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On a wave-induced turbulence and a wave-mixed upper ocean layer

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[1] A concept of wave-amplitude-based Reynolds number is suggested which is hypothesised to indicate a transition from laminarity to turbulence for the wave-induced motion. If the hypothesis is correct, the wave-induced motion can be turbulent and the depth of upper ocean mixing due to such wave-generated turbulence can be predicted based on knowledge of the wave climate. Estimates of the critical wave Reynolds number provide an approximate value of Re_{cr} = 3000. This number was tested on mechanically-generated laboratory waves and was confirmed. Once this number is used for ocean conditions when mixing due to heating and cooling is less important than that due to the waves, quantitative and qualitative characteristics of the ocean's Mixed Layer Depth (MLD) are shown to be predicted with a satisfactory degree of agreement with observations. Testing the hypothesis against other known results in turbulence generation and wave attenuation is also conducted. **Citation:** Babanin, A. V. (2006), On a wave-induced turbulence and a wave-mixed upper ocean layer, *Geophys. Res. Lett.*, *33*, L20605, doi:10.1029/2006GL027308.

1. Introduction

[2] In this paper, the wave motion considered is that due to the wind-generated waves at the ocean surface. Non-linear corrections are unimportant for the purpose of the paper, and the linear wave theory is used to scale the mean wave motion as a function of water depth. Although such approach is routinely used and is regarded reliable on average, it is necessary to comment that very significant deviations from the linear theory predictions, both overestimations and underestimations of anticipated depth-dependent wave characteristics, have been reported (see, e.g., *Cavaleri et al.* [1978] for a review).

[3] At this stage, a few further caution words with regard to the linear theory have to be mentioned. Ocean waves of lengths $\lambda > 1$ m, if described by the perturbation theory based on Euler equation, are considered free, with no viscosity and surface tension [e.g., *Komen and Hasselmann*, 1994]). Such assumptions further lead to a conjecture that the waves happen to be irrotational. Although the former assumptions are a mere approximation and justified in most of the cases (with noticeable faults in other cases, as mentioned above, though), the latter feature of irrotationality imposes a serious limitation on the wave motion because it basically bans the wave-induced turbulence (although some theoretical mechanisms of generating the turbulence by irrotational waves are still possible [e.g., *Ardhuin and Jenkins*, 2006]).

[4] The wave-induced turbulence, however, is the topic of the present paper. A hypothesis of wave-amplitude-based Reynolds number is suggested. Critical value of this number, based on wave amplitude and orbital velocity, would indicate a transition from laminar to turbulent wave motion.

[5] There is accumulating evidence, both direct and indirect, that such turbulence does exist. Historically, as far as 35 years ago *Yefimov and Khristoforov* [1971] concluded that their measurements provide "a basis for assuming that small-scale turbulence is generated by the motion of waves of fundamental dimensions". They did not provide account on whether those waves were breaking or not, and we conducted estimates based on the breaking threshold criteria of *Babanin et al.* [2001]. Our conclusion is that, for the two records analysed by *Yefimov and Khristoforov* [1971, Figure 5], breaking rates of dominant waves were 0.4% and 0.01%. Both rates are marginal, second one being negligible, and we have to conclude that the substantial levels of wave-associated turbulence could not have been brought about by the wave breaking, but were induced by mean wave motion.

[6] *Cavaleri and Zecchetto* [1987] in their dedicated and thorough measurements of wave-induced Reynolds stresses gave explicit accounts for the wave breaking. One set of their data correspond to active wind-forcing conditions (many breakers present are mentioned), whereas the other set describes steep swell (no breaking). Non-zero vertical momentum fluxes in absence of breaking are evident. Magnitude of the fluxes appears to depend quadratically on the height of individual waves which is consistent with our wave-amplitude-based Reynolds number hypothesis. Cavaleri and Zecchetto concluded that "in the water boundary layer there can occur an additional mechanism of generation of turbulence… full, correct description of the phenomenon is still lacking". Further analysis of the Cavaleri and Zecchetto results will be done in the Discussion section below.

[7] Lately, *Babanin et al.* [2005] conducted simultaneous measurements of the surface wind energy input rate and the wave energy dissipation in water column of a finite-depth lake. They showed that, once the waves were present and even in absence of wave breaking, turbulence persisted through the entire water column, not only in the shear boundary layers near the surface and bottom.

[8] A need for the wave-induced turbulence has also been felt by theoreticians in their search for mechanisms to fill the gaps in explanations of ocean mixing and non-breaking wave attenuation. *Qiao et al.* [2004] brought in wave-induced turbulent viscosity and applied it in a global ocean circulation model to predict the upper-ocean mixing. To solve the closure problem, they introduced the mixing

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length of the wave-induced turbulence proportional to the range of the wave particle displacement. This is quite the same idea as that of the present paper where the problem is approached on the basis of experimental evidence and the wave-amplitude-proportional turbulent scale follows from our Reynolds number hypothesis.

[9] *Ardhuin and Jenkins* [2006] compared swell-attenuation rates measured in ocean observations, as well as those empirically inferred in operational wave models, with viscous and wind-caused [*Kudryavtsev and Makin*, 2004] damping and concluded that wave interaction with oceanic turbulence can be one of the reasons of such attenuation, albeit small to justify the observed decay. They remained within the frame of zero-vorticity for wave solutions, but introduced, in a way, a wave-induced turbulence. Source of their turbulence is Stokes drift, the zero-frequency solution, which turbulence then interacts with the surface waves. If the viscosity is allowed in theory, and obviously the water is a viscous fluid even though that can be neglected in majority of applications, it can further promote shear instability [*Balmforth*, 1999]. This issue is beyond the scope of the present paper. Whether the wave-induced turbulence is due to a rotational or irrotational phenomenon (i.e. Ardhuin-Jenkins mechanism), there must be a respective Reynolds number which identifies the balance of viscous and inertia forces and the transition from laminar to turbulent wave-caused motion.

[10] We will be using the linear theory to find our wave-based Reynolds number, to describe its distribution along the water depth and to approach a possible upper-ocean mixing mechanism due to such waves. We would like to emphasise that this is not a compromise with the criticism of the limitations of the linear theory above. To estimate the Reynolds numbers we only need a scaling of mean wave orbital motion at different depths which the linear theory approximates well enough.

[11] The hypothesis of Wave Reynolds Number, if proved, will have three consequences which we will try to address in the paper. First, the wave motion should be able to generate turbulence even in absence of wave breaking. As discussed above, this is not a completely unexpected conclusion as such turbulence has been observed for a while. What has not been appreciated, however, is the potential significance of such turbulence source as the waves in the ocean are ever present and wave-caused speeds of water motion are at least an order of magnitude greater than those of shear currents and Langmuir circulations which are usually attributed with turbulence supply.

[12] Second consequence is decoupling of the wave-induced non-breaking turbulence from analogies with the wall-layer law tradition which are often employed to describe such turbulence [e.g., *Agrawal et al.*, 1992]. According to the present hypothesis, the principal difference of the wave turbulence is the existence of characteristic length scale (radius of the wave orbit) as opposed to the wall turbulence which does not have a characteristic length.

[13] Third, such wave-induced turbulence would enhance the upper ocean mixing on behalf of the normal component of the wind stress. The wind stress plays a dual role in the upper-ocean dynamics. Tangential component of the stress generates the surface shear currents which further induce turbulence and promote mixing. Under moderate and strong winds, however, normal component of the wind stress dominates, which is supported by the momentum flux from wind to waves [e.g., *Kudryavtsev and Makin*, 2002]. This means that, before the momentum is received by the upper ocean in the form of turbulence and mean currents, and thus enters the further cycle of air-sea interaction, it goes through a stage of surface wave motion. This motion can directly affect or influence the upper-ocean mixing and other processes, and thus skipping the wave phase of momentum transformation undermines accuracy and perhaps validity of approaches based on the assumption of direct mixing of the upper ocean due to the wind.

[14] The hypothesis thus attempts to link together three ocean features which are routinely treated as separate properties: the wind waves, the near-surface turbulence and the upper ocean mixed layer. Mechanisms of deepening the Mixed Layer Depth (MLD) are believed to be affected by a number of ocean properties and processes: i.e. wind stress, heating and cooling, advection, wave breaking, Langmuir circulation, internal waves, with the surface wind forcing being the major factor at many circumstances [e.g., *Martin*, 1985]. If our hypothesis finds support, the role of the wind stress, acting at the ocean surface, may need to be reconsidered in terms of the mixing throughout the water column due to wind-generated wave orbital motion.

2. Hypothesis of the Wave Reynolds Number and Depth of the Mixed Layer

[15] The surface wave elevation η, propagating in time t and space x,

$$\eta(x,t) = a_0 \cos(\omega t + kx) \tag{1}$$

where $\omega = 2\pi f$ is the radian frequency and $k = 2\pi/\lambda$ is the wave number, has two characteristic length scales: wave length λ and wave amplitude a. In deep water, the amplitude a decays exponentially away from the surface:

$$a(z) = a_0 \exp(-kz) \tag{2}$$

where z is vertical distance from the mean water level.

[16] The wave length λ does not depend on depth z and defines the horizontal scale over which the wave oscillations change phase. It also characterises the depth of penetration of the wave oscillations (distance from the surface where the oscillations can still be sensed is approximately $\lambda/2$). This scale, however, does not comprise physical motion of the water particles. The motion of physical particles involved in the wave oscillations is depicted by the other scale, a, which is also the radius of wave orbits.

[17] It is the hypothesis of this paper that the a-based Reynolds number

$$\mathrm{Re} = \frac{aV}{\nu} = \frac{a^2\omega}{\nu} \tag{3}$$

where $V = \omega a$ is orbital velocity, and ν is kinematic viscosity of the ocean water, indicates transition from laminar orbital motion to turbulent. All the subsequent results are based on this single hypothesis. It is interesting to notice that the wave Reynolds number can have the velocity scale eliminated if the dispersion relationship $\omega^2 = gk$ (g is the gravitational constant) is used, and be expressed in terms of the two length scales:

$$\mathrm{Re} = \frac{a^2\sqrt{gk}}{\nu} = \sqrt{2g\pi}\frac{a^2}{\nu\sqrt{\lambda}}. \quad (4)$$

Thus, according to (2), for a given wave length λ (wave frequency ω) Reynolds number Re decays rapidly as a function of depth:

$$\mathrm{Re}_\lambda(z) \sim a(z)^2 \sim \exp(-2kz). \quad (5)$$

At the surface (z = 0, $\exp(-2kz)$ = 1), longer waves of the same amplitude a_0 will produce smaller Reynolds numbers.

[18] Let $\mathrm{Re}_{critical}$ be the critical value of the wave Reynolds number (3)–(4). If near the surface $\mathrm{Re}_\lambda(z) > \mathrm{Re}_{critical}$, then the corresponding wave orbital motion will be turbulent. At some depth $z_{critical}$, dependence (5) will lead to $\mathrm{Re}_\lambda(z_{critical}) = \mathrm{Re}_{critical}$, and from that depth down the orbiting will become laminar. $z_{critical}$, therefore, will define the mixed layer depth - depth of the upper ocean layer mixed due to the turbulence generated by orbital movement produced by the surface waves. Obviously, in reality the convection, advection, heating and other processes can alter this value. Also, the background ocean waters are nearly always turbulent, and therefore $z_{critical}$ is not the depth below which turbulence is absent, but is rather a depth below which we do not expect presence of wave-induced turbulence.

[19] From (3)–(4), the Reynolds number at a given $\lambda(\omega)$, as a function of z, is

$$\mathrm{Re} = \frac{\omega}{\nu}a_0^2\exp(-2kz) = \frac{\omega}{\nu}a_0^2\exp\left(-2\frac{\omega^2}{g}z\right). \quad (6)$$

Therefore, if the critical Reynolds number Re_{cr} were known, the critical depth, which is also wave-induced MLD, would be readily available:

$$z_{cr} = -\frac{1}{2k}\ln\left(\frac{\mathrm{Re}_{cr}\nu}{a_0^2\omega}\right) = \frac{g}{2\omega^2}\ln\left(\frac{a_0^2\omega}{\mathrm{Re}_{cr}\nu}\right). \quad (7)$$

As seen, for a wave frequency ω, if the wave height grows, MLD will increase. If a few waves of the same height but different scales are present at a time, the mixed layer will be mostly determined by the lowest frequency ω (longest length λ) as its z_{cr} will be the largest.

[20] Real wind-generated waves are spectral, and, apart from rare cases of pure swell, multiple wave scales are always superposed at the ocean surface. The wave spectrum, however, has a sharp peak, with spectral density (wave height) decaying very rapidly away from the peak frequency ω_p both towards smaller scales (higher frequencies) and larger scales (lower frequencies). Therefore, ω_p and associated wave height can be chosen as characteristic wave scales which determine MLD-z_{cr} in case of wind-wave spectrum. As the representative amplitude of spectral waves, half of the significant wave height $a_s = H_s/2$ will be used in this paper where significant wave height H_s is the height of one third of the highest waves and is typically used as a descriptive value for wave magnitude.

3. Verifications

[21] Comprehensive experimental validation of the hypothesis is left to further studies. Preliminary qualitative and quantitative verifications of the hypothesis, and consistency checks, however, can be performed by testing conclusion (7) on the basis of known values of MLD (z_{cr}) for different water bodies and different wave conditions. Additionally, a laboratory feasibility check was conducted.

[22] Based on the Black Sea depth of mixed layer ($z_{cr} \approx$ 25 m in April [*Kara et al.*, 2005]), we will determine the value of critical wave Reynolds number Re_{cr} according to (6) by having use of the typical extreme values of peak frequency and amplitude of the wind-generated waves in this sea in April. This dimensionless number should be universal for the wave motion, according to the hypothesis (3), and therefore be equally applicable to such outermost extremes as the deep ocean and much lower laboratory mechanically-generated waves. Thus, once this critical Reynolds number is known, we should be able to predict transition of non-forced wave motion from laminarity to turbulence in the laboratory and to predict MLD in the ocean for different wave circumstances on the surface in cases when wind stresses (waves) dominate over other mixing mechanisms.

[23] April in Northern Hemisphere is chosen for our estimates because the combined effect of surface cooling and heating on MLD is expected to be minimal in early spring [e.g., *Martin*, 1985], and therefore April data will provide the cleanest material for investigation of mixing due to wind (waves). An extensive wind-wave data set collected in a deep water region of the Black Sea throughout April, 1986 is available to the author (it is described in detail and partially tabulated by *Babanin and Soloviev* [1998]). The minimal value of the peak frequency f_p = 0.175 Hz is encountered three times in Table I of Babanin and Soloviev and is thus chosen as representative of the typical extreme wave conditions at the Black Sea in April. Corresponding values of variance m ranged from 0.379 m^2 to 0.500 m^2.

[24] Water temperatures recorded at the measurement site in April were around 10°C, and therefore, for the Black Sea whose salinity is half of that in the ocean, the kinematic viscosity was chosen as $\nu = 1.35 \cdot 10^{-6}$ m^2/s. The wave amplitude used is $a_s = H_s/2 = 2\sqrt{m}$. Finally, equation (6) leads to the critical Reynolds number in the range $\mathrm{Re}_{cr} = 2602 \div 3433$. Given the approximate nature of the estimates done, we chose Re_{cr} = 3000 close to the center of the range as the critical Reynolds number for wave orbital motion.

[25] Such Reynolds number is in a very good accord with critical numbers for other fluid flows. Typical Reynolds numbers for a great variety of engineering applications outside of the boundary layer are $\mathrm{Re}_{cr} = 2000 \div 4000$ [e.g., *Cengel and Cimbala*, 2006].

[26] Laboratory test was conducted in wave tank of the Monash University to check a feasibility of the found critical number. Regardless its relation with the upper ocean mixing, critical wave Reynolds number Re_{cr} = 3000 should be able to predict onset of turbulence for a simple case of monochromatic waves. Mechanically-generated gently-

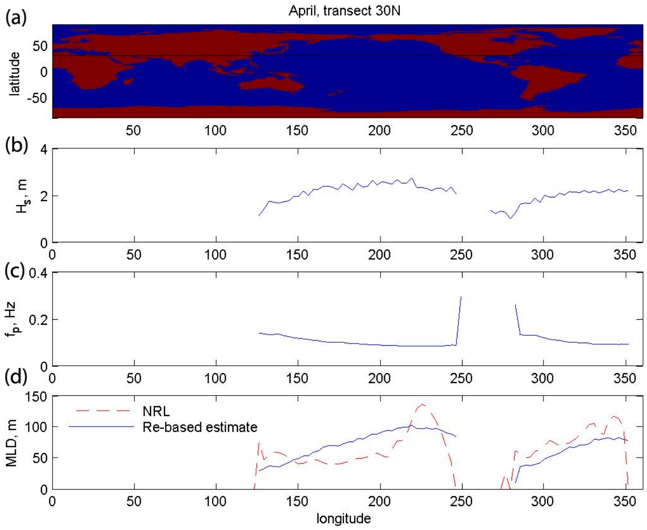

Figure 1. (a) World map. The transect latitude 30°N is shown with solid line. (b) Mean April significant wave height H_s along the 30°N latitude. (c) Mean April peak frequency f_p along the 30°N latitude. (d) Reynolds-number-based estimate of MLD (solid line) and NRL estimate of MLD (dashed line) along the 30°N latitude.

sloped (ak = 0.035 ÷ 0.075) waves were run which therefore involve no additional forcing superposed over wave orbital motion.

[27] A single set of measurements was conducted for waves of 0.667 Hz frequency, with the wave amplitude gradually been changed from 2 cm up to 4 cm, then down to 2 cm and up to 4 cm again. For such frequency and water depth of approximately d = 1 m, the waves are in a finite depth environment (kd = 1.9), and the wave orbits are somewhat elliptical. For simplicity, the vertical amplitude of the wave orbit was chosen to estimate the Reynolds number. Traces were injected into the water in the centre of the tank at 10 cm depth below the surface ($\exp(-kz)$ = 0.83), far away from the bottom, surface and wall boundary layers.

[28] At the lower margin of a = 0.02 m, wave Reynolds number (3) is Re = 1150 and the orbital motion was expected to be fully laminar. At the other end of a = 0.04 m, Re = 4600 and fully turbulent motion was expected. Transition was anticipated at $a \approx 0.03 \div 0.035$ m where Re = 2600 ÷ 3500.

[29] At a = 0.02 m, the motion remained clearly laminar, with patterns of injected ink, while moving along the orbits, stayed unchanged for minutes. At a = 0.03 m, some vortexes became visible which eroded the upper parts of ink patterns. At a = 0.04 m, the motion was obviously turbulent, with the ink being completely dissolved within seconds after injection. When the amplitude was reduced down to a = 0.02 m again, laminar behavior of the traces was immediately restored as the source of turbulence was apparently removed. On the return way up to a = 0.04 m, onset of turbulence was observed at approximately the same wave amplitude as previously.

[30] Before we apply the newly found Re_{cr} to find z_{cr} (7) in the ocean, we would like to emphasise that at this stage such estimates will only have an approximate value. Indeed, for the ocean wave conditions only mean (rather than mean extreme) magnitudes of wave height and peak period are available to us [*Young and Holland*, 1996]. Mean extremes would need a designated definition for this kind of estimates anyway, even if the raw wave data of interest were available, because those would have to be not only high waves, but also waves persisting long enough (normally tens of hours within the month, e.g., *Martin* [1985]) for MLD to settle. Besides, readings of mean waves and mean MLDs used here, even though taken for the same month of April, were done in different years and may be out of synch to some degree due to interannual variability. Also, the thermal, advective and other effects, although assumed to be relatively small, may not be negligible. And importantly, Sea Surface Temperature (SST) is used here to estimate kinematic viscosity which is further applied in equations (6) and (7). The temperature and therefore the viscosity are different below MLD (temperature is lower and viscosity is greater), and thus Re will be slightly smaller down there. This fact can lead us to some overestimation of MLD in the current exercise. Such precision details are beyond our scope, and what we would like to see in this section is an approximate quantitative and reasonable qualitative agreement of our predictions, based on equations (6)–(7), with ocean observations.

[31] For the ocean estimations, let us begin from well-documented directly-measured April values of MLD and SST in the Pacific at ocean stations *November* (140°W, 30°N) and *Papa* (145°W, 50°N). Values of $z_{cr} \approx 104$ m and $z_{cr} \approx 75$ m were read for the stations N and P respectively from Figures 5 and 6 of *Martin* [1985]. Corresponding surface temperatures were 19.6°C and 5.3°C, which lead us to $\nu \approx 1.07 \cdot 10^{-6}$ m^2/s and $\nu \approx 1.58 \cdot 10^{-6}$ m^2/s. Mean significant wave height and mean peak period, provided by the atlas of *Young and Holland* [1996] for April, were H_s = 2.33 m and f_p = 0.084 Hz at N, and H_s = 3.44 m and f_p = 0.097 Hz at P.

[32] Estimates, obtained for Re_{cr} = 3000 using equation (7), are $z_{cr} \approx 95$ m for N and $z_{cr} \approx 78$ m for P. Given the uncertainties mentioned above, they are in good quantitative (9% and 4% deviations respectively) and very good qualitative agreement with the measurements. One and a half times higher waves at *Papa* did not produce a deeper mixed layer because the excessive height was compensated by a more rapid depth attenuation of the wave motion due to the higher peak frequencies.

[33] To further investigate these agreements we conducted calculations for a transect across the Pacific and Atlantic oceans in April at the 30°N latitude which provides a significant variety of wave conditions along (Figures 1b and 1c). Wave climatology was taken from *Young and Holland* [1996], MLD climatology – from the US Naval Research Laboratory (NRL) site http://www7320.nrlssc.navy.mil/nmld/nmld.html. The latter was "constructed from the 1° monthly-mean temperature and salinity climatologies of the World Ocean Atlas 1994 [*Levitus and Boyer*, 1994; *Levitus et al.*, 1994] using a method for determining layer depth that can accommodate the wide variety of temperature and density profiles that occur within the global ocean." The satellite-measured SST for April were taken from US NOAA site http://www.osdpd.noaa.gov/PSB/EPS/SST/al_climo_mon.html

and were assumed of approximately constant value of 25°C, thus giving $\nu \approx 0.96 \cdot 10^{-6}$ m^2/s at 30°N.

[34] Results are shown in Figure 1. Along the transect, significant wave height varies 4 times, from 0.65 m to 2.7 m, and so does the peak frequency which varies from 0.08 Hz to 0.3 Hz (Figures 1b and 1c, respectively). Comparison of the NRL results with the present Reynolds-number-based estimate of MLD is shown in Figure 1d. The *Re*-based estimate reproduces the trend well, and given the fact that both the methods are results of modelling rather than in situ measurements, the quantitative agreement is also satisfactory.

4. Summary and Discussion

[35] In this paper, a concept of wave-amplitude-based Reynolds number is suggested which is hypothesised to indicate a transition from laminarity to turbulence for the wave-induced motion. If the hypothesis is correct, the wave orbital motion can generate turbulence and the Mixed Layer Depth can be predicted based on knowledge of the wave climate. It should be pointed out that possible rates of development of the mixed layer cannot be addressed on the basis of this hypothesis: the predicted MLD will be achieved if duration of the wave-carrying storms, or duration of combined succession of such storms, is long enough for the turbulent diffusion to take its course.

[36] Estimates of the critical wave Reynolds number provide an approximate value of Re_{cr} = 3000. This number was tested on mechanically-generated laboratory waves and was confirmed to describe onset of turbulence. If this value is used, combined with known wave climate data, quantitative and qualitative characteristics of MLD are shown to be predicted with a good degree of agreement with observations for the month of April when the combined effect of other processes is expected to be minimal.

[37] Obviously, prediction of MLD is not the only potential outcome of the introduction of wave-induced turbulence. It is therefore interesting to check our results across a range of oceanographic applications. The two records described by *Yefimov and Khristoforov* [1971] (assuming the water temperatures were between 5°C and 20°C) produce *Re* between 120000 and 180000 and between 50000 and 70000 respectively and therefore should have induced the turbulence as they did. The swell, propagating across the Pacific from New Zealand to Alaska without much of attenuation [*Snodgrass et al.*, 1966], had *Re* of 310, 260, 140 and 50 at stations Tutuila, Palmyra, Honolulu and Yakutat respectively and therefore could not have spent any energy on producing the turbulence which would be accompanied by a rapid damping (here, our estimates are done for the case described in Figure 30 of Snodgrass et al. and the water temperature is taken 20°C along the entire swell's path which makes the Reynolds numbers an upper limit values).

[38] More material for checking are the results of *Cavaleri and Zecchetto* [1987] where multiple unexpected features were measured which could not have been explained: 1) wave-induced turbulence existed, even in the case of non-breaking swell; 2) momentum fluxes below the water surface did not match (exceeded) the wind stress; 3) in case of wind-forcing, the wave orbits were tilted. Consequences 1) and 2) appear to be directly related to expectations of the present paper. 1) The observed swell was quite steep, with peak steepness of 0.054 just below the breaking threshold of 0.055 of *Babanin et al.* [2001]. Corresponding $Re \sim 200000$ clearly indicate values well-above the onset of turbulence and therefore non-zero vertical momentum fluxes from such a source of turbulence are to be expected. 2) Turbulent stresses in the water and in the air do not have to match as the wave turbulence can be generated even in absence of the wind if the waves are steep enough. Obviously, the two-order magnitude difference observed in the paper under conditions of active generation is mostly due to the phase shift between the horisontal-vertical components of the orbital velocities, but the present wave-induced turbulence would also contribute towards excessive vertical momentum fluxes. 3) Tilting of the orbits does not appear to relate to the wave-induced turbulence as it was only present in case of the wind-generated waves. Since those waves were also actively breaking (dominant breaking rate of some 13% according to our estimate based on *Babanin et al.* [2001] dependence), we would suggest that the tilting was perhaps due to asymmetry of the wave shape which is a prominent feature of the breaking waves [e.g., *Young and Babanin*, 2006]. We must emphasise at this stage that the suggested critical Wave Reynolds Number only signifies onset of turbulence generation and does not bear information on rates of this generation. Such rates, which would be of most interest for the turbulence, dissipation and mixing modelling, cannot be inferred from the Reynolds number and would have to be approached by other theoretical and experimental means.

[39] To conclude this paper, we would like to say that our hypothesis needs further elaboration and more experimental verifications. If, however, it is proved, it may have significant implications on a variety of oceanographic, meteorological and climate modelling applications. For example, for the depth-dependent turbulent scale, which is a crucial property of the models for oceanic turbulence [e.g., *Craig*, 2005], a wave-based turbulent length may need to be introduced.

[40] In particular, the hypothesis may have important consequences for the modelling of climate. If the wave length and height, rather than wind speed directly, are responsible for the magnitude of MLD, the climate models would have to incorporate wave properties.

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