Note 7

Wind in wave modelling

1. Introduction

Accurate knowledge of the wind speed and wind direction in the atmospheric boundary layer (ABL) above the ocean surface is important for the prediction of wind generated waves and for the analysis of wave and wind measurements. In this note, two kinds of wind data are considered: those that are generated by numerical models and those that are obtained by measurements. When numerical methods are used, a correct modelling of the ABL is important. When wind data from measurements are used, one has to be careful with the processing and interpretation because many disturbing effects should be accounted for, such as the variation of wind speed and wind direction with height, measurement errors due to flow distortions near the wind anemometer, and rapid fluctuations of the wind vector.

Since wind speed and wind direction vary with height, it is essential to state these data in combination with the height at which they are measured or to transform them to some standard height. If this is not done, interpretation errors are easily made when comparing wind data obtained at different stations having different anemometer heights. A standard height of 10 m is recommended by the World Meteorological Organization (Dobson, 1983). This, however, does not mean that all anemometers should be placed at the standard height of 10 m; good exposure to the wind to avoid flow distortions and good maintenance prevail.

In recent years most wind measurements at sea are made at the top of offshore structures with typical anemometer heights of 100 m. Near such structures the effect of flow distortions on the wind measurements can not always be neglected. Normally, in the direct vicinity of the basic platform structure, undisturbed wind measurements are not possible. At the top of the tower, however, the platform structure may cause only minor deviations. Before such wind measurements are used it is desirable that the magnitude of the flow distortions is determined by model investigation (e.g. Vermeulen et al., 1985).

Since the air flow in the boundary layer is turbulent, smoothing out the effect of short term statistical variability of the wind vector is required. Therefore, averaging techniques have to be used in the processing of the wind data. An averaging time of 10 minutes is recommended by the World Meteorological Organization (Dobson, 1983). The use of an averaging time of less than 10 minutes produces scattered results whereas averaging times of more than 20 minutes are likely to destroy information on intermediate scale motions.

For the conversion of the measured wind speed and direction to some standard height it is necessary to know which parameters play a role in the description of the atmospheric boundary layer. A related problem is the choice between the wind speed at a certain height and the friction velocity, as a parameter to scale wave growth.

In order to clarify the above mentioned points, a short description is given of the ABL as far as it is relevant for wave modelling. Attention is given to the variation of wind speed with height and to the variation of wind direction with height.

2. The atmospheric boundary layer

The atmospheric boundary layer can be defined as that part of the atmosphere that is influenced by frictional effects due to the underlying land or sea surface. The ABL is normally divided in two parts. The first and lowest part of the ABL, is called the surface layer and is characterised by turbulent fluctuations of mean values of wind speed, temperature and other parameters. Since vertical fluxes of e.g. momentum and heat are found to be nearly constant with height, it is also usually referred to as the constant flux layer. Another feature of the surface layer is that the wind direction remains nearly constant with height and that the Coriolis force is normally neglected in the analysis of this layer. The surface layer has a height of about 100 m. In the second and upper part of the ABL, the wind speed and wind direction vary relatively slowly with height in comparison with the rate of variation in the surface layer. In general most of the wind veering with height occurs in the upper layer, while most of the reduction of the wind speed occurs in the surface layer.

In the upper layer three forces are of significance, the Coriolis force, the pressure gradient force and a turbulent friction force. The principal mechanism that is responsible for the wind veering is described in section 4. The upper layer is sometimes referred to as the Ekman layer.

A measure for the turbulent momentum transport in the surface layer is the friction velocity u_*, which is defined as the square root of the ratio of the surface stress (τ) caused by the wind and the density of air (ρ_a):

$$u_* = \sqrt{\frac{\tau}{\rho_a}} \tag{2.1}$$

Knowledge of u_* is important for wave modelling, since u_* is assumed to be directly related to the downward flux of horizontal momentum from the surface layer to the wave field. Based on this consideration preference for scaling wave growth with friction velocity rather than wind speed at a certain height is argued by various authors (e.g. Miles, 1959, Komen et al. ,1984). Empirical support favouring the scaling with u_* is given in Janssen et al. (1987).

3. Variation of wind speed with height

The most common method to compute the variation of wind speed with height assumes a logarithmic velocity profile in the surface layer, given by:

$$U_z = \frac{u_*}{\kappa} \ln\left(\frac{z}{z_0}\right) \tag{3.1}$$

in which U_z is the wind speed at height z and κ is the Von Karman constant (κ » 0.41), and z_0 is a roughness length.

The logarithmic velocity profile is based on the assumption of a constant stress in the surface layer. In expression (3.1) the friction velocity and roughness length are scaling parameters that give characteristic velocity and length scales in the surface layer. For neutral atmospheric conditions the measured velocity profiles are almost in agreement with the logarithmic velocity profile (e.g. Charnock, 1955). For turbulent boundary layers the value of the roughness length z_0 depends on the physical characteristics of the surface. Normally the roughness length z_0 above land is related to the height and shape of characteristic elements on the surface. Typical values of the ratio of the roughness length z_0 to the height of these roughness elements vary from 1/30 to 1/5 (Holton, 1979). Above sea the situation is more complicated because the roughness elements, i.e. the waves, are moving. From observations it has become clear that the ratio of the surface roughness length to the root mean square wave height is much smaller than characteristic values of the corresponding ratio above land. Various studies showed that the roughness of the sea is almost entirely caused by the capillary waves (Munk, 1955).

Charnock (1955) argued that the roughness length should be a function of u_* and the gravitational acceleration g only. Consequently, on dimensional considerations he proposed the form:

$$z_0 = a\frac{u_*^2}{g} \tag{3.2}$$

The constant a was measured by Garratt (1977) who found a mean value of 0.0144. If the wind speed at a certain elevation has been measured, equations (3.1) and (3.2) can be solved by iteration yielding u_*, z_0 and the wind speed at any elevation within the surface layer.

In relating wind speeds at different heights using the logarithmic velocity profile use is made of the drag coefficient C_d. The drag coefficient $C_d(z)$ is defined as:

$$C_d\left(z\right)=\frac{\tau}{\rho_a U_z^2} \tag{3.3}$$

Normally, the drag coefficient C_d is used as a parameter relating the wind speed U_z and the friction velocity u . Combining the Eqs. (2.1) and (3.3) results in:

$$C_d\left(z\right)=\left(\frac{u_*}{U_z}\right)^2 \tag{3.4}$$

Combining the Eqs. (3.1) and (3.2) to eliminate z , followed by using (3.4) to eliminate u gives an expression for the dependence of the drag coefficient on wind speed and anemometer height, which can be solved by iteration (Wu, 1982):

$$\sqrt{C_d\left(z\right)}=\frac{\kappa}{\ln\left(\frac{zg}{aC_d\left(z\right)U_z^2}\right)} \tag{3.5}$$

The drag coefficient is not constant but varies with height z 1 and the wind speed U_z. For z =10 the relation (3.5), with a = 0.0144 and κ = 0.41, can be well approximated by the much more simple formula (Wu, 1982):

$$C_d\left(10\right)=\left(0.8+0.065U_{10}\right)\times 10^{-3} \tag{3.6}$$

From the Eqs. (3.4) and (3.6) it follows that the ratio u_*/U_{10} varies as much as a factor 1.41 if U_{10} varies from 5 to 25 m/s. This has important consequences for wave modelling (Holthuijsen, 1980; Janssen et al., 1987), which, however, is not discussed here.

The wind speed at a height of 10 m can be computed from the wind speed at height z by the following expression. Using Eq. (3.1) to eliminate z_0 gives:

$$U_{10}=U_z+\frac{u_*}{\kappa}\ln\left(\frac{10}{z}\right) \tag{3.7}$$

Using (3.4) to eliminate u_* gives (with $C_d\left(z\right)$ replaced by $C_d\left(10\right)$):

$$U_{10}=U_z\left(1+\frac{\sqrt{C_d\left(10\right)}}{\kappa}\ln\left(\frac{z}{10}\right)\right) \tag{3.8}$$

in which $C_d(10)$ is a function of U_{10} as given by (3.6). Equation (3.8) is solved by iteration. Equation (3.8) is often used in wave modelling studies.

4. Variation of wind direction with height

The variation of wind direction with height in the upper part of the ABL is easily explained in a qualitative way by means of a simple model. Above the ABL the air flow is geostrophic and determined by the pressure gradient force and the Coriolis force only. In the ABL the surface drag causes a reduction of the wind speed with decreasing height. With decreasing wind speed there is also a decrease of the Coriolis force. The combined effect of these reductions is a counter-clockwise rotation of the wind direction with decreasing height (Northern Hemisphere). This effect is often called the Ekman effect. Above open sea the magnitude of this effect can be as much as 15° (Brown and Liu, 1982). Typical values for the wind veering at sea over a height interval between 100 and 10 m can be found in Riissanen (1975) to be 5°. This value is supported by Wieringa (1987, personal communication).

References

Brown, R.A. and W.T. Liu, 1982: An operational large-scale marine planetary boundary layer model, *J. Appl. Meteor.*, 21, No. 3, 261-269.

Charnock, H., 1955: Wind stress on a water surface. *Quart. J. Roy. Meteor. Soc.*, 81, 639-640.

Dobson, F.W., 1983: Review of reference height for and averaging time of surface wind measurements at sea. WMO Marine Meteorology Rep. No. 3.

Garratt, J.R., 1977: Review of drag coefficients over oceans and continents, *Mon. Weather Rev.*, 105, No. 7, 915-929.

Holthuijsen, L.H, 1980: Methoden voor golfvoorspelling, Technische Advies Commissie voor de Waterkeringen, Staatsdrukkerij, 2 delen.

Holton, J.R., 1979: An introduction to dynamic meteorology, Sec. Ed., Academic Press.

Janssen, P.A.E.M., G.J. Komen and W.J.P. de Voogt, 1987: Friction velocity scaling in wind wave generation, *Bound.-Layer Meteor*., 38, 29-35.

Komen, G.J., S. Hasselmann and K. Hasselmann, 1984: On the existence of a fully developed wind-sea spectrum, *J. Phys. Oceanogr*., 14, No. 8, 1271-1285.

Miles, J.W., 1959: On the generation of surface waves by shear flows. Part 2. *J. Fluid Mech*., 6, No. 4, 568-582.

Munk, W.H., 1955: Wind stress on water: an hypothesis, *Quart. J. Roy. Meteo. Soc.*, Vol. 81, 320-332.

Riissanen, J., 1975: Some features of wind variation in the friction layer at Helsinki airport. Finnish Met. Inst. Contr., 80.

Vermeulen, P.E.J., B. Oemraw and J. Wieringa, 1985: Wind tunnel measurements of the flow distortion near the anemometer positions on PENNZOIL K-13A platform, Netherlands Organization for Applied and Scientific Research, Fluid Dynamics Department, Apeldoorn, the Netherlands, Rep. No. 85-01246.

Wu, J., 1982: Wind-stress coefficients over sea surface from breeze to hurricane, *J. Geophys. Res*., 87, No. C12, 9704-9706.

List of symbols

a 2 constant
$C_d(z)$ 3 drag coefficient at height z 4
g 5 gravitational acceleration
z 6 height
U_z 7 wind speed at height z 8
u_* 9 friction velocity
z_0 10 roughness length

κ 11 von Karman constant
ρ_a 12 density of air