The influence of humidity fluxes on offshore wind speed profiles

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ABSTRACT

This research focuses on quantifying the effect of humidity fluxes on stability corrected wind speed profiles. The effect on wind speed profiles is found to be important in only stable conditions where including humidity fluxes forces conditions towards neutral. Excluding humidity fluxes leads to over-estimation of the wind speed profile at 150 m by approximately 5%.

INTRODUCTION

Global wind capacity by the end of 2006 was 74 GW (www.gwec.net) and annual average growth rates for installed capacity were almost 30% with a growing role for offshore installations [1]. This move towards increased emphasis on harnessing of the wind resource offshore is also manifest at the national level. By 2006 Danish offshore wind farms had a capacity of over 400 MW, relative to a total national wind capacity of over 3000 MW. As described elsewhere this move provides both new challenges [2],[3] and opportunities [4],[5]. One of the challenges is that developments offshore coupled with the increase in turbine hub-heights means that wind energy is extending into a historically under-sampled region of the atmosphere – the marine boundary layer between 100-200 m above the surface. It is now accepted that there is a need to account for deviations of stability conditions from near-neutral in predicting vertical profiles of wind speed, we extend this to examine the role of partitioning of the surface energy flux between sensible heat and latent heat transfer. Here we specifically examine the role of humidity fluxes from the sea-surface caused by evaporation/ condensation of water vapour on the vertical profile of wind speeds offshore. Understanding the impact of humidity fluxes may have other applications such as a consideration of how wind speed profiles may change in areas which currently experience sea ice during winter but may become ice free under global climate change or the retrieval of wind speeds from satellite images.

STABILITY CORRECTIONS TO WIND SPEED PROFILES

There is increasing evidence that the constant flux layer assumption inherent in Monin-Obukhov similarity theory used to derive stability corrections for predicting wind speed profiles to current turbine hub-heights is inadequate [3, 6]. However, using a stability correction to the logarithmic profile improves predictions relative to observations [7, 8] and is used here in the absence of proven alternatives. The change of wind speed with height depends on surface roughness which varies with wind speed offshore according to [9]. This change in surface roughness has been found to have a negligible impact on the mean wind shear offshore due to the very low values for the roughness length, except where the initial prediction height is very close to the surface (e.g. less than 10 m as may be the case with buoy data) [10, 11].

The stability corrected or diabatic wind speed profile is defined by [12]:

$$U_z = \frac{u_*}{\kappa}\left[\ln\frac{z}{z_0} + \Psi_m\left(\frac{z}{L}\right)\right] \tag{1}$$

where U_z is the wind speed at height z.

u_* is the friction velocity which is related to generation of waves at the sea surface. ($u_*^2 = \frac{\tau}{\rho}$, where τ is the shear stress and ρ is the air density) and the dimensionless drag coefficient (C_D) ($u_*^2 = C_D U_{10}^2$, where U_{10} is the wind speed at 10 m above the surface).

κ is the von-Karman constant.

z_0 is the roughness length.

$\Psi_m(z/L)$ is the stability function for momentum that describes the deviation from the neutral wind speed profile [12] and depends on the stability index z/L where z is the height above the surface and L is the Monin-Obukhov length.

The Monin-Obukhov length (*L*) is a metric of atmospheric stability and is approximately the height at which buoyancy starts to dominate over mechanically generated turbulence [12]:

$$L = \frac{-u_*^3}{(\kappa(g/\overline{\theta_v})(\overline{w'\theta_v'}))} \tag{2}$$

Where the over-bar indicates a time average.

g is acceleration due to gravity.
$w'\theta_v'$ is the virtual kinematic heat flux.
θ_v is the virtual potential temperature.

Potential temperature is used to correct temperature to a standard reference pressure (e.g. 1000 mb) making the assumption that the air parcel is unsaturated and assuming adiabatic (close to near-neutral) conditions:

$$\theta = T\left(\frac{1000}{P}\right)^{R/cp} \tag{3}$$

where T is temperature.
P is the atmospheric pressure.
R is the universal gas constant.
c_p is the specific heat capacity of air.

Virtual temperature is used to approximate the temperature that dry air would have assuming the same pressure and density as a parcel of moist air:

$$T_v = T(1 + 0.61q) \tag{4}$$

where q is the specific humidity.

The stability function Ψ_m (z/L) in equation (1) in stable conditions, where L is positive, is given as [12]:

$$\Psi_m = 4.7\frac{z}{L} \tag{5}$$

And L is positive giving a positive correction to wind speed profiles (see equation 1). For unstable conditions L is negative and the correction is:

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

Where $x = \left(1 - 15\frac{z}{L}\right)^{1/4}$

As shown in (2) the virtual kinematic heat flux is related to the combined effect of sensible and latent heat flux that derives from transfer of sensible and latent heat which may have different boundary conditions offshore because, at the top of the mixed layer, dry air is entrained into the boundary layer and the surface is a source of humidity whereas the air-sea temperature gradient which drives the sensible heat flux at the bottom of the boundary-layer maybe either positive or negative. Over land the role of the latent heat flux is generally much smaller because the surface does not act as an infinite source of water.

Inserting equation (3) in (2) it can be seen that:

$$\frac{z}{L} = -\frac{g\kappa z}{u_*^3\theta}\overline{w'\theta'} - 0.61\frac{g\kappa z\theta}{u_*^3\theta_v}\overline{w'q'} \tag{7}$$

By defining

$$\frac{z}{L_T} = -\frac{g\kappa z}{u_*^3\theta_v}w'\theta' \tag{8}$$

and

$$\frac{z}{L_q} = -0.61\frac{g\kappa z\theta}{u_*^3\theta_v}w'q' \tag{9}$$

then

$$\frac{z}{L} = \frac{z}{L_T} + \frac{z}{L_q} \tag{10}$$

where z/L_T accounts for sensible heat fluxes and z/L_q account for humidity fluxes.

Considering that, in the marine boundary layer, humidity fluxes are always positive due to saturated conditions at the surface as a consequence when estimating the Monin Obhukov stability parameter, z/L in equation (10) we have the following:

Unstable conditions: both sensible heat and humidity fluxes are positive (upwards)

from (7) z/L (without $\overline{w'q'}$) < z/L (with $\overline{w'q'}$)

therefore adding humidity fluxes results in more unstable conditions

Stable conditions: sensible fluxes are negative and humidity fluxes are positive

from (7) z/L (without $\overline{w'q'}$) > z/L (with $\overline{w'q'}$)

therefore adding humidity fluxes results in more neutral conditions.

Figure 1 illustrates this for a case where z/L_q is set to -0.2 and z/L_T varies between -5 and 5. In unstable conditions, inclusion of humidity fluxes increases z/L indicating increased instability. This is also intuitive – both sensible heat and latent heat fluxes are upwards, thus the lower atmosphere is made increasingly unstable. In stable conditions, z/L_T tends towards zero but is negative, while z/L_q remains negative and thus the fluxes are in opposing directions and drive conditions towards near-neutral.

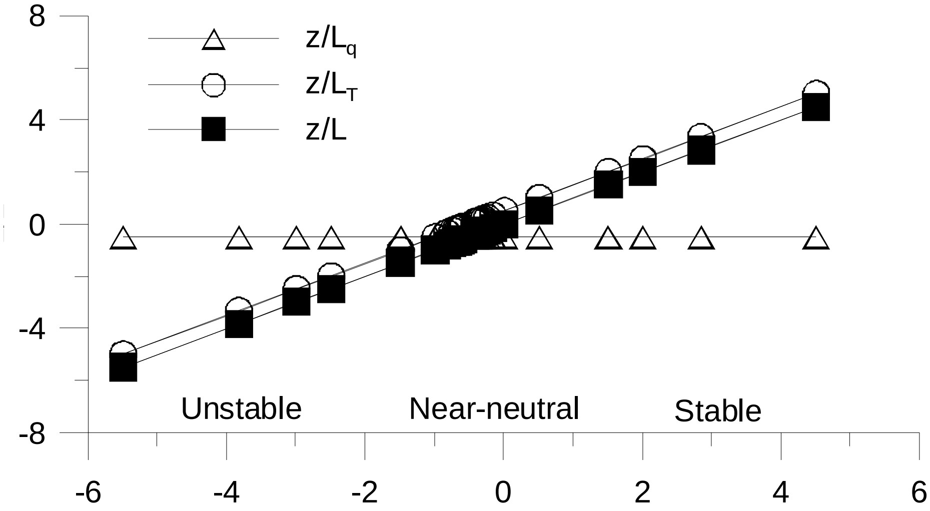

Figure 1. Using equation (7) to determine the influence of z/L_q (latent heat flux contribution) and z/L_T (sensible heat flux contribution) on the total value of z/L.

EXAMPLES OF THE EFFECT ON VERTICAL WIND SPEED PROFILES FROM DATA ANALYSIS

Humidity fluxes are notoriously difficult to measure [13] and bulk formulations are frequently used to obtain fluxes [14], [15] although errors are relatively large in comparison with observations particularly in stable conditions [16], [17]. The data presented here are the results of a measurement campaign on the Danish island of Anholt from September 1990 to October 1992. Details of the measurement campaign are given in [18]. A 22 m meteorological mast was equipped with standard meteorological equipment in addition to a 3-D sonic anemometer and a fast response hygrometer. Figure 2 shows the partitioning of the stability parameter z/L between the contributions from sensible (z/L_T) and latent heat fluxes (z/L_q). Over land, additional scatter is due to internal boundary layer growth from the coastline in some directions. For offshore conditions, z/L_q can account for up to 30% of the total z/L. As shown, the contribution of z/L_q is relatively large in unstable conditions and is also negative which means that adding humidity fluxes enhances unstable conditions.

The effect on wind speed profiles of including humidity fluxes is illustrated in Figure 3. In stable conditions, inclusion of humidity fluxes reduces the wind speed at 150 m height by approximately 5% relative to a profile computed without their inclusion (Figure 3). In unstable conditions, latent heat fluxes play a much smaller role in dictating the overall stability and hence there is a negligible effect on wind speed profiles. This is because a change in the value of L in unstable conditions due to humidity increases fluxes forcing conditions to be more unstable and, as shown in [7], the stability correction to the wind speed profile in unstable conditions tends to be smaller and much less sensitive to the value of L than in stable conditions. Additionally we consider that the influence of humidity flux must be small unless the value of the sensible heat flux is small [19] and this is also consistent with the influence of humidity fluxes being important only in stable conditions.

CONCLUSIONS

As wind energy develops offshore and resource predictions are required for greater heights, better understanding of the wind speed profile over the sea is required. One of the major differences between on- and offshore is the constant presence of strong humidity fluxes offshore. In stable conditions, the presence of humidity fluxes forces conditions towards near-neutral thus causing the vertical wind speed profile to approach log-normal. Using this as the criteria, differences in diabatic corrections to the wind speed profile were up to 5% at 150 m height in moderate and very stable conditions. However it should be acknowledged that Monin-Obukhov scaling is not valid to these heights in stable conditions. In unstable conditions, the inclusion of humidity fluxes increases instability but due to the relatively large sensible heat flux and the moderate correction to the wind speed profile in unstable conditions, the effect was not detectable. Hence the conclusion is that correction to wind speed profiles for humidity influences is only important in stable conditions and probably can be made in moderate conditions using a standard correction for the virtual potential temperature.

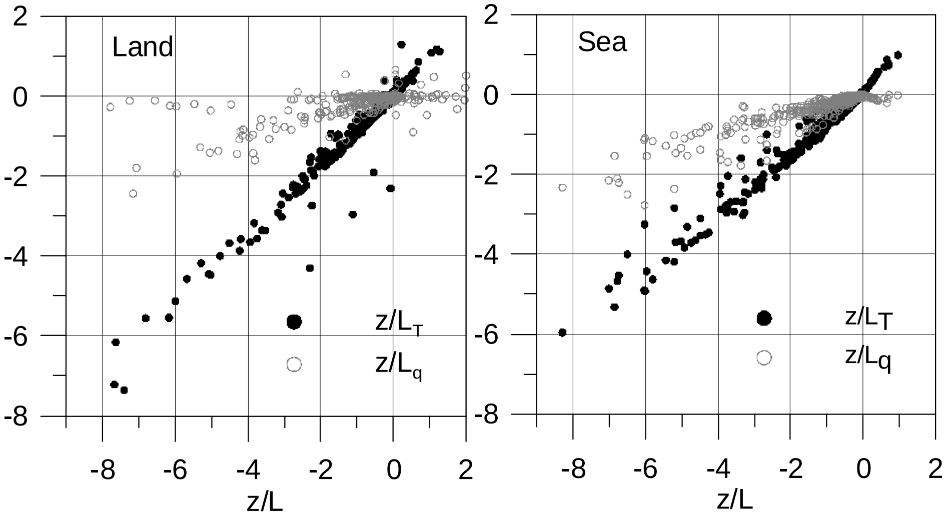

Figure 2. Partitioning of z/L between z/L_T and z/L_q for land (left) and offshore (right) conditions. Over land the influence of humidity fluxes (z/L_q) is small and z/L is dominated by z/L_T.

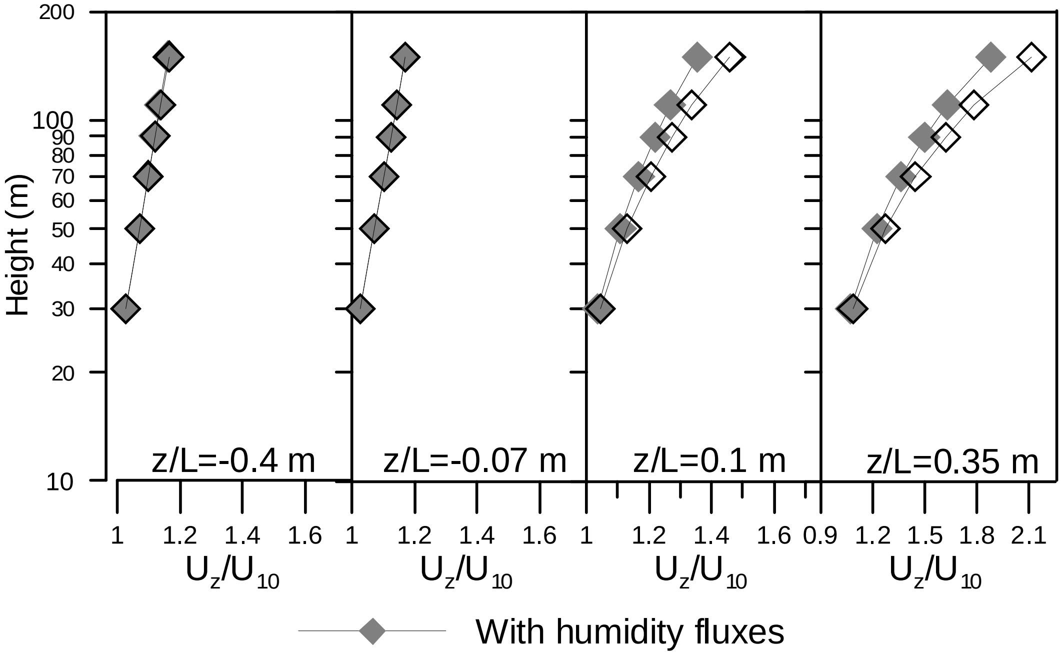

Figure 3. The influence of humidity fluxes on the wind profiles using data from the Anholt experiment [18]. The filled symbols are the normalised wind speed profile predicted using (1) and the experimental z/L including humidity fluxes; the open symbols are the wind speed profile without humidity fluxes. Inclusion of humidity fluxes does not influence the profile for unstable conditions but in stable conditions the profile is driven towards neutral.

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