a wave dependent surface stress, wave breaking, and shear due to the wave motion on the structure of the upper ocean. We concluded that accounting for the dependence of the surface stress on the wave spectrum increases the depth of the ocean mixed layer. The effect is larger at high wind speed and for a short duration. Wave breaking is an important source of turbulence in the ocean and it can explain the large turbulence dissipation that is observed in the uppermost part of the mixed layer. However, the turbulence generated by wave breaking does not give an effective contribution to the deepening of the mixed layer because it does not penetrate far from the sea surface. The inclusion of the source of turbulence associated with the shear of the wave motion in the second order closure model gives unrealistic results for the depth of the mixed layer.

1 Introduction

In this study we address the following questions. Does the energy and momentum transfer from the atmosphere to the ocean depend on surface waves? How large is the transfer of energy from the waves to the turbulence? Is it possible, in the framework of the models that are presently available, to represent these quantities? Are they important according to the model dynamics?

In order to answer we have to consider the interactions among three systems: the atmospheric boundary layer, the air-sea interface (i.e. the surface wave motion and the associated amount of energy and momentum), and the upper ocean (say the surface mixed layer). The three systems exchange energy and momentum as is sketched in Fig.1. There are fluxes of energy, denoted with ϕ, and of momentum, denoted with τ. Subscripts a, w, o denote respectively the atmosphere, the wave and the ocean (for instance $\tau_{\rm aw}$ is the flux of momentum from the atmosphere to the wave, $\tau_{\rm ao}$ the flux of momentum from the atmosphere to the ocean and so on). The overall downward fluxes from the atmosphere are $\tau_{\rm a} = \tau_{\rm aw} + \tau_{\rm ao}$, $\phi_{\rm a} = \phi_{\rm aw} + \phi_{\rm ao}$. The overall fluxes to the ocean are $\tau_{\rm o} = \tau_{\rm ao} + \tau_{\rm wo}$, $\phi_{\rm o} = \phi_{\rm ao} + \phi_{\rm wo}$. The net flux to the waves, resulting in their growth is $\tau_{\rm w} = \partial M_w/\partial t = \tau_{\rm aw} - \tau_{\rm wo}$, $\phi_{\rm w} = \partial E_w/\partial t = \phi_{\rm aw} - \phi_{\rm wo}$. The purpose of this study is to determine the importance of the effect of the waves on $\tau_o$and ϕ_o.

The waves are described using the WAM model (The WAMDI group, 1987). The WAM model solves the wave energy transport equation without making any parameterization of the low frequency part of the wave spectrum:

$$\frac{DF(t,\sigma,\theta)}{Dt} = S_{\rm in} + S_{\rm nl} + S_{\rm ds}. \tag{1}$$

Here $S_{\rm in}$ represents the input of energy from the wind, $S_{\rm nl}$ the nonlinear interactions, $S_{\rm ds}$ the dissipation due to the wave breaking.

The characteristics of the model and the expressions of the source function are described in detail in the original paper. With respect to the initial version of the model $S_{\rm in}$ and $S_{\rm ds}$ have been modified by adopting the quasi-linear theory of the wave generation (Janssen, 1991) according to which a dependence of the roughness length z_0 on the wave induced stress is introduced:

$$z_0 = \frac{\alpha\tau}{g\sqrt{1-\tau_{\rm aw}/\tau}}. \tag{2}$$

The model provides a computation of the friction velocity as a function of the wind speed U_{10} and of the wave spectrum F. Consequently the overall flux of momentum from the atmosphere is computed as $\tau_{\rm a} = \rho_{\rm a} u_*^2$. Moreover, the source functions allow the computation of the fluxes to and from the wave. For instance the energy flux

$$\phi_{\rm wo} = \frac{1}{2}\rho_{\rm w} g \int S_{\rm ds} dk \tag{3}$$

corresponds to the energy lost by the wave through breaking.

The TCM (Turbulence Closure Model) computes the velocity, the temperature, the salinity and the TKE (Turbulence Kinetic Energy). The closure adopted is the so called 2.5 level (Mellor and Yamada, 1982), which retains the anisotropy of the Reynolds stress only at the lowest order and neglects the material derivatives and the diffusion terms in the equation for the temperature and salinity variances (i.e. it assumes a local balance between generation and dissipation of the variance). This simplification is not applied to the equations for the TKE q and the mixing length l:

$$\frac{D}{Dt}\left(\frac{q^2}{2}\right) - \frac{\partial}{\partial z}\left[lqS_q\frac{\partial}{\partial z}\left(\frac{q^2}{2}\right)\right] = P_s + P_b - \epsilon\,, \tag{4}$$

$$\frac{D}{Dt}\left(\frac{lq^2}{2}\right) - \frac{\partial}{\partial z}\left[lqS_q\frac{\partial}{\partial z}\left(\frac{lq^2}{2}\right)\right] = lE_1(P_s + P_b) - \frac{q^3}{B_1}\left[1 + E_2\frac{1}{(kl)^2}\right]\,, \tag{5}$$

where P_s is the shear production of TKE, P_b is the buoyant production and ϵ is the dissipation. The constants B_1, E_1, E_2 are determined from neutral flow data, and the coefficient S_q is a function of the Richardson number. Therefore the model is capable of describing the diffusion of TKE along the vertical away from the place where it has been generated.

2 The effect of the wave on the surface stress

The rate of growth of wave momentum τ_{w} is much smaller than τ_{o}. This conclusion can be reached using the WAM model or using observations of wave development from which an estimate of τ_{w} can be derived. In fact assuming a Toba spectrum $F(\sigma,\theta) = \alpha_{\mathrm{T}} u_* g I(\theta)\sigma^{-4}$ (α_{T} is the Toba constant, g is the constant of gravity, $I(\theta)$ is the angular distribution, and σ is the frequency) one obtains

$$M_w = \rho_{\mathrm{w}} g \int d\theta I(\theta) \int d\sigma \alpha_{\mathrm{T}} u_* \sigma^{-3} = \rho_{\mathrm{w}} \frac{u_*^3}{g} \frac{\alpha_{\mathrm{T}}}{2} \frac{1}{\sigma_*^2}, \tag{6}$$

where $\sigma^* = \sigma u_*/g$. The wave momentum M_w divided by the characteristic time required for wave growth $T = u_* T_*/g$, $T_* \approx 10^6$, gives the net momentum flux into the wave : $\tau_{\mathrm{a}} - \tau_{\mathrm{o}} = \tau_{\mathrm{w}} \approx M_w/T = \rho_{\mathrm{w}} u_*^2 \alpha_{\mathrm{T}}/(T_* \sigma_*^2)$, which typically is few percent of τ_{a}.

This does not mean that the direct transfer of momentum from the atmosphere to the wave is small, on the contrary one has, according to our estimate (Lionello, 1993), $\tau_{\mathrm{aw}} \approx \tau_{\mathrm{a}}$ or a substantial fraction of it, say $0.15\ \tau_{\mathrm{a}} \leq \tau_{\mathrm{aw}} \leq \tau_{\mathrm{a}}$. It happens that the waves transfer to the underlying upper ocean almost all the momentum that they receive from the atmosphere. The present uncertainty on the magnitude of the overall momentum τ_{a} is much larger than the net flux to the waves τ_{w}. The direct effect of the waves on the momentum balance in the open ocean is consequently small, and presently it is not important to account for it using a sophisticated wave model.

To take the waves into account, however, may still be important because the flux of momentum from the atmosphere to the ocean is modulated by the waves. The drag coefficient is observed to reach a maximum when the sea is young and to decrease in its following development (Donelan, 1982). Presently two explanations of the observed dependence have been proposed: one is based on the effect of the waves on the surface roughness over the sea (e.g. Donelan, 1982). The second is based on the presence of a large wave stress, associated with wave induced fluctuations in the air flow (Janssen, 1991). Both approaches give a qualitative explanation of the observed behaviour of the drag coefficient in a growing windsea and the basic outcome of this coupling experiment does not depend on the approach that is chosen. The WAM model adopts the quasi-linear theory (Janssen, 1991).

To investigate if the dependence of the surface stress on the wave spectrum is relevant for the depth of the ocean mixed layer, we compared two numerical experiments. A coupled experiment in which the upper ocean model is driven by the variable stress computed by WAM and a reference experiment in which the stress has a constant value[1], which is in substantial agreement with the bulk formulas commonly found in the literature. The comparison has been carried out for $U_{10} = 10, 15, 20, 25$ m/s. The initial stratification was $1C^0/10$m, there was no buoyancy flux, and simulations were carried out at $60^0 N$. The resulting

[1] The time average of the stress computed by the WAM model has been used

depth of the mixed layer[2] is shown in Fig.2.

The effect of the dependence of the drag coefficient is an increment of the depth of the mixed layer. In fact the initially higher stress produces a faster deepening of the mixed layer in the coupled experiment. In the following development the differences in the stress between coupled and reference run decreases progressively, with a time scale that is proportional to the time needed for the wave to growth, i.e. proportional to the wind speed. Consequently at high wind speed the increment in the stress persists for a long time and the difference between coupled and reference experiments is relatively large. However, since the time scale of the wave growth is shorter than the time scale of the mixed layer deepening, the increment of the mixed layer depth is not large in spite of an initial large increment of the stress.

3 The role of waves on the energy flux into the upper ocean

Wave breaking is a major source of turbulence in the upper ocean. The amount of energy lost by the wave through breaking can be derived from the output of WAM, but also from general considerations. Since the energy E_w and the momentum M_w of the wave are related as $E_w = C_{ph}M_w$, where C_{ph} is the phase speed, one has $\phi_{\mathrm{aw}} \approx C_{ph}\tau_{\mathrm{aw}}$. The direct flux of momentum to the upper ocean from the air is $\phi_{\mathrm{ao}} \approx \tau_{\mathrm{ao}}u_*$. Consequently, assuming $\tau_{\mathrm{aw}} \approx \tau_{\mathrm{ao}}$, as has been briefly discussed in the previous section, then $\phi_{\mathrm{aw}}/\phi_{\mathrm{ao}} \approx C_{ph}/u_* \gg 1$. Therefore, if waves and upper ocean receive a comparable amount of momentum from the air then the waves receive a much larger amount of energy. Very little of the energy received by the wind is kept by the wave. One can estimate the overall wave energy $E = \rho_w\, u^4{}_* E_*/g$, $E_* \approx 10^3$ which, when divided by the time required for wave growth, gives an estimate of the energy flux $\phi_{\mathrm{w}} = E_*\rho_w u_*^3/T_*$ of order $\phi_{\mathrm{a}} = \rho_a u_*^3$. Therefore $\phi_{\mathrm{aw}} \approx \phi_{\mathrm{wo}} \gg \phi_{\mathrm{ao}}$.

The energy lost by the waves is not transferred to the mean motion and it consequently generates turbulence. This is deduced from the following argument. A wave energy loss $C_{ph}\Delta M_w$ corresponds to an amount of momentum ΔM_w lost by the wave field. The whole momentum is acquired by the mean motion, with an associated energy gain $(\Delta M_w)^2/2\rho\delta V$. The energy available to the turbulence is consequently $\Delta E_t = \Delta M_w(C_{ph} - \Delta M_w/(\rho)) = \Delta E_w$ as $\Delta M_w \to 0$. Therefore we conclude that the wave breaking is a major source of TKE (Turbulent Kinetic Energy) in the vicinity of the air-sea interface and that the modelling of the wave evolution and the use of (3) is presently the only way to compute such a source.

Our numerical experiments were designed to find out whether the TKE produced by wave breaking would be diffused far away from the sea surface, contributing significantly to the turbulence field in the ocean and producing some deepening of the mixed layer. A flux of TKE corresponding to the energy dissipated by the waves according to the WAM model is imposed at the uppermost layer of the TCM. The penetration of the turbulence produced by wave breaking

[2] The depth of the mixed layer is estimated as the level at which the temperature is $0.2C^0$ lower than the surface value.

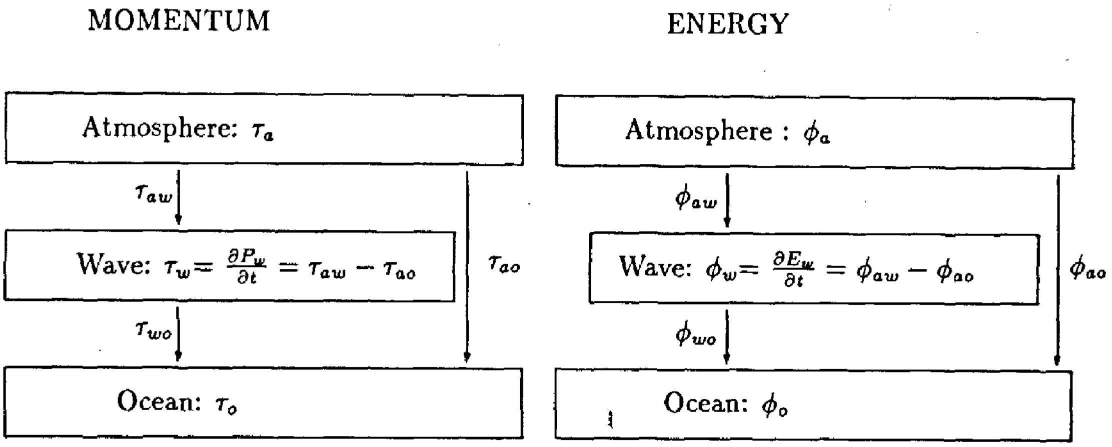

Fig. 1. Exchanges of energy and momentum among the atmosphere, the wave and the ocean.

depends on the value of the mixing length $L(1)$ that is imposed at the uppermost level $Z(1)$ of the TCM. If $L(1) \leq kZ(1)$, where k is the von Karman constant, there is little penetration, because the TKE dissipation is very large in the vicinity of the air-sea interface. On the other hand, a larger value of $L(1)$ allows the TKE to reach a deeper level before being dissipated. The data in Fig. 3 have been obtained by imposing $L(1) = 10u_*^2\sqrt{\alpha_p}/g$, where α_p is the Phillips constant. This means a mixing length proportional to the size of the high frequency waves in the tail of the spectrum, whose phase speed is lower than u_*. With this choice we have $L(1) \approx kZ(1)/10$. The large TKE dissipation shown in Fig. 3 in the vicinity of the sea surface is associated with the penetration of the TKE generated by wave breaking, which disappears at a larger depth, where the dissipation agrees with the prediction of the wall-layer theory. Fig. 3 shows results produced at various stages of the development of the windsea, showing the effect of the variation of the level of the spectral tail and of the flux ϕ_{wo}. The figure is meant to be compared against some measurements that showed in the presence of wave breaking a TKE dissipation 10 to 100 times larger than the prediction of the wall-layer theory (Agrawal *et al.*, 1992). The magnitude of the dissipation that we computed is comparable with the measurements but the depth does not agree – it is farther away from the surface in the measurements. The TKE produced by wave breaking gave no significant contribution to the deepening of the mixed layer in this numerical experiment.

4 Conclusion

The introduction of a wave dependent surface stress produces a deeper mixed layer. The increment of the depth is of the order of a metre. The effect is larger

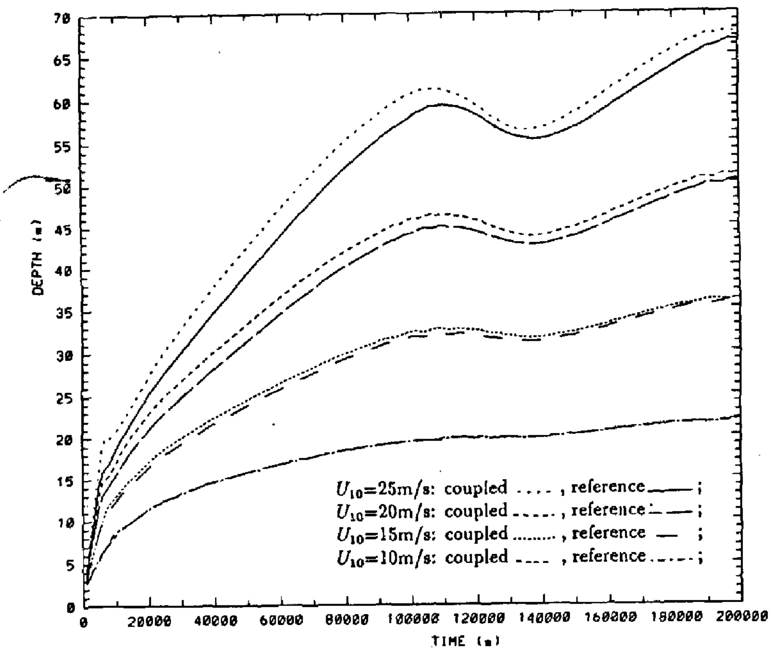

Fig. 2. The depth of the mixed layer as function of time.

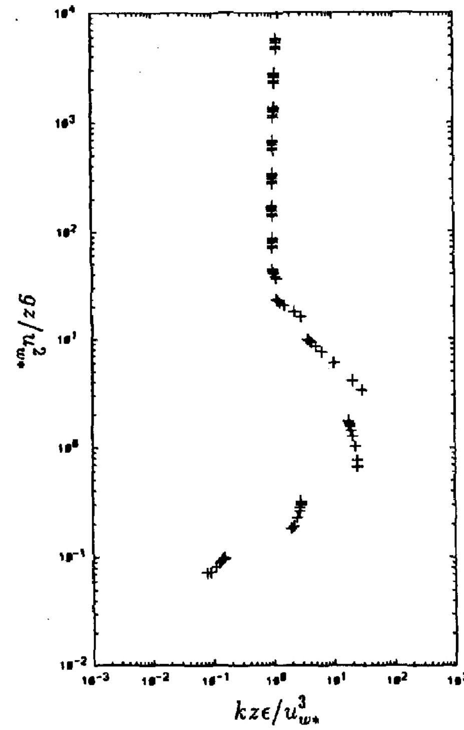

Fig. 3. The TKE dissipation (x-axis) as function of the depth (y-axis). Note that depth is normalized with u_{w*}^2/g, where u_{w*} is the friction velocity in the water, and dissipation with u_{w*}^3/kz, where k is the von Karman constant and z the depth.

when the mixed layer begins its formation and it decreases progressively. It persists longer at high wind speeds, but it is anyway not large because the time scale of the variation of the surface stress is much shorter than the time scale of the development of the mixed layer.

Wave breaking is a major source of TKE in the upper part of the mixed layer and its presence can explain the large dissipation observed near the air-sea interface (Agrawal *et al.*, 1992), but it is not relevant for the deepening of the mixed layer. In fact the large amount of TKE generated by the wave breaking is dissipated near the sea surface and it does not penetrate to the bottom of the mixed layer where it could contribute to its deepening.

We also introduced in the term P_s of (4) an extra contribution accounting for the shear of the wave motion. The sensitivity of the ocean model to the introduction of the wave shear is very large and it results in a very deep mixed layer – a 100% increase has been observed. The impact is too large to be realistic and we conclude that the modelling set-up is not adequate for a reliable analysis of this effect.

References

Agrawal, Y. C., Terray, E. A., Donelan, M. A., Hwang, P. A., Williams, A. J., Drennan, W. M., Kahma, K. K., and Kitaigorodskii, S. A.: 1992, 'Enhanced Dissipation of Kinetic Energy beneath Surface Waves', *Nature* **359**, 219-220.

Donelan, M. A.: 1982, 'The dependence of the aerodynamic drag coefficient on wave parameters', in *Proc. First International Conference on Meteorology and Air-Sea Interaction of the Coastal Zone*, American Meteorological Society, Boston, Mass., pp. 381-387.

Janssen, P. A. E. M.: 1991, 'The Quasi-linear Theory of WIND Wavve Generation Applied to Wave Forecasting', *J. Phys. Oceanogr.* **21**, 1631-1642.

Lionello, P.: 1993, 'The use of a wave model for the computation of the energy and momentum fluxes from the atmosphere to the ocean', *ISDGM T.R. 169*, pp. 20.

Mellor, G. L. and Yamada, T.: 1982, 'Development of a turbulence closure model for geophysical fluid problems', *Rev. of Geophys. and Space Phys.* **20**, 851-875.

The WAMDI Group (S. Hasselmann, K. Hasselmann, E. Bauer, P. Janssen, G. Komen, L. Bertotti, P. Lionello, M. Reistad, V. Cardone, J. Greenwood, A. Guillaume, L. Zambreski, J. Ewing): 1988, 'The WAM model - A third generation ocean waves prediction model', *J. Phys. Oceanogr.* **18**, 1776-1810.