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Fluxes of Sensible and Latent Heat and Carbon Dioxide in the Marine Atmospheric Boundary Layer

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Abstract

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Oceans cover about 70% of the earth's surface. They are the largest source of the atmospheric water vapour and act as enormous heat reservoirs. Thus in order to predict the future weather and climate it is of great importance to understand the processes governing the exchange of water vapour and heat between the ocean and atmosphere. This exchange is to a large extent mediated by turbulent eddies. Current numerical climate and weather forecast models are unable to resolve the turbulence, which means that the turbulent exchange needs to be simplified by using parameterizations.

Tower based measurements at the Östergarnsholm Island in the Baltic Sea have been used to study the air-sea turbulent exchange of latent and sensible heat and the heat flux parameterizations. Although the measurements are made at an island, data obtained at this site is shown to represent open ocean conditions during most situations for winds coming from the east-south sector. It is found that during conditions with small air-sea temperature differences and wind speeds above 10 m s^{-1} , the structure of the turbulence is re-organized. Drier and colder air from aloft is transported to the surface by detached eddies, which considerably enhance the turbulent heat fluxes. The fluxes were observed to be much larger than predicted by current state-of-the-art parameterizations. The turbulence regime during these conditions is termed the Unstable Very Close to Neutral Regime, the UVCN-regime.

The global increase of the latent and sensible heat fluxes due to the UVCN-regime is calculated to 2.4 W m^{-2} and 0.8 W m^{-2} respectively. This is comparable to the current increase of the radiative forcing due to anthropogenic emissions of greenhouse gases, reported in Intergovernmental Panel on Climate Change fourth assessment report (IPCC AR4). Thus the UVCN-effect could have a significant influence when predicting the future weather and climate.

Keywords: Sensible heat flux, Latent heat flux, Carbon dioxide, Turbulent exchange, Global heat fluxes, Marine boundary layer, Air-sea interaction

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List of Papers

This thesis is based on the following papers, which are referred to in the thesis by their Roman numerals:

- I** Högström, U., Sahlée, E., Drennan, W.M., Kahma, K.K., Smedman, A., Johansson, C., Pettersson, H., Rutgersson, A., Tuomi, L., Zhang, F., and Johansson, M., 2007. To what extent can we believe measurements on a land-based tower to represent upwind open sea conditions? Submitted 2007-07-03 to Boreal Env. Res.
- II** Smedman, A., Högström, U., Hunt, J.C.R., and Sahlée, E., 2007. Heat/mass transfer in the slightly unstable atmospheric surface layer. *Quart. J. Roy. Meteorol. Soc.*, 133, 37-41.
- III** Smedman, A., Högström, U., Sahlée, E., and Johansson, C., 2007. Critical re-evaluation of the bulk transfer coefficient for sensible heat over the ocean during unstable and neutral conditions. *Quart. J. Roy. Meteorol. Soc.*, 133, 227-250.
- IV** Sahlée, E., Smedman, A., Högström, U., and Rutgersson, A., 2007. Re-evaluation of the bulk exchange coefficient for humidity at sea during unstable and neutral conditions. Accepted in *J. Phys. Ocean.*
- V** Sahlée, E., Smedman, A., Rutgersson, A., and Högström, U., 2007. Influence of a new turbulence regime on the global air-sea heat fluxes. To be submitted to *J. Climate*.
- VI** Sahlée, E., Smedman, A., Rutgersson, A., and Högström, U., 2007. Spectra of CO₂ and water vapour in the marine atmospheric surface layer. Accepted in *Boundary-Layer Meteorol.*

The author was responsible for the analysis and the writing of papers IV-VI. In paper I the author participated in the field work, evaluated and prepared the data-set and took part in the analysis of the data. In papers II and III the author took part in the analysis of the data. Papers II-III, Copyright Royal

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Introduction

Correct predictions of the future climate are of great importance for the entire society. A good scientific basis is needed on which decision makers can make long term planning. This is one of the reasons the IPCC (Intergovernmental Panel on Climate Change) was founded; to summarize the current climate related research and produce assessment reports. Since the foundation in 1988 four reports have been published (1990, 1995, 2001 and 2007) where the past and current climate has been described as well as presenting predictions of the future climate. The public has primarily focused on how the global mean temperature will change in the future. In the latest IPCC report, the increase of the global mean surface temperature in the next 100 years, assuming a doubling of the CO₂ concentration, is predicted to likely be in the range between 2°C and 4.5°C (Meehl et al., 2007). This is a significant increase, but the predicted range of increase is also quite large, i.e. there are uncertainties in the calculations.

What is the explanation of the large uncertainty range, and the expression 'likely' in the IPCC estimation? Uncertainty in the predictions depend on a number of factors, some has to do with how the different climate models are constructed and their resolution. Other aspects are how and what kind of physical processes that are included in the models. For instance, how clouds are formed and their total effect on the incoming and outgoing radiation is poorly described in the current climate models (Randall et al., 2007). Another uncertainty arises from how the models treat processes in the boundary layer.

The boundary layer is the lower part of the atmosphere. It is directly influenced by the surface where heating and friction creates turbulence. Turbulence exists at many scales but even the largest turbulent eddies are too small to be resolved by the coarse resolution in the climate models. Since turbulence is a very efficient way of transporting heat, water vapor and other gases to and from the surface it needs to be dealt with in some way in the models. This means simplifications. Usually the models approximate the turbulence using other parameters. How this approximation is done strongly influences the turbulent fluxes and thus also the global energy balance and water cycle.

Turbulence and turbulent fluxes have been extensively investigated over land through measurements. However, over the sea boundary layer processes are less well studied. One of the reasons for this is the harsher environment,

which makes high quality measurements harder to obtain. For instance, the presence of waves on the ocean surface will induce motions on floating platforms, like ships, almost in the same frequency range as that of the turbulence thus disturbing the measurements. Salt particles originating from water drops can stick to the instruments, which deteriorate the measurements. Another problem when measuring from ships is that the vessel structure to some extent induces a flow distortion, which needs to be accounted for. Ship measurements are usually made during certain dedicated field experiments. Thus, the results are valid only for the limited environmental conditions during which the cruise was performed.

However, there are ways around these problems. Influence from wave motions can be removed by the use of the ID-method (Inertial Dissipation) (Edson et al. 1991). Alternatively, motion sensors (see e.g. Edson et al., 1998) can be employed to directly remove the wave motions from the measurements. Contamination of the instruments by salt particles is mainly a problem for temperature sensors using hot wire or hot film technology. By using temperature measurements from sonic instruments instead, salt particles are less of a problem (Larsen et al. 1993). However, as discussed in Paper III, the temperature measurements from the sonic suffer from other problems at high wind speeds.

The flow distortion errors can be reduced by specially designed masts. They can also to some extent be modeled and corrected for in the post processing of the data. Also, the ID-method is less sensitive to flow distortion compared to the eddy-correlation method (Edson et al., 1991). However, the ID-method rests on certain critical assumptions and has been shown to have problems when measuring over swells, i.e. waves moving faster than the wind (Sjöblom and Smedman, 2003, 2004).

The data used in this thesis were obtained from tower measurements on the island of Östergarnsholm in the Baltic Sea. Using tower measurements solve the problem of platform motion and minimize the flow distortion. In addition, due to the low salinity in the Baltic Sea, there is no major problem with salt contamination. That this site indeed represents marine or open ocean conditions is shown in Paper I. An advantage by using a tower platform compared to ship borne measurements is that the tower platform enables the possibility of long term measurements. This allows for studies of the air-sea interaction during all kinds of atmospheric and oceanic conditions.

The main objective of this work was to study the turbulent fluxes of heat, water vapor and CO₂ over the oceans using the Östergarnsholm tower data. Oceans cover about 70% of the earth's surface and are huge energy reservoirs. Thus it is of great importance to understand how the transport processes governing the exchange between the sea and the atmosphere work.

Turbulence measurements

The eddy correlation technique

Turbulence exists at a wide range of scales. Most of the turbulent energy in the atmospheric surface layer exists at time scales from tens of minutes down to fractions of a second. Luckily, the instrument doesn't need to resolve the entire range when measuring turbulent fluxes. The important scales are satisfactorily covered at a sampling frequency of 10-20 Hz.

The measured signal, x , can be divided into two parts: one mean part and one part representing the deviations from the mean:

$$x = \bar{x} + x' \quad (1)$$

where the overbar denotes the mean part and the prime denotes the deviations from the mean, the turbulent part (in the stable boundary layer wave motions are also included in this part). This is also known as the Reynolds' decomposition of the signal. By decomposing the product of two measurements, for instance measurements of the vertical velocity, w (m s^{-1}), and potential temperature, θ (K), in this manner, and averaging using the Reynolds' averaging laws, we get:

$$\overline{w\theta} = \overline{w'\theta'} + \overline{w\bar{\theta}} \quad (2)$$

where the first term on the right hand side denotes the vertical kinematic flux of potential temperature and the second part is the transport due to the mean flow. Over a flat surface like the ocean the mean vertical velocity is very small, close to zero, i.e. $\overline{w\bar{\theta}} = \overline{w'\bar{\theta}}$. Thus by measuring the vertical velocity and the temperature, humidity or carbon dioxide, we can calculate the vertical turbulent fluxes by averaging the product of the two fluctuating parts over some time period, usually 10-60 minutes. Since the product $\overline{w'\theta'}$ represents a covariance this approach is called the eddy covariance or eddy correlation technique.

Sensor separation

Although the eddy correlation method seems easy in theory, there are some practical problems which need to be dealt with. One of these problems arises when two instruments are needed for the flux calculations. This is the case with the humidity and carbon dioxide fluxes, where one instrument measures the vertical velocity and another the humidity and carbon dioxide. These instruments are separated, i.e. they measure at two different locations. The sensor displacement will inevitably cause an attenuation of the correlation between the measurements, which in turn will result in a flux loss. Kristensen et al. (1997) note that the flux attenuation is a function of the ratio between the sensor separation distance and the scale of the turbulence. Thus since the turbulence scale gets smaller close to the surface, a given horizontal sensor displacement results in a larger attenuation at height of 1 m compared to 10 m. The attenuation will also be larger during stable stratification compared to unstable stratification due to the smaller scale turbulence.

The flux attenuations of humidity and CO₂ due to sensor separation at the Östergarnsholm site were investigated in Papers IV and VI. By using the Horst (2006) model the mean attenuation for the measurements at 10 m height due to the 0.3 m horizontal sensor displacement was 2% during unstable stratification and 5% during stable stratification. The attenuation was calculated and corrected for, for each individual measurement.

Instrumentation

Wind speed and temperature

At the Östergarnsholm site, most measurements of turbulent fluctuations in the three wind components and temperature are made using a sonic anemometer. The instrument basically measures the time between the emission and detection of a sound pulse over two times a given distance, which then is converted to wind speed. Since the speed of sound is a function of temperature, this instrument can also be used for temperature measurements. By using the same instrument for both the vertical velocity and temperature, no errors related to sensor separation need to be considered when calculating the kinematic heat flux. However other corrections need to be applied. The measured signal is not linearly related to the air temperature, it is also influenced by the wind fluctuations. The following correction has been used for the Östergarnsholm measurements (Kaimal and Gaynor, 1991):

$$\overline{w'\theta'_{so}} = \overline{w'\theta'_{so,uncorr}} + \frac{2\overline{U}\overline{u'w'}}{403} \quad (3)$$

where $\overline{w'\theta'_{so}}$ is the sonic kinematic heat flux (m s⁻¹ K), $\overline{w'\theta'_{so,uncorr}}$ is the uncorrected sonic kinematic heat flux, \overline{U} is the mean wind speed and $\overline{u'w'}$

is the kinematic momentum flux (m^2s^{-2}). A second correction is needed to get the true kinematic heat flux since the speed of sound is affected also by the humidity (Schotanus et al., 1983):

$$\overline{w'\theta'} = \overline{w'\theta'_{so}} - 0.51\overline{w'q'} \quad (4)$$

where $\overline{w'q'}$ is the kinematic humidity flux ($\text{m s}^{-1} \text{ kg kg}^{-1}$). Both of these corrections have been applied throughout this thesis whenever heat fluxes measured by the sonic are presented.

During intensive field campaigns the MIUU (Meteorological Institute at Uppsala University)-turbulence instrument of in-house design has also been employed. It uses hot film technology for measurement of the three wind components and platinum wires for measurement of temperature and wet-bulb temperature. The high accuracy and good long-term stability of this instrument, as demonstrated by Högström (1988), makes it well suited for high quality turbulence measurements. A detailed description can be found in the Appendix of Paper III.

During a field inter-comparison test between three MIUU-instruments and two kinds of sonic anemometers (Solent R2 and Solent R3) the MIUU-instrument demonstrated a very high performance (Högström 2001; Högström and Smedman, 2004). It is justified to call it a ‘reference instrument’. The turbulence statistics measured by two sonic instruments showed much worse performance compared to the MIUU-instrument, despite being calibrated individually in a wind tunnel prior to the experiment.

In Paper III, the field inter-comparison test is revisited and it is found that the heat flux measured by the sonic is unreliable at wind speeds above 10 m s^{-1} , except during conditions with a large air-sea temperature difference, ca. 4-5 K. The sonic measurements had to be corrected to match the MIUU measurements according to the following expression:

$$y = 1.0 + 0.2(U_{10} - 10) \quad (5)$$

where $y = \overline{(w'\theta')}_{MIUU} / \overline{(w'\theta')}_{sonic}$. However the fit is tentative and large scatter was observed. At 13 m s^{-1} the correction is about 60%. Data at wind speeds larger than 13 m s^{-1} were unavailable i.e. equation (5) is valid only in the wind speed range $10\text{-}13 \text{ m s}^{-1}$.

It is speculated that the prongs of the sonic are deformed at high wind speeds, which increases the sound flight time, thus influencing the measured heat flux. Similar results have been observed also by Grelle and Lindroth (1996) in a field inter-comparison test between a Sonic R2 anemometer and a platinum sensor.

Humidity and carbon dioxide

A Licor LI-7500, open path gas analyzer is used at the Östergarnsholm station. The densities of water vapor and CO₂ are measured by emitting an infrared beam and register the absorption over a known volume.

The Webb correction

A well known problem when performing density measurements are related to the fact that the measured density of any gas is affected by fluctuations in the density of the ambient air. Thus, by ignoring pressure fluctuations, the measured density fluctuations of water vapor is influenced by simultaneous fluctuations of temperature and the measured CO₂ density is influenced by fluctuations of both temperature and humidity. Webb et al. (1980) were the first to derive equations for correction of density flux measurements:

$$F_c = \overline{w' \rho'_c} + \mu \frac{\overline{\rho_c}}{\overline{\rho_d}} \overline{w' \rho'_v} + (1 + \mu \sigma) \frac{\overline{\rho_c}}{\overline{T}} \overline{w' T'} \quad (6)$$

$$F_v = (1 + \mu \sigma) (\overline{w' \rho'_v} + \frac{\overline{\rho_v}}{\overline{T}} \overline{w' T'}) \quad (7)$$

where F is the corrected mass flux (kg m⁻² s⁻¹), ρ is the mass density (kg m⁻³), subscripts v and c refer to water vapour and CO₂ respectively. σ is the

ratio $\frac{\overline{\rho_v}}{\overline{\rho_a}}$ (subscript a refers to the ambient air), T is the absolute air tempera-

ture (K), $\overline{w' T'}$ is the vertical turbulent temperature flux (ms⁻¹K) and μ is the

ratio $\frac{M_d}{M_v}$ where M is molar mass (kg mol⁻¹) and subscript d refers to the dry

air. These equations are standard practice today in post processing of density fluxes measurements.

In Paper VI an alternative method for density corrections, termed the Direct Conversion (DC)–method, is evaluated. By converting the 20 Hz density measurements directly to mixing ratios relative to dry air, the simultaneous fluctuations of the ambient air density are removed. It was found that the two methods are equivalent as shown in Figures 1a and 1b where a comparison of the resulting fluxes is presented.

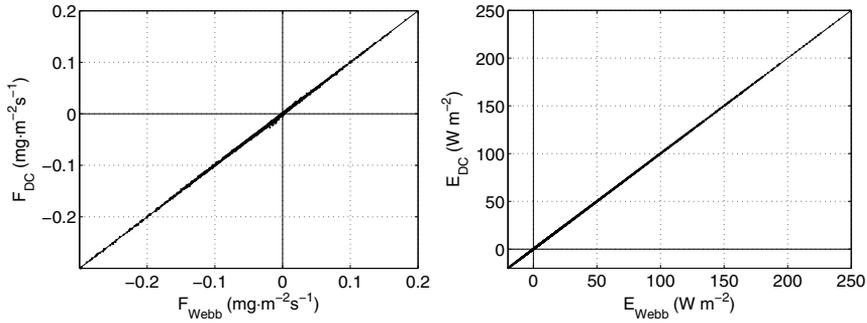


Figure 1 a) (left figure) Comparison between the CO₂ flux corrected with the Webb et al. (1980) method (F_{Webb}) and the CO₂ flux corrected with the DC method (F_{DC}). The solid line denotes a 1:1 relation. b) (right figure) as a) but showing a comparison of the latent heat fluxes (E).

Since the DC method corrects the 20 Hz measurements it is more versatile than the original Webb correction. Any turbulent moment containing a density variable calculated from the converted 20 Hz measurement is automatically corrected. A similar method derived by Dettu and Katul (2007), where external fluctuations (fluctuations in the ambient air density) are removed from the measured signal, also has this advantage. However, long equations are needed in their corrections. The simplicity of the DC-method is more appealing.

In addition, spectra calculated from the converted time series, the mixing ratio time series, are also corrected. In Figure 2 the $w\text{-CO}_2$ and $w\text{-}q$ cospectra based on the original measurements are converted to mixing ratio using mean variables to enable comparison with the corrected cospectra. Typically, the converted $w\text{-}q$ cospectrum displays almost no difference compared to the uncorrected cospectrum. Integration of the converted $w\text{-CO}_2$ cospectrum equals the corrected CO₂ mass flux in Equation (6) (divided by the air density), i.e. this is a way to study the frequency distribution of the Webb correction. Although the magnitude of the CO₂ flux is changed, the cospectral shape is well preserved.

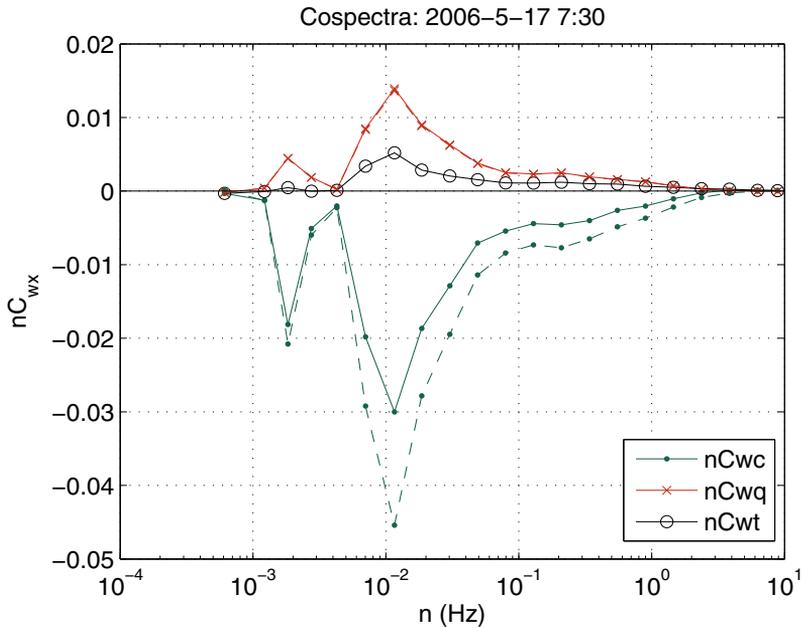


Figure 2 Half hour cospectra from 2006-05-17, 07:30 (LST). Lines with dots represent w-CO₂ cospectra, lines with crosses represent w-humidity cospectra and the line with circles represents the w-temperature cospectrum. Dashed lines show cospectra calculated directly from the measured signal, $nC_{wx}^{raw}(n)$, however the $nC_{wq}^{raw}(n)$ overlaps the DC-corrected $nC_{wq}^{DC}(n)$ cospectrum completely. Solid lines represent cospectra calculated from mixing ratio time series $nC_{wx}^{DC}(n)$.

Methods similar to the DC-method have been used by others before (e.g. Bergeron et al., 2007), although no evaluation of this method has been presented prior to Paper VI. It is interesting to read Leuning (2004) who states that this method works fine with closed path instruments but cannot be used with open path instruments. Figure 1 proves that this is an erroneous statement.

Parameterization of heat fluxes

Bulk formulations

Since the turbulent fluxes is governed by motions on a relatively small scale, which present numerical climate and weather forecast models can't resolve due to computational cost, the fluxes need to be parameterized. This is commonly done by using bulk formulas. These formulas relate the flux to more easily accessible parameters and an exchange coefficient. The sensible and latent heat fluxes can be expressed with the following equations:

$$E = \lambda \rho_a \overline{w'q'} = \lambda \rho_a C_E U_{10} (q_s - q_{10}) \quad (8)$$

$$H = c_p \rho_a \overline{w'\theta'} = c_p \rho_a C_H U_{10} (\theta_s - \theta_{10}) \quad (9)$$

where λ is the heat of evaporation (J kg^{-1}), H is the sensible heat flux (W m^{-2}), C_E is the exchange coefficient for latent heat (the Dalton number), c_p is the specific heat capacity for air at constant pressure ($\text{J kg}^{-1} \text{K}^{-1}$) and C_H is the exchange coefficient for sensible heat (the Stanton number). Subscript 10 refers to 10 m height and subscript s refers to surface values.

Applying Monin-Obukhov similarity theory by expressing the wind and scalar gradients through non-dimensional profile functions, the exchange coefficients can be expressed as:

$$C_E = \left[\frac{k}{\ln(z/z_0) - \psi_m} \right] \left[\frac{k}{\ln(z/z_{0q}) - \psi_q} \right] \quad (10)$$

$$C_H = \left[\frac{k}{\ln(z/z_0) - \psi_m} \right] \left[\frac{k}{\ln(z/z_{0t}) - \psi_h} \right] \quad (11)$$

where k is the von Karmán constant (0.4, Andreas et al., 2006), z_0 , z_{0q} and z_{0t} are the roughness lengths for momentum, humidity and temperature respectively (m). ψ_m , ψ_q and ψ_h are the integrated non-dimensional profile functions:

$$\psi_x = \int_0^{z/L} (1 - \phi_x(\zeta)) / \zeta d(\zeta) \quad (12)$$

where $x=q, h, m$ and $\zeta=z/L$. The non-dimensional profile functions are expressed as:

$$\phi_q(z/L) = \frac{\partial q}{\partial z} \frac{kz}{q_*} \quad (13)$$

$$\phi_h(z/L) = \frac{\partial \theta}{\partial z} \frac{kz}{\theta_*} \quad (14)$$

$$\phi_m(z/L) = \frac{\partial u}{\partial z} \frac{kz}{u_*} \quad (15)$$

where u_* is the friction velocity (m s^{-1}), $q_* = \frac{\overline{w'q'}}{u_*}$ and $\theta_* = \frac{\overline{w'\theta'}}{u_*}$ are the scaling parameters for humidity and temperature. z/L is a stability parameter where L is the Obukhov length (m):

$$L = \frac{-u_*^3 \theta_0}{gk \overline{w'\theta'_v}} \quad (16)$$

where g is the acceleration of gravity (m s^{-2}) and $\overline{w'\theta'_v}$ is the kinematic flux of virtual potential temperature ($\text{m s}^{-1}\text{K}$).

Most investigations remove the influence of atmospheric stratification from C_E and C_H to enable comparison with other experiments. These are called the neutral exchange coefficients, C_{EN} and C_{HN} , and have the form:

$$C_{EN} = \frac{k^2}{\left[\ln(z/z_0)\right]\left[\ln(z/z_{0q})\right]} \quad (17)$$

$$C_{HN} = \frac{k^2}{\left[\ln(z/z_0)\right]\left[\ln(z/z_{0t})\right]} \quad (18)$$

The influence of the sea state on the heat exchange

Guo Larsén et al. (2004) showed that the sea state needs to be considered when calculating the ϕ_m -function. In the presence of waves moving faster than the wind, swells, the wind gradient is modified due to wave induced upward directed momentum flux (Smedman et al. 2003, Sullivan et al. 2007). During unstable stratification the ϕ_m -function was found to decrease as the proportion of swells increased. Since this function is used to calculate

z_0 , which enters in Equations (10)-(11) and (17)-(18), the behavior of this function is important when parameterizing the heat fluxes. In Paper IV C_{EN} calculated using the ϕ_m -function of Guo Larsén et al. (2004) was compared to C_{EN} calculated using a ϕ_m -function derived from land measurements (Högström, 1996). It was found that during conditions dominated by swells, C_{EN} was reduced by 10% when using the wave dependent ϕ_m -function.

A similar result was obtained for C_{HN} in Paper III. The mean value of C_{HN} during swell conditions was shown to be 0.91×10^{-3} , for wind speeds in the range $3-10 \text{ m s}^{-1}$. During growing sea, C_{HN} was considerably larger having a mean value of 1.09×10^{-3} .

Surface renewal theory

One way of calculating the exchange coefficients is by using the so called surface renewal theory. Originally presented in Brutsaert (1975), it describes the physical processes governing the exchange at the interfacial sub-layers. This is the thin layers in the ocean and atmosphere directly in contact with the surface, where the fluxes of latent and sensible heat are assumed to be controlled solely by molecular diffusion. The diffusive transfer takes place between small-scale eddies and the surface. These eddies originate from the turbulent outer layer and they randomly enter and leave the interfacial layer. Eddies exchange heat during the time they are in contact with the surface. When an eddy loses contact with the surface and leaves the interfacial sub-layer, another eddy replaces it and the diffusion resumes. The rate at which this process takes place is called renewal rate. Diffusive transport is less efficient than turbulent transport, which means that the interfacial sub layer is a bottleneck in the transport processes at the air-sea interface.

A couple of models are more or less based on surface renewal theory, for instance the LKB model (Liu et al. 1979), the COARE (Coupled-Ocean-Atmosphere Response Experiment) algorithm (Fairall et. al. 1996, describe version 2.5) and the model by Clayson et al. (1996). A general result from these is that the exchange coefficients decrease slightly with increasing wind speed. The theoretical explanation for this decrease is related to the influence from the waves: A rougher surface will result in sheltering effects and thus suppressing the renewal rate, reducing the heat transfer. Countering the sheltering effect is the increased turbulent transfer. From the LKB model it was found that these two effects balanced each other at $4-5 \text{ m s}^{-1}$ at which the sheltering effect starts to dominate, reducing the exchange coefficients at higher wind speeds.

This modelled decrease is somewhat in line with field measurements at low to moderate wind speeds. As shown in Table 1 in Paper III or Table 4 in Paper IV, the exchange coefficients seem to be constant for wind speeds below 10 m s^{-1} . However, above 10 m s^{-1} the deviations from surface re-

new theory and measurements are larger. Some previous field experiments presented in the tables show that the exchange coefficients continue to be constant or slightly increasing also for wind speeds above 10 m s⁻¹. This brings one to the conclusion that surface renewal theory might only be used up to a certain wind speed.

COARE 3.0

In the latest version (3.0) of the COARE algorithm, presented in Fairall et al. (2003), a number of modifications have been made. One of the changes is that the parameterization of the roughness length, z_0 , has been altered. Previous versions of COARE used Charnock's formula (plus a limit for smooth flow) with a constant Charnock parameter, α , to obtain z_0 :

$$z_0 = \frac{\alpha u_*^2}{g} + \frac{0.11\nu}{u_*} \quad (19)$$

where ν is the viscosity of air. In version 3.0, α is increasing slightly for wind speeds between 10 and 18 m s⁻¹. Two alternative expressions for z_0 are also offered as an option. These are taken from the papers by Taylor and Yelland, (2001) and Oost et al. (2002), where z_0 is derived from data describing the wave properties (i.e. significant wave height, slope and wave age).

Fairall et al. (2003) also make a reanalysis of the HEXMAX data (HEXOS Main Experiment, where HEXOS was the Humidity Exchange Over the Sea program, results from HEXMAX are presented in DeCosmo et al. 1996). Using the modified HEXMAX data and another extensive data set including data from four field experiments, Fairall et al. (2003) derived a parameterization describing the roughness length for humidity, $z_{0q}=z_{0t}$, which is dependent on the roughness Reynolds number, $R_r = \frac{z_0 u_*}{\nu}$:

$$z_{0T} = z_{0q} = \min(1.1 \times 10^{-4}, 5.5 \times 10^{-5} R_r^{-0.6}) \quad (20)$$

This approach follows from the surface renewal theory and is also used by Liu et al. (1979). z_{0q} and z_{0t} , decrease with R_r but the rate of the decrease is less in the COARE 3.0 algorithm compared to the LKB model.

Since the exchange coefficients are dependent on z_0 , z_{0q} and z_{0t} , the new parameterizations of these makes C_{EN} and C_{HN} respond differently to an increase in wind speed. With the LKB representation of the roughness lengths (also used in COARE 2.5) the exchange coefficients decrease with increasing wind speed. The COARE 3.0 parameterization instead results in a small increase of C_{EN} and C_{HN} with increasing wind speed. However, a physical explanation for this behaviour is lacking.

The COARE 3.0 algorithm is considered to be ‘state-of-the-art’ by the flux community. However, as demonstrated by Brunke et al. (2003) in a comparison between 12 bulk flux algorithms, there are still large uncertainties in the present flux parameterizations. We demonstrate in Papers III-IV that this is likely to be due to lack of inclusion of important physical processes.

Measurements of the marine boundary layer from a land based tower

The Östergarnsholm site

This thesis is based on measurements made at the Östergarnsholm site. Östergarnsholm is a small island in the Baltic Sea ($57^{\circ}27'N$, $18^{\circ}59'E$) ca. 4 km east of the larger island of Gotland (see Figure 3).

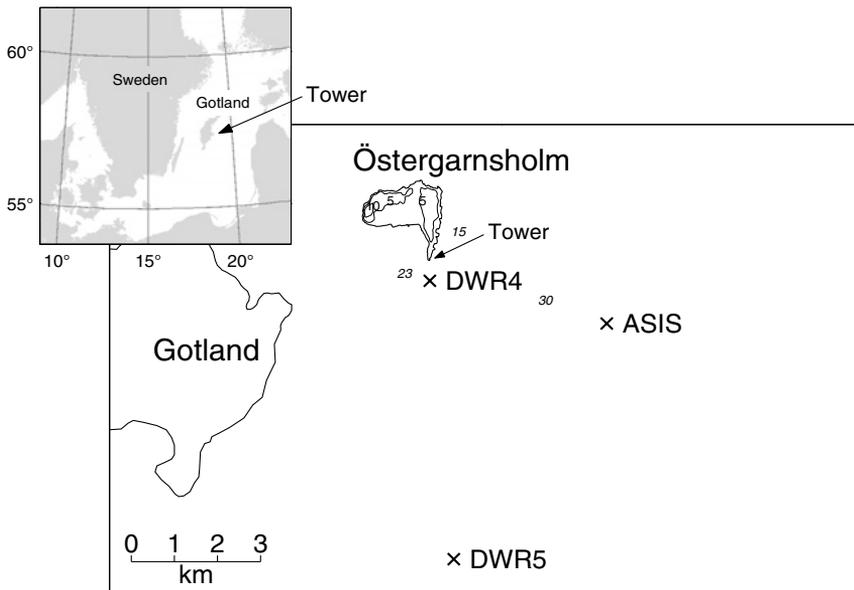


Figure 3 Map of Östergarnsholm and surroundings (modified from Johansson 2003a). Crosses mark buoy locations during the BASE experiment (see next section).

The island stretches about 2 km in the W-E and N-S direction respectively. The southern peninsula is very flat, rising only a few meters above the sea surface. Since 1995 a 30 m tower is situated at the southernmost tip of this peninsula, only a few tens of meters from the shoreline. However, the distance to the shoreline is dependent on the sea level which is constantly changing as a response to the meteorological conditions (tidal effects are negligible). Typically, the tower base is 1 ± 0.5 m above the sea surface. Sea

level data is recorded by SMHI (Swedish Meteorological and Hydrological Institute) on the west coast of Gotland in Visby harbor, and are used to get correct instrumental heights in relation to the sea level.

Only a few single trees grow on the island. These are situated at the NE corner, which is in a sector in which no tower measurements are used. Other vegetation consists primarily of low herbs and grass together with small bushes.

The seafloor is sloping approximately 1:30 out to a distance of 500 m from shore. At a distance of 10 km the depth is 50 m.

Instrumentation

The tower is instrumented with both slow response sensors and instruments for measurements of rapid turbulent fluctuations. Slow response measurements of temperature, wind speed and wind direction are made at five levels, 6.9, 11.9, 14.3, 20.2 and 20.8 m above the tower base. Two sets of instrumentations have been used since the measurements commenced in 1995. From 1995 to 2006 the wind speed and wind direction were measured using light cup anemometers and wind vanes of in-house design. These instruments were replaced in 2006 with Wind monitor propeller anemometers (Young, MI, USA). Temperature was measured using Pt-500 sensors in ventilated radiation shields during 1995-2003. In 2003 the Pt-500 sensors were replaced by copper-constantan thermocouples. Slow response measurements are also made of humidity at 7 m (Rotronic AG, Basserdorf, Switzerland) and air pressure at 1 m (144SC0811, Sensortechnics GmbH, Puchheim, Germany).

Turbulent fluctuations of the three wind components and temperature are made at three levels, 9, 16.5 and 25 m above the tower base, using Solent 1012R2 sonic anemometers (Gill instruments, Lymington, UK). As mentioned previously the sonic temperature measurements have been corrected for crosswind contamination (Kaimal and Gaynor, 1991), also the humidity influence has been removed (Schotanus et al., 1983). Since 2001 a Licor LI-7500 (LI-COR Inc. Lincoln, NE, USA), measuring turbulent fluctuations of humidity and CO₂, is placed at the 9 m level at 0.3 m distance from the sonic. During intensive field campaigns, the MIUU turbulence instrument has also been employed.

A wave rider buoy (run and maintained by the Finnish Institute of Marine Research, FIMR) is moored at the position marked as ASIS in Figure 3. The wave rider buoy is regularly removed during part of the winter to avoid damage from ice. During the BASE experiment it changed location temporarily (see below).

The BASE experiment

In the period September-October 2003 a joint Swedish, Finnish and US field experiment called BASE, the Baltic Sea Swell Experiment, were performed at the Östergarnsholm site. The main goal was to study the response of the atmosphere to swell waves. A second goal was to investigate to what extent the tower represents upwind open ocean/marine conditions (presented in Paper I).

During the experiment an ASIS (Air-Sea Interaction Spar buoy) was deployed (run and maintained by Rosenstiel School of Marine and Atmospheric Sciences, University of Miami, USA) as well as two directional wave rider buoys and a current profiler (run and maintained by FIMR). Measurements were also made onboard the Finnish research vessel R/V Aranda, which made several cruises in BASE area.

Besides instrumentation for measurements of the wave field and water temperature, the ASIS was deployed with two R2A Sonic anemometers (Gill instruments, Lymington, UK) mounted at 2.56 and 5.3 m height. A third sonic at 4.0 m failed. Two cup anemometer and vanes (WM301, Vaisala Oyj, Vantaa, Finland) were mounted at 1.18 and 2.42 m height. Ventilated and radiation shielded copper-constantan thermocouples were deployed at 0.7 and 1.7 m height above the sea surface. For details concerning the ASIS platform see Paper I and Graber et al. (2000).

The locations of the buoys are marked on the map in Figure 3. The ASIS buoy is situated about 4 km ESE of the island in 36 m deep water. Measurements made from this platform are assumed to represent ocean conditions for winds (and waves) coming from the open sea sector. Thus, this allows for an extensive evaluation to what extent the tower measurements really represent open ocean conditions by comparing them to the ASIS data.

Results

Footprint analysis

In general, any measurement made at a certain height is influenced by the surface at some distance upwind, called the footprint area. As shown in Appendix A of Smedman et al. (1999) the footprint area can be calculated using equations originally developed for atmospheric dispersion. The footprint is highly influenced by the atmospheric stability. During BASE the effective midpoint of the footprint, the distance at which 50 % of the flux originates from shorter distances and 50% from longer distances, was calculated for three cases, stable, neutral and unstable. The mean value for the 10 m level was 760 m, i.e. judging by this result, for winds coming from the sea sector the footprint is well over water and far upwind so that we expect that shallow water effects near the tower not to influence the measured fluxes. It was concluded that measurements during winds from the sea sector, 50-220°,

represent marine conditions. However for very stable cases a small part of the footprint may be over land for situations with winds from the 210-220° sector.

The wave field was evaluated using a refraction model of FIMR and a wave model (the WAM model, Komen et al. 1994) and the wave measurements from the three buoys. It was found that for winds and waves from the 90-190° sector, the flux footprint truly represent open ocean conditions, i.e. the waves are unaffected by the coast and don't 'feel' the sea floor. In absence of swell this is true also for the winds and waves coming from the 50-80° sector. However, during swell a shoal in this area influences the wave field, i.e. during situations with swells this sector does not represent open ocean conditions. In the sector 190-220°, the proximity of the Gotland Island increasingly influences and disturbs the wave field.

If the tower footprint truly represents open ocean conditions, then fluxes measured at ASIS and at the tower should show a close correlation. Such a comparison of the friction velocity is shown in Figure 4.

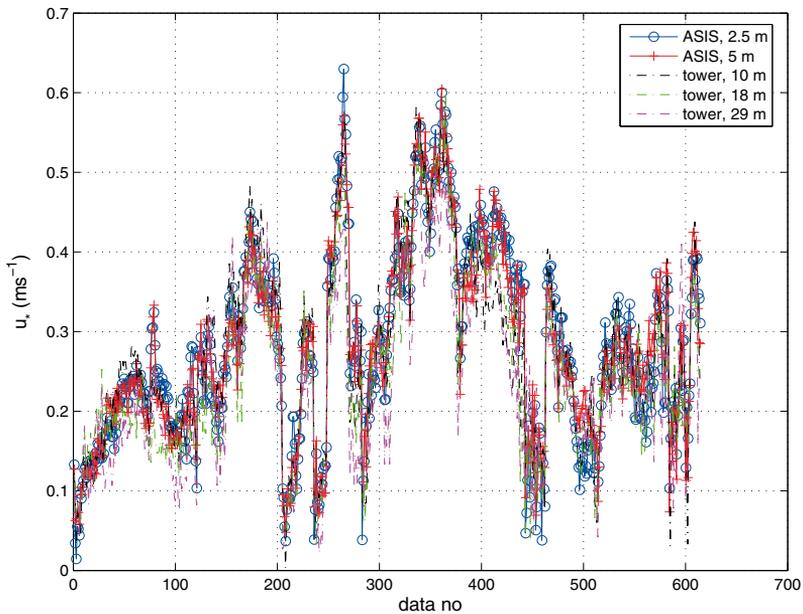


Figure 4 Friction velocities from ASIS at two levels and from three levels on the tower, plotted against data number.

It is clearly seen that the data follow each other closely. In mean, the ratio of the friction velocities between the ASIS and the tower at 10 m, is close to unity. Similar results were obtained for the sensible heat fluxes although with larger scatter due to small fluxes close to the noise level. Thus, the sea surface in the flux footprint areas, at 10 m level on the tower and at the ASIS 4 km to the SE, can be said to be very similar.

Wind profiles

The ϕ_m -function was evaluated from the wind speed measurements on both ASIS and the tower during BASE. For stable stratification it is in fair agreement with previous findings (e.g. Bergström and Smedman, 1995; Cheng and Brutsaert, 2005):

$$\phi_m = 1 + 6.0z/L \quad (21)$$

However, for unstable stratification, the BASE data are systematically lower than what has been previously obtained from land measurements (see e.g. Högström, 1996). Two straight lines provided the best fit to the data:

$$\phi_m = 1 + 7.5z/L, \quad 0 > z/L > -0.12 \quad (22)$$

$$\phi_m = 0.1, \quad -0.12 > z/L > -1 \quad (23)$$

This result was interpreted as an effect of the low boundary layer height during the BASE campaign. The influence of the boundary layer height on the ϕ_m -function has previously been investigated using Östergarnsholm data (Johansson et al., 2003b), with similar results.

In Paper I it was shown that the drag coefficient, C_D , was sensitive to the choice of ϕ_m -function when converted from the neutral drag coefficient C_{DN} . The variation could be as much as 36% depending on the choice of ϕ_m . This is most likely also the case when converting C_{EN} to C_E and C_{HN} to C_H . Thus, the boundary-layer height could play an unexpected important role in the air-sea exchange. However, it is still unknown how frequent these situations occur. Should a study show that these situations are common, the climate models ability to calculate the boundary-layer height must be assessed.

Using the integrated ϕ_m -function the wind profiles can be calculated by the following expression:

$$u = \frac{u_*}{k} (\ln(z/z_0) - \psi_m) \quad (24)$$

provided that the friction velocity is approximately height constant. In the mean during BASE this requirement was fulfilled. In Paper I, the wind profile was calculated from Equation (24) using mean values of u_* , L and z_0 (from ASIS measurements at 2.42 m) in combination with the integrated ϕ_m -function (Equations (21)-(23)). This way of calculating the wind profile showed excellent agreement with the directly measured wind profiles all the way from the anemometer at 2.42 m at ASIS to the top level on the tower at 30 m, as demonstrated in Figure 5.

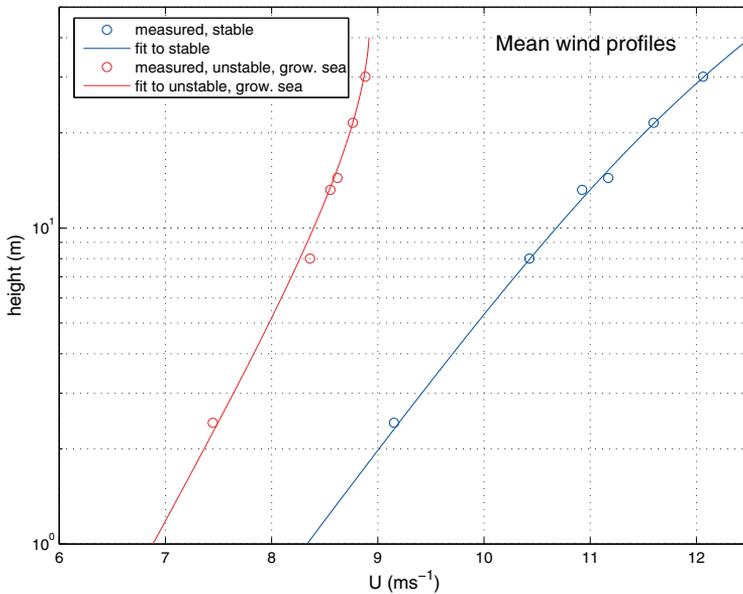


Figure 5 Mean wind profiles during base for unstable (red) and stable (blue) conditions. Solid line are calculated wind profiles from Equation (24), open circles are averages of 236 half hour wind speed measurements. Notice the lin-log representation.

However, the individual ASIS wind profiles didn't match the tower profiles in most cases. The mismatch was typically 0.5 m s^{-1} , and was shown to be a random process most likely an effect of the dynamical structure of the boundary layer.

To conclude: The results from the BASE experiment showed that the tower measurements at Östergarnsholm truly represent open ocean conditions with winds from $90\text{-}190^\circ$ and marine conditions (footprint over sea but waves affected by the coastline or sea floor) in the $50\text{-}220^\circ$ sector. In absence of swell, the $50\text{-}90^\circ$ sector also represents open ocean conditions.

Air-sea heat fluxes in the near neutral boundary layer

Exchange coefficients

The exchange coefficients for sensible and latent heat during unstable stratification were investigated using Östergarnsholm data in Papers III and IV. Surprisingly it was found that at wind speeds above about 10 m s^{-1} , C_{EN} and C_{HN} increased with decreasing air-sea temperature difference, ΔT . Figure 6 shows C_{HN} as a function of wind speed. Measurements with both the MIUU instrument and the R2 Sonic are included where the sonic data have been corrected according to Equation (5). Wave dependent ϕ_m -functions have been used in the calculations.

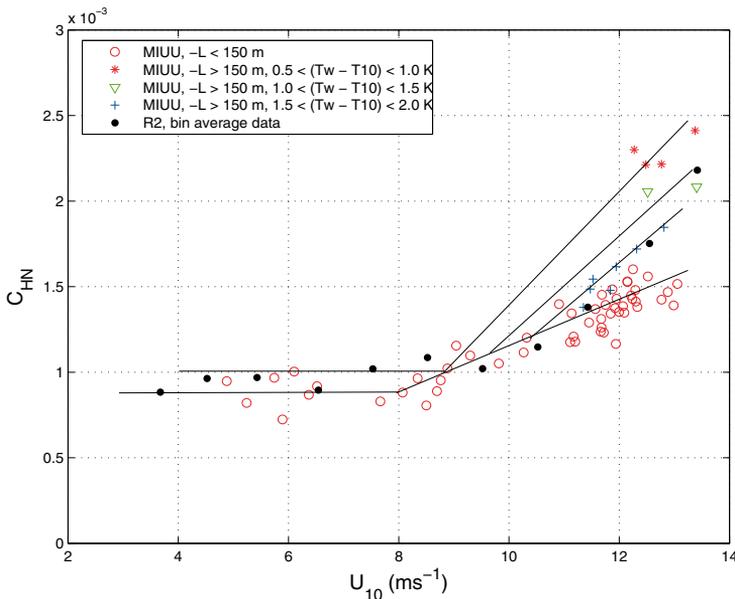


Figure 6 C_{HN} as a function of wind speed at 10 m. Symbols explained by legend. Solid lines are tentative fits to MIUU data. R2 data are bin averages (bin size, 1 m s^{-1}), MIUU data represent individual measurements.

For cases with $L > -150 \text{ m}$ and wind speeds above 9 m s^{-1} , C_{HN} is observed to increase with increasing wind speed. The increase is most promi-

ment for the smallest air-sea temperature difference. Wind speed dependence has been found also by other authors, however there is a large divergence between different experimental findings (see e.g. Table I in Paper III).

A similar result was found also for C_{EN} in Paper IV, shown in Figure 7. Based on a data set of 925 half hours, the exchange coefficient was observed to increase with increasing wind speed for situations with small ΔT . The behavior of C_{EN} differs from that of C_{HN} ; while C_{HN} is observed to increase with wind speed for all values of ΔT , C_{EN} is constant during conditions with $\Delta T > 3$ K.

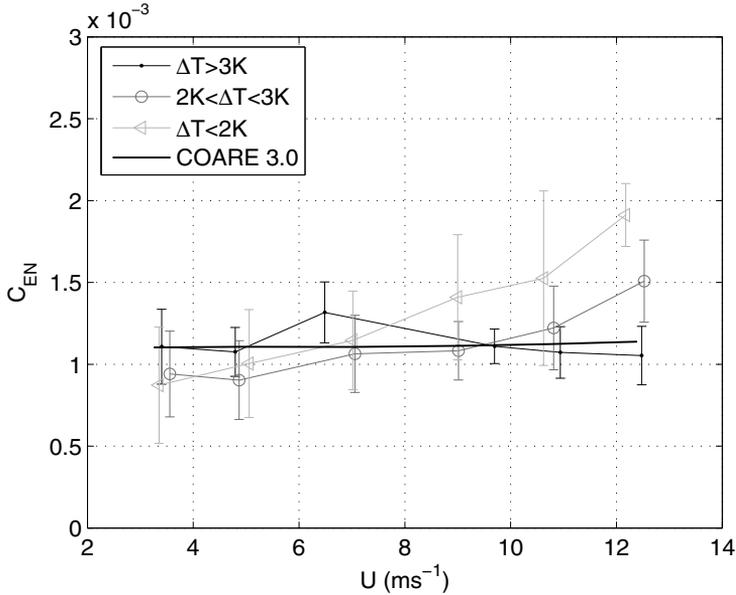


Figure 7 C_{EN} as a function of wind speed at 10 m. Wave dependent ϕ_m -functions have been used. Symbols represent bin averages (bin size 2 m s^{-1}), error bars show ± 1 standard deviation.

C_{EN} from the COARE 3.0 algorithm is also plotted in Figure 7. For $\Delta T > 3$ K the COARE algorithm agrees well with the experimental results. However, for smaller air-sea temperature differences and wind speeds above 9 m s^{-1} , C_{EN} predicted by COARE is significantly smaller. The difference between the observations and the COARE predictions increases as ΔT decreases and the wind speed increases. This indicates that some physical process is lacking in the COARE algorithm.

Sea spray

Sea spray is not included in the COARE algorithm and it could possibly influence the air-sea heat fluxes. Breaking waves produce sea spray droplets,

which enhance the latent heat flux by evaporation from the droplets' surfaces. This process in turn cools the air thus also enhancing the sensible heat flux. However, in lack of direct measurements of spray an indirect method was utilized in Paper IV to quantify the spray influence by employing the flux model by Andreas (2004), which incorporates sea spray in the calculations. For the cases with $\Delta T < 2 \text{ K}$ and $U > 10 \text{ m s}^{-1}$, it was found that sea spray mediated fluxes enhanced C_{EN} by 10%. However, the observed increase was 56.3%, i.e. sea spray alone is not enough to explain the full increase of C_{EN} .

Turbulence reorganization

In Papers II-IV spectral analysis was performed to study the turbulent fluctuations in the unstable boundary layer. It was found that temperature spectra, cospectra of vertical velocity-temperature and vertical velocity-humidity, developed a second peak at a high frequency as the instability decreased. This peak represents an increase in the energy of the turbulence at scales an order of magnitude smaller than found in an ordinary convective boundary layer. Cases exist where the smaller scale turbulence dominates, i.e. the spectral peak shifts completely.

The unstable boundary layer in moderate winds is characterized by the longitudinal roll type of eddy structure. However, the results from the spectral analysis indicate that these roll structures break down and that the turbulence is reorganized in the near neutral slightly unstable boundary layer.

This new turbulence structure was termed the Unstable Very Close to Neutral regime, the UVCN-regime (Paper II). As shown in Figure 8 these results are observed also at a land site i.e. the formation of these structures is not dependent on the nature of the underlying surface.

Quadrant analysis (Lu and Willmarth, 1973, Willmarth and Lu, 1974 and Raupach, 1981) showed that during UVCN conditions a large part of the upward latent and sensible heat fluxes originated from cold and dry air brought down to the surface from layers aloft. This is quite different from the moderately unstable boundary layer in which the quadrant analysis showed that the heat fluxes were dominated by warm/humid air moving upwards, as intuitively expected (Papers II and IV).

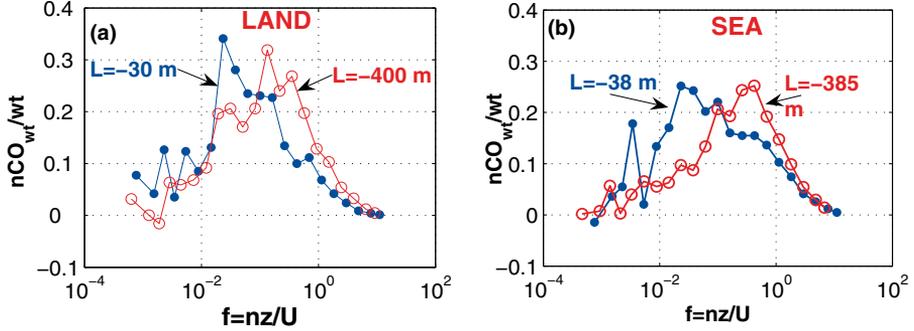


Figure 8 Examples of dimensionless $w-\theta$ cospectra as a function of normalized frequency. Filled circles represent unstable conditions, open circles represent UVCN-conditions. Subfigure a) shows measurements at a land site, b) shows measurements from the marine site Östergårnsholm.

A theoretical framework explaining the experimental results can be found in the studies of the very high Reynolds number boundary layers by Hunt and Morrison (2000) and Hunt and Carloti (2001). These theories have been confirmed by measurements in the neutral boundary layer by Höglström et al. (2002). In the neutral boundary layer detached eddies are created in the shear in the upper part of the surface layer. As these eddies move downward they become distorted due to blocking by the surface and stretching in the presence of the strong shear. In other words, the eddies are compressed and elongated. The horizontal length scale is of the boundary-layer height but the vertical length scale only approximately 1/30 of the boundary-layer height. The strong shear near the surface also initiates a replication of the eddies, a feature often called cat paws or honami waves. A series of photographs in Hunt and Morrison (2000) illustrates the effects on a sea surface due to these structures (their Figure 6). A sketch of the above described mechanism is shown in Höglström et al. (2002) in their Figure 1.

The theory of the detached eddies was linked with the horizontal roll structures in Paper II. It was suggested that they are two parts, one steady and one unsteady, of the same dynamical system. As the thermal forcing decreases, the system bifurcates into two branches, the horizontal rolls being the steady branch and the detached eddies the unsteady. The likelihood of formation of detached eddies increases with decreasing buoyancy force.

The observed high frequency peak of the spectra was interpreted as a signature of the detached eddies, thus the double peaked ('camel shaped') spectra and cospectra could be explained as a transition regime with both the steady and unsteady branch present.

In Papers III-IV the UVCN regime was observed to ensue for L values < -150 m. Expressed in bulk variables the limit is probably close to conditions with $U > 10$ m s^{-1} and an air-sea temperature difference, ΔT , smaller than 2

K. These are approximate limits, and in Paper II it is suggested that a correct threshold should instead be expressed in terms of h/L where h is the height of the boundary-layer. However, in the absence of continuous measurements of the boundary-layer height, no threshold value has been presented.

As shown in Figures 6 and 7, the UVCN regime enhances the air-sea heat fluxes. In Papers III and IV, the reason for the increase is linked to the detached eddies. As the eddies move downward they bring down drier and colder air as shown by the quadrant analysis, thus enhancing the surface fluxes. This is also in agreement with the results by Maitani and Ohtaki (1989) who observed that downdrafts of dry air were a more efficient mechanism for the upward humidity flux compared to updrafts of humid air.

In Paper II it is speculated that the intensification of downbursts during the UVCN conditions also enhances micro scale wave breaking. This would disrupt the diffusive sub-layers at the air-sea interface, thus reducing the resistance and enhancing the heat transfer. Since flux algorithms based on surface renewal theory do not include these processes, they underestimate the heat fluxes during neutral stratification as shown to be the case with COARE 3.0 in Papers III and IV.

Parameterization of the UVCN-regime

In Rutgersson et al. (2007) a parameterization was presented which incorporates the influence of the UVCN-regime. This parameterization is based on the experimental data from Papers III and IV. For unstable, non-UVCN conditions, the following is suggested:

$$C_{HNa} = \begin{cases} 1.0 \times 10^{-3} & 0 < U_{10} < 9.0 \text{ m s}^{-1} \\ (0.12U_{10} - 0.08) \times 10^{-3} & 9.0 < U_{10} < 14 \text{ m s}^{-1} \end{cases} \quad (25)$$

$$C_{EN} = \left(1.12 - \frac{1.5}{6 + (U_{10} - 4.5)^2} + 0.002U_{10} \right) \times 10^{-3} \quad 0 < U_{10} < 14 \text{ m s}^{-1} \quad (26)$$

For UVCN conditions the following expressions are used:

$$C_{HN} = \begin{cases} [(0.63 - 0.08\Delta T)(U_{10} - 9) + 0.45] \times 10^{-3} \\ \max(C_{HN}, C_{HNa}) \end{cases} \quad (27)$$

$$C_{EN} = \left[\left(0.31 - \frac{\Delta T^2}{45} \right) U_{10} + 0.2\Delta T^2 - 1.7 \right] \times 10^{-3} \quad (28)$$

The global heat fluxes

The strong increase of the exchange coefficient during near neutral conditions, as indicated in Figures 7 and 8, suggests that the UVCN regime possibly has a large influence when calculating the global heat fluxes. This was investigated in Paper V, where data from the European Centre for Medium-Range Forecasts (ECMWF) Re-Analyses (ERA-40) database for the years 1979-2001 were used together with the Rutgersson et al. (2007) parameterizations. A description of the ERA-40 data can be found in Uppala et al. (2005).

The Rutgersson et al. (2007) parameterization was based on our experimental data from Östergarnsholm, which only allowed the exchange coefficients to be evaluated for wind speeds up to 14 m s^{-1} . Thus, the conservative assumption was made of no further increase of C_{EN} and C_{HN} for higher wind speeds, i.e. U_{10} is set to 14 m s^{-1} in Equations (25)-(28).

In Figure 9 is shown the mean annual frequency of occurrence of UVCN conditions based on data every 6th hour for the years 1979-2001. Here, UVCN conditions is defined as wind speed above 10 m s^{-1} and ΔT smaller than 2 K.

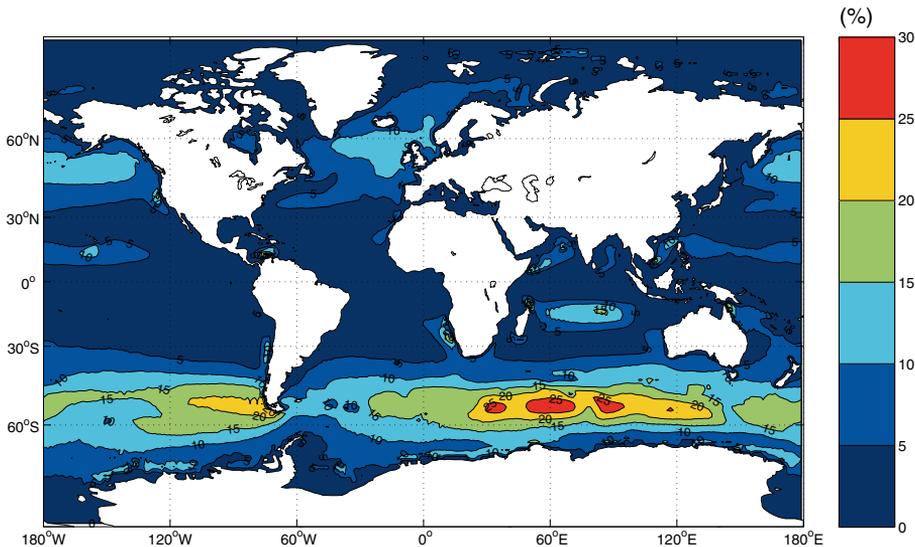


Figure 9 Annual relative frequency of occurrence of UVCN conditions, averaged over the 1979-2001 period.

UVCN conditions are most common in the southern hemisphere oceans, with maxima of around 30% of the time. Over the northern hemisphere, UVCN is not quite as frequent, reaching about 15% in the north Atlantic and north Pacific. UVCN conditions over the southern oceans are mainly con-

strained by the air-sea temperature difference since the mean wind speed is above 10 m s^{-1} the year around. The northern oceans have weaker mean wind speed, with a minimum during northern hemisphere summer. In addition, during winter, situations with cold air outbreaks over the oceans are more common in the northern hemisphere compared to the southern hemisphere due to larger land masses, which act as source regions for cold and dry air. These outbreaks create large temperature differences thus preventing formation of the UVCN regime.

The seasonal variation is large for some regions. In the Arabian Sea, UVCN conditions have a maximum of about 50% during June-August due to the intensive monsoon circulation.

The influence of the UVCN regime on the air-sea latent heat flux is shown in Figure 10, positive numbers indicate upwards fluxes. In the southern oceans, the latent heat flux enhancement range between $2.5\text{-}12.5 \text{ W m}^{-2}$. The increase is somewhat smaller over the northern oceans, reaching about 10 W m^{-2} in the north Atlantic. Due to the monsoon winds in the northern hemisphere summer, the Arabian Sea displays a maximum of over 20 W m^{-2} .

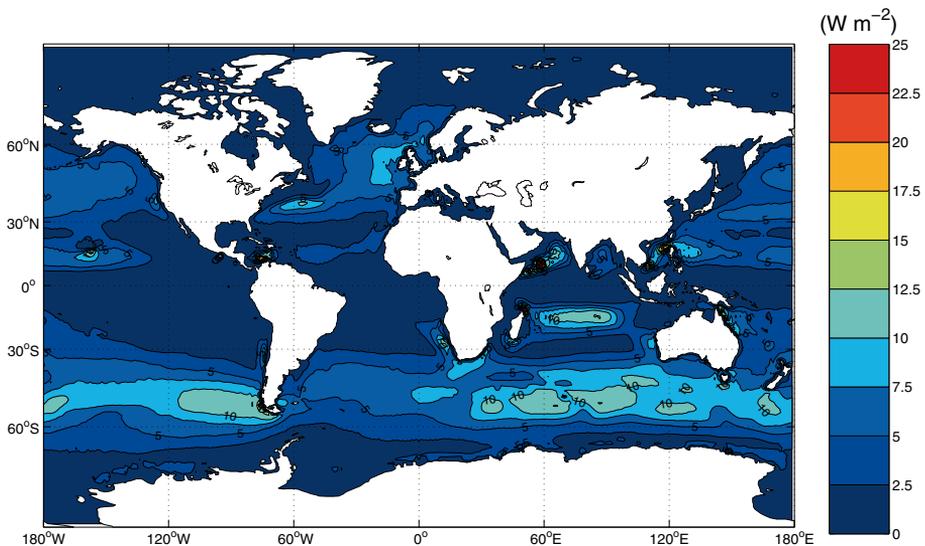


Figure 10 Mean annual increase of the latent heat flux due to the UVCN regime.

The increase of the sensible heat flux, shown in Figure 11, follows the same pattern as the increase of the latent heat flux. However, the magnitude of the increase is smaller, reaching about 5 W m^{-2} in the north Pacific and north Atlantic and 7 W m^{-2} in the southern oceans. This is due to the small air-sea temperature gradient in the near neutral boundary layer, which prevents strong sensible heat fluxes but does not constrain the latent heat flux.

The mean annual global increase of the sensible heat flux due to the UVCN regime is about 0.8 W m^{-2} , or a relative increase of 8%. The increase

of the latent heat flux is three times as large, 2.4 W m^{-2} . However, the relative increase is smaller compared to the sensible heat flux, 4.2%. If the fluxes are averaged over the global ice free oceans instead of averaging over the entire globe, the increase of the sensible heat flux is 1.2 W m^{-2} and the increase of the latent heat flux is 3.6 W m^{-2} .

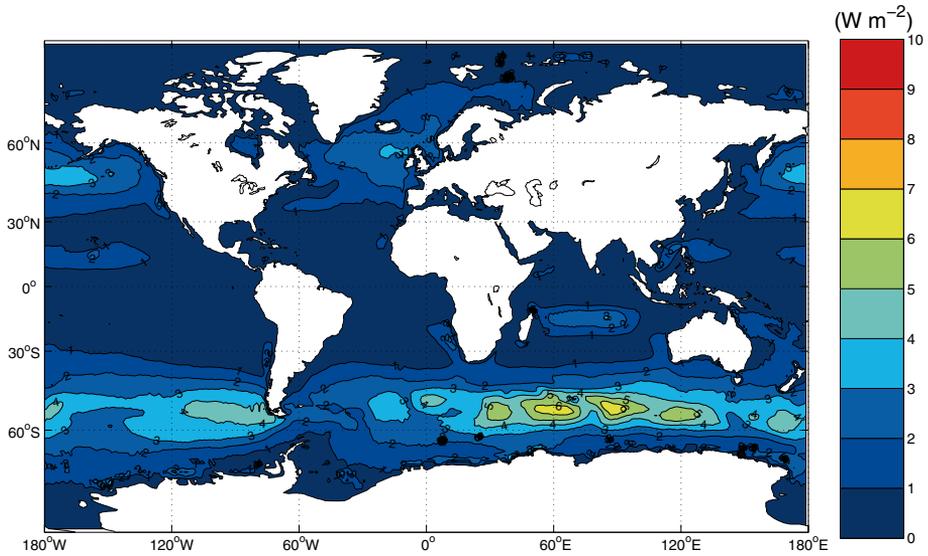


Figure 11 Mean annual increase of the sensible heat flux due to the UVCN regime

In a global context, the influence of the UVCN regime on the air-sea heat fluxes is quite large. It is of similar magnitude as the increase of the radiative forcing due to the anthropogenic emissions of greenhouse gases, which according to IPCC AR4 amounts to $2.63 \pm 0.26 \text{ W m}^{-2}$ (Forster et al., 2007). Thus the UVCN regime could potentially have significant influence on the predictions of the future climate.

However, the wind speed dependence of the exchange coefficients at wind speeds above 14 m s^{-1} is unknown, which adds some uncertainty to the estimations. In addition, feedback effects are not included in the calculations. It is expected that increased heat fluxes would lead to cooling of the sea surface, which in turn would constrain the heat fluxes and affect the mixing in the ocean. This has been shown to be the case when the UVCN effects over the Baltic Sea were studied using a regional climate model and an ocean model (Rutgersson et al., 2007). The boundary-layer over the Baltic Sea also got slightly more humid and slightly warmer.

Summary and conclusions

The marine atmospheric boundary layer was studied using tower based measurements from Östergarnsholm Island in the Baltic Sea. Result from an extensive field experiment showed that measurements made at this site indeed represents open ocean conditions for winds coming from the 90-190° sector. Small parts of the flux footprint may be over land during winds from 190-220°. Swells from 50-90° are affected by a shoal close to the island. Measurements during these situations are said to represent marine conditions.

The air-sea flux of sensible heat was found to increase with increasing wind speed and decreasing air-sea temperature difference, ΔT . The latent heat flux was found to increase with wind speed only for cases with small ΔT . These results can not be explained by traditional theory. From spectral and quadrant analyses it was found that during winds above 10 m s^{-1} and $\Delta T < 2 \text{ K}$ the turbulence in the boundary layer was reorganized. The horizontal roll structures often present in the moderately unstable boundary layer breaks down as the thermal forcing decreases. Instead, the transfer processes is dominated by smaller scale turbulence, which seem to originate as detached eddies from layers above. It was suggested that they are two parts of the same dynamical system; one stable part, the rolls, and one unstable, the detached eddies. The turbulence structure dominated by the detached eddies was termed the Unstable Very Close to Neutral regime, the UVCN-regime. The UVCN-regime was observed to exist also over land i.e. it is not dependent on the nature of the underlying surface.

Quadrant analysis revealed that the enhancement of the surface heat fluxes was due to transport of dry and cold air from above. This is most likely a result of the detached eddies. Another source of enhancement could possibly be from the intensive downbursts present during UVCN conditions. It is speculated that these would increase micro scale wave breaking. This would disrupt the thin diffusive sublayer at the air-sea interface thereby reducing the resistance and enhancing the heat transport.

Using ERA-40 reanalysis data the global impact of the UVCN regime was studied for the year 1979-2001. It was found that the global sensible heat flux was enhanced by 0.8 W m^{-2} and the latent heat flux was enhanced by 2.4 W m^{-2} . Thus, the UVCN regime is an important physical process on a global scale, which could be of large importance when predicting the future weather and climate.

It was found that the transfer coefficients for sensible and latent heat used in the flux parameterizations are indirectly affected by the sea state. During swell C_{EN} is reduced by 10% compared to growing sea conditions when using a wave dependent ϕ_m -function in the calculation. A similar result was obtained for C_{HN} .

A method for correcting density measurements made with an open path instrument was evaluated. Converting the high frequency density measurements to a time series of mixing ratio was shown to be equivalent to the Webb-correction. Using this so called DC-method makes it possible to calculate corrected spectra and cospectra from the density measurements.

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Summary in Swedish

Förståelsen av de transportprocesser som styr avdunstningen och värmeutbytet är av yttersta vikt då man vill ha bra förutsägelser av det framtida vädret och klimatet. Eftersom haven täcker 70 % av jordytan och är den primära källan till atmosfärens vattenånga samt fungerar som en gigantisk värmere-servoar är det av speciell betydelse att förstå samspelet mellan haven och atmosfären.

Turbulensen i gränsskiktet, det lägsta skiktet av atmosfären, är den mekanism som står för det största direkta utbytet av värme och vattenånga mellan jordytan och atmosfären. De turbulenta virvlarnas storlek spänner över en mängd av skalor; från små virvlar av millimeterstorlek upp till virvlar som täcker hela gränsskiktet dvs. ca 1000 m. Energitransporten sker främst med de största turbulenselementen men även dessa är för små för att kunna simuleras explicit i de globala klimatmodellerna. För att beskriva den turbulenta transporten måste man därför göra förenklingar. Grunden till förenklingarna baseras på mätningar av de turbulenta flödena.

Över land har de turbulenta flödena kunnat studeras noggrant under det senaste halvsekle, över hav däremot så är det betydligt svårare att mäta, varför informationen fortfarande är ofullständig. T.ex. kan saltpartiklar från havet skada eller förstöra instrumenten. Dessa saltpartiklar påverkar främst instrument som använder sig av s.k. ”hot film” eller ”hot wire” teknik. Man kan därmed undvika instrumentskador om ljudanemometrar används istället. Dock har denna studie visat att ljudanemometrar ger felaktiga mätningar vid vindar över 10 m s^{-1} vilket komplicerar proceduren.

Att mäta turbulensen i det marina gränsskiktet från en mast på land är ett fördelaktigt alternativ. Vågornas rörelse påverkar inte masten och genom att montera instrumenten på långa bommar minimeras störningar från masten. Ytterligare en fördel med mastmätningar är att dessa kan pågå under lång tid, flera år. Information kan då samlas in under en stor mängd förhållanden. Detta är inte möjligt med mätningar ifrån skepp då dessa är ute under en begränsad period. I det här arbetet har mätningar ifrån en sådan fast installation använts. Masten är placerad på ön Östergarnsholm, en liten låg ö ca 4 km öster om Gotland. Östersjöns salthalt är låg vilket underlättar mätningarna eftersom man minimerar risken att instrumenten skadas av saltpartiklar.

Av avgörande betydelse är då att mastmätningarna verkligen är representativa för de förhållanden som finns över de öppna haven. Just detta studerades vid Östergarnsholm i ett stort experiment 2003. Finska marininstitutet

deltog med ett forskningsfartyg, R/V Aranda, samt två bojar för mätningar av vågor och vattentemperatur. Universitet i Miami deltog med en boj för vågmätningar och även instrumenterad för atmosfäriska mätningar bl.a. av turbulensen nära havsytan. Resultatet av detta experiment visade att mätningarna gjorda från Östergarnsholmsmasten till största del verkligen är samma som om de skulle ha varit gjorda över det öppna havet. Vissa begränsningar finns dock då vinden kommer från sektorer nära kusten, eller då långa vågor har passerat ett grund i närheten.

Genom att använda Östergarnsholmsmätningarna kunde de turbulenta flödena av sensibelt och latent värme mellan havet och atmosfären studeras. Vi har upptäckt att de utbyteskoefficienter som används vid parameterisering av det sensibla och latent värmeflödet i klimat- och vädermodeller är funktioner av både vindhastighet och temperaturskillnaden mellan luften och vattnet. Vid små temperaturskillnader blir värmeutbytet mer effektivt vid vindstyrkor över 10 m s^{-1} än vad det är för samma vindstyrkor men stora temperaturskillnader.

Genom att använda en flödesmodell som inkluderar effekter av brytande vågor undersöktes det om dessa kunde vara orsaken till den observerade ökningen av värmeflödena. Det visade sig att brytande vågor bara kunde förklara ca 1/5 av ökningen.

Effektivitetshöjningen av värmetransporten är istället en dynamisk effekt. Genom att i detalj studera turbulensen med hjälp av spektra och kvadrantanalys upptäcktes det att de turbulenta virvlarna organiserar sig på ett annat sätt vid höga vindar och små temperaturskillnader mellan havet och atmosfären. Normalt organiseras de största turbulenta virvlarna i det instabila gränsskiktet vid måttliga till friska vindar i långa horisontella rullar. På så sätt samverkar de och ger upphov till band med uppåtstigande respektive nedåtsjunkande luft. Det tydligaste kännetecknet för detta är de s.k. molngatorna. Då temperaturskillnaden minskar och gränsskiktet närmar sig neutral skiktning bryts denna struktur successivt ner. Istället bildas virvlar i övre skikt som, då de närmar sig ytan, transporterar ner kallare och torrare luft och på så sätt förstärker flödena. Denna mekanism, som kallas "Unstable Very Close to Neutral – regime" eller UVCN-regimen, har även observerats över land och är alltså generell för det atmosfäriska gränsskiktet.

UVCN-regimens påverkan på de globala värmeflödena beräknades vara 2.4 W m^{-2} för det latent värmeflödet och 0.8 W m^{-2} för det sensibla värme-flödet. Detta är i paritet med förstärkningen av växthuseffekten pga. antropogena utsläpp av växthusgaser som enligt FNs klimatpanels senaste rapport idag beräknas vara 2.6 W m^{-2} . Förekomsten av UVCN-regimen är olika för olika regioner och årstider. Vid beräkningar av det framtida regionala och globala klimatet samt vid väderprognosberäkningar kan det därför vara av stor vikt att inkludera dessa effekter på värmeflödena.

Vågor kan också indirekt påverka transporten av värme och vattenånga. Då långa vågor som färdas snabbar vinden, dymning, dominerar havsytan på-

verkas vindprofilen. Detta påverkar även värmeflödena indirekt eftersom den normaliserade vindgradienten används i parameteriseringarna. Det latent värmeflödet reduceras med ca 10 % vid dymning jämfört med uppbyggande sjö. Vågorna reducerar det sensibla värmeflödet med liknande storleksordning.

I avhandlingen har även en metod för korrigerings av snabba densitetsmätningar utvärderats. Genom att den uppmätta densiteten direkt omvandlas till blandningsförhållande relativt torr luft behövs ingen korrigerings i efterhand som man traditionellt använder. Dessutom blir spektra, kvadrantanalyser, etc. samt alla turbulenta moment där densitetsmätningen ingår korrigerade; inga nya ekvationer behöver härledas.

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