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An experimental investigation of the role of atmospheric stability in wind wave growth

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Abstract

Differences in temperature between air and water influence the stability of the atmospheric boundary layer. The altered structure of the boundary can affect the rate of growth of surface gravity waves. An extensive data set, collected under a wide range of well-documented atmospheric stability conditions, is investigated to quantify this effect. It is found that in unstable conditions, wind wave growth is enhanced, whereas in stable conditions, it is reduced. A correction, which can be applied to common fetch-limited growth curves, is developed which can account for atmospheric stability. In extreme cases, the error (in terms of energy) introduced by the neglect of stability effects can be as much as 50%.

Keywords: Waves; Finite depth; Stability; Boundary layer

1. Introduction

The relationship among wind speed, fetch, duration, water depth and the resulting wave energy seems qualitatively obvious. Numerous experiments have shown that the total wave energy increases with increasing wind speed, fetch, duration and water depth. Quantitative relationships are, however, difficult to define accurately, even for apparently simple and well-defined cases such as fetch-limited growth. Data invariably show significant scatter within their own data set, as well as significant differences in magnitude between the data sets.

Such inconsistencies possibly indicate that other parameters are also important in determining wind wave evolution. One such parameter, which has been considered in

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the past, is the detailed structure of the atmospheric boundary layer. The structure (shape and turbulence) of the atmospheric boundary layer is affected by the air–water temperature difference. Previous data have indicated that the air–water temperature difference, as a measure of the atmospheric stability, appears to be correlated with the growth rate of wind waves. Data, which quantify this effect over a range of stability conditions, have, however, been unavailable. This paper presents an analysis of the well-known Lake George (Young and Verhagen, 1996) data set aimed at quantifying the influence of atmospheric stability on the magnitude of the observed wind wave growth.

The arrangement of the paper is as follows. In Section 2, a review of existing knowledge on the structure of the atmospheric boundary layer and the affect of stability on this structure is presented. Section 3 critically summarizes previous wind wave growth experiments and the potential influence of stability on the reported findings. The present Lake George data set and the instrumentation used to record the data are described in Section 4, followed by a quantitative assessment of the influence on the observed evolution in Section 5. Section 5 also contains a comparison of the magnitude of the effect reported here with other data sets. Finally, Section 6 contains conclusions of this study.

2. The atmospheric boundary layer

Under neutral conditions (i.e., no vertical density gradient), air flow over the ocean is reasonably well-approximated by the logarithmic profile often used to represent flow over a rough surface:

$$\frac{u(z)}{u_*} = \frac{1}{\kappa} \ln\left(\frac{z}{z_0}\right), \tag{1}$$

where u is the wind speed measured at a height, z, above the surface, z_0 is a surface roughness length, u_* is the friction velocity and $\kappa \approx 0.4$ is the von Kármán constant.

Should the air and water temperatures differ, there will be a temperature gradient in the vertical, and hence, the air density will vary with height. The resulting stratification of the atmospheric boundary layer may be important because density differences, in conjunction with gravity, will affect the magnitude of turbulent mixing (e.g., Geernaert and Plant, 1990). Measurements over the ocean by Large and Pond (1982) have shown that the effect can be represented in terms of the Obukhov scale height (Arya, 1988):

$$L = -\frac{u_*^3 T_v}{\kappa g \langle w' T_v' \rangle}, \tag{2}$$

which defines the height at which the production of turbulence by shear equals the production by buoyancy. In Eq. (2), T_v is the mean virtual temperature at a height of 10 m, g is the gravitational acceleration, w is the vertical component of velocity and primed quantities denote fluctuations.

When the water is warmer than the air, there is an upward flux of heat, an atmospheric density gradient with denser air overlying less dense air and the value of L

is negative. Such cases are described as unstable. Conversely, when the water is colder than the air, L is positive and conditions are described as stable. Neutral stratification occurs for large $|L|$. An alternative indicator of stability is the Bulk Richardson Number (Kahma and Calkoen, 1992)

$$R_{\mathrm{b}} = \frac{g(T_{\mathrm{a}} - T_{\mathrm{w}})}{z_{\mathrm{t}} T_{\mathrm{a}} (u/z)^2}, \tag{3}$$

where T_{a} and T_{w} are the air and water temperatures (K) respectively, z_{t} is the height at which the temperature is measured, and z is the height at which the wind speed is measured. Again negative values of R_{b} represent unstable conditions.

The vertical heat flux and associated buoyancy modify the logarithmic profile (Eq. (1)) (e.g., Priestly, 1959; Lumley and Panofsky, 1964; Webb, 1970) yielding:

$$\frac{u(z)}{u_*} = \frac{1}{\kappa}\left[\ln\left(\frac{z}{z_0}\right) - \psi(z/L)\right], \tag{4}$$

where ψ is called the integrated universal function and is based on the Monin and Obukhov (1954) formalism (Monin and Yaglom, 1971). It has been determined by Webb (1970), Dyer and Hicks (1970) and Businger et al. (1971). For stable conditions ($L > 0$), ψ becomes:

$$\psi = -\frac{5z}{L}, \tag{5}$$

and for unstable conditions ($L < 0$):

$$\psi = 2\ln\left(\frac{1+x}{2}\right) + \ln\left(\frac{1+x^2}{2}\right) - 2\tan^{-1}x + \pi/2, \tag{6}$$

where $x = (1 - 16z/L)^{1/4}$.

The effect of atmospheric stability on the mean velocity profile is shown in Fig. 1. This figure shows the mean velocity profile normalized in terms of the wind speed measured at a reference height of 10 m (U_{10}) for values of L between -100 m and $+100$ m. As the magnitude of the stability effect increases (i.e., $|L|$ decreases), the profiles progressively deviate from the neutral logarithmic profile predicted by Eq. (1). The enhanced vertical mixing, which occurs for unstable conditions, results in a more uniform profile with height. Thus, in unstable conditions, higher velocities are expected closer to the ground than in stable conditions.

The shape of the boundary layer has a number of potential effects on wind wave evolution. Debate still occurs as to the most appropriate measure of wind speed to be used in wind wave studies. The wind speed at a reference height of 10 m, U_{10}, the friction velocity, u_*, and the wind speed at a height of half the ocean wave length, $U_{\lambda/2}$, have all been proposed. The example shown in Fig. 1 illustrates the influence on each of these quantities. In all cases shown, the value of U_{10} will be the same. The shear stress, and hence, u_*, are related to the curvature of the boundary layer. Due to the higher values of mean wind speed closer to the water surface which result in unstable conditions, the gradient of wind velocity in such cases must be larger than in stable

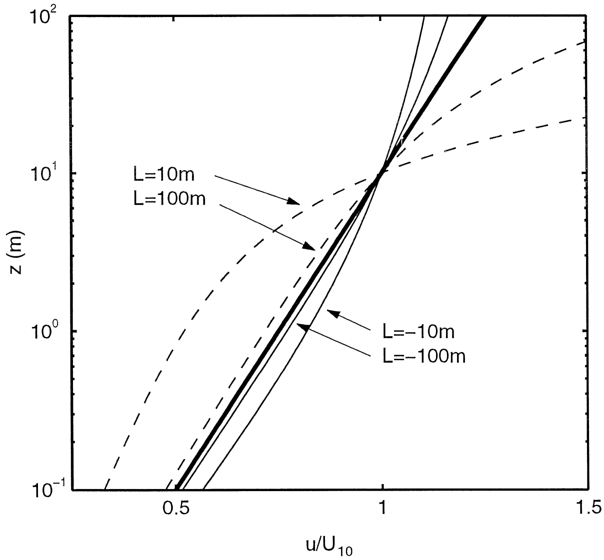

Fig. 1. Shape of the atmospheric boundary layer for different values of the Obukhov scale height, L, as calculated from Eq. (4). Stable conditions are represented by positive values of L, and unstable conditions by negative values of L. The thick, solid line represents the logarithmic relationship (Eq. (1)) which occurs for neutral conditions.

conditions. Hence, the surface shear stress and u_* will be larger in unstable than in stable conditions. The relative values of $U_{\lambda/2}$ will depend on the value of λ under consideration.

In addition to the influence that stability has on the value of the scaling wind speed, the atmospheric turbulence is also affected. Monahan and Armendariz (1971), Sethuraman (1979) and Cavaleri and Cardone (1994) have all shown enhanced turbulent velocity fluctuations in unstable conditions. These are to be expected as the upwards heat flux implies turbulence with strong vertical mixing, as already characterized by the larger drag at the surface. In a series of numerical experiments, Voorrips et al. (1994) have shown that an increase in the turbulent intensity results in enhanced wave growth.

3. Observations of wind wave growth

The two most commonly referenced deep water fetch-limited data sets are those of JONSWAP (Hasselmann et al., 1973) and Lake Ontario (Donelan et al., 1985). Adopting the Kitaigorodskii (1962) similarity relationships, such studies typically consider the growth of the nondimensional energy, $\varepsilon = g^2E/U_{10}^4$, as a function of nondimensional fetch, $\chi = gx/U_{10}^2$, where E is the total energy of the wave field and x is the fetch. Alternatively, the nondimensional variables may be formed in terms of the friction velocity, u_*: $\varepsilon * = g^2E/u_*^4$ and $\chi * = gx/u_*^2$. Kahma and Calkoen (1992) examined data from the above experiments, together with data from the Bothnian Sea (Kahma, 1981). They concluded that all data sets show more rapid growth in unstable than in stable conditions. The effects of stability were only partly removed when u_* scaling was used in preference to U_{10}. As u_* is typically not directly measured in such experiments, it must be calculated indirectly, thus introducing another source of potential error. They finally summarized the composite data set as:

$$\varepsilon = \begin{cases} 9.3 \times 10^{-7} \chi^{0.76} & \text{Stable stratification} \\ 5.4 \times 10^{-7} \chi^{0.94} & \text{Unstable stratification} \end{cases} \tag{7}$$

Kahma and Calkoen (1992) suggested that it would be advantageous to investigate ε as a function of χ and a measure of the stability (they suggested z/L). The available data were, however, too sparse for this purpose. Instead, they assumed that the transition was confined to a relatively small range of values of z/L near neutral conditions and proposed an empirical relation which transitioned between the two branches of Eq. (7).

Fetch-limited wind wave growth experiments in finite-depth conditions are extremely rare, the only significant data set being from Lake George (Young and Verhagen, 1996). To account for the effects of finite depth, Young and Verhagen (1996) introduced the nondimensional depth $\delta = gd/U_{10}^2$ where d is the water depth, in addition to ε and χ. No account of stability effects were made in the analysis which yielded the result:

$$\varepsilon = 3.64 \times 10^{-3} \left\{ \tanh A_1 \tanh\left[\frac{B_1}{\tanh A_1} \right] \right\}^{1.74}, \tag{8}$$

where

$$A_1 = 0.493 \delta^{0.75}, \tag{9}$$

$$B_1 = 3.13 \times 10^{-3} \chi^{0.57}. \tag{10}$$

4. The present data set

The Lake George experiment has been described in detail by Young and Verhagen (1996) and Young et al. (1997). The experiment consisted of the measurements of water surface elevation and wind speed at a total of eight stations in a shallow lake of average water depth of 2 m. The stations were at fetches ranging between 1.3 km and 15.7 km. In addition, one station (Station 6) also measured air temperature and humidity using a Rotronic YA-100 probe, and water temperature using a platinum resistance thermometer. The data analysis reported by Young and Verhagen (1996), from which Eq. (8) was developed, divided the data into two groups: the North–South data set and the East–West data set. The array of eight measurement stations was aligned North–South, and hence, the North–South data set yielded a wide range of nondimensional fetches, χ. The East–West data set was confined to a smaller range of χ, however, the amount of data was significantly larger than for the North–South data set. The measurements of air and water temperatures were not maintained for the full period of the experiment, significant data only being available for the East–West data set. Hence, the present analysis has been confined to this data set.

Data were limited to cases where the wind speed and direction were relatively constant and the wind direction was approximately normal to the shoreline [see Young

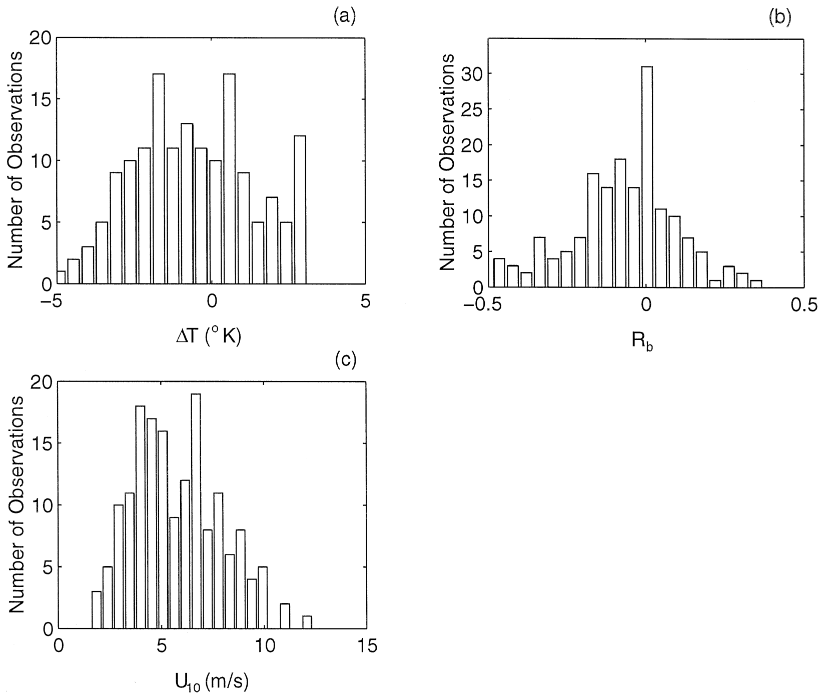

Fig. 2. Distribution of values within the Lake George data set. (a) $\Delta T = T_a - T_w$. (b) Bulk Richardson number, R_b. (c) Wind speed, U_{10}.

and Verhagen (1996) for details]. The resulting data set consists of 184 observations. The distribution of quantities within this data set is shown in Fig. 2. This figure shows that the data set contains wind speeds between 3 m/s and 12 m/s, and air–water temperature differences, $\Delta T = T_a - T_w$, between -5°C and $+3$°C. As mentioned earlier, the Bulk Richardson number, R_b, is a better measure of atmospheric stability than ΔT. Values of R_b range between -0.5 and $+0.4$, indicating a wide range of both stable and unstable conditions. There are, however, more data for neutral conditions than one would assume by simply considering values of ΔT. The range of values of R_b represented in Fig. 2 is significantly larger than reported in previous experiments. As an example, the Lake Ontario data of Donelan et al. (1985) only contain data with R_b in the range -0.15 to $+0.22$.

5. The influence of stability

In deep-water applications, the effects of stability are traditionally investigated by plotting nondimensional growth curves (ε vs. χ), with the data partitioned according to stability (Kahma and Calkoen, 1992). The present data set has been collected in finite depth water, and hence, an additional nondimensional parameter in δ is also important. Rather than a single growth curve existing, as in deep water, a family of curves (one for each value of δ) is required to describe the data (Young and Verhagen, 1996). A simple partitioning of the data, based on stability within this three-dimensional parameter space, would require a data set far more comprehensive than that available.

In order to assess the effects of stability on the present data set, the deviation of the data from the result predicted by Eq. (8) has been investigated. This deviation, ξ, is defined as:

$$\xi = \frac{\varepsilon - \varepsilon_m}{\varepsilon_m}, \tag{11}$$

where ε is the nondimensional energy calculated for each of the data, and ε_m is the corresponding value of nondimensional energy predicted by Eq. (8) for these same data, using the measured values of χ and δ. Hence, Eq. (11) is a measure of the deviation of the measured nondimensional energy from the mean result.

Fig. 3 shows ξ as a function of R_b. Positive values of ξ indicate growth rates larger than the mean, and negative values of χ indicate growth rates smaller than the mean. Although there is significant scatter in the data, a clear trend showing enhanced growth for unstable conditions and reduced growth for stable conditions is apparent. In contrast to the findings of Kahma and Calkoen (1992), this effect is not confined to a small region near neutral stability. The deviation from the mean result continues to increase in magnitude as the magnitude of R_b increases.

Although there is significant scatter in the data, a linear regression yields:

$$\xi = -1.22 R_b + 0.01. \tag{12}$$

Also shown on Fig. 3 are 95% confidence limits on the linear regression (Eq. (12)). Although there is significant scatter within the data, these confidence limits indicate that Eq. (12) is a reasonable representation of the observable trend. The results indicate that

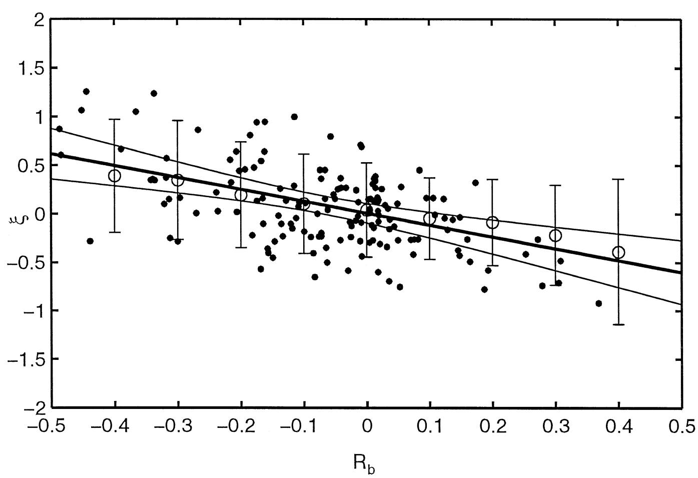

Fig. 3. Correction to the mean energy, ξ, as a function of Bulk Richardson number, R_b. The solid, thick line represents the linear regression to the data of Eq. (12). Also shown are the 95% confidence limits for Eq. (12). The open circles represent mean values of ξ over a bin of extent ± 0.2 in R_b space. The error bars represent ± 1 S.D. for the data in each bin.

significant deviations from the mean result can occur due to the effects of atmospheric stability. As an example, for a relatively large value of $|R_b| = 0.5$, the energy deviates from the mean by approximately 50%.

Fig. 3 also presents a running mean in which data have been allocated to bins in R_b space of size ± 0.2. The mean value of ξ for each bin, together with ± 1 S.D. about this mean, are shown for each bin. The same general trends, as observed for the regression analysis, are again evident.

It should be noted that R_b is a function of the heights at which the wind (z) and temperature (z_t) are measured (see Eq. (3)). In this case, the values, $z = 10$ m and $z_t = 2$ m, have been used. Kahma and Calkoen (1992) used R_b to solve for the Obukhov scale length, L, and then adopted this as a measure of stability. Such an analysis requires knowledge of the surface roughness. No direct measurements of the surface roughness were available and, in addition, the dependence of the surface roughness on wind and wave conditions in finite depth water is unknown. Therefore, R_b, which can be determined directly from measured quantities, was adopted as the most appropriate available measure of stability.

In the above analysis, it has been implicitly assumed that all of the observed deviation from the mean growth rate, as defined by Eq. (11), can be attributed to the effects of stability. Other influences could also contribute to this effect. The deviation could, for example, also be a function of the nondimensional fetch, χ and the nondimensional water depth, δ, if Eq. (8) contains any systematic bias. Eq. (8) was, however, derived from data obtained from this same site, and at values of χ and δ which span the present data. Also, the conditions under which the data used to develop Eq. (8) were collected, were carefully selected with only small variations in wind speed and direction being tolerated (Young and Verhagen, 1996). Therefore, it is believed that any spurious correlation with χ or δ will be small. Nevertheless, the scatter apparent in Fig. 3 is significant, and hence, Eq. (12) can only be considered as a low-order correction to the mean result. Other processes may also contribute to the apparent scatter.

It is interesting to compare the magnitude of the effect determined from the present data with previous studies. Resio and Vincent (1977) have investigated the influence of stability on estimates of wind speed. They proposed a modification factor to correct wind speed estimates which was presented in graphical form as a function of ΔT. This factor is approximately proportional to $\sqrt{|\Delta T|}$. Most deep water fetch-limited experiments (e.g., Hasselmann et al., 1973; Donelan et al., 1985) find that ε is linearly dependent on χ. In addition, for deep water conditions (δ large, χ small), Eq. (8) reduces to a linear dependence between ε and χ. Such a linear relationship implies that the energy, E, is proportional to U_{10}^2. Hence, the correction for wind speed proposed by Resio and Vincent (1977) translates to a linear dependence between ξ and ΔT. This is consistent with the form proposed by Eq. (12) from the present data.

A quantitative comparison between the results of Resio and Vincent (1977) and Eq. (12) requires the selection of specific examples. For comparative purposes, consider: $T_a = 20°C$, $z_t = 2$ m, $z = 10$ m and $U_{10} = 10$ m/s, values representative of the present data set. The corrections to E for a range of values of ΔT are shown in Table 1. The

Table 1
Comparison between the results of the predictions of Resio and Vincent (1977) and Eq. (12)

[1] ΔT (°C)	[2] $(U+\Delta U)/U$ (Resio and Vincent, 1977)	[3] $(E+\Delta E)/E$ ($\propto$ [2]^2)	[4] R_b	[5] $(E+\Delta E)/E$ Eq. (12)
5	1.11	1.23	0.08	1.10
10	1.15	1.32	0.17	1.21
15	1.19	1.42	0.25	1.31
20	1.22	1.49	0.33	1.40

Results are calculated for a case with: $T_a = 20$°C, $z_t = 2$ m, $z = 10$ m and $U_{10} = 10$ m/s. Results are considered for a variety of different air–water temperature differences, ΔT (Column 1). The wind speed corrections of Resio and Vincent (1977) are shown in Column 2. The resulting influence on energy, assuming this is proportional to the wind speed squared, is shown in Column 3. The Bulk Richardson number, (Eq. (3)) is shown in Column 4. The influence on energy predicted by Eq. (12) is shown in Column 5. The values in Column 5 can be compared with Column 3.

comparison between Resio and Vincent (1977) and Eq. (12) is surprisingly good, with the results of the two approaches being within 10% of each other.

Kahma and Calkoen (1992) partitioned the data of Donelan et al. (1985) into stable and unstable domains. Linear form growth curves, which represented these two data groups, were: $\varepsilon = 2.8 \times 10^{-7}\chi$ (stable data) and $\varepsilon = 3.8 \times 10^{-7}\chi$ (unstable data). The ratio of these two relationships indicates growth rates 1.36 times greater in unstable conditions than in stable conditions. As stated earlier, the Donelan et al. (1985) data were in the range $-0.15 < R_b < 0.22$. Hence, mean values of R_b for the unstable and stable data of -0.08 and 0.11, respectively, can be assumed. The use of such values in Eq. (12) yields a difference between the growth rates of a factor of 1.22. Again, the difference between the two data sets is of order 10%.

6. Conclusions

In contrast to previous data sets, the present data contain a very wide range of atmospheric stability conditions. Bulk Richardson numbers, R_b, range between -0.5 and $+0.4$. As a result, it has been possible to quantify the effect of atmospheric stability on wind wave growth. The results clearly show enhanced growth in unstable conditions and reduced growth in stable conditions. The effects appear to be approximately linearly related to R_b, and a simple linear regression analysis has been used to develop a correction factor for atmospheric stability. This result can be applied to both deep water and finite depth cases. Although the data exhibit scatter, the linear correction appears to be a reasonable first-order correction for stability effects.

The data clearly show that at large values of $|R_b|$, the influence of atmospheric stability on the resulting growth rate can be very significant. For $|R_b| \approx 0.5$, neglect of the effects of stability could result in wave energy estimates in error of 50%. Such extreme values of R_b would generally only exist over small bodies of water such as lakes. In the ocean, very large air–water temperature differences generally do not occur, and hence, the effects of atmospheric stability would be less pronounced. Effects as large as 10% to 20% could, however, still occur in such cases, particularly in coastal regions.

The present result expands previous investigations of the influence of atmospheric stability, providing an easily calculated correction for such effects. The present correction is consistent with both the present data and these previous studies.

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