BOREAL ENVIRONMENT RESEARCH 3: 127–136 ISSN 1239-6095
Helsinki 6 November 1998

Wind wave breaking and aerodynamic roughness of the air-sea interface as seen from above and from below

Sergei A. Kitaigorodskii

Finnish Meteorological Institute, P.O. Box 503, FIN-00101 Helsinki, Finland; on leave from the Russian Academy of Sciences, Institute of Oceanology, Moscow

Kitaigorodskii, S. A. 1998. Wind wave breaking and aerodynamic roughness of the air-sea interface as seen from above and from below. *Boreal Env. Res.* 3: 127–136. ISSN 1239-6095

In this paper an attempt is made to explain some of the features of classical aerodynamic classification based on the effect of mobility of roughness elements (Kitaigorodskii 1968) for both sides of the air-sea interface by incorporating into such a classification the effect of breaking wind waves. Also the "fluctuating" regime of the sea surface roughness discussed by Toba *et al*. (1991) is explained by taking into account the existence of the dissipation subrange (Kitaigorodskii 1983, 1986, 1992a, 1992b, 1997) and the inverse cascade of energy in the well developed wind wave field (Zaharoff and Zaslavskii 1982, and Kitaigorodskii 1983). In the last section of the paper a theory is proposed on how to incorporate the effect of wave breaking in the parametrization of the sea surface roughness as seen from below. It is shown that aerodynamically smooth conditions in the classical sense never exist below the sea surface, since effective turbulent viscosity due to shear free turbulence generated by wave breaking is much larger than molecular viscosity.

Brief review

In the recent summary of air-sea interaction studies by Donelan (1990), the data analysis led to rather interesting conclusions about the variability of sea surface roughness. Together with the study of Geernaert *et al*. (1986) it demonstrated the relative validity of the fluid dynamical analogy between momentum transfer in the vicinity of a solid rough surface and at the air-sea interface with moving waves considered as roughness elements. The latter analogy was first suggested and used by Kitaigorodskii (1968) as an effort to explain both the successes and limitations of Charnock's (1955) prediction for the variability and the wind dependence of the surface roughness parameter. When applying the Kitaigorodskii (1968, 1973) classification of aerodynamic properties of the air-sea interface, two questions are of primary importance:

a) what is the limitation of the fluid dynamical analogy between the actual sea surface and solid rough surfaces? and,
b) what is the effective roughness height for different wind-wave generation conditions?

Charnock's (1955) idea that all the stress is supported by the high-wave number part of the wave displacement spectrum, leading to a strictly wind speed or (friction velocity) dependent roughness length, was both very attractive and adequate for many ocean cases. According to this idea the effective height of roughness elements of the sea surface h_s depends only on the friction velocity u_* and gravity g, i.e.,

$$h_s \approx u_*^2/g \tag{1}$$

This leads to the typical value of $h_s \approx 0.01$ m being much larger than the typical value of the thickness of the viscous sublayer $\delta_v \sim \nu/u_* \approx 5.10^{-5}$ m with a typical value of $u_* \approx 0.30$ m s^{-1}. The Reynolds roughness number Re_s based on these two length scales is equal to:

$$\mathrm{Re}_s = h_s/\delta_v = u_*^3/g\nu \tag{2}$$

Practically, for all reasonable values of u_*, it is larger than 100. Therefore, the sea surface (from the view point of an observer at a fixed point above it) looks generally aerodynamically rough. The fact that $\mathrm{Re}_s \geq 10^2$ was a justification to simply write that

$$z_o = m\frac{u_*^2}{g} \tag{3}$$

where m was later on called the Charnock's constant. For the range of wind speeds 7.5–20 m s^{-1}, the average value of $m \approx 0.014$ shows that the viscous stresses cannot be very important (note that the Nikuradse's (1933) grain roughness concept $z_o = h_s/30$ would correspond to $m = 0.03$). More precise descriptions of the critical values of Re_s can be found from the classical relationships identifying the aerodynamically smooth and transitional regimes:

$$z_o = 0.1\delta_v = 0.1\frac{\nu}{u_*} \qquad \text{for } \mathrm{Re}_s \leq \mathrm{Re}_s' \approx 5 \tag{4}$$

$$z_o = A_s h_s \quad \text{with } A_s = \frac{1}{30} \quad \text{for } \mathrm{Re}_s \geq \mathrm{Re}_s'' \approx 90 \tag{5}$$

There were a number of studies stressing the variability of m with sea state, starting from Kitaigorodskii (1968). Among them, one must mentioned also the relatively recent works of Geernaert *et al.* (1986), Donelan (1990), Toba *et al.* (1990) and Donelan *et al.* (1993). However, all these works do not shed much light on the physical mechanisms behind the observed "variability" of the numerical value of m, which underlines one of the deficiencies of Charnock's approach. The range of observed variability for $m \approx 2.10^{-3}$–2.10^{-1} was probably one of the reasons to look at its wave age dependence. Most of the authors cited in Donelan *et al.* (1993) share a common view that the roughness decreases with increasing wave age. This is more or less in agreement with the concept of "mobility" of the roughness elements of the sea surface developed in Kitaigorodskii and Volkov (1965) and Kitaigorodskii (1968). The classification of the latter author leads to the unavoidable conclusion that the ratio of the roughness height to the mean square wave height (or significant wave height) is a diminishing function of the ratio C_p/U_a (or C_p/u_*) where C_p is the phase velocity of the spectrum peak. However, the collection of both field and laboratory data do not produce the universal relationships of this kind, but rather looks like either two different parallel lines (like in fig. 1. in Donelan *et al.* 1993) or as a curve with extremum at point $U_a/C_p \approx 10^{-1}$ (Ebuchi *et al.* 1990, Toba and Ebuchi 1991). These facts demonstrate one of the most definite difficulties in the application of the strict fluid dynamical analogy between a rough solid surface and the sea surface with "mobile" roughness elements with respect to the drag law — the concept suggested by Kitaigorodskii as early as in 1968. We shall argue in the next sections that one of the main reasons for this is related to the differences in the *wind wave breaking* conditions at different stages of wave development, as well as to the appearance of "aging" waves, moving faster than the wind and producing the "feedback" mechanism of transferring momentum from waves to the air flow. In order to take these effects into account, the useful tool for parametrisation of wave breaking can be Kitaigorodskii's (1983, 1992a, 1992b) theory of the dissipation subrange and the inverse energy cascade model (Kitaigorodskii 1983, Zaharoff and Zaslavski 1982, 1983, Zaharoff 1992).

Influence of wind wave breaking and dissipation subrange on the "variability" of the sea surface roughness as seen from above.

As we have already mentioned, the normalisation of experimental data on the "variability" of the sea surface roughness according to Kitaigorodskii (1968) does not produce the expected collapse of the field and laboratory data, rather, it produces two different groups of experimental points. We will argue below that the reason for the difference in laboratory and field data is due to the difference in wave breaking conditions or in other words in the different geometrical structures of the water surface under these two regimes. In laboratory conditions, microscale wave breaking does not reduce sufficiently h_s, and practically all wave components contribute to the "mobile" roughness elements. This is probably due to the fact that for this range of U_a/C_p (≥ 2.5), directionality becomes so important (Banner 1990) that only waves moving into the direction of wind can break, and they all belong to the dissipation subrange in the whole rear face of spectra. In other words these strongly nonlinear waves constitute the dissipation subrange.

Very recently, on the other hand, the prediction about the existence of a dissipation subrange in presence of a quasi-equilibrium form of the spectra dominated by direct energy cascade was verified and results were presented in Kitaigorodskii (1992a: fig. 1) and in Kitaigorodskii (1992b: table 1). They rather clearly demonstrate how the lower frequency boundary of the dissipation subrange depends on the stage of wave development (the peak frequency). We were recently able to add some additional data to these results, and construct the relationship between the transitional frequency ω_g and the peak frequency ω_p, also including laboratory data of Jahne and Riemer (1990). This composite picture is reproduced here in Fig. 1.

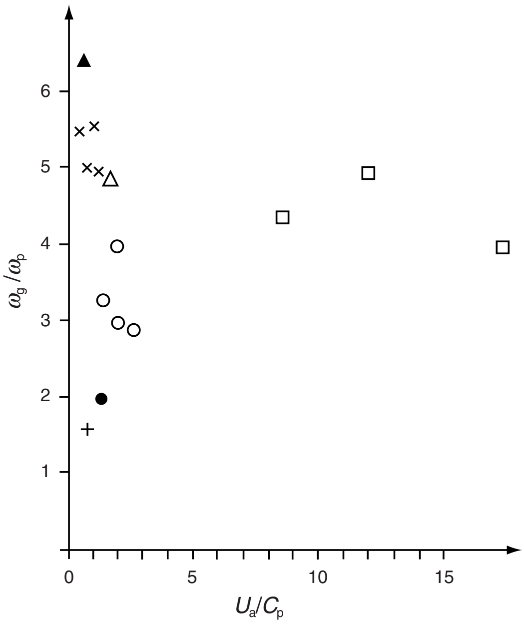

Fig. 1. Relationship between the ratio of the transitional frequency to peak frequency ω_g/ω_p and "wave age" U_a/C_p. Key to the data: × = slope frequency spectra (Tang and Shemdin 1983), △ = spatial short wave number spectra in upwind and downwind direction (Banner *et al*. 1989), + = stereowave observation project (SWOP, 1960), ○ = frequency spectra on lake Washington (Hansen *et al*. 1990), ● = spatial spectra of developed wind waves (Lupyan and Sharkov 1989), □ = spatial upwind and down wind spectra in laboratory conditions with short fetches (Jahne and Riemer 1990). The detailed description of the sources used above are given in Kitaigorodskii (1992a, 1992b).

The transitional frequency ω_g in Fig. 1 can be related to the effective value of transitional wave number $\hat{k}_g$ equal to

$$\hat{k}_g = \int k_g(\Theta)\mathrm{d}\Theta = Bg\varepsilon_o^{-2/3} \qquad (6)$$

where ε_o is the nonlinear energy flux towards high wave numbers, B an absolute constant and θ the angle. The results presented in Fig. 1 thus demonstrate that ε_o increases with the shift of the peak frequency towards lower frequency, i.e. with

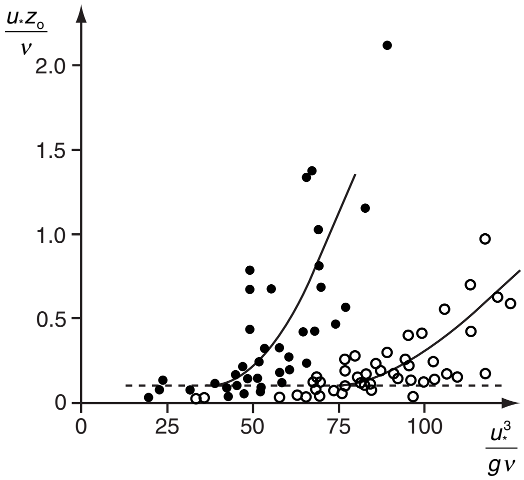

Fig. 2. Dimensionless dependence of the roughness parameter z_o on friction velocity u_* for large values of $\frac{g\sigma_\eta}{u_*^2} = 1.2 \times 10^2$, where σ_η is the mean square surface displacement (redrawn from Kitaigorodskii 1973). (●): experimental data for $\frac{g\sigma_\eta}{u_*^2} = 1.2 \times 10^2$, (○): experimental data for $\frac{g\sigma_\eta}{u_*^2} = 0.6 \times 10^2$, and the dashed line is $u_* z_o / \nu = 0.11$ for the aerodynamically smooth wall.

wave growth. This interpretation seems to be rather natural, even though we do not have a direct proof of this. The experimentally derived values ω_g in Fig. 1 were defined (Hansen *et al.* 1990) as the beginning of the rapid spectral fall-off on the rear face of the frequency spectra (or through $\omega_g = (gk_g)^{0.5}$) in spatial one-dimensional spectra, associated with the asymptotic approach to the saturation form:

$$\Psi_s(k,\theta) = Bk^{-4}\, \varphi(\theta) \quad (7)$$

$$S(\omega) = \beta g^2 \omega^{-5} \quad (8)$$

Here $B = 2\beta$, while $\Psi_s(k,\theta)$ is a symmetrical wave number spectrum, and $S(\omega)$ the frequency spectrum. Actually, due to the mismatch between the equilibrium forms of the spectra $\Psi_s(k,\theta)$ and $S(w)$ (with k_{in}, ω_{in} typical wind energy input wave number and frequency, respectively)

$$\Psi_s(k,\theta) = A\varepsilon_o^{1/3} g^{1/2} k^{-7/2} \Phi(\theta) \quad \text{for } k_{in} \le k \le k_g \quad (9)$$

$$S(\omega) = 2A\varepsilon_o^{1/3} g \omega^{-4} \quad \text{for } \omega_{in} \le \omega \le \omega_g \quad (10)$$

and the asymptotic saturation form (Eqs. 7 and 8), the spectral fall-off can identify a more rapid decrease of energy with k or ω than k^{-4} (or ω^{-5}). As a first approximation for a description of the dissipation subrange we can now accept those given by Fig. 1 and Eq. 6.

It is clear from Fig. 1 that in laboratory conditions (high values of ω_p) the transition to the dissipation subrange, if it exists, occurs at very high frequencies (or wave numbers). The width of the dissipation subrange is rather narrow and thus the roughness parameter is smaller than for more developed waves. The "variability" of boundaries of the dissipation subrange must be reflected in the "variability" of the aerodynamic roughness of the sea surface (Kitaigorodskii *et al.* 1995). Indeed the width of the dissipation subrange and its location in the wave-number space indicate at what range of scales the wave breaking is a dominant process in limiting the growth of wave components. But simultaneously, wave breaking can be associated with the separation of air flow behind the crests of the waves, the process responsible for creating an effective roughness height h_s and the analogy with solid surfaces. Therefore, according to the relationship $\omega_g = \omega_g(\omega_p)$, we can expect that with movement of ω_g towards lower frequencies, the width of the dissipation subrange must increase and the effective roughness, proportional to the contribution for mean square surface displacement from the dissipation subrange, must also increase in its turn. This is possibly the reason for the difference between laboratory and field data presented in Donelan *et al.* (1993; Figs. 1 and 2).

The latter conclusion is very important; it shows that the popular method for scaling sea surface roughness as z_o/H_s or z_o/σ_η versus U_a/C_p (with $H_s = 4(\bar{\varsigma}^2)^{0.5}$ the significant wave height) is not clarifying much, since H_s or σ_η also depend on U_a/C_p, and neither H_s nor σ_η are necessarily a fixed portion of the contribution to $(\bar{\varsigma}^2)^{0.5}$ from the dissipation subrange.

As a rule, plots representing the dependence of a non-dimensional roughness length versus wave age (as was proposed by Kitaigorodskii, 1968) show an increase of z_o/H_s by 2 or 3 orders of magnitude from order of 10^{-2} to 10^0 (more or less a linear increase). However, z_o can of course behave in a different way since such plots can either just reflect a dependence of H_s on U_a/C_p (spurious correlation), or which is most relevant, just reflect the decrease of z_o with a decrease of U_a/C_p, the interesting circumstance, which we will address later.

Therefore, it is more relevant to look at the relationship between roughness and some more refined characteristics of underlying surfaces. This was probably noticed for the first time in the paper by Toba and Ebuchi (1991). The most interesting feature of the time series of simultaneous wind and wave records presented in their fig. 5 and 6 was the observed correlation between sea surface roughness and deviations of wind wave spectra from the so-called Toba's ω^{-4}-wind dependent frequency spectra, which can be interpreted also in the framework of Kolmogoroff's direct energy cascade as in Eqs. 9 and 10. Toba and Ebuchi (1991) introduced what they called the degree of undersaturation $\tilde{\alpha}_s$ defined as

$$\tilde{\alpha}_s = 1 - \frac{\alpha_s}{<\alpha_s>} \qquad (11)$$

where $<\alpha_s>$ is the average value of the coefficient of the energy level in the equilibrium Toba-spectrum given by:

$$S(\omega) = <\alpha_s> g u_* \omega^{-4} \qquad (12)$$

The value of $<\alpha_s>$ is 0.062 according to Jones and Toba (1991). The Toba-spectrum, if interpreted in the framework of a direct cascade theory in the wind wave field (Kitaigorodskii 1983, 1986, Zaharoff 1992, Hansen *et al.* 1990), simply requires replacement of $<\alpha_s>$ either by a value α_u (Hansen *et al.* 1990), or by A, with another non-dimensional combination relating ε_o with u_* and g (Kitaigorodskii 1983).

According to Toba and Ebuchi (1991) the fluctuations in $\tilde{\alpha}_s$ are caused by the delay for the adjustment of the energy level in the spectrum (Eq. 12) to the wind variations. The physical interpretation of the correlation between $\tilde{\alpha}_s$ and z_o given by Toba and Ebuchi (1991) were the following: the variation of U_{10} on time scales of several minutes to one hour causes the variations of ω_p and $\tilde{\alpha}_s$. The large $\tilde{\alpha}_s$-values correspond to the situation where the level of the equilibrium range increases rapidly. Such a condition is associated with larger z_o. This probably means that the wave components of the very high frequency part of the wave spectra outside the equilibrium range are steeper in this situation, and small scale processes related to viscosity and surface tension will be more important in the momentum flux. According to Toba-Ebuchi's opinion, while $\tilde{\alpha}_s$ is changing, the shape of the spectra remains still close to ω^{-4}, which can be interpreted as important for wave-wave interactions in maintaining the equilibrium form (12). Consequently, the air momentum entering the waves due to the increase in roughness at the very high frequency part of the spectrum is quickly transferred to the whole equilibrium range "up" to near ω_p.

Toba and Ebuchi (1991) also mentioned the delicate variation of the wave spectral shape near ω_p with $\tilde{\alpha}_s$. However, all their guesses, though probably correct, neither give a clear enough answer to what are the main reasons behind the strong correlation between $\tilde{\alpha}_s$ and surface roughness, nor give a key to the solution of the problem of how to parameterise surface roughness with respect to wind wave field properties. Recently, Kitaigorodskii *et al.* (1995) reanalysed some of the data presented in Toba and Ebuchi (1991) and demonstrated that the variability of the width of the dissipation subrange is indeed of primary importance in the observed correlation between $\tilde{\alpha}_s$ and the roughness height h_s of the sea surface.

The influence of wind wave breaking on the aerodynamic roughness as seen from below

When considering the wind-induced drift current just below the sea surface it is usually assumed that the mean velocity profile can be described by the logarithmic law in a specific range of depth. That is how the concept of the sea surface roughness as seen from below is usually introduced. Roughness as seen from below has been used in parameterising the fluxes across air-sea interface of gases whose resistance is in the water phase (Kitaigorodskii and Mälkki 1979). However, Kitaigorodskii (1984) demonstrated that in the presence of wind wave breaking the mechanism of turbulent transfer in the vicinity of the gas-liquid interface (air-sea interface) is quite different. We repeat here a sentence from this paper: "In real wind wave generation conditions, even with only small scale wave breaking, the transformation of wave energy into turbulence can dramatically change the character of turbulence just below the wavy surface. Because of this the whole concept of classical momentum viscous sublayer, based solely on the analogy with flow above a rigid surface, can be quite inappropriate for the description of the damping effect of the wavy free surface on the dynamics of turbulence generated by breaking waves". Nevertheless, the analogy with the model for surface roughness as seen from above is still used in modelling of the upper ocean layer structure in connection to the dynamics of pure drift currents (Zilitinkevich *et al.* 1991).

When considering the drift current below the sea surface it is usually assumed that the velocity profile can be described as

$$U_s - U(z) = (u_*^w/\kappa) \times \ln (z/z_{od}) \tag{13}$$

where U_s is the velocity of the surface drift, u_*^w is the friction velocity in water ($u_*^w \approx u_*/30$ and $\kappa = 0.4$) and z_{od} is the surface roughness as seen from below. The basic contribution to U_s is due to drift current, the share associated with Stokes drift U_{sd} can be assumed small (Wu 1975). However, under strong wind conditions of the open ocean and large fetches this is not necessarily true. Anyway, laboratory experiments (Wu 1975, Kreiman and Karlin 1979) show that Eq. 13 is valid, especially under light wind conditions. Of course, for the existence of a logarithmic portion in wind drift velocity profile, there must exist a range of depth where neither wave breaking, nor rotation and density stratification are important. Since the Ekman depth for drift current, of order u_*^w/f (f is the Coriolis parameter) is almost always much larger than the amplitude of breakers, we can expect the existence of a constant flux logarithmic region for most of hydrometeorological situations. Zilitinkevich *et al.* (1991), suggested to describe z_{od} by the similar formula as in the case of roughness as seen from above (ν^w is the viscosity of water), i.e.

$$z_{od} = m_2 \frac{\nu^w}{u_*^w} \tag{14}$$

$$z_{od} = m_3 \frac{\left(u_*^w\right)^2}{g} \tag{15}$$

The fluid dynamical analogy for roughness as seen from above gives reasonable values for the constants m_2 and m_3 ($m_2 = 0.135$; $m_3 = 0.014 - 0.035$). The value of $m_3 = 0.034$ using the concept of moving roughness elements (Kitaigorodskii 1973) leads to $z_o/h_s \approx 1/30$ where h_s is the sand roughness, in good agreement with the classical results of Nikuradse (1933). Data from laboratory experiments performed by Kreiman and Karlin (1979) and Zilitinkevich and Kreiman (1991) together with field data from lakes yielded the following values (Zilitinkevich *et al.* 1991):

$$m_2 = 33 \text{ and } m_3 = 3.10^3 \tag{16}$$

Zilitinkevich *et al.* (1991) tried to construct the same model for z_{od} as in Kitaigorodskii (1973) for the dependence of $z_{od}u_*/\nu$ on the non-dimensional roughness number $Re_s = h_s/\delta_\nu = (u_*^w)^3/g\nu$. This is illustrated in Figs. 2 and 3, which we have reproduced here. Fig. 2 is taken from Kitaigorodskii (1973) which at this time represented the first effort to describe the "variability" of z_o with different wind wave conditions. Fig. 3 is taken from Zilitinkevich *et al.* (1991) where the authors were trying to use a similar scaling as in Fig. 2, but where u_* was taken as the friction velocity in water (u_*^w). However, the deficiency of the analogy between z_{od} and z_o can be clearly seen from the implied unreasonable high values for m_2 and m_3 of Eq. 16.

This can be explained by considering first the analysis of rough conditions (Eq. 15). The first thing to mention is that the choice of u_*^w together with g for characterising the roughness length z_{od} does not make much sense. The main reason is that the scale $(u_*^w)^2/g$ is even smaller than the Kolmogoroff's microscale l_ν. According to the latest measurements of Drennan *et al.* (1991), the dissipation ε^w in the presence of breaking wind waves close to the sea surface is in the range $(0.5 - 10) \times 10^{-4}$ m^2 s^{-3} which leads to

$$l_\nu = \left[\frac{\left(\nu^w\right)^3}{\varepsilon^w}\right]^{0.25} \gg \frac{\left(u_*^w\right)^2}{g}$$

Hence if the breaking is important for introducing the scale for roughness from below, then it still must be related to the amplitudes of breaking waves, which must be related to $\frac{u_*^2}{g}$ (but not $\frac{(u_*^w)^2}{g}$). Formula (15) can therefore be rewritten as:

$$z_{od} = m_3{}' u_*^2/g \quad \text{with } m_3{}' = 3 \tag{17}$$

This indicates that it makes more sense in choosing u_*^2/g as a scale for z_{od} in presence of breaking waves, than $(u_*^w)^2/g$. This means, if one wants to use Charnock's idea for the parameterisation of roughness as seen from below, that it is reasonable to have the usual scale u_*^2/g measuring the amplitude of high frequency wave. Nevertheless, it is still important to notice that, for moderate wind conditions, the sea surface roughness as seen from above $z_o << z_{od}$ as given by (Eq. 17).

An explanation to this can be given by considering wave breaking as a mechanism similar to the generation of shear free turbulence by a grid (Kitaigorodskii, 1984). The amplitude of the oscillations of the grid can be related to the amplitude of breaking waves. Then, it becomes no more impossible that the sea surface roughness as seen from below is much larger than the one seen from above. For the latter roughness parameter, the important length scale is related to the amplitude of short waves, responsible for flow separation behind their crests. For the roughness below the sea surface, the important length scales are associated with the amplitude of breakers which can be larger then the amplitude of short waves responsible for flow separation in the air. This becomes even more evident if one remembers that for the surface as seen from below the effects of flow separation are highly unlikely (due to relatively low values of U_s compared to the phase velocity of short gravity waves). Thus, the use of the length scale u_*^2/g has a different meaning for the roughness as seen from above as to the roughness as seen from below. This is possibly the main reason for the enormous differences in the value of the Charnock's constant in each case.

To explain the high values of m_2 in Eqs. 14 and 16, which supposedly must represent the aerodynamically smooth regime below the sea surface, we must prove that the scale $\frac{\nu^w}{u_*^w}$ is not a proper measure of roughness in this case (remember that the value of $m_2 = 33$ cited in Zilitinkevich *et al.* (1991) is about 300 (!) times larger than the value of the similar constant for aerodynamically smooth conditions above a solid surface). The most obvious way to prove this is to use the fact that, even for light wind conditions, microscale wave breaking, which is difficult to observe directly, can produce close to the surface the diffusion of turbulence downward as observed in Drennan *et al.* (1991) and Terray and Bliven (1985).

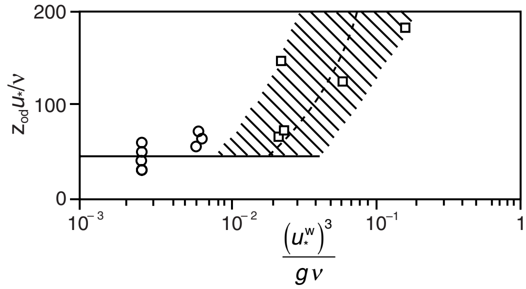

Fig. 3. Dimensionless dependence of the drift-roughness parameter z_{od} on the friction velocity u_*^w in water from laboratory experiments: ○ = according to Zilitinkevich and Kreiman (1988), □ = according to Kranenburg (1984). The horizontal solid line corresponds to Eq.(14) with $m_2 = 33$, the hatched region to Eq. 15 at $10^3 < m_3 < 10^4$, and the dashed line to Eq. 15 with $m_3 = 3 \times 10^3$. (Figure redrawn from Zilitinkevich and Kreiman 1991).

This diffusion can be better characterised using the theory of shear free turbulence with a constant eddy viscosity K (Long, 1978). The value of K is, as we will show below, at least 10^2 times larger than ν^w. Thus, instead of Eq. 14, we can propose another expression for z_{od} which is also based on dimensional considerations, i.e.

$$z_{od} \approx \frac{K}{u_*^w} \tag{18}$$

Eq. 18 just reflects the fact that in the presence of a momentum flux and drift current the shear free turbulence approximation for effective turbulent viscosity coefficient K is more appropriate than for example eddy viscosity associated with shear instability which makes K z-dependent. Then the constant of proportionality in (18) will be of order one and since $\frac{K}{\nu^{\mathrm{w}}} = 10^2$ that will explains the high value of m_2 in (16).

Thus, close to the surface, in presence of a constant momentum flux, the first to develop must be a linear velocity profile based on K = const., and it must be continued by a logarithmic part. Then, with some justification, we can choose as constant of proportionality in (18) the same value as in typical shear flows, i.e. to write instead of Eq. 18

$$z_{\mathrm{od}} \approx a_1 \frac{K}{u_*^{\mathrm{w}}} \quad \text{with } a_1 = 0.11 \tag{19}$$

This value of a_1 is now taken by analogy with the viscous sublayer but for region where the eddy viscosity K = const. There are some direct evidence of this since carefully designed laboratory measurements (Terray and Bliven 1985) of wind induced current below waves indicate the existence of a thin slab above the logarithmic region. This slab can be attributed to the well mixed region moving with a surface drift velocity U_s. This can be considered as a supporting argument that a thin region below the sea surface looks like shear-free. Of course, in natural field conditions, including lakes, the presence of wave breaking can lead to the formation of a linear-log profile (this should not be mixed with the log-linear profile in a stratified surface layer). Formula (19) just reflects the situation when, still in presence of a momentum flux $\rho_{\mathrm{w}}\left(u_*^{\mathrm{w}}\right)^2$ and a drift current, the shear free turbulent approximation K = const. is a better approximation for turbulence in the vicinity of the sea surface in presence of breaking wind waves than for example the eddy viscosity $K = k u_*^{\mathrm{w}} z$.

Now it remains to be proved that $K >> \nu^{\mathrm{w}}$. Estimating K in a shear-free turbulence model, and applied to the turbulence below wavy surface, definitely presents some difficulties. The natural way to estimate K is to write it in the form

$$K = a\Omega L \tag{20}$$

where a is the amplitude of the oscillations of the grid which can be related to the amplitude of breaking waves a_{br}, Ω is the frequency of the oscillation of the grid which can be related to the periodicity of breaking events. Setting $\Omega = 2\pi(10T)^{-1} \approx \Omega_{\mathrm{br}}$, where T is a typical period of breaking waves, the equality $\Omega_{\mathrm{br}} = 2\pi(10T)^{-1}$ means that Ω_{br} represents the periodicity of breaking events (Kitaigorodskii 1984), but not the period of breaking wave $T = 2\pi/\Omega_{\mathrm{br}}$ (here for simplicity we just accept that only one out of ten waves breaks). A more detailed discussion can be found in a recent review paper (Kitaigorodskii 1997a). In Eq. 20, L is a length scale which can be proportional either to a, i.e. yielding,

$$K = a^2\Omega \tag{21}$$

or to some geometrical characteristics of the grid, which can possibly be related to the distance between breakers. In the latter case, we can also allow for the existence of an additional dependence of the form $K = a^2\Omega F[a/L]$. Taking $\nu^{\mathrm{w}} \approx 10^{-6}\ \mathrm{m^2\,s^{-1}}$, it is enough to have $a_{\mathrm{br}} \approx 10^{-2}$ m, and $T \approx 1$ s to get $\frac{K}{\nu^{\mathrm{w}}} \approx 10^2$. Now according to our new interpretation of the sea surface roughness z_{od}, we must replace the nondimensional parameter $\frac{\left(u_*^{\mathrm{w}}\right)^3}{g\nu^{\mathrm{w}}}$ used by Zilitinkevich *et al.* (1991) by the ratio $\frac{K}{\nu^{\mathrm{w}}}$, which we can consider as a new Reynolds roughness number Re_s^k for the turbulence regime below the sea surface in the presence of breaking wind waves. With K defined as in Eq. 21, and

$$\Omega \approx \frac{2\pi}{10T} \approx \frac{1}{T} \approx \left[\frac{2\pi\lambda_{\mathrm{br}}}{g}\right]^{-0.5}$$

where λ_{br} is the wavelength of breaking waves (this is probably correct only for laboratory and light wind conditions, *see* Kitaigorodskii, 1997a), we will get

$$\mathrm{Re}_s^k \cong \left[\frac{a_{\mathrm{br}}}{\lambda_{\mathrm{br}}}\right]^{0.5} \frac{a_{\mathrm{br}}^{\frac{3}{2}} g^{\frac{1}{2}}}{\nu^{\mathrm{w}}} \left[\frac{1}{2\pi}\right] \cong \frac{\left(u_*\right)^3}{g\nu^{\mathrm{w}}} \left[\frac{a_{\mathrm{br}}}{2\pi\lambda_{\mathrm{br}}}\right]^{0.5} \tag{22}$$

since $a_{\mathrm{br}} \approx \frac{u_*^2}{g}$. Thus, the new Reynolds roughness number for the sea surface as seen from below with the constant value $a_{br}/\lambda_{br} \approx 0.1$ is approximately similar to the one used for the sea surface from above, but with one but very important distinction: instead of ν in the air we have ν^{w} the viscosity in the water in the nondimensional ratio $\frac{u^3}{g\nu^{w}}$. This leads to values of Re_s^k larger or even much larger than $\mathrm{Re}_s = \frac{u_*^3}{g\nu^{a}}$ (*see* Eqs. 2–5). Thus, the sea surface as seen from below is rougher in the classical sense than the sea surface as seen from above. This is an unexpected and interesting conclusion which deserved to be accurately checked against observations. Thus, the asymptotic limits for the behaviour of z_{od} will be:

$$z_{od} \approx 0.1\frac{\nu^{w}}{u_*^{w}} \qquad \text{as } \mathrm{Re}_s^k \to 0 \qquad (23)$$

$$z_{od} \approx 0.1\frac{K}{u_*^{w}} \qquad \text{as } \mathrm{Re}_s^k \to \infty \qquad (24)$$

Of course, the unknown proportionality constants in the Eqs. 20 and 21 for K and for $a_{br} \approx \frac{u_*^2}{g}$ make the range of variability of the Reynolds roughness number still not precisely known (Kitaigorodskii 1997b).

Concluding remarks

In this paper, we have comparatively considered the problematic of the roughness concept above and below the sea surface. We have demonstrated that in describing the mean velocity distribution above and below the air-sea interface, the effective roughness parameters can strongly depend on the process of wind wave breaking and its characteristics. To explain the variability of the sea surface roughness as seen from above, the existence and width of the so-called dissipation subrange (Kitaigorodskii 1983, 1992b) must be taken into account.

For the description of roughness conditions below the sea surface, the introduction of an effective eddy viscosity typical for shear-free turbulence can be useful to explain the observed variability of the roughness parameter z_{od} under certain constrained conditions. However, the exact range of changes in the new Reynolds roughness number Re_s^k for roughness from below remains still unknown, thus yet not permitting to find out what are the prevailing aerodynamic conditions below the sea surface.

Acknowledgments: Comments by the reviewers were much appreciated. The financial support by the Academy of Finland to the author while visiting the Finnish Institute of Marine Research in Helsinki is gratefully acknowledged.

References

Banner M.L. 1990. Equilibrium spectra of wind waves. *J. Phys. Oceanog*. 20: 966–984.

Charnock H. 1955. Wind stress on a water surface. *Quart. J. Royal Meteorol. Soc*. 81: 639–640.

Donelan M.A. 1990. Air-sea Interaction. In: LeMehaute B. and D. Hanes (eds.), *The Sea: Ocean Engineering Science*, Vol. 9. Publ. J. Wiley & Sons, Inc., pp. 239–292.

Donelan M.A., Dobson F.W., Smith S D. & Anderson, R. L. 1993. On the dependence of sea surface roughness on wave development. *J. Phys. Oceanogr*. 23: 2143–2149.

Drennan W.M., Kahma K.K., Terray E.A., Donelan M.A. & Kitaigorodskii S.A. 1991. Observations of the enhancement of kinetic energy dissipation beneath breaking — wind wave. *Proceedings of IUTAM Breaking Waves Symposium Sydney, July 15–19*, pp. 95–101.

Ebuchi N., Toba Y. & Kawamura H. 1990. On the wave dependence of sea surface wind stress. A study by using data from an ocean data buoy station. *J. Oceanogr. Soc. Japan* 46: 177–183.

Geernaert G.L., Katsaros K.B. & Richter K. 1986. Variation of the drag coefficient and its dependence on sea state. *J. Geophys. Res*. 92: 13 127–13 139.

Hansen C., Katsaros K.B., Kitaigorodskii S.A. & Larsen S.E. 1990: The dissipation range of wind wave spectra observed on a lake. *J. Phys. Oceanogr*. 20: 1264–1277.

Jahne B. & Riemer K.S. 1990. Two dimensional wavemember spectra of small scale water surface waves. *J. Geophys. Res*. 95 (C7): 11531–11546.

Kitaigorodskii S.A. & Volkov Y.A. 1965. On the roughness parameter of the sea surface and the calculation of momentum flux in the near water layer of the atmosphere. *Izv. Acad. Sci. USSR, Atmospheric and Oceanic Physics* 1: 973–988.

Kitaigorodskii, S.A. 1968: On the calculation of the aerodynamic roughness of the sea surface. *Izv. Acad. Sci. USSR, Atmospheric and Oceanic Physics* 4: 498–502.

Kitaigorodskii S.A. 1973. *The physics of air-sea interaction*. Israel Program for Scientific Translations, Jerusalem. 237 pp (Available as TT 72-50062 from U.S. National Technical Information Service, Springfield, VA 22151, USA).

Kitaigorodskii S.A. & Mälkki P. 1979. Note on the parametrization of turbulent gas transfer across an air-water interface. *Finnish Marine Research* 246: 111–124.
Kitaigorodskii S.A. 1983. On the theory of the equilibrium range in the spectrum of wind generated gravity waves. *J. Phys. Oceanogr.* 13: 816–827.
Kitaigorodskii S.A. 1984. On the fluid dynamical theory of turbulent gas transfer across on air-sea interface in the presence of breaking wind waves. *J. Phys. Oceanogr.* 14: 960–972.
Kitaigorodskii S.A. 1986. The equilibrium ranges in wind-wave spectra (Physical Arguments and Experimental evidence for and against their existence). In: Phillips O.M. & Hasselman K. (eds.), *Wave dynamics and radio probing of the ocean surface*. Plenum Publishing Corporation, pp. 9–40.
Kitaigorodskii S.A. 1992a. The dissipation subrange of wind wave spectra. In: Banner M.L. & Grimshaw R.H.J. (eds.), *Breaking Waves IUTAM Symposium,* Sydney, Australia, July 15–19, 1991. Springer-Verlag, Berlin, Heidelberg, pp. 199–204.
Kitaigorodskii S.A. 1992b. The dissipation subrange in wind wave spectra. *Matematisk-fysiske Meddelelser* Det Kongelige Danske Videnskabernes Selskab. The Royal Danish Academy of Sciences and Letters, Munksgaard, Copenhagen, 42(5). pp. 3–24.
Kitaigorodskii S.A. 1994. A note on the influence of breaking wind waves on the aerodynamics roughness of the sea surface as seen from below. *Tellus* 46A: 681–685.
Kitaigorodskii S.A., Volkov Yu.A. & Grachev A.A. 1995. A note on the analogy between momentum transfer across a rough solid surface and the air-sea interface. *Boundary-Layer Meteorol.* 76: 181–197.
Kitaigorodskii S.A., 1997a. The influence of wind wave breaking on the local atmospher-ocean interaction. *Izv. Russian Acad. Sci., Physics of Atmosphere and Ocean* 33: 1–9.
Kitaigorodskii S.A. 1997b. The influence of breaking wind waves on the aerodynamic roughness of the sea surface as seen from below. Proceedings of Symposium on the air-sea interface, radio and acoustic sensing, turbulence and wave dynamics, Marseille (F), June 24–30, 1993, pp. 177–186.
Kreiman R.D. & Karlin L.N. 1979. Laboratory modelling of thermal structure of upper ocean. *Oceanology* 19: 49–52.
Long R.R. 1978. Theory of turbulence in a homogeneous fluid induced by an oscillating grid. *Phys. Fluid* 21: 1887–1888.
Nikuradse I. 1933. Strömungsgesetze in rauhen Röhten. *VDI-Forschungscheft* No. 361.
Terray E.A. & Bliven L.B. 1985. The vertical structure of turbulence beneath gently breaking wind waves. In: Toba Y. & Mitsuyasu H. (eds.), *The ocean surface*. D. Reidel Publishing Company, pp. 395–400.
Toba Y., Iida N., Kawamura H., Ebuchi N. & Jones S.F. 1990. Wave dependence of sea surface wind stress. *J. Phys. Oceanogr.* 20: 705–721.
Toba Y. & Ebuchi N. 1991. Sea surface roughness length fluctuating in concert with wind and waves, *J. Oceanogr. Soc. Japan* 47: 63–79.
Wu J. 1975. Wind-induced drift currents. *J. Fluid Mech.*. 65: 49–70.
Zakharoff V.E. & Zaslavskii M.M. 1982. The kinetic equation and Kolmogoroff spectra in the weak turbulence theory of wind waves. *Izv. Acad. Sci. USSR, Atmos. Oceanic Phys.* 18: 745–753.
Zakharoff V.E. & Zaslavskii M.M. 1983. Shape of the spectrum of energy carrying components of a water surface in weak turbulence theory of wind waves. *Izv. Acad. Sci. USSR, Atmos. Oceanic. Phys.* 19: 207–212.
Zakharoff V.E. 1992. Inverse and direct cascade in the wind driven surface wave turbulence and wave breaking. In: Banner M.L. & Grimshaw R.H.J. (eds.), *Breaking waves IUTAM Symposium*, Sydney, Australia, July 15–19, 1991. Springer Verlag, Berlin Heidelberg, pp. 70–91.
Zilitinkevich S.S. (ed.) 1991. *Modelling air-lake interaction. Physical Background*, Springer Verlag, Berlin, 129 pp.
Zilitinkevich S.S., & Kreiman K. D. 1991. Wind induced drift of surface films. In: Zilitinkevich S. (ed.), *Modelling air-lake interaction, Physical background*, Springer-Verlag. pp. 63–73.

Received 4 March 1997, accepted 30 December 1997