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Article *in* Journal of Geophysical Research: Oceans · February 2014

DOI: 10.1002/2013JC009318

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Journal of Geophysical Research: Oceans

COMMENT

10.1002/2013JC009318

This article is a comment on *Huang and Qiao* [2010], doi:10.1029/2009JC005853.

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Citation:
Kantha, L., H. Tamura, and Y. Miyazawa (2014), Comment on "Wave-turbulence interaction and its induced mixing in the upper ocean" by Huang and Qiao, *J. Geophys. Res. Oceans*, *119*, doi:10.1002/2013JC009318.

Received 4 AUG 2013
Accepted 29 SEP 2013
Accepted article online 19 FEB 2014

Comment on "Wave-turbulence interaction and its induced mixing in the upper ocean" by Huang and Qiao

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1. Introduction

The interaction of surface gravity waves with the ocean below has been the subject of numerous investigations over the past century. However, for the most part, studies focused on wave/mean current interactions. Wave-turbulence interactions were more or less ignored (with a few exceptions mentioned below), because of the mistaken belief that the principally irrotational wave motions cannot possibly generate vortical turbulence except through processes such as wave breaking. However, pioneering theoretical studies by *Kitaigorodskii and Lumley* [1983] and *Kitaigorodskii et al.* [1983] and field observations by *Anis and Moum* [1995] brought wave-turbulence interactions to the forefront. Turbulence kinetic energy (TKE) dissipation measurements in the upper ocean made by *Anis and Moum* [1995] conclusively demonstrated the enhanced dissipation rates by wave breaking near the surface and by wave-turbulence interactions well below the surface, which were a significant departure from dissipation rates obtained from the traditional law of the wall. While they did suggest a model for wave-turbulence interaction-driven dissipation rate (more on this below), no mention was ever made of Langmuir circulation or Langmuir turbulence as the mechanism for such interactions.

It was however the study of Langmuir turbulence by *McWilliams et al.* [1997] that rekindled significant interest in wave-turbulence interactions. They demonstrated the large potential impact of "Langmuir turbulence" on mixing in the oceanic mixed layer and Stokes-Coriolis force on Ekman turning in the upper ocean. Based on this study, *Kantha and Clayson* [2004] showed how to include TKE production by wave-turbulence interactions in their second moment closure-based, two-equation model of the oceanic mixed layer. Their study confirmed the importance of this additional source of energy to turbulent mixing in the oceanic mixed layer, which arises due to the action of Reynolds stresses on the shear of the wave-induced Stokes drift current. *Kantha* [2006] followed up by demonstrating that this TKE source also constitutes a significant sink of wave motions and could be important for accurately quantifying the decay of swell in the open ocean, a problem that has defied solution. *Ardhuin and Jenkins* [2006] used generalized Lagrangian mean to show that as far as turbulence is concerned, in spite of the zero-vorticity nature of wave motions, the vertical shear of the wave-induced Stokes drift current behaves very similar to the vertical shear of the mean current, as far as the extraction of energy by the Reynolds stress is concerned. *Kantha et al.* [2009] showed that extraction of energy from wave motions by turbulence constitutes an important sink of wind waves in the global ocean, dissipating as much as 2.5 TW of wave energy, comparable to dissipation rates in the surf zones around ocean basins. In fact, by ignoring the air side of the ocean-atmosphere interface, they might very well have underestimated the dissipation rate by more than a factor of two (more on that elsewhere). *Kantha et al.* [2009] also coined the term Stokes production of TKE and Stokes dissipation of wave energy to denote this wave-turbulence interaction mechanism. *Carniel et al.* [2005] studied the effect of Stokes production on Ekman turning in the water column. *Teixeira and Belcher* [2002] have also studied the influence of surface waves on turbulence. For a recent review of turbulence in the upper oceanic mixed layer, from an observational point of view, see *D'Asaro* [2014].

Kantha et al. [2010] used the *Kantha and Clayson* [2004] model and comprehensive observations in the Baltic Sea to show that Stokes production of TKE is as important as the shear production. It is now widely recognized that a mixed layer model must include Stokes production of TKE for a realistic simulation of properties in the mixed layer. For example, *Smith et al.* [2013] recognize the importance of including the Stokes drift and Langmuir turbulence in their ocean-wave coupled model. However, the pendulum appears to have swung too far with some studies overestimating Stokes production. *Huang and Qiao* [2010] is an example and this comment is to caution against such overestimation.

Table 1. Parametric Values Calculated From *Anis and Moum* [1995] Observations as Reported in *Huang and Qiao* [2010][a]

Parameter	90d1	90d2	90n1	90n2	90n3	89n2	Wuest
τ_0 (Nm^{-2})	0.05	0.14	0.23	0.12	0.03	0.11	
Hs (m)	1.2	1.8	2.0	1.8	1.8	2.0	0.25
T (s)	6	6	5	6	7	8	4
$L=\frac{gT^2}{2\pi}$ (m)	56.2	56.2	39.0	56.2	76.5	100	25
β	0.75	0.15	0.75	0.40	1.0	1.0	1.0
$3.75\pi\beta\sqrt{\frac{H_s}{L}}$	1.29	0.32	2.00	0.84	1.81	1.67	2.34
$1-e^{-2kd}$	0.97	0.97	0.99	0.97	0.91	0.85	
U_{S0} (m s^{-1})	0.042	0.094	0.202	0.094	0.060	0.049	0.006
u^* (m s^{-1})	0.0067	0.0118	0.0149	0.0108	0.0056	0.0104	0.003
$Ln=(U_{S0}/u^*)^{1/3}$	1.85	2.00	2.39	2.06	2.20	1.68	1.27

[a]The turbulence Langmuir number Ln is calculated according to *Kantha and Clayson* [2000], *Carniel et al*. [2005], and *Kantha et al*. [2009]. Note that 89n1 is omitted because of two different swells propagating though the region.

2. *Huang and Qiao* [2010] and *Huang et al*. [2011, 2012] Studies

Using data collected by *Anis and Moum* [1995], *Huang and Qiao* [2010] parameterized the TKE dissipation rate due to wave-turbulence interactions as

$$\varepsilon_w = \frac{15\pi^2\beta}{L}\sqrt{\frac{H_s}{L}}\left(U_{S0}u_*^2\right)e^{2kz} \tag{1}$$

and found the value of the empirical constant β for various measurements of *Anis and Moum* [1995] by simply fitting equation (1) to the microstructure profiler-measured TKE dissipation rate profiles in the upper layer. The inferred value for β ranges between 0.15 and 1.0 (see Table 1). This parameterization has been included in a version of the *Mellor and Yamada* [1982] mixed layer (ML) model by *Huang et al*. [2011] to show that the inclusion of this additional source of TKE improves the performance of the model. However, *Huang et al*. [2011] did not modify the $q^2\ell$ equation in the two-equation second moment closure model, which *Kantha and Clayson* [2004] showed is essential to proper inclusion of Stokes production in such models. *Huang et al*. [2012] also deployed a microstructure profiler in the South China Sea to measure dissipation rates under moderate (4.7–8.9 m s^{-1}) winds and assert that these measurements agree with the parameterization in equation (1). Once again β values were found by just fitting equation (1) to observed dissipation rate profiles and the values range between 0.25 and 1.5 (see Table 2).

Interestingly, the parameterization in equation (1) was not proposed by *Anis and Moum* [1995], although they propose parameterizations for what they call the transport of wave-breaking turbulence downward by swell (their equation (30)) and for "indirect" production of turbulence by wave stresses (their equation (35)). In order to see if the parameterization in equation (1) does a good job of representing Stokes production, it is necessary to examine it closely. Integrating equation (1) w.r.t. z over a mixed layer of depth d, the total dissipation rate can be shown to be

$$\varepsilon_{\text{total}} = \alpha\left(U_{S0}u_*^2\right); \quad \alpha = 3.75\pi\beta\sqrt{\frac{H_s}{L}}\left(1-e^{-2kd}\right) \tag{2}$$

Now the Stokes production of TKE and hence the dissipation rate of TKE under equilibrium conditions [*Kantha and Clayson*, 2004; *Kantha*, 2006; *Kantha et al*., 2009] is

Table 2. Parametric Values Calculated From Data Collected During the Observational Campaign of *Huang et al*. [2012]

Parameter	S1	S2	S3	S4	S5	S6	S7	S8	S9	S10	S11	S12	H
Hs (m)	2.1	1.9	2.0	2.0	2.2	2.7	2.1	1.9	1.6	1.3	1.9	2.5	1.9
T (s)	9.7	10.0	9.9	9.8	10.6	10.6	10.0	9.6	8.9	8.2	8.4	10.7	7.1
$L=\frac{gT^2}{2\pi}$ (m)	147	156	153	150	175	175	156	144	124	105	110	179	79
β	0.25	0.20	0.45	0.20	1.0	0.25	0.6	1.5	0.5	1.5	1.0	1.0	1.0
$3.75\pi\beta\sqrt{\frac{H_s}{L}}$	0.35	0.26	0.61	0.27	1.32	0.37	0.82	2.03	0.67	1.97	1.55	1.39	1.83
$1-e^{-2kd}$	0.99	0.98	0.98	0.99	0.97	0.97	0.98	0.99	0.99	1.00	1.00	0.97	1.00

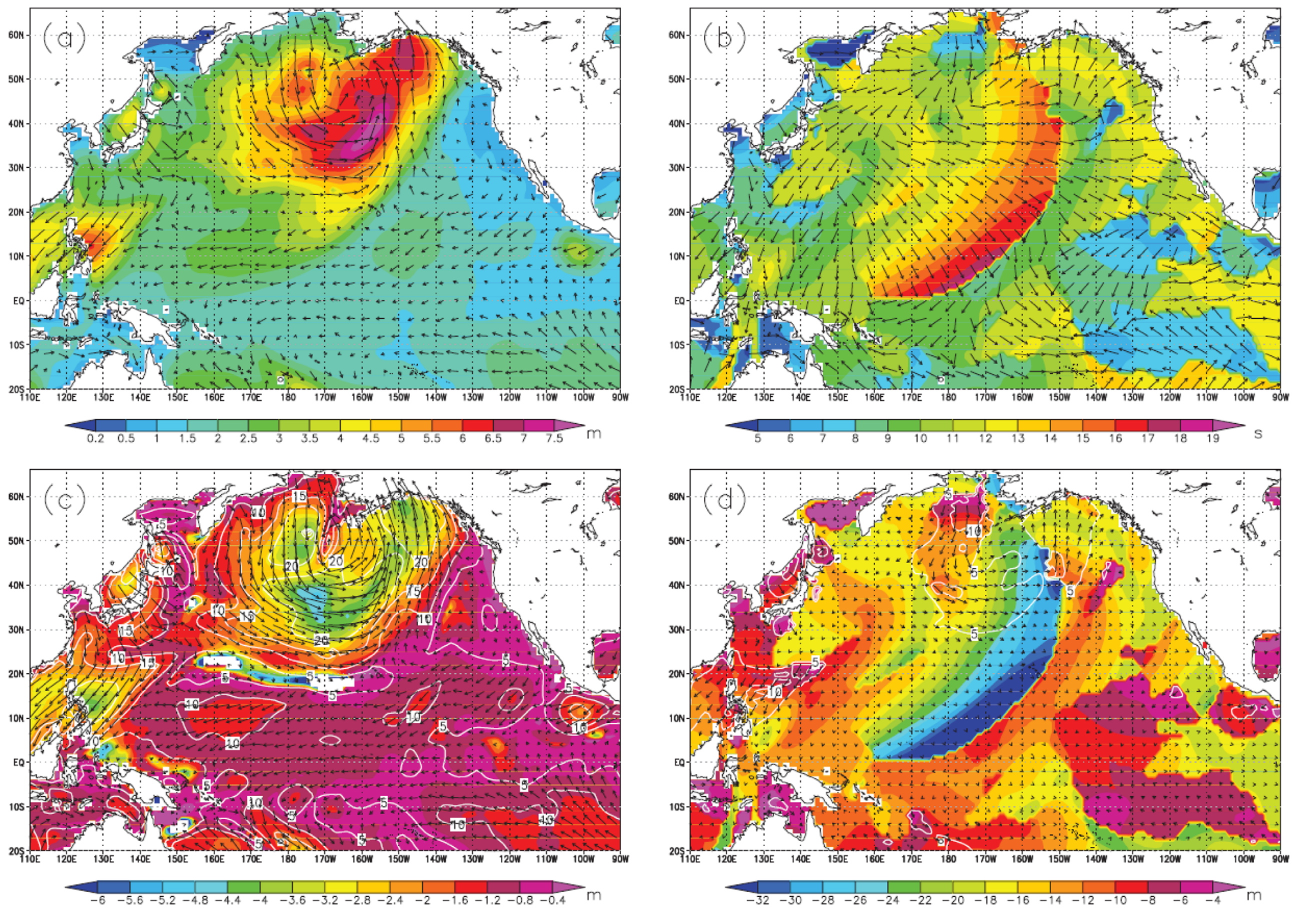

Figure 1. A snap shot of wave and the Stokes drift field (00z 15 January 2009) by wave hindcast [from *Tamura et al*., 2012]: (a) surface wind (vector) and Hs (color), (b) peak wave direction (vector) and peak wave period (color), (c) surface Stokes drift estimated from wave spectra (vector) and the Stokes *e*-folding depth (color), and (d) same as Figure 1c but estimated by wave bulk parameters.

$$\varepsilon_{St}=P_{St}=-\left(\vec{\tau}\cdot\frac{d\vec{U}_S}{dz}\right) \tag{3}$$

Integrating this in the vertical, we get

$$\varepsilon_{\text{total}}=\alpha\left(U_{S0}u_*^2\right) \tag{4}$$

where α is a constant. It is clear that the theoretical upper bound on α is 1.0, the value that prevails for a deep slab mixed layer in which the shear stress is aligned with the wind stress and constant with depth, and the swell is propagating in the direction of the wind itself. As *Kantha* [2006] demonstrated using *Kantha and Clayson* [2004] ML model that included the Stokes production term in both the q^2 and $q^2\ell$ equations, α is below 1.0, the precise value depending very much on the ratio of the ML depth to the swell wavelength L (or equivalently its period T). *Kantha et al*. [2009] used a value of 0.65 to make preliminary estimates of the dissipation rate of wind waves in the global ocean using WWIII model output for the year 2007.

Under the best-case scenario of a slab ML model with swell aligned with the wind, the upper bound on α in equation (2) is

$$\alpha=\left(1-e^{-2kd}\right) \tag{5}$$

Ignoring the exponential term, the upper bound becomes 1.0. This also means $3.75\pi\beta\sqrt{\frac{H_s}{L}}$ must be less than 1.0, and for swell of period greater than about 8 s, even smaller. Table 1 shows the values [*Huang and*

Qiao, 2010] found by fitting equation (1) to observed dissipation rate profiles. It is clear that β is being overestimated by a factor of 2–3 at times. *Huang et al*. [2011] use equation (1) with $\beta=1$ to augment mixing in their 3-D ocean model.

Traditionally, *Mellor and Yamada* [1982] type second moment closure models have underestimated mixing. Now it appears that with the inclusion of overestimated β values, *Huang and Qiao* [2010] and *Huang et al*. [2011, 2012] are overestimating mixing. Needless to say, neither is acceptable.

Overestimation happens because *Huang and Qiao* [2010] assume all of the observed dissipation (and hence TKE production) occurs from wave-turbulence interactions (more appropriately due to Stokes production), and fitted the observed epsilon profiles with equation (1). This is incorrect. A procedure similar to *Kantha et al*. [2009] is much better, since conventional shear (and even buoyancy) production may not have been negligible.

The relative importance of Stokes production in upper layer turbulence is indicated by the turbulence Langmuir number $Ln=\left(\frac{\vec{\tau}_w\cdot\vec{U}_S(0)}{\rho u_*^3}\right)^{1/3}$ as defined by Kantha [see *Kantha*, 2012; *Kantha and Clayson*, 2000; *Carniel et al*., 2005; *Kantha et al*., 2009, 2010], where $\vec{U}_S(0)$ is the Stokes drift current at the ocean surface, $\vec{\tau}_w$ is the wind stress, and u_* is the corresponding friction velocity. When the wind stress and the Stokes drift vector are aligned with each other, $Ln=\left(\frac{|\vec{U}_S(0)|}{u_*}\right)^{1/3}$ and these *Ln* values are shown in Table 1. It is clear that while Stokes production is large, shear production is not negligible. Moreover, equation (1), even with proper value of β overestimates Stokes production, since no ML is slab like and waves may not be propagating in the same direction as the wind stress.

Huang et al. [2012] repeated the *Huang and Qiao* [2010] exercise but with epsilon data they themselves collected in the South China Sea. But with no wave measurements, they used ERA-interim analyses to compute H_s and T_s. Consequently, this study is not as useful as the *Anis and Moum* [1995] study. Nevertheless, it is clear from Table 2 that the value of β is at times being overestimated, since equation (1) is again being simply fitted to observed profiles without properly removing shear production. This is evident especially at Station H in the coastal ocean, where a wave rider buoy was used to measure wave parameters, which are therefore more reliable.

There is yet another problem in the use of equation (1) for estimating the dissipation rate and hence Stokes production. Only for a monochromatic wave is the *e*-folding scale for the Stokes drift equal to $D_s=1/2k$. For a general wind wave spectrum, Stokes drift (both its magnitude and its depth of influence) is best computed from the spectrum itself [e.g., *Kenyon*, 1969; *Tamura et al*., 2012]:

$$\vec{U}_S(z)=\frac{2}{g}\int\int\sigma^3 S(\sigma,\theta)\frac{\vec{k}}{|k|}\exp(2kz)d\sigma d\theta \tag{6}$$

Equation (6) shows that the magnitude of the Stokes drift depends on the third moment of the spectrum, and therefore, shorter, higher frequency components contribute proportionately more. H_s and T_s, by definition, refer to longer, lower frequency components of the spectrum and generate a Stokes drift with a surface value of $U_{S0}=\frac{\pi^3 H_s^2}{gT_s^3}$ and an *e*-folding scale $Ds=\frac{gT_s^2}{8\pi^2}$. The actual *e*-folding depth scale is less than this value. What this means to the Stokes production depends very much on the ratio d/D_S, where d is the ML depth. For the same value of the wind stress and the U_{S0}, the Reynolds stress variation with depth depends on the value of d, while the shear of the Stokes drift current depends on D_S, and the magnitude of Stokes production is the dot product of the local Reynolds stress and the local shear of the Stokes drift velocity, integrated over the ML depth.

Tamura et al. [2012] examined the applicability of estimating the Stokes drift using bulk parameters, namely the significant wave height H_s and the peak period T_p, and assuming a monochromatic wave. Such bulk-parameter usage underestimates the surface Stokes velocity somewhat but, more importantly, it significantly overestimates the Stokes *e*-folding depth. The bulk parameter *e*-folding depth has little resemblance to the *e*-folding depth computed using the wave spectrum. Figure 1a shows the wind vectors and wave heights, and the resulting surface Stokes drift magnitude and Figure 1c shows the depth scale. Figure 1b shows the wave direction and the peak period. The surface Stokes drift magnitude and the depth scale calculated using bulk parameters is shown in Figure 1d. Comparison of the bottom two plots clearly shows that the bulk parameter usage overestimates the Stokes drift depth scale by an order of magnitude, with a spatial distribution more correlated with the peak period distribution, whereas the distribution computed from the wave spectrum (Figure 1c) bears no resemblance to the peak period distribution.

If the wave spectrum is not available, reliable estimates can be made using empirical correlations such as the one suggested by *Ardhuin et al.* [2009] for the surface value of Stokes drift:

$$U_{S0}(\sigma_c) \simeq 5\times10^{-4}\left[1.25-0.25\left(\frac{0.5}{\sigma_c}\right)\right]U_{10}\times\min\left(U_{10},14.5\right)+0.025(H_s-0.4) \tag{7}$$

which appears to be quite robust. σ_c is the cutoff frequency, which can be assumed to be infinity. The Stokes drift profile can itself be assumed to be

$$U_S(z)=U_{S0}\exp\left(\frac{z}{D_S}\right) \tag{8}$$

where D_s is the *e*-folding depth. *Li and Garrett* [1993] suggest

$$U_{S0} \simeq (0.014-0.015)U_{10};\, D_S=0.12\left(\frac{U_{10}^2}{g}\right) \tag{9}$$

whereas *Harcourt and D'Asaro* [2008] suggest

$$U_{S0} \simeq 0.0175U_{10};\, D_S=\frac{0.23}{2k_p} \tag{10}$$

but how well equations (9) and (10) perform for a general wind wave spectrum, especially with a significant swell component needs to be assessed before their use in determining Stokes production.

Huang et al. [2011] included Stokes production in their OGCM and concluded that subsurface thermal structures are improved by including the Stokes production. The Stokes production does enhance vertical mixing. However, as discussed above, they are overestimating Stokes production in several ways.

3. Concluding Remarks

While there is no longer any doubt that Stokes production of TKE contributes significantly to mixing in the oceanic mixed layer, it is important to include it properly in ML models, instead of using ad hoc formulations such as equation (1), especially at the risk of overestimating it, as *Huang and Qiao* [2010] and *Huang et al.* [2011, 2012] appear to have done. It is interesting to note that in contrast, *Zhang et al.* [2012] followed *Kantha and Clayson* [2004] approach to include Stokes production in their 3-D Princeton Ocean Model of the Yellow Sea and also came to the same conclusion that the surface mixed layer cannot be simulated properly without wave breaking and Stokes production, with the latter playing a more prominent role. Also, more recently, based on LES studies of "Langmuir turbulence," *Harcourt* [2013] has suggested that it may be necessary to modify the stability functions as well in two-equation second moment closure models to account for the Craik-Leibovich vortex force term [*Craik and Leibovich*, 1976] responsible for Langmuir circulation, but whether the resulting marginal improvements warrant the added complication (plus 2 more constants) is not yet clear.

Acknowledgments
This work is part of the Japan Coastal Ocean Predictability Experiment (JCOPE) supported by the Japan Agency for Marine-Earth Science and Technology (JAMSTEC). L.K. thanks JAMSTEC for hosting him during this study.

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