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Turbulent Structure of the Marine Atmospheric Boundary Layer and Its Implications for the Inertial Dissipation Method

BY

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ABSTRACT

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In order to improve climate- and weather forecasting models, a better knowledge of the physical processes taking place in the lowest part of the atmosphere over the oceans is essential. In these models it is often assumed that the atmospheric boundary layer over sea behaves in the same way as that over land. But, the results show that the processes over sea are significantly different, which has to be accounted for in the models.

By using long term measurements it is shown that the surface waves play a very important role for the turbulent structure in the marine atmospheric boundary layer. For example, they give rise to a height structure that can not be found over land. A consequence of this is that measurements from a buoy (at a few meters above the surface) need to be treated different than measurements on a ship (at 10-30 m above the surface).

The wave influence affects the turbulent kinetic energy budget. Besides the height dependency, the imbalance between local production and local dissipation is a function of stability, wave age and wind speed, and the commonly assumed balance can therefore be questioned. This has direct implications for the so called inertial dissipation method, a method often used to determine turbulent fluxes over sea with the aid of measurements from ships and buoys. A comparison with the more direct eddy-correlation method at 10 m height gives that the inertial dissipation method works best for near neutral conditions and growing sea.

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List of papers

This thesis consists of the present summary and the following papers. In the summary, the papers are referred to by their Roman numerals.

Paper I:

Sjöblom Anna and Ann-Sofi Smedman, 2002: The turbulent kinetic energy budget in the marine atmospheric surface layer. *J. Geophys. Res., accepted for publication.*

Paper II:

Sjöblom Anna and Ann-Sofi Smedman, 2002: Vertical structure in the marine atmospheric boundary layer and its implication for the inertial dissipation method. *Boundary-Layer Meteorol., submitted.*

Paper III:

Sjöblom Anna and Ann-Sofi Smedman, 2002: Comparison between eddy-correlation and inertial dissipation methods in the marine atmospheric surface layer. *Boundary-Layer Meteorol., submitted.*

Comments:

In Paper I and II, the author of this thesis is responsible for the data analysis and the writing, except for the Appendix in Paper I. In Paper III, the inertial dissipation method calculations were performed by M. Yelland at SOC, U.K., and S. Cloche at CETP, France, whereas the author of this thesis is responsible for the analysis and the writing. Papers II and III are reproduced with kind permission from Kluwer Academic Publishers.

Contents

1. Introduction

The marine atmospheric boundary layer is not at all as well understood as that over land. There are several reasons for this lack of knowledge; a major problem is for example that the sea surface is mobile due to waves. Ships and buoys have to be used for the measurements, and the waves are causing movements, which have to be accounted for. The flow distortion can also be quite significant on these types of platforms. Furthermore, measurements from ships and buoys are generally very expensive and must thus be limited in duration. If tower measurements are used, the tower has to be placed at near-shore sites, thereby raising the question of effect of limited water depth and shoaling waves. Another problem with ship and buoy measurements is that they are usually limited to only one measurement height, i.e. assuming validity of Monin-Obukhov similarity theory. It has therefore often been assumed that the boundary layer over sea behaves in the same way as the boundary layer over land.

However, oceans cover about 70% of the earth's surface and play a decisive role in climate models. A better knowledge of the marine boundary layer is essential in order to improve the models, since there are major differences between air-sea and air-land interaction processes that can not be ignored.

A possible wave influence on the turbulent structure is a question that has been very controversial in the past. Already in the 70s this problem was identified (Volkov, 1970; Benilov et al., 1973; Makova, 1975), and it is nowadays more or less accepted that swell (i.e. waves travelling faster than the wind) alters the turbulence characteristics (e.g. Smedman et al., 1999). The wave influence during swell can be seen up to considerable heights and occasionally it influences the whole boundary layer (e.g. Smedman et al., 1994). Problems that then arise are for example: is Monin-Obukhov similarity theory valid and can we use parameterisations that have been developed over land? Unfortunately these questions have not all been answered completely satisfactory yet, and there still remains a lot of effort to do so.

The turbulent kinetic energy (TKE) budget describes the physical processes that generates, destroys and transports turbulence in the boundary layer. One important application of the TKE-budget is the so called inertial dissipation method; a method that has been used for over 30 years as a way to determine the fluxes over sea with the aid of measurements from ships and buoys (e.g. Fairall et al., 1990; Yelland and Taylor, 1996). The inertial dissipation method is usually preferable over the eddy-correlation method in marine conditions since it is not as sensitive to wave induced motions and flow distortion as the eddy-correlation method; only measurements in the high frequency part of the wind spectrum are used (e.g. Edson et al., 1991). But, the inertial dissipation method is unfortunately not free of problems. It relies on several assumptions and constants that have to be verified if the method shall be used during all conditions. A special problem is that the direction of the momentum flux is always assumed positive, which is not always the case during swell (e.g. Grachev and Fairall, 2001).

Some comparisons of the inertial dissipation method and the eddy-correlation method have been done before (e.g. Large and Pond, 1981; Fairall et al., 1990). These investigations showed good agreement between the two methods, at least for pure wind sea, but as Drennan et al. (1999) pointed out, the two methods might disagree a lot during strong swell conditions. Unfortunately, the influence of waves on the inertial dissipation method has not been considered much when testing the method in the past, and here much more research is needed.

The intention of this study is to examine the turbulent characteristics of the marine atmospheric boundary layer and its implications for the TKE-budget and thereby also for the inertial dissipation method.

2. Measurements at Östergarnsholm

Semi-continuous measurements have been performed since May 1995 at the air-sea interaction station Östergarnsholm. Östergarnsholm is situated about 4 km east of Gotland in the middle of the Baltic Sea (Figure 1). A 30 m tower is erected at the southernmost tip, with turbulence instruments (20 Hz sampling rate) at three levels and slow response ("profile") instruments (1 Hz sampling rate) at five levels, measuring wind speed, wind direction and temperature.

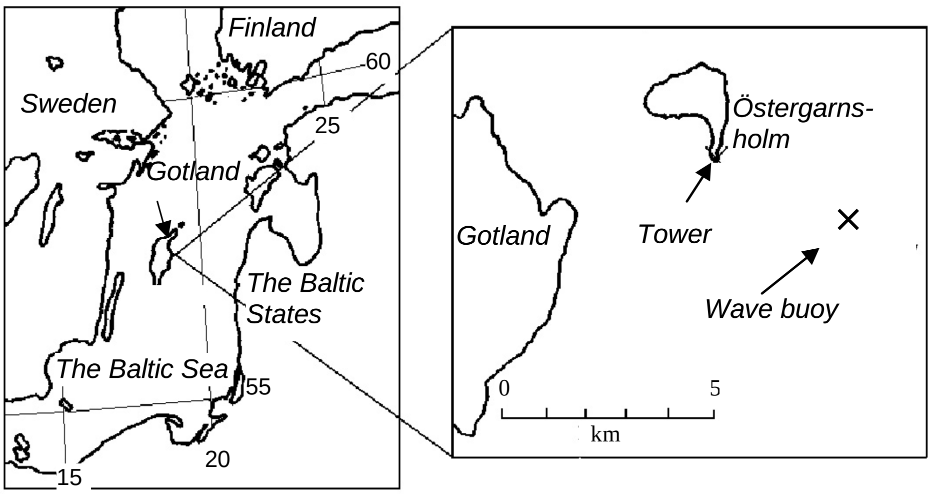

Figure 1: *Map of the Baltic Sea, with a close up on Östergarnsholm. The location of the wave buoy is also indicated.*

In the sector with wind coming from approximately 100-220°, there is a long (more than 150 km) undisturbed over water fetch, and the distance from the tower to the shoreline is normally a few tens of meters. The sea floor slope outside Östergarnsholm permits an undisturbed wave field for most conditions. However, during high wind speeds, a minor correction for limited water depth has to be applied (Smedman et al., 1999).

Since the sea level varies, the height from the water level to the instruments on the tower varies. Daily averages of the height were calculated with the aid of sea level measurements performed in Visby harbour at the west coast of Gotland (by the Swedish Meteorological and Hydrological Institute, SMHI) and a calibration procedure described in Paper I was applied. This gave mean heights of 10.4, 17.9 and 26.4 m above the sea surface for the three turbulence instruments (hereinafter refereed to as Level 1, Level 2 and Level 3), and 8.3, 13.3, 15.7, 21.6 and 30.2 m for the slow response instruments. Note, however, that in the analysis, the actual heights above the water were always used.

Both the turbulence (Solent Ultrasonic Anemometer 1012R2, Gill Instruments, Lymington, United Kingdom) and the profile instruments were calibrated individually in a big wind tunnel before they were installed on the tower (Paper I). The temperature measured by the sonic anemometers T_s is very close (about 0.20%) to the virtual temperature T_v (Dupuis et al., 1997; Paper I). The virtual heat flux $\overline{w'\theta_v'}$ obtained by the sonic anemometers has been corrected for "cross-wind" velocity contamination, since the signal is contaminated by the wind components normal to and along the path (Kaimal and Gaynor, 1991).

In addition to the tower instruments, a Wave-Rider Buoy (owned and run by the Finnish Institute for Marine Research) is deployed about 4 km from Östergarnsholm (direction 115°, Figure 1), measuring sea surface (bucket) temperature, significant wave height, wave direction and the spectra of the wave field (Smedman et al., 1999).

The buoy is moored at 36 m water depth, and is placed in the upwind fetch of the tower measurements, thereby representing the wave conditions in the "footprint area" outside Östergarnsholm. The wave measurements have been performed semi-continuously during the same period as the tower measurements, with the exception of wintertime periods with risk for ice damage.

Most of the swell encountered at Östergarnsholm is produced in the southern part of the Baltic Sea and then propagated northwards, giving swell at Östergarnsholm a more unidirectional appearance than usually observed in the open ocean. Thus, there may be a difference regarding the wave field in this study compared to the open ocean.

3. Features of the marine atmospheric boundary layer

The surface waves create boundary conditions for the atmospheric flow, and the varying characteristics of the wave field exert profound influence on the structure in the marine atmospheric boundary layer, for example the atmospheric stability and the wind profile. The atmospheric layer where surface waves have direct influence on the atmospheric flow is usually called the "wave boundary layer" (WBL).

3.1 Waves and stability

A common way to describe how well developed the wave field is, is to use the so called wave age, for which two definitions can be used

$$\text{Wave age} = c_0/(U_{10}\cos\theta) \tag{1}$$

$$\text{Wave age} = c_0/u_* \tag{2}$$

c_0 is the phase velocity at the peak of the spectrum, U_{10} the wind speed at 10 m, θ the angle between wind- and wave directions and u_* the friction velocity, defined as

$$u_*^2 = \left[(-\overline{u'w'})^2 + (-\overline{v'w'})^2\right]^{1/2} \tag{3}$$

where $-\overline{u'w'}$ and $-\overline{v'w'}$ are the components of the kinematic momentum flux in the along- and cross wind direction. Swell, i.e. waves travelling faster than the wind is then defined as $c_0/(U_{10}\cos\theta)>1.2$ (Pierson and Moskowitz, 1964) or $c_0/u_*>30$ (Volkov, 1970).

As shown in Paper I, it is not possible to exclude $-\overline{v'w'}$ in Equation (3) during swell conditions, i.e. the direction of the swell plays an important role.

The degree of wave development can be divided into three wave age intervals (Paper II):

- Growing waves, $c_0/(U_{10}\cos\theta)<0.5\text{-}0.9$
- Mature- or saturated waves, $0.5\text{-}0.9<c_0/(U_{10}\cos\theta)<1.2$
- Swell, $c_0/(U_{10}\cos\theta)>1.2$

As discussed in Paper I, the wave age and the stability is closely connected if the stability is defined as z/L:

$$\frac{z}{L} = -\frac{zgk\overline{w'\theta_v'}}{u_*^3 T_0} \tag{4}$$

where L is the Monin-Obukhov length and $\overline{w'\theta_v'}$ the flux of virtual potential temperature. T_0 is a reference temperature in the surface layer, k the von Karman constant [equal to 0.40 (Högström, 1996)], g the acceleration due to gravity and z the height of the measurements.

During swell, u_* is typically observed to be very small, causing large values of $|z/L|$. This originates from the definition of the stress, τ:

$$\tau = \rho u_*^2 \tag{5}$$

where τ can be expressed as a sum of three terms:

$$\tau = \tau_{turb} + \tau_{wave} + \tau_{visc} \tag{6}$$

τ_{turb} is the turbulent shear stress, τ_{wave} the wave induced stress and τ_{visc} the viscous stress. During swell, τ_{wave} will become negative and hence reduce or change the sign of the stress (e.g. Volkov, 1970; Smedman et al., 1994; Grachev and Fairall, 2001). Since the heat flux is often small over the sea, it is mainly u_* rather than the heat flux that determines z/L. Thus, there is a clear distinction from the boundary layer over land.

3.2 The wind profile

For growing sea, the wind profile over sea resembles that over land, i.e. it is logarithmic for neutral conditions. But, for mature sea and swell, the wind profile can be divided into three height intervals, and the depth of these layers varies due to the wave state, see Figure 2.

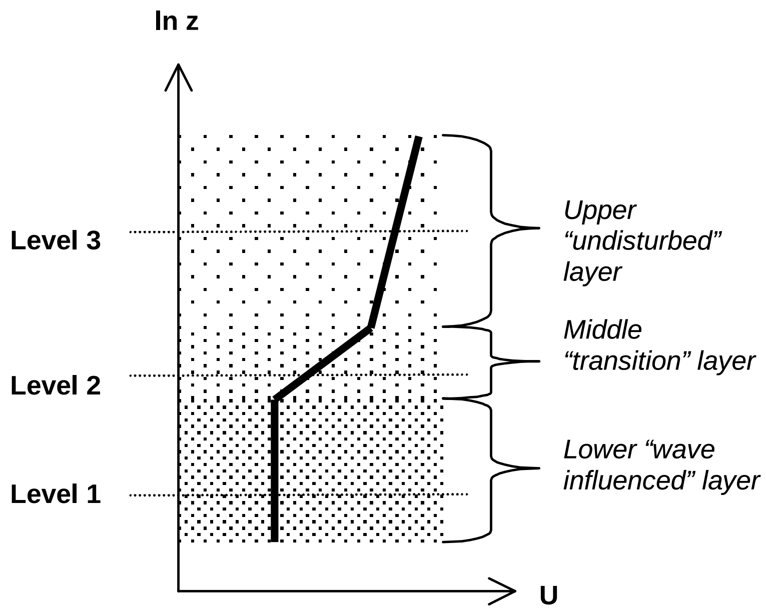

Figure 2: *Typical appearance of a wind profile over the sea.*

The lowest layer is a directly wave influenced layer, where a so called "wave driven wind" is not an uncommon feature (Smedman et al., 1999). The highest layer is the undisturbed "constant flux" layer, i.e. the wind profile is logarithmic during neutral conditions. Between these two layers, a "transition layer" appears, and the wind gradient may become very large, thereby questioning the validity of Monin-Obukhov similarity theory (Paper II).

3.3 Validity of Monin-Obukhov similarity theory

In Paper II it is shown that the validity of Monin-Obukhov similarity theory can be questioned not only for swell conditions, but also for mature- or saturated sea conditions.

According to Monin-Obukhov similarity theory, the turbulent fluxes should be approximately constant with height within the surface layer (e.g. Obukhov, 1971) and the mean wind gradient made non-dimensional with u_* and z should be a function of z/L only:

$$\phi_m\left(\frac{z}{L}\right)=\frac{kz}{(u_*)_{z=0}}\frac{\partial U}{\partial z} \tag{7}$$

As shown in Paper I, also the cross wind component has to be accounted for

$$\phi_m=-\frac{kz}{u_*^3}\left(\overline{u'w'}\frac{\partial U}{\partial z}+\overline{v'w'}\frac{\partial V}{\partial z}\right) \tag{8a}$$

where U and V are the mean wind speeds in along- and cross wind direction. But, since $\overline{v'w'}(\partial V/\partial z)$ is much smaller than $\overline{u'w'}(\partial U/\partial z)$, ϕ_m can be approximated to

$$\phi_m=-\frac{kz}{u_*^3}\overline{u'w'}\frac{\partial U}{\partial z} \tag{8b}$$

The cross wind component can, however, not be neglected when calculating u_* [Equation (3)].

Figure 3 shows measurements of ϕ_m plotted against z/L for the three measuring heights. All values are locally determined, i.e. also u_* [Equation (3)] at the three measuring heights individually. Dots (•) are the measurements, the solid line bin-averaged mean values, and dashed line is after Högström (1996), showing results obtained over land. Figure 3a shows Level 1 (~10m), Figure 3b Level 2 (~18m) and Figure 3c Level 3 (~26m).

As seen in Figure 3b, ϕ_m at Level 2 is larger than at the other heights at neutral stability, about 1.5 rather than 1.0. This can be explained by the fact that Level 2 is mostly situated in the "transition layer" described above, thereby having a too large wind gradient.

The conditions at Level 1 and 3 resemble land conditions quite well for neutral conditions. For unstable conditions, Level 1 shows the lowest individual values, indicating that this level is most directly influenced by swell.

As shown in Paper I and II, ϕ_m also has a clear wave age dependency. For growing sea [$c_0/(U_{10}\cos\theta)<0.5$], ϕ_m is constant with height, but for mature sea

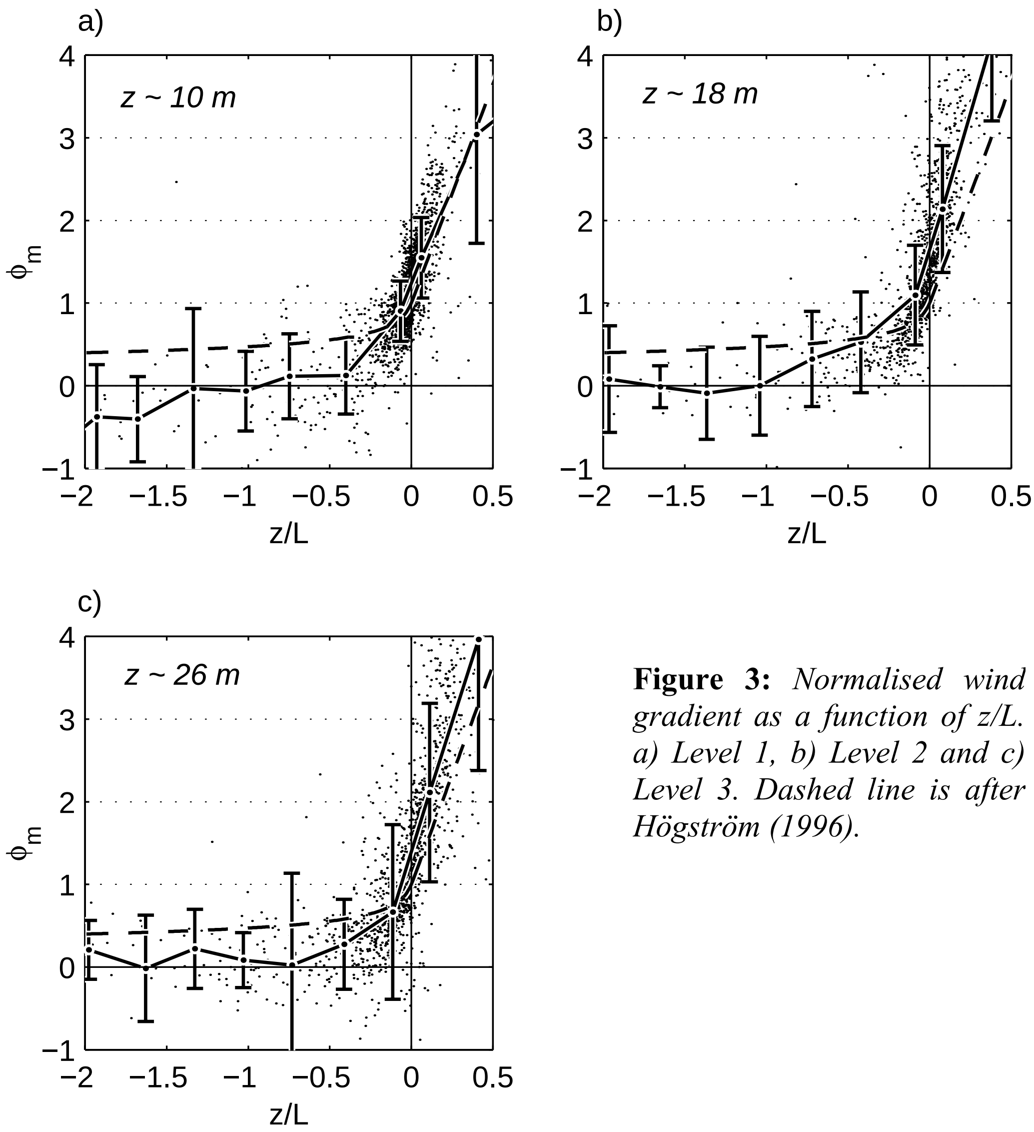

Figure 3: *Normalised wind gradient as a function of z/L. a) Level 1, b) Level 2 and c) Level 3. Dashed line is after Högström (1996).*

and swell this is not the case. The validity of Monin-Obukhov similarity theory for these conditions can therefore be questioned.

The validity of Monin-Obukhov similarity theory can also be examined with the aid of the normalised dissipation, ϕ_ε, since the inertial subrange of wind spectra should scale with height. ϕ_ε is defined as

$$\phi_\varepsilon = \frac{\varepsilon kz}{u_*^3} \tag{9}$$

where ε is the dissipation, determined from the inertial subrange by assuming Kolmogorov similarity and that Taylor's hypothesis is valid:

$$\phi_\varepsilon = \frac{kz}{u_*^3}\left[\frac{nS_u(n)}{\alpha}\right]^{3/2} \cdot n \cdot \frac{2\pi}{U} \tag{10}$$

where α is Kolmogorovs constant, n the frequency and $nS_u(n)$ the spectral value times the frequency. Here, α=0.52 has been used accordingly to Högström (1996). In Figure 4, ϕ_ε is plotted as a function of z/L (symbols as in Figure 3) for all measuring heights. Dashed line is after Högström (1990), showing results obtained over land. ϕ_ε is constant with height at neutral stability on average, but as shown in Paper II, this height constancy is evident only for $c_0/(U_{10}cos\theta)$<0.9. This means however, that ϕ_ε reacts on an increase in wave age later than ϕ_m. For swell, neither ϕ_ε nor ϕ_m are height constant.

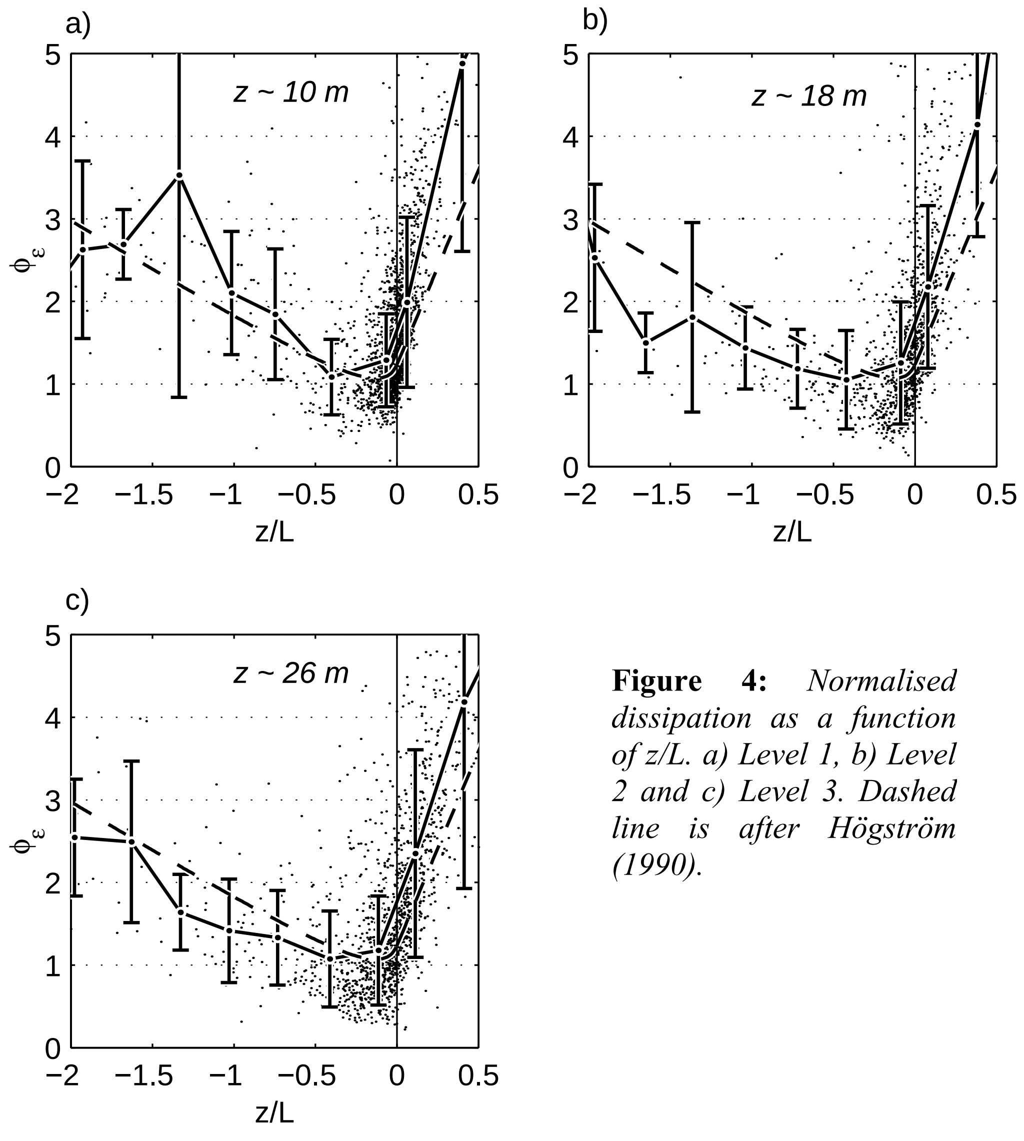

Figure 4: *Normalised dissipation as a function of z/L. a) Level 1, b) Level 2 and c) Level 3. Dashed line is after Högström (1990).*

A minimum of ϕ_ε is found at all heights at approximately z/L = –0.25, which is similar to findings over land (dashed line; Högström, 1990) and also over sea (Schacher et al., 1981). ϕ_ε is quite close to land equivalents at Level 1 for neutral and unstable conditions. But, on the stable side the measurements at all heights are slightly higher than over land.

Thus, ϕ_ε is not only a function of z/L, the actual height z also has to be considered over sea. This is not only the case for swell conditions, but also during mature sea. The applicability of Monin-Obukhovs similarity theory can therefore again be questioned. This is important, since measurements are usually taken at one height and then assumed to scale with height, but as shown above this is only the case for growing sea.

4. The turbulent kinetic energy budget

The problems discussed above have obvious implications for the TKE-budget. There have not been so many investigations of all the individual terms in the TKE-budget over the sea, normally only production and dissipation were measured (e.g. Edson and Fairall, 1998). The influence of waves on the TKE-budget also needs further research.

4.1 Theory

The TKE-budget and all the individual terms are described in detail in Paper I. Assuming stationary and horizontally homogeneous conditions, the normalised TKE-budget is defined as

$$\underbrace{-\frac{kz}{u_*^3}\left(\overline{u'w'}\frac{\partial \overline{u}}{\partial z}+\overline{v'w'}\frac{\partial \overline{v}}{\partial z}\right)}_{(I)}+\underbrace{\frac{kz}{u_*^3}\frac{g}{T_0}\overline{w'\theta_v'}}_{(II)}-\underbrace{\frac{kz}{u_*^3}\frac{\partial \overline{w'e}}{\partial z}}_{(III)}-\underbrace{\frac{kz}{u_*^3}\frac{1}{\rho}\frac{\partial \overline{p'w'}}{\partial z}}_{(IV)}-\underbrace{\frac{kz}{u_*^3}\varepsilon}_{(V)}=0 \qquad (11)$$

where $e = 0.5(u'^2 + v'^2 + w'^2)$. (*I*) corresponds to normalised mechanical production of TKE from the mean flow, (*II*) normalised buoyant production or loss, (*III*) normalised turbulent transport, (*IV*) normalised pressure transport and (*V*) normalised molecular dissipation of TKE. Equation (11) can also be written as

$$\underbrace{\phi_m}_{(I)} - \underbrace{\frac{z}{L}}_{(II)} - \underbrace{\phi_t}_{(III)} - \underbrace{\phi_p}_{(IV)} - \underbrace{\phi_\varepsilon}_{(V)} = 0 \qquad (12)$$

Term (*III*) and (*IV*) do not create nor destroy TKE, they just move it from one height to another. It is common to assume that these transport terms are small or cancel each other, at lest in near-neutral conditions. They are therefore often neglected, and a balance between local production and dissipation is assumed (Wyngaard and Coté, 1971; Hicks and Dyer, 1972; Large and Pond, 1981; Fairall and Larsen, 1986; Edson et al., 1991; Smith et al., 1992).

As shown in Equation (11), the terms are normalised with kz/u_*^3. This will of course make them very sensitive to the actual value of u_*. And, as discussed above, u_* can become very small during swell conditions, thereby influencing the normalised terms very much! However, the different terms are not influenced by the waves in the same way and below all terms will be discussed individually. Henceforth, positive values indicate a gain in turbulent energy, and negative values a loss.

4.2 Mechanical production

The normalised mechanical production [term (*I*) in Equation (11 and 12)] is by definition the same as the normalised wind gradient, ϕ_m, discussed in section 3.3. Over land, this function is quite well determined (e.g. Businger et al., 1971, Högström, 1996), but the same relationships are not always valid over sea as shown in Paper I and II.

ϕ_m is not only a function of stability but is also very sensitive to the wave age and to some degree wind speed. This wave age dependency will also give rise to a dependence on the actual height, not only z/L, a phenomenon that can not be found over land.

Using parameterisations that have been developed over land will therefore in most cases be erroneous. Except for the stability, also the wave age and the height dependence have to be included in the parameterisation.

4.3 Buoyancy

The normalised buoyancy is by definition the same as minus the stability z/L [Equation (4)]. This term will then be a gain in energy during unstable conditions, and a loss during stable conditions. As discussed in section 3, it is mainly u_* that determines this parameter over sea.

4.4 Dissipation

The normalised dissipation was defined and discussed in section 3.3. This term is always an energy loss (i.e. Figure 4 actually shows minus the dissipation). As shown in Paper I and II, ϕ_ε not only has a stability dependence, but also a wave age dependency. This, as for ϕ_m, give rise to a dependence of the actual height, not only z/L for mature sea and swell, and the dissipation, ε, will no longer scale with height.

4.5 Imbalance between production and dissipation

When considering the whole boundary layer, the normalised production equals the normalised dissipation by definition, i.e. the two transport terms are on average zero. But this is seldom the case locally, even if it is often assumed (see references above). The imbalance is defined as

$$\text{Imbalance} = \phi_m - z/L - \phi_\varepsilon \tag{13}$$

In Figure 5, the imbalance at the three heights is plotted as a function of stability, z/L (symbols as in Figure 3).

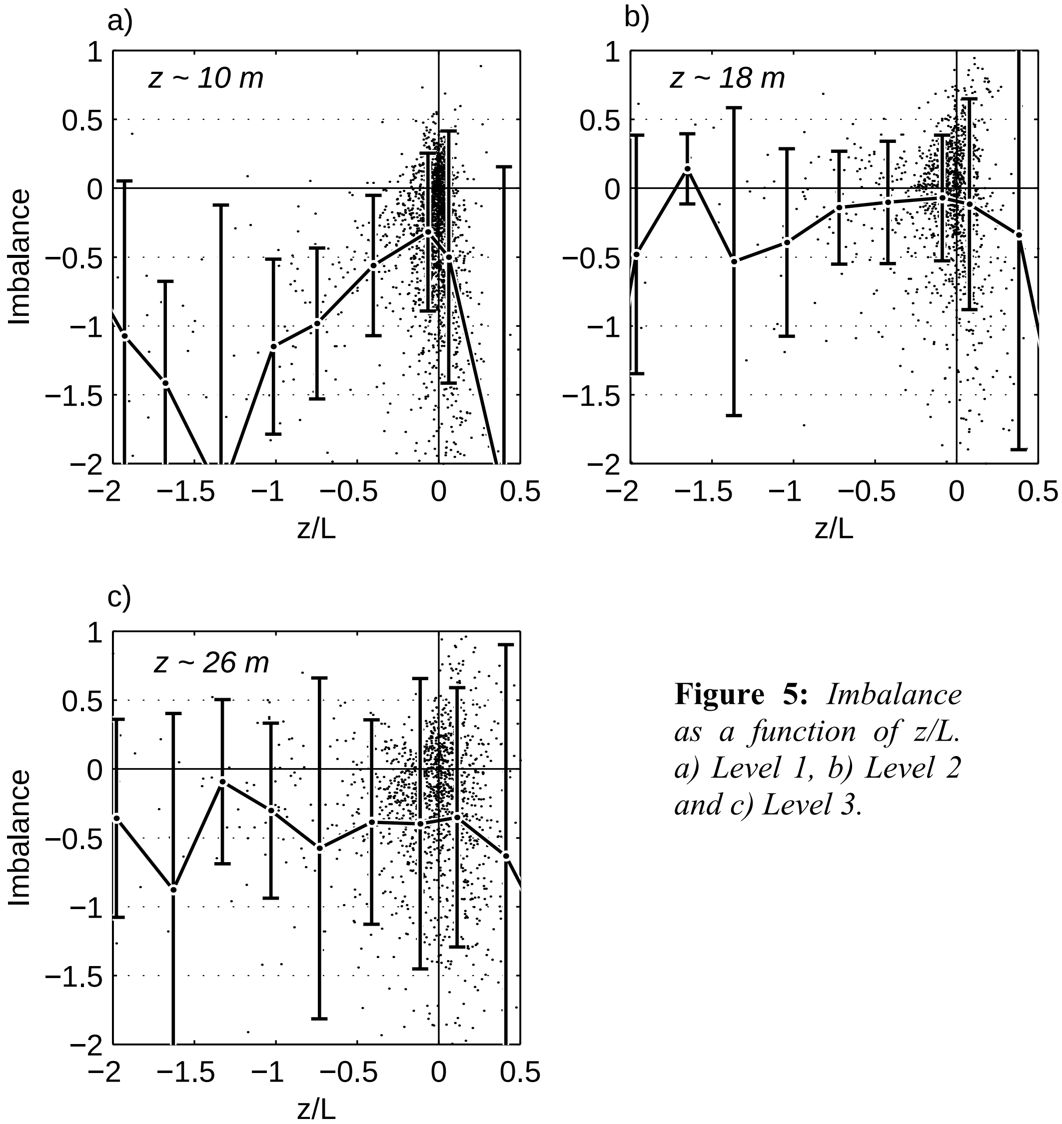

Figure 5: *Imbalance as a function of z/L. a) Level 1, b) Level 2 and c) Level 3.*

For unstable conditions, normalised dissipation is much larger than normalised production, especially at Level 1. The imbalance is smallest at Level 2 for neutral conditions, since ϕ_m is largest at this height (see discussion above). At Level 1 and 3 the imbalance is approximately –0.40; dissipation exceeding production. But as discussed above, the terms are dependent not only on stability but also on the wave age, and in Figure 6, the near neutral data of the imbalance is plotted as a function of both stability (z/L) and wave age [$c_0/(U_{10}\cos\theta)$] at the three heights. The data has been bin-averaged in two dimensions, and the numbers inside Figure 6 are the averaged values of the imbalance. The dashed line is drawn subjectively, showing imbalance ~ 0.

As discussed in Paper I, the imbalance at Level 1 (Figure 6a) for neutral conditions, can be divided into three different wave age intervals. For growing sea [$c_0/(U_{10}\cos\theta)$<0.5], the imbalance is positive, i.e. production exceeds dissipation. This is very likely to be the result of excess turbulent energy going into the growing waves (e.g. Edson and Fairall, 1998). For moderate wave ages

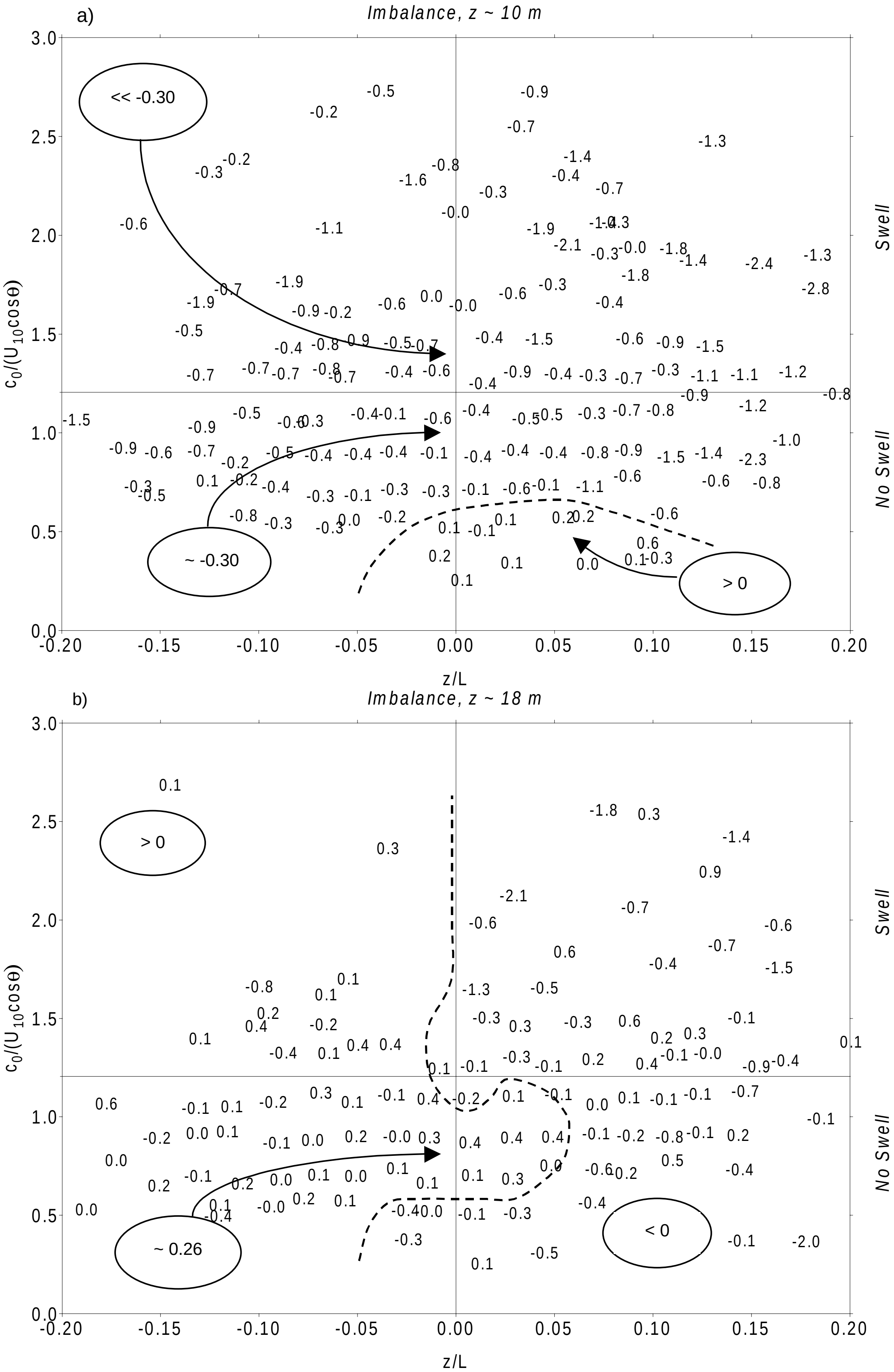

Figure 6: *Imbalance as a function of z/L and $c_0/(U_{10}\cos\theta)$. a) Level 1, b) Level 2 and c) Level 3. Dashed line (--) shows imbalance ~ 0.*

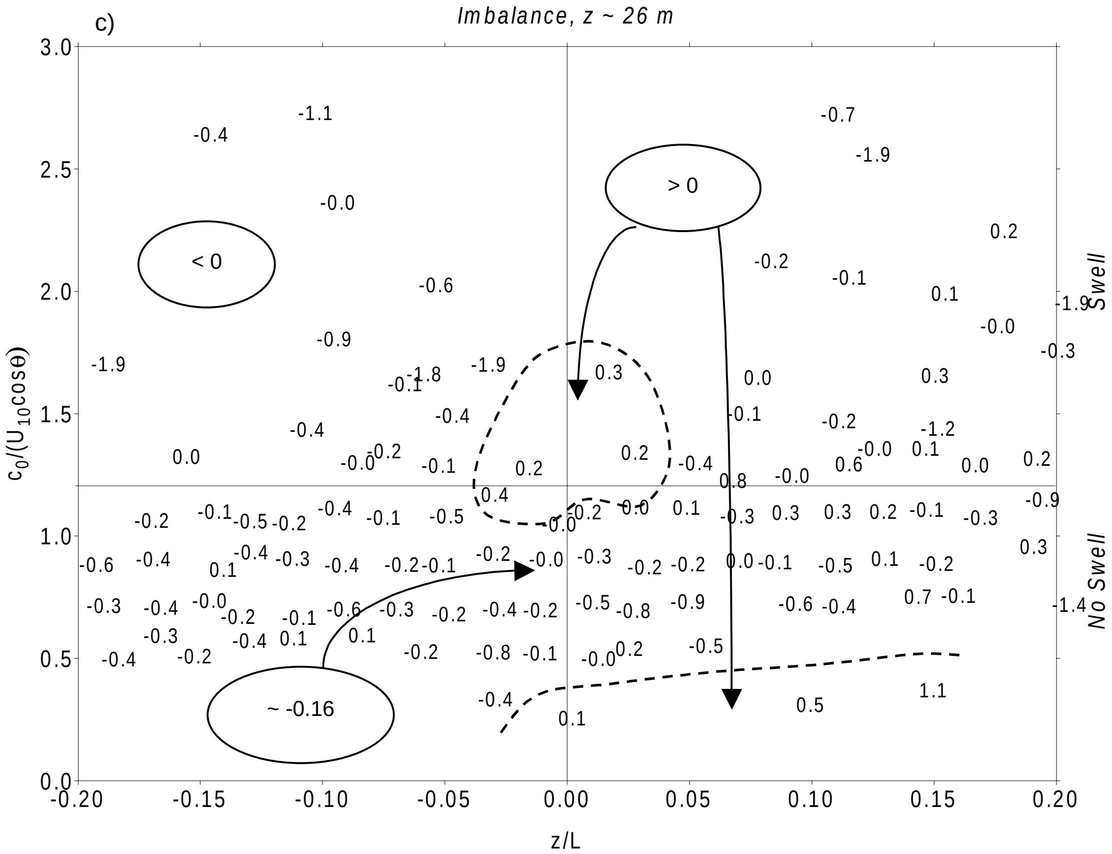

Figure 6, continued

[$0.5<c_0/(U_{10}\cos\theta)<1.2$], the imbalance is about -0.30, which is close to what has been found over land (e.g. Högström, 1996). For swell [$c_0/(U_{10}\cos\theta)>1.2$], the imbalance becomes much larger; dissipation exceeding production.

At Level 2 (Figure 6b), the imbalance for moderate wave ages is instead positive (about 0.26), indicating that the transition layer is present at this height for these wave ages. The positive imbalance therefore mainly originates from the large values of ϕ_m (Figure 3b).

At Level 3 (Figure 6c), the imbalance again becomes negative for moderate wave ages, although not as large as at Level 1, indicating that the measurements are now above the transition layer.

The imbalance is also the same as the sum of the transport terms [term *III* and *IV* in Equation (11 and 12)], and as shown above, this sum is seldom zero.

4.5.1 Turbulent transport

The normalised turbulent transport is defined as

$$\phi_t = \frac{kz}{u_*^3}\frac{\partial \overline{w'e}}{\partial z} \tag{14}$$

This term was found to be small (Paper I), even though a small increase with height can be found (not shown here). It can however be concluded that ϕ_t is not mainly responsible for the imbalance.

4.5.2 *Pressure transport*

The normalised pressure transport is defined as

$$\phi_p = \frac{kz}{u_*^3} \frac{1}{\rho} \frac{\partial \overline{p'w'}}{\partial z} \tag{15}$$

This term is very difficult to measure directly due to the small variations of pressure, even if some results of successful measurements have been reported (e.g. Wilczak et al., 1999). Here, ϕ_p was determined as a residual from the other terms. Since it is a residual, all errors from determination of the other terms will also end up in this term and add an uncertainty to the specific values.

It can however be noted that the normalised pressure term is quite close in magnitude to the imbalance. It can therefore be concluded that it is mainly ϕ_p that is responsible for the imbalance between production and dissipation since ϕ_t was found to be small (Paper I and II).

As discussed in Paper I, ϕ_p at Level 1 for neutral conditions is about 0.25 for moderate waves, which is in accordance with Högström (1996), found over land. This was explained by the process of inactive turbulence, where energy is being brought down from higher levels to be dissipated in the surface layer (e.g. Högström, 1990).

4.5.3 *Effective Kolmogorov constant*

A possible way to correct for the imbalance between production and dissipation is to use a so called "apparent" or "effective" Kolmogorov constant (e.g. Deacon, 1988), which was discussed in Paper I and III. According to Högström (1996) the "real" Kolmogorov constant should be $\alpha = 0.52 \pm 0.02$. The effective constant, α_a, is more uncertain, but values often suggested are $\alpha_a = 0.55$ (Large and Pond, 1981; Anderson, 1993; Donelan et al., 1997) or $\alpha_a = 0.59$ (Deacon, 1988; Högström, 1996).

In Paper III, α_a was calculated at Level 1. In Figure 7, α_a is calculated at the three measuring heights for $-0.025<z/L<0.025$ and shown as a function of wave age (symbols as in Figure 3). The dashed lines show respectively: $\alpha = 0.52$, 0.55 and 0.59.

It is clear from Figure 7 that α_a at Level 1 has a wave age dependency; α_a increasing with wave age. The overall mean is however 0.59, which is in accordance with land studies (e.g. Högström, 1996). Level 2 and 3 have not the same wave age dependency, α_a at Level 2 is quite small and at Level 3 it is 0.55 on average, which is in accordance with previous suggestions (see references above).

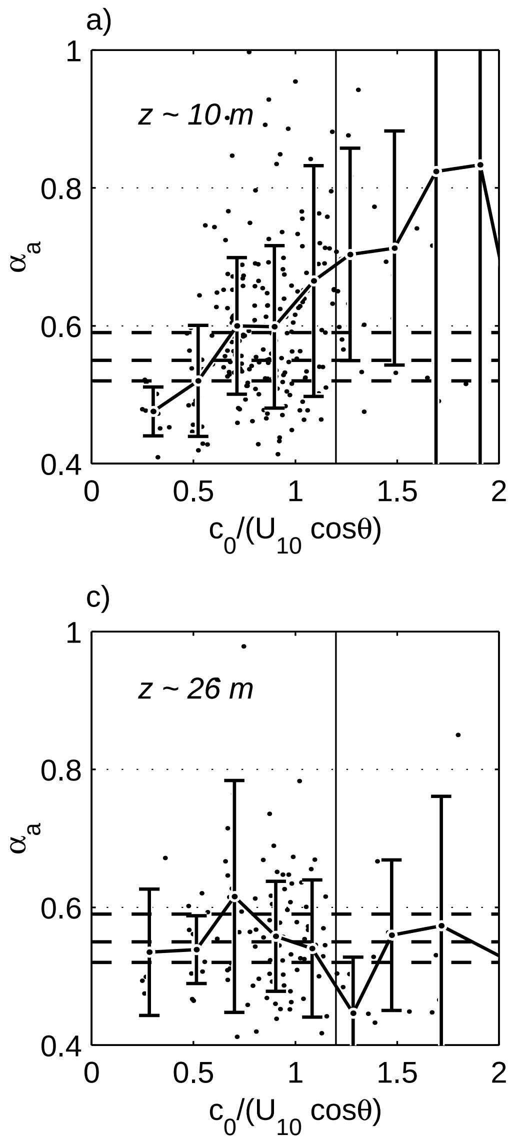

Figure 7: *Effective values of the Kolmogorov constant α_a, as a function of wave age for near-neutral data (-0.025<z/L<0.025). a) Level 1, b) Level 2 c) Level 3. Dashed lines (--) shows respectively: α = 0.52, 0.55 and 0.59.*

This means that a fixed value of α_a can not be used in the whole surface layer. α_a is a function of height, and at Level 1 it is also a function of the wave state, which has to be accounted for.

5. Measuring techniques

There are several techniques for determining the fluxes in the atmosphere over the ocean, and the most common are: the eddy-correlation method (see below), the profile method (e.g. Dupuis et al., 1995), the bulk aerodynamical method (e.g. Rugersson et al., 2001) and the inertial dissipation method (see below). The eddy-correlation and the inertial dissipation method will now be discussed more in detail.

5.1 Eddy-correlation method

This method was used to obtain the results shown above, and is the most direct way to measure, since the fluctuations are measured directly. The total stress, τ, can then be determined from Equation (3 and 5). However, this method can be quite difficult to use over sea, corrections for flow distortion and platform motion have to be made. Unfortunately, instrumentation for measuring all six motion components was until recently quite expensive, and even if successful results have been obtained (e.g. Anctil et al., 1994; Edson et al., 1998) there is still a long way to go before this method is easily available on a regular basis on ships and buoys.

5.2 Inertial dissipation method

An easier and therefore more common method to measure fluxes over sea, is the so called inertial dissipation method, which is described in Paper III. This method uses the normalised TKE-budget, assuming stationary and horizontally homogeneous conditions [Equation (11 and 12)]. From this the total stress can be calculated

$$|\tau| = \rho u_*^2 = \rho \left(\left[\frac{nS_u(n)}{\alpha} \right]^{1/2} \left[\frac{2\pi}{U_{app.}} \cdot \frac{kz}{\phi_m\left(\frac{z}{L}\right) - z/L - \phi_t\left(\frac{z}{L}\right) - \phi_p\left(\frac{z}{L}\right)} \cdot n \right]^{1/3} \right)^2 \tag{16}$$

where $U_{app.}$ is the apparent wind speed, i.e. the flow past the sensor.

The main problem is the uncertainties concerning the ϕ-functions. Also, an iterative method has to be applied, or an estimation of the stability has to be made, since z/L is needed for determination of the ϕ-values (e.g. Dupuis et al., 1997). Kolmogorovs constant, α, also has to be known.

The advantages with this method is that it is not as sensitive to flow distortion as the eddy-correlation method, and the measurements (i.e. within in the inertial subrange) are not in the same frequency interval as the movement of the ship or buoy. The wind speed needed is only the apparent wind speed, which can be different from the real wind speed if the ship is moving. Also, any measurements of the vertical wind speed are avoided.

5.3 Comparison between the eddy-correlation and the inertial dissipation method

The inertial dissipation method has been used for about 30 years as a way to determine the fluxes over sea (Smith et al., 1996), but not so many comparisons against the eddy-correlation method have been made. However, during the HEXOS experiment a comparison gave an uncertainty for the inertial dissipation method of 10% for stress, 20% for sensible heat flux and 25% for latent heat flux (Fairall et al., 1990).

As discussed above, the determination of the stability with the inertial dissipation method is not obvious. As shown in Paper III, this is especially a problem for unstable conditions. The correlation coefficient between the stability determined from the eddy-correlation method and the inertial dissipation method for z/L<-2 is as low as 0.10. For $-1<z/L<0.5$ it is much better, about 0.69, even if some individual values still have large deviations. Figure 8 (from Paper III) shows z/L determined with the eddy-correlation method (x-axis) and the inertial dissipation method (y-axis).

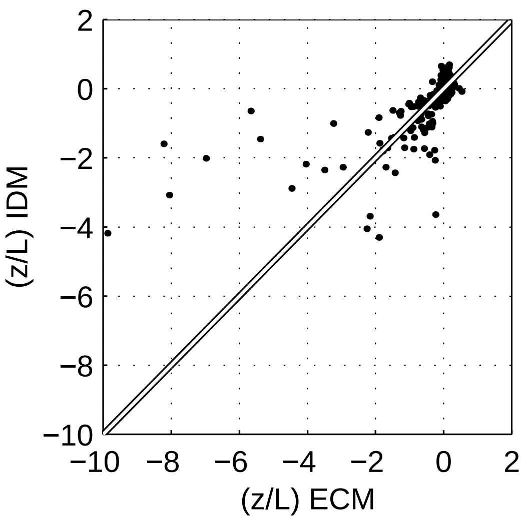

Figure 8: *The stability parameter z/L from the eddy-correlation method (ECM) and the inertial dissipation method (IDM). The dots (•) are the measurements (from Paper III).*

The large deviations during unstable conditions can probably be explained by the influence of swell, which is almost always influencing at these stabilities (see section 3.1). The influence can be seen directly in u_* determined from the eddy-correlation method, but not with the inertial dissipation method. The consequence is that the unstable values are actually treated as being more neutral than they actually are. If these erroneous z/L-values are used to make stability corrections for example of the drag coefficient, C_D, the errors can be quite large.

In Paper III, the friction velocity u_* determined from the two methods were also examined, and the agreement is best for young wave age conditions; the difference increasing with wave age. In Figure 9 (from Paper III), near-neutral data ($-0.02<z/L<0.02$) of u_* from the inertial dissipation method divided by u_* from the eddy-correlation method (both at Level 1) is plotted as a function of wave age. The value for the effective Kolmogrov constant in the inertial dissipation method has been set to $\alpha_a = 0.55$.

The wave age dependency in Figure 9 is quite clear. For $c_0/(U_{10}\cos\theta)<1$, the agreement between the two methods is fairly good, but at higher wave ages, the inertial dissipation method gives higher values than the eddy-correlation method. This again indicates that the inertial dissipation method woks best for growing sea, and that the wave influence can not be captured completely with the inertial dissipation method.

Above, it was shown that most turbulent parameters have a wave age dependence and it is probably this that gives the effect in Figure 9. New parameterisations for the inertial dissipation method therefore have to be applied,

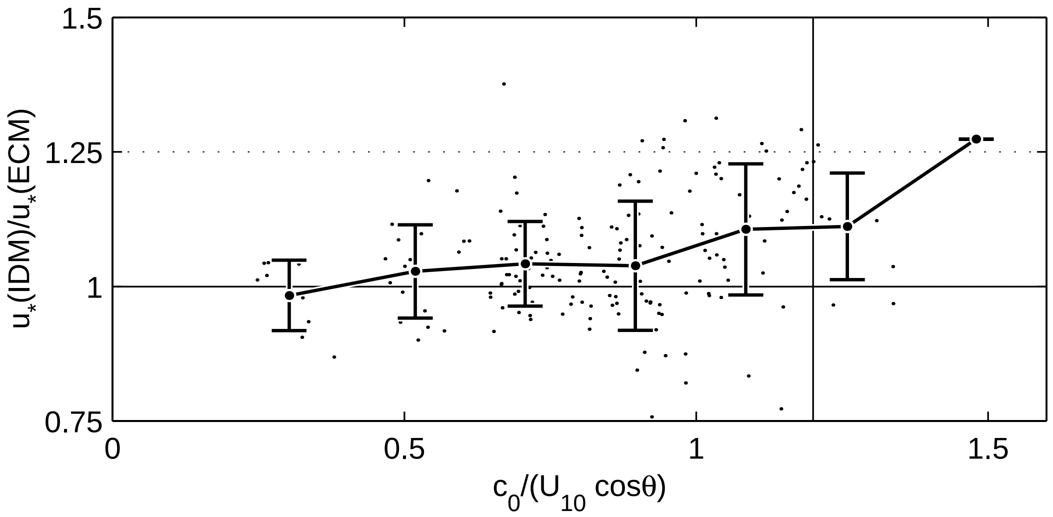

Figure 9: *u_* from the inertial dissipation method divided by u_* from the eddy-correlation method as a function of wave age. The dots (•) are all the measurements (from Paper III).*

where the wave influence is included. A height dependence also has to be considered.

6. Conclusions

By using long term measurements at three levels from an undisturbed land based air-sea interaction station with long over water fetch, it has been shown that the boundary layer over sea is in many ways different from the boundary layer over land.

The main difference is of course the influence of waves, which not only gives a wave age dependence, but also gives rise to a height dependence that can not be found over land. Due to this height dependence, Monin-Obukhovs similarity theory can be questioned during mature sea and swell.

This will also affect the turbulent kinetic energy budget, and thereby the parameterisations used in the inertial dissipation method. The commonly assumed balance between local production and local dissipation in the inertial dissipation method can be questioned during most conditions; the imbalance is not only a function of stability which has been previously suggested, but also a function of wave age and to some degree wind speed. It is mainly the normalised pressure transport that is responsible for this imbalance; the normalised turbulent transport was found to be small or negligible during the conditions encountered.

In addition, a height dependence has to be accounted for, since the waves give rise to at least three height layers with different characteristics. The depth of these layers is determined by wave age, i.e. a “directly wave disturbed” layer, a “transition” layer, and an “undisturbed constant flux” layer.

Completely different parameterisations have to be used depending on in which height layer the measurements are taking place. For example, measurements from a buoy with a measurement height of a few meters need a different parameterisation than measurements on a ship with a measurement height between 10 and 30 m. To be able to decide in which layer the measurements are taking place, wave measurements have to be performed simultaneously with the turbulence measurements. A better knowledge of what determines the depth of the layers is also needed.

A comparison of momentum fluxes determined by the inertial dissipation method and the eddy-correlation method gives about 15% higher values for the inertial dissipation method on average for $-1<z/L<0.5$. For more unstable conditions, the main problem lies in determining stability correctly.

The inertial dissipation method works best for growing sea during near neutral conditions, while for mature sea and swell, it gives higher values than the eddy correlation method.

To conclude, much further work is needed in this research area if measurements of turbulent fluxes over sea are to be performed on a more regular basis. Especially important is to come up with new parameterisations for the turbulent kinetic energy budget and thereby also the inertial dissipation method. These should be based not only on stability, but also have a wave age dependency, where the vertical structure is taken into account.

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