Air-Sea Exchange of Momentum and Sensible Heat Over the Baltic Sea

BY

XIAOLI GUO LARSEN
Contents

1 Introduction ..............................................................................................................1

2 Measuring site and measurements ........................................................................3
  2.1 The sites ........................................................................................................3
  2.2 Instrumentation at Östergarnsholm ..............................................................4
  2.3 Wave data .....................................................................................................5

3 Boundary Layers .....................................................................................................6
  3.1 Atmospheric constant-flux surface layer ..................................................7
  3.2 Oceanic boundary layer .............................................................................8
      Life cycle of wind-induced waves ............................................................8
      Wave spectrum and wave parameters ...................................................8
      Statistics of the wave field in the Baltic Sea ....................................10
  3.3 Wave boundary layer ..............................................................................11
      Vertical distribution of stress .................................................................11
      Wind profiles .........................................................................................12
      Temperature profiles .............................................................................14
      Wave boundary layer height $h_w$ ..............................................................15

4 Methods of obtaining turbulence fluxes ...............................................................16

5 Wave impact on drag ............................................................................................18
  5.1 Pure wind-sea ............................................................................................18
  5.2 Following mixed sea and following swell ..............................................20
  5.3 Cross mixed sea and cross swell .............................................................22
  5.4 Non-stationary wind field .........................................................................24

6 Sensible heat transfer .........................................................................................26

7 Summary and conclusions ...............................................................................28

8 Acknowledgement ...............................................................................................30

9 Reference: ............................................................................................................31
1 Introduction

The interaction between atmosphere and ocean has substantial influence on global weather and climate. More than 70% of the earth surface is covered by water. Oceans absorb more solar radiation than land does, thus heat the atmosphere from below. Accompanying the earth’s rotation, zonal differences in heating drive large-scale atmospheric circulation, which in turn has significant impact on the oceanic currents and waves. As a coupled system of exchanging momentum and energy, surface waves of the oceans will inevitably influence the structure of the marine atmospheric boundary layer (MABL), thus making it different from that of the ABL over land.

In modeling atmospheric as well as oceanic movements of different scales (weather prediction, climate simulation, ocean circulation, wave growth etc.), it is important to determine the boundary conditions at the air-water interface. The surface turbulence fluxes are the key boundary parameters.

Surface fluxes can be estimated with various methods; eddy correlation method, profile method, inertial dissipation method and bulk method are the four fundamental methods. Except for the first one, the other three are all based on similarity theories that have been developed under certain restricted conditions over land. Over the wavy surface, the classical similarity theories have been found to be questionable when the surface is dominated by swell (Volkov 1970; Smedman et al. 1994, 1999; Mahrt et al. 1998 and Drennan et al. 2002). This indicates that the impact from swell has to be taken into account if one wants to use methods based on similarity theories.

Measuring the turbulence fluxes in open ocean with any of the eddy correlation-, profile- or inertial dissipation method is both costly and technically difficult. The bulk method therefore appears practical and attractive since it needs only mean variables at two levels and determination of the exchange coefficients. Studying how the exchange coefficients are influenced by waves is hence important and it is the focus of this study.

Long term semi-continuous meteorological measurements from an air-sea interaction station, Östergarnsholm, and wave measurements from a wave buoy in the Baltic Sea are used to study the impact of surface waves on the turbulence statistics in the marine atmospheric surface layer (MASL). The climatological characteristics of the Baltic Sea are investigated (in Paper IV). The vertical variation of mean wind speed and temperature in the MASL is
explored as a function of the surface waves. The validation of similarity theory for the surface layer is studied for long wave conditions (Paper I & II).

With the aid of measurements with the eddy correlation method, the exchange of momentum is studied for different wave conditions, namely, wind-sea, following-swell/mixed sea and cross-swell/mixed sea conditions (Paper I & II). In addition, in Paper II, a wind-over-wave coupled model of Makin and Kudryavtsev (2002) (hereinafter MK2002) is used to explore the impact of swell on the sea drag. The model assumes that the mechanism of momentum exchange is the same for pure wind-sea and swell. The discrepancies will clearly reflect the difference in impact from wind sea and from swell.

The air-sea exchange of sensible heat is investigated with a comprehensive data set (Paper III).

In this summary, sites and measurements will be introduced in Section 2. Characteristics of different boundary layers are given in Section 3. Different methods of obtaining surface fluxes are introduced in Section 4. Wave impact on exchange coefficients for momentum and sensible heat are analyzed in Sections 5 and 6, respectively. Summary and conclusions are given in Section 7.
2 Sites and measurements

2.1 The sites

Figure 1. Map of the Baltic Sea (upper left), with a close-up of Östergarnsholm. Dashed lines are contours of water depth. Solid lines on Östergarnsholm are contours of height. (From Johansson 2003)

Östergarnsholm is an island situated about 4 km east of the main island of Gotland in the Baltic Sea (Figure 1). It is small (~ 2×1 km²), low and flat with no trees. The peninsula in the southeast part of Östergarnsholm is about 1 km long and no more than a couple of meters above the mean sea level. At the southernmost tip of the peninsula there is a 30 m high tower. The distance from the tower to the shoreline is some tens of meters in the sector
80-220°. The tower base is situated about 1 m above the mean sea level. The actual heights of measurements are corrected with the aid of water level measurements at Visby harbor, situated at the western coast of the island of Gotland (Sjöblom and Smedman, 2002).

Östergarnsholm is exposed to open-ocean conditions when winds are coming from an easterly or southerly direction (about 80°-220°). The undisturbed water fetch is over 150 km in this sector. The slope of the sea floor right outside the peninsula is approximately 1:30 down to a depth of 19 m, and 1:17 further out to the south and the ratio becomes even smaller further out. The possible influence of limited water depth on the tower measurements has been carefully studied in Smedman et al. (1999).

In the study of sensible heat transfer, some data are obtained from another site Nässkär. It is a flat rock with no vegetation, situated in the outer part of the Stockholm archipelago. It is about 20×40 m² with a maximum height 2 m. More details about this site can be found in Paper III.

2.2 Instrumentation at Östergarnsholm

The tower is instrumented with slow response sensors (with a sampling frequency of 1 Hz) of in-house design for temperature (Högström, 1988) as well as for wind speed and direction (Lundin et al., 1990) at 5 levels (7, 12, 14, 20 and 29 m above the tower base). The wind speed and direction sensor is a combination of a small, lightweight cup anemometer and a styrofoam wind vane. The accuracy is 0.2 m s⁻¹ with no over-speeding (Lundin et al., 1990). Air temperature is measured with 500 Ω resistance platinum sensors in aspirated radiation shields. An additional sensor is placed at the lowest level for absolute temperature. The accuracy is 0.02 K (Högström 1988). At the first level, relative humidity is measured with the Rotronic sensor, with uncertainty ±2%.

Turbulent fluctuations are measured with SOLENT 1012R2 sonic anemometers (Gill Instruments, Lymington, UK) at three levels (9, 17 and 25 m above the tower base). From the sonic signals the three orthogonal components of the wind (u, v, w) and the so-called ‘sonic temperature’ are obtained. The ‘sonic temperature’ is very close to the virtual temperature. Turbulence signals are sampled at 20 Hz. In order to remove possible trends, a high-pass filter based on a 10-min running average was applied to the turbulence time series before calculating the variances and co-variances. This is equivalent to using a high-pass filter with a low cutoff frequency of about 10⁻³ Hz. On special occasions, the cutoff frequency needs to be determined from the co-spectra.
The virtual sensible heat flux ($\overline{w'\theta'}$) measured by the sonics has been corrected for crosswind contamination in the way described by Kaimal and Gaynor (1991).

During some periods the MIUU (Meteorological Institute, Uppsala University) turbulence instrument was also used for measurements of momentum, sensible and latent heat (Paper III). The MIUU-instrument is basically a wind-vane-mounted hot-film instrument with additional platinum sensors for dry and wet bulb temperature. There is a negligible systematic effect from flow distortion, sensor separation and from other possible sources on the fluxes of momentum and sensible heat (Högström, 1988). The MIUU-instrument is thus regarded as a reference instrument (Högström and Smedman, 2003).

2.3 Wave data

Together with the meteorological measurements, wave measurements have been taken continuously except for breaks in wintertime to avoid possible ice damage. Wave data are recorded with a Directional Wave-Rider Buoy, which is owned and run by the Finnish Institute for Marine Research. The buoy is moored at a water depth of 36 m about 4 km southeast of the tower in the direction of 115°. The position represents the wave conditions in the ‘footprint area’ (Smedman et al. 1999). Wave data are recorded every hour as a 1600 s time series on board the buoy. A directional spectrum is calculated from the time series. The spectrum contains 64 frequency bands ranging from 0.025 to 0.58 Hz, with the peak frequency determined by a parabolic fit. The significant wave height $H_s$ is calculated by trapezoid method from 0.05 to 0.58 Hz.

In many studies, the bucket sea surface temperature, $T_w$, measured by the wave buoy at a depth of 0.5 m is used. The difference between $T_w$ and the real sea surface temperature $\theta_s$ is related to the heat transfer in the limited water layer (here $z = 0$ to –0.5 m). Rutgersson et al. (2001) applied the procedure of Fairall et al. (1996a) and found that the difference is on average 0.15 K. The accuracy of $T_w$ is 0.1 K.
3 Boundary Layers

In the atmosphere-ocean interacting system different boundary layers could be defined. Figure 2 is a general scheme of the system (similar to that suggested by Chalikov and Belevich, 1993). The system contains the oceanic boundary layer and marine atmospheric boundary layer (MABL). The MABL usually includes a surface micro-layer, a wave boundary layer (WBL), a constant-flux surface layer or Monin-Obukhov boundary layer (MOBL), a matching layer and an outer boundary layer. Mahrt et al. (1998) show that the positions of different boundary layers vary with both atmospheric and oceanic forces (Figure 2). With 10 m as a reference height, the first example in Figure 2 (denoted ‘Classical’) shows the ideal case where Monin-Obukhov Similarity Theory (MOST) applies. For Case II, the reference level is in the WBL and scales of the waves are relevant parameters. For Case III, the reference level is above the MOBL and bulk boundary layer scaling is required. For Case IV, the influence of boundary layer height extends down to the wave boundary layer.

The Baltic Sea region provides atmospheric phenomena of wide scales, and the four cases given in Figure 2 can all be found. The present study concentrates on studying the influence of waves in the surface layer, i.e. Classical and Case II.

Figure 2. Idealized layering of the boundary layer, where $z$ is the height above the mean sea level, $\lambda$ is wavelength parameter and $L$ is the Obukhov length, see next Section. $\delta$, $h_\nu$, $h_M$ and $z_i$ are the heights of the surface micro-layer, WBL, MOBL and Outer layer respectively (After Mahrt et al. 1998, slightly modified).
3.1 Atmospheric constant-flux surface layer

In a Monin-Obukhov boundary layer, turbulence fluxes are approximately constant with height; the normalized flux-gradients for wind and temperature \( (\phi_u \text{ and } \phi_h) \) are independent of wave state and they are universal functions of the stability parameter \( \zeta = z/L \), where \( L \) is the Obukhov length scale:

\[
L = \frac{-u^* T_0}{\kappa g w \theta_s}
\]

where \( g \) is the gravitational acceleration, \( T_0 \) is the reference absolute temperature in K, \( \kappa \) is the von-kármán constant, \( u^* \) is the friction velocity and it is calculated in terms of the downstream momentum flux \( -\overline{u'w'} \) and the cross-wind momentum flux \( -\overline{v'w'} \):

\[
u_* = \left\{ (-u^* w)^2 + (-v^* w)^2 \right\}^{1/4}
\]

This is the essence of MOST. If MOST is valid then \( \overline{v'w'} \) is negligible. The normalized flux-gradients \( \phi_u \) and \( \phi_h \) are defined as

\[
\phi_u(z / L) = (\kappa z / u_*) \left( \partial U / \partial z \right)
\]

\[
\phi_h(z / L) = (\kappa z / T_*) \left( \partial \theta / \partial z \right)
\]

where \( T_* = -\overline{w' \theta' / u^*} \) is the temperature scale. Assuming the turbulent fluxes to be independent of height, the mean quantities at height \( z \) could be obtained by integrating the profiles (3.1-3.2) from \( z_0 \) \((z_0 T)\) to \( z \):

\[
U(z) - U_s = (u^* / \kappa) \left[ \ln(z / z_0) - \psi_u \right]
\]

\[
\theta(z) - \theta_s = (T^* / \kappa) \left[ \ln(z / z_{0T}) - \psi_h \right]
\]

where \( z_0 \) and \( z_{0T} \) are the roughness lengths for momentum and sensible heat, at which the extrapolated wind speed and temperature profiles approach their surface values \( U_s \) and \( \theta_s \). The surface drift \( U_s \) does not exceed centimeters per second even in gale conditions and it is set to be 0 in the present study. \( \psi_u \) and \( \psi_h \) are the integrated analytical forms of \( \phi_u \) and \( \phi_h \):

\[
\psi(z / L) = \int_{z_0 / L}^{z / L} \left[ 1 - \phi(\zeta) \right] / \zeta \cdot d\zeta
\]

The most common expressions of \( \phi \)-functions in the interval of \(-2<z/L<1\) are empirical equations based on measurements over land (Businger et al. 1971; Dyer 1974; Högström 1996 etc):
\[ \phi_{\alpha} = 1 + C_1 z / L \quad \phi_{b} = 1 + C_2 z / L \quad z/L > 0 \quad (6.1), \]

\[ \phi_{\alpha} = (1 - C_3 z / L)^{1/4} \quad \phi_{b} = C_4 (1 - C_4 z / L)^{1/2} \quad z/L < 0 \quad (6.2), \]

Different values of the constants have been reported: \( C_1 \in [4.7, 7] \), \( C_2 \in [4.7, 8] \), \( C_3 \in [15, 19] \), \( C_4 \in [9, 16] \) and \( C_5 \in [0.74, 1] \). For ‘very stable conditions’, Holtslag and De Bruin (1988) suggested an extended stability function:

\[ \phi_{\alpha} = \phi_{b} = 1 + a z / L + (1 - c_1 - d_1 z / L) \cdot z / L \cdot \exp(-d_1 z / L), \quad (6.3) \]

where \( a_1 = 0.7 \), \( b_1 = 0.75 \), \( c_1 = 5 \) and \( d_1 = 0.35 \). For free convection condition (\( z/L \to -\infty \)), Godfrey and Beljaars (1991) added to the mean wind speed a gustiness factor, which is related to the height of boundary layer (\( z_i \)).

Over land, usually the bottom 10% of the boundary layer is defined as the ‘constant-flux surface layer’ and the reference height of 10 m is usually within this layer (e.g. Stull 1988), see also Figure 2, ‘classical’. But it is not necessarily the case in a MABL.

### 3.2 Oceanic boundary layer

**Life cycle of wind-induced waves**

When wind begins to blow over still water, the whirls in the flow cause pressure differences that break the still water surface and capillary waves appear. The capillary waves develop into wind waves when they become larger than 1.73 cm. The most important factors that control the wave growth are wind speed, wind duration and fetch. Other factors such as geometry and water-basin depth also have their impact on longer waves (e.g. Mitsuyasu, 1982). When wind calms down and becomes slower than the wave propagation speed, it can no longer feed energy to the waves. Long waves that do not scale with the local wind are called swell, which include also those waves propagating from distant storms.

**Wave spectrum and wave parameters**

As in the ABL, motions in the oceanic boundary layer are often described in terms of a spectrum of eddies. Figure 3 is a scheme given by Kitaigorodskii (1962) who divided the wave spectrum into various frequency bands, each with different controlling factors, in analogy to the atmospheric spectrum of turbulence.
Figure 3. Schematic representation of a spectrum of wind waves. Energy density $S(n)$ (m$^2$Hz) is plotted as a function of frequency $n$ (Hz). The spectrum is divided into seven frequency bands (From Kitaigorodskii 1962).

Under the restrictive conditions of stationarity and homogeneity with clearly defined fetch it is expected that the similarity of the spectrum shape can be described by a small set of parameters. The wave spectrum is usually approximated by simple parameters such as the so-called wave age parameters:

$$\frac{c_p}{u^*} \quad \text{or} \quad \frac{c_p}{U_c},$$

(7)

where $U_c$ is the mean wind speed (usually at 10m) component in the wave propagation direction. $U_c = U_{10} \cos \alpha$, with $\alpha$ the direction difference between wind and dominant waves. $c_p$ is the wave phase velocity of the spectral peak.

Sufficiently long waves ‘feel’ the presence of the bottom, implying the calculation of $c_p$ should follow the wave dispersion relation

$$c_p = \left( \frac{g}{k_p} \tanh(k_p h) \right)^{1/2},$$

(8.1)

where $k_p$ is the wave number of the spectral peak and $h$ is the water depth.

To quantify the influence of the shallow water, a weighted mean phase velocity is defined by using the weighting function $F(x, z)$ in Smedman et al. (1999):

$$\langle c_p \rangle = \int_0^\infty F(x, z)c_p(x)dx$$

(8.2)

In the present study, $c_p$ has been calculated with Eq. (8.2).
Pierson and Moskowitz (1964) suggested that full wave development corresponds to wave age \( cp/U_* = 1.2 \). Typically young wind sea corresponds to a wave age \( cp/U_* \) of the order 10, while old wind sea corresponds to an order 25 (Komen et al. 1994). Using \( cp \) implies that the wave spectrum has a consistent shape for a given wave age (Donelan et al. 1993), and of course there should be only one peak in the spectrum.

In Paper I, a new wave parameter \( E_1/E_2 \) is derived. The 1-D wave spectrum is divided into two parts

\[
E_1 = \int_{n_1}^{n} S(n)dn \quad \quad \quad E_2 = \int_{n}^{\infty} S(n)dn \quad \quad \quad (9)
\]

at \( n_1 = g/(2\pi U_*) \). At this frequency the phase velocity \( c \) for deep water equals the wind speed component in the wave propagation direction at 10 m. The separation thus results in a spectral part \( E_1 \) where waves propagate faster than wind, and a short wave spectral part \( E_2 \) where waves are generated by the wind. See figure 2 in Paper I for an example. When plotted against \( U_{10}^2 \), \( E_1 \) does not show any dependence on the local wind, while there is a strong linear relation between \( E_2 \) and \( U_{10}^2 \) for \( U_{10} > 3 \) m s\(^{-1} \). This suggests that it is reasonable to refer to \( E_2 \) as the wind sea part of the spectrum. The \( E_1/E_2 \) parameter does not require a consistent spectral form and is not sensitive to the numbers of spectral peaks. This analysis method is a modified version of that proposed by Dobson et al. (1994). \( E_1/E_2 \) is used to classify data according to wave state and the details are given in Table I.

Statistics of the wave field in the Baltic Sea Proper

The wave field in the Baltic Sea proper is closely related to the wind field (Paper IV). 44% of the time wind is from the sector 80 – 220\(^\circ\). Waves passing the buoy are mostly from the sector of south to southwest. The highest waves are also from this sector. There are almost no waves from the direction of Gotland. Waves from the north are mostly pure wind sea wave.

Based on the definition of ‘single-peak’ in Paper IV, 69% of the cases (for wind from 80 – 220\(^\circ\)) are single-peak. For cases with wind from 80 – 220\(^\circ\), the frequency distribution of the angle between the wind direction and dominant wave direction, \( \alpha \), is given in Table II. In Paper II, \( \alpha < 30^\circ \) is used for definition of ‘wave following wind’, while \( 30^\circ < \alpha < 90^\circ \) is used for ‘wind-cross-wave’. When \( \alpha \) is larger than 90\(^\circ\), it can be classified as ‘wind-against-wave’. The frequency distribution of different wave age parameters in the intervals of common interest is presented in Table III.

Spray effect accompanying wave breaking becomes significant when wind speed at 10 m is larger than \(-10 \) m s\(^{-1} \). For wind in the sector of 80 – 220\(^\circ\), for 17% of the time, \( U_{10} > 10 \) m s\(^{-1} \).
Table I: Classification of data according to $E_1/E_2$.

<table>
<thead>
<tr>
<th>Interval of $E_1/E_2$</th>
<th>Wave state</th>
</tr>
</thead>
<tbody>
<tr>
<td>$E_1/E_2 &lt; 0.2$</td>
<td>Pure wind sea</td>
</tr>
<tr>
<td>$0.2 \leq E_1/E_2 &lt; 4$</td>
<td>Mature or mixed sea *</td>
</tr>
<tr>
<td>$E_1/E_2 \geq 4$</td>
<td>Swell *</td>
</tr>
</tbody>
</table>

* In Paper II, it is also required for swell-dominant cases that $n_p < n_1$.

Table II. Frequency distribution of wave-wind direction difference $\alpha$.

<table>
<thead>
<tr>
<th>$\alpha$</th>
<th>0-20°</th>
<th>20-30°</th>
<th>30-40°</th>
<th>40-90°</th>
<th>90-180°</th>
</tr>
</thead>
<tbody>
<tr>
<td>Freq.</td>
<td>45%</td>
<td>15%</td>
<td>11%</td>
<td>20%</td>
<td>9%</td>
</tr>
</tbody>
</table>

Table III. Frequency distribution of different wave age parameters in the intervals of common interest.

<table>
<thead>
<tr>
<th>$c_p/U_c$ / $u_*$</th>
<th>Freq.</th>
<th>$c_p/u_<em>$ / $u_</em>$</th>
<th>Freq.</th>
<th>$E_1/E_2$ / $E_2$</th>
<th>Freq.</th>
</tr>
</thead>
<tbody>
<tr>
<td>$c_p/U_c &lt; 0.8$</td>
<td>16%</td>
<td>$c_p/u_* &lt; 20$</td>
<td>12%</td>
<td>$E_1/E_2 &lt; 0.2$</td>
<td>20%</td>
</tr>
<tr>
<td>$0.8 \leq c_p/U_c &lt; 1.2$</td>
<td>34%</td>
<td>$20 \leq c_p/u_* &lt; 25$</td>
<td>17%</td>
<td>$0.2 \leq E_1/E_2 &lt; 4$</td>
<td>47%</td>
</tr>
<tr>
<td>$c_p/U_c \geq 1.2$</td>
<td>50%</td>
<td>$c_p/u_* \geq 25$</td>
<td>71%</td>
<td>$E_1/E_2 \geq 4$</td>
<td>33%</td>
</tr>
</tbody>
</table>

3.3 Wave boundary layer

Vertical distribution of stress

Under the assumptions of horizontally homogeneous and stationary flow with no subsidence, the momentum equation results in the conservation of the stress vector in the surface layer:

$$\frac{\partial \vec{\tau}}{\partial z} = 0$$

(10)

where the cap of arrow means a vector. Eq. (10) is the characteristic condition of MOST for land condition.

Theory (e.g. Steward 1974) and laboratory measurements (e.g. Anisimova et al. 1974) suggest that in the WBL, there is a pressure field induced by the waves in addition to the purely turbulent pressure field. Thus, over waves due to the wave-induced components of velocity and pressure, the total stress contains a wave-induced part ($\vec{\tau}_w$) in addition to the viscous molecular ($\vec{\tau}_v$) and the turbulent ($\vec{\tau}_t$) parts:

$$\vec{\tau} = \vec{\tau}_v + \vec{\tau}_t + \vec{\tau}_w$$

(11)
The viscous stress ($\nu \partial U/\partial z$, where $\nu$ is the kinematic viscosity) is important only in the surface micro-layer ($z<\delta_v$, $\delta_v$ is the height of the viscous sublayer), which is the lowest millimeter above the surface.

Across the interface between the atmosphere and ocean, the momentum transfer is accomplished by pressure/wave slope correlation ($p \partial \eta / \partial z$, where $\eta$ is the surface elevation, $p_0$ is the pressure at the surface), which, according to the continuity of stress, equals the wave-induced stress in magnitude. $\tau_w$ is significant in the WBL ($z<h_w$, $h_w$ is the wave boundary layer height) but its magnitude decreases with height and becomes negligible for $z>h_w$. Consequently, in the WBL, the turbulent stress increases with height and becomes equal to the total stress at $h_w$.

Wind profiles

For neutral stratification the logarithmic wind law is valid over land. Over water during wind sea conditions, the wave-induced stress $\tau_w$ is a significant portion of the total stress $\tau$ only in the lowest meter (Belcher and Hunt, 1994). Above it, $\tau=\tau_t$ and the log-wind law is expected to be valid. Measurements by Drennan et al. (1999a) at 2 m above the mean water level indicate no measurable effect from waves during pure wind sea conditions.

As shown in Paper I, during neutral conditions, when $E_1/E_2 \to 0$ (very young waves), $z_0$ calculated from $U_{10}$ and $u^*$ at one level, i.e. Eq. (4.1), is very close to $z_0$ obtained by extending the $U$-profiles to $U_i=0$. This strongly suggests that during pure wind-sea conditions, the wind profile ($\sim 8 – 30$ m) is logarithmic and the influence of wave-induced stress is small. This is also supported by the climatological study, see the profile for $E_1/E_2<0.2$ in Figure 4 (from Paper IV).

The stress anomalies (including negative stress) obtained during swell conditions at a level around 10 m indicate that $\tau_w$ is a considerable portion of $\tau$ (e.g. Smedman et al. 1994, 1999; Drennan et al. 1999a; Grachev et al. 2001, 2003). The long-term measurements used in Paper IV show that for 3% of data, $u^* \bar{w}>0$, and 96% of these positive flux have wave age $c_p/U_c>1.2$ (swell). With momentum flux directed from waves to the air, the airflow is accelerated. The so-called ‘wave-driven-wind’ has been found by e.g. Harris (1966), Davidson and Frank (1973), Holland et al. (1981) and Smedman et al. (1999).

Paper I shows that when $E_1/E_2$ increases, $z_0$ calculated with Eq. (4.1) deviates from that obtained from the $U$-profile, indicating that the profile deviates from being logarithmic. Both the case study (Paper I) and the climatological study (Paper IV) show the variation of the wind profile shape with the wave state. Figure 4 shows that, on average, the profile during swell conditions contains three layers, an uppermost normal layer (corresponding
to a normal magnitude of $z_0$ derived from the profile), the lower layer modified by swell (a much lower $z_0$ from the profile) and a transition layer in between (a much larger value of $z_0$ from the profile).

Figure 4. Mean wind profiles for near neutral stability, wind from 80 – 220° and $\alpha<40^\circ$, averaged for categories of $E_1/E_2$.

The normalized wind gradient $\phi_m$ (Eq. (3.1)) is found to be dependent on the wave parameter $E_1/E_2$, in addition to the stability parameter $z/L$ (Paper III). The result is reproduced in Figure 5. For $z/L<0$, the group of $E_1/E_2<0.2$ shows good agreement with the classical curve of (6.2) with $C_3 = 19$. But this classical estimation of $\phi_m$ considerably overestimates the measured $\phi_m$-values for $E_1/E_2>0.2$. An empirical expression of $\phi_m$ as a function of both $z/L$ and $E_1/E_2$ is given for $E_1/E_2>0.2$:

$$\phi_m = 1 - \left(-\beta z / L\right)^{1/2}, \quad \text{for } (z/L)_c < z/L < 0, \quad (12.1)$$

where $\beta$ is a wave parameter that varies with $E_1/E_2$, and it is determined empirically from measurements. $(z/L)_c$ is the critical value where $\phi_m$ levels out. For $z/L<(z/L)_c$, the $\phi_m$-curve seems to become independent of the stability parameter $z/L$ and can be described by a constant $C_{\beta}$, which is also dependent on $E_1/E_2$:

$$\phi_m = C_{\beta} \quad \text{for } z/L < (z/L)_c \quad (12.2)$$

The integration form can be derived straightforwardly:

$$\psi_m = 2\left(-\beta z / L\right)^{1/2} \quad \text{for } (z/L)_c < z/L < 0 \quad (13.1)$$

$$\psi_m = (1 - C_{\beta}) \ln\left[z / L\right] \bigg|_{(z/L)_c}^{z/L}, \quad \text{for } z/L < (z/L)_c, \quad (13.2)$$

The quantities $\beta$, $(z/L)_c$, $C_{\beta}$ and specified $\psi_m$-functions of Eq. (13.2) are given in Table IV for mixed sea and swell. Integrating (12.1) requires that $\beta$
is constant with height, which is not fulfilled. How \( \beta \) varies with height beneath 10 m is unknown. This means that the integration of Eq.s (12) is an approximation.

For \( z/L > 0 \), the difference between the three \( E_1/E_2 \)-groups in Figure 5 is not significant. The empirical equation of Holtslag and De Bruin (1988), i.e. Eq. (6.3) with \( b_1 \) now tuned to 1.2 instead of 0.75, seems to be an adequate description of \( \phi_m \) as a function of \( z/L \).

![Figure 5](image)

**Figure 5.** Mean values and one standard deviation of \( \phi_m \) are plotted in \( z/L \)-bins (\( z \approx 10 \) m) in categories of \( E_1/E_2 \). Dashed curves are new \( \phi_m \) functions.

**Table IV.** \( \phi_m \)-function for \( z/L < 0 \) and \( z/L \geq 0 \).

<table>
<thead>
<tr>
<th>( z/L )</th>
<th>( E_1/E_2 )</th>
<th>( \phi_m )-function</th>
<th>( \beta )</th>
<th>( (z/L)_c )</th>
<th>( C_\beta )</th>
<th>Eq. (13.2) for ( z/L &lt; (z/L)_c )</th>
</tr>
</thead>
<tbody>
<tr>
<td>( z/L &lt; 0 )</td>
<td>( E_1/E_2 \leq 0.2 )</td>
<td>MOST</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>( 0.2 \leq E_1/E_2 &lt; 4 )</td>
<td>Eq. (12)</td>
<td>2</td>
<td>-0.5</td>
<td>0</td>
<td>( \psi_a = \ln(z/L) - \ln 0.5 )</td>
<td></td>
</tr>
<tr>
<td>( E_1/E_2 \geq 4 )</td>
<td>Eq. (13.2)</td>
<td>3</td>
<td>-1</td>
<td>-0.73</td>
<td>( \psi_a = (1 + 0.73) \ln(z/L) )</td>
<td></td>
</tr>
<tr>
<td>( z/L \geq 0 )</td>
<td>All ( E_1/E_2 ) ranges</td>
<td>Eq. (6.3) with ( b = 1.2 ).</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
</tbody>
</table>

**Temperature profiles**

The transfer of momentum close to the surface is caused by viscosity and by pressure perturbations. But the transfer of heat is ultimately caused by diffusivity. The temperature profiles do not show similar variation with wave state as the wind profiles, which has also been found by Holland (1981).

The variation of \( \phi_h \) (Eq. (3.2)) at 10 m with stability \( (z/L) \) and wave condition \( (E_1/E_2) \) is explored in Paper III. The result is presented in Figure 6. For \( z/L < 0 \), the overall relation of \( \phi_h \) with \( z/L \) follows the classical prediction of MOST (Högnström 1996). For \( z/L > 0 \), \( \phi_h = 1 + 5.3 z/L \), and Eq. (6.3) with \( b_1 = 1.2 \) are both reasonable descriptions. The scatter of \( \phi_h \) around \( z/L = 0 \) is
artificially induced by the near zero values of heat flux as denominator. The result implies no systematic wave impact on \( \phi_h \). At least at 10 m, the stability function of MOST can still be applied. The analysis of \( \phi_h \) at 17 m and 25 m in Johansson (2003) indicates that the similarity theory is not valid at higher levels.

\[ \phi_h(z/L) \]

Figure 6. \( \phi_h \) is plotted against \( z/L(z\sim10\,m) \) in categories of \( E_1/E_2 \). Classical \( \phi_h-z/L \) relation is also given.

Wave boundary layer height \( h_w \)

The wave boundary height \( h_w \) is defined as the height where the influences from water waves become negligible. However, the influences of waves could be different on different variables, such as pressure perturbation, stress, wind profile, aerosols and heat fluxes. The wave influences on different variables may decay with height at different rates, thus giving different \( h_w \).

The commonly used ‘exponential decay’ rate proportional to the wave number \( k_p \) gives: \( B_{hw} = B_0 \cdot \exp(-ak_ph_w) \), where \( B_0 \) is the wave-induced effect at the surface and \( B_{hw} \) is the effect at \( h_w \). It indicates a larger \( h_w \) for swell (smaller \( k_p \)) than for wind sea condition. Although the decay with height is consistent with measurements, Hare et al. (1997) observed much more complicated profiles of the pressure perturbation than the exponential decay.

For the wave-induced stress, measurements have suggested that its influence is restricted to a few meters during wind-sea conditions (e.g. Drennen et al. 1999a; Paper I). During swell conditions, stress measurements indicate that \( h_w \) can reach at least 26 m (Smedman et al. 1999; Paper I), and sometimes as high as the whole boundary layer (Smedman et al. 1994). Hare et al. (1997) observed the wave influence up to \( kz = 4 \) for \( c_p/u* > 25 \), i.e. the effects of waves can be felt at a height of 30 m if the typical wavelength is 50 m.
4 Methods of obtaining turbulence fluxes

The *eddy-correlation method* is the most accurate method for measuring turbulence fluxes. It is therefore used in the present study as a standard method. The instrumentation is described in Section 2.2.

One way to avoid the difficulties with flow distortion and/or moving platforms (a problem accompanying the eddy-correlation method) is to use the so-called *inertial dissipation method*. But as shown in Sjöblom and Smedman (2003) there are again problems during swell condition, for example, the transport terms in the turbulent kinetic budget must be taken into account.

*Using wind and wave coupled models* is also an alternative to obtain the momentum flux. In Paper II, a wind over wave coupled (WOWC) model (MK2002) is run to estimate the surface stress and to help reveal the impact of swell. This model is built on the conservation of momentum in the surface layer as shown by Eq. (10). Right at the surface, stress includes the viscous stress ($\tau_v$) and wave-induced stress supported by non-separated airflow over non-breaking waves ($\tau_w$) and another part that is supported by airflow separation associated with wave breaking ($\tau_s$). Far above the surface outside the WBL, the impact of waves on the stress vanishes and the stress is supported only by turbulent stress ($\rho u^2$). A model of wave spectrum developed by Kudryavtsev et al. (1999) is used for the high wave number range $k>>k_p$. The lower wave number part of the spectrum, $k_p<k<~10k_p$, is described by the empirical model of Donelan et al. (1985). For swell cases, the model of Donelan et al. is not valid and therefore measurements of wave spectra are used. The model only deals with the wind-sea part of the swell dominated cases $S(n>n_1)$ and therefore assumes no influence from the pure swell part. The discrepancy between model and measurements in this case will clearly reflect the difference in impact on the stress given by a pure wind-sea and by swell.

The *bulk method* is based on a model that describes the physical processes in the interfacial layers of ocean and atmosphere. One example is the ‘*surface renewal theory*’ postulated by Brutsaert (1965) and its developed versions (e.g. Liu et al. 1979; Fairall et al. 1996b; Clayson et al. 1996). This kind of model calculates the surface roughness lengths ($z_0$ and $z_{or}$) and then uses
MOST to estimate surface fluxes. It describes the interaction between sub-layers and the surface layer. Therefore it is only valid for generally low wind speed conditions when a sublayer exists and the effect of sea spray is negligible.

Another choice of bulk method is to relate the surface fluxes to the mean variables via the exchange coefficients: $C_D$, the drag coefficient for momentum, $C_H$ the Stanton number for sensible heat:

\[
\left( \frac{\overline{u'u'}}{U^2} + \frac{\overline{v'v'}}{U^2} \right)^{1/2} = C_D U^2
\]  

(14.1)

\[
\overline{w'\theta} = C_H U (\theta_* - \theta),
\]  

(14.2)

These coefficients can be determined from Eqs (14) with the help of measurements. If Eqs (4) are valid, combining Eqs (14) and (4) leads to $C_D = f(z, z_0, \psi_m)$ and $C_H = f(z, z_0, z_0 T, \psi_m, \psi_h)$:

\[
C_D = \left[ \kappa / (\ln(z / z_0) - \psi_m) \right]^2
\]  

(15.1)

\[
C_H = \left[ \kappa / (\ln(z / z_0) - \psi_m) \right] \cdot \left[ \kappa / (\ln(z / z_{0T}) - \psi_h) \right]
\]  

(15.2)

In practice, $C_D$ and $C_H$ are usually corrected to a reference height of 10 m and to neutral stability, so Eqs (15) become:

\[
C_{DN} = \left[ \kappa / \ln(z / z_0) \right]^2
\]  

(16.1)

\[
C_{HN} = \left[ \kappa / \ln(z / z_0) \right] \cdot \left[ \kappa / \ln(z / z_{0T}) \right]
\]  

(16.2)

where $z \sim 10$ m. Hereinafter, $C_D$, $C_H$, $C_{DN}$ and $C_{HN}$ are referred to a height of 10 m.

It is important to study how waves influence the transfer of momentum and heat when estimating the surface turbulent fluxes with the bulk method.
5 Wave impact on drag

5.1 Pure wind-sea

Over pure wind-sea when the flow is smooth, roughness length $z_0$ is determined by the viscousity. When the surface is dominated by wind-driven waves, which are in equilibrium with the wind, wind speed is expected to describe the variation of $z_0$ or $C_{DW}$. This is in consistence with the idea of Charnock (1955). Charnock’s idea that stress is governed by short gravity waves leads to a strict wind speed or friction velocity dependent roughness:

$$z_0 = \alpha_{ch} \frac{u^2}{g}$$  \hspace{1cm} (17.1)

where $\alpha_{ch}$ is the Charnock’s parameter. Eq. (17.1) is often used in large-scale synoptic and climatic models with a constant value of $\alpha_{ch}$. Thus with (17.1) substituted into (16.1), $C_D$ can actually be expressed as an increasing function of $U_{10}$ only. However, it is not the wind itself but the wind-driven waves that determine the roughness of the surface. $\alpha_{ch}$ has also been found to vary with the inverse wave age $u/c_p$ in many field experiments (Maat et al. 1991; Smith et al. 1992; Komen et al. 1998; Drennan et al. 2002), lab experiments and models (Komen et al. 1998):

$$\alpha_{ch} = C \left( \frac{u_*}{c_p} \right)^D$$  \hspace{1cm} (17.2)

where $C$ and $D$ are empirical constants. For very young tank waves $u/c_p > 0.2$, it is found that $C > 0$ and $D < 0$. In the field experiments the inverse wave age is much smaller than 0.2 and it is found that $C > 0$ and $D > 0$.

In Paper I it is shown that for pure wind-sea, deep water and rough flow conditions, $z_0$ scales with the surface elevation $\sigma$ and $z_0/\sigma$ increases with $u/c_p$. The result is presented in Figure 7 and it is in reasonable agreement.

---

1 The roughness of flow is defined according to the roughness Reynold’s number $Re = z_0 u_*/\nu$. The flow is defined as smooth when $Re < Re_1$ and it is defined as rough when $Re > Re_2$. The usually taken values of $Re_1$ and $Re_2$ are 0.11 and 2.3 (e.g. Donelan 1990, Drennan et al. 2002). Flow with $Re_1 < Re < Re_2$ is transitional.
with the results of Donelan (1990) and Drennan et al. (2002). The surface elevation is defined as

$$\sigma = \left( \int S(n) \, dn \right)^{1/2}, \quad (18)$$

where \( S(n) \) is the energy density of the wave spectrum as a function of frequency \( n \).

**Figure 7.** \( \sigma/\sigma_p \) plotted as a function of \( u^*/\sigma_p \) for pure wind sea, rough flow and deep water conditions, grouped in \( u^* \)-bins. Also presented are Donelan (1990) and Drennan et al. (2002) curves and current regression curve for the whole data set.

With \( \alpha_{ch} \) given as a function of inverse wave age (17.2) and the definition of \( C_D \) (14.1) substituted into (16.1), \( C_D \) can be represented as a slightly non-linear increasing function of \( U_{10} \) at different values of \( u^*/\sigma_p \). However for waves that are in equilibrium with wind, \( U_{10} \) and \( u^*/\sigma_p \) are inter-related.\(^2\) Using the extended Charnock’s formula is therefore equivalent to

$$C_D = f(u^*/\sigma_p) \quad (19)$$

\( C_D \) as a function of \( u^*/\sigma_p \) for pure wind-sea is plotted in Figure 8 (squares) (from Paper I). In brief, when using Eq.s (17) and (19) the following conditions have to be satisfied:

- Rough flow,

\(^2\) When plotting \( C_{DN} \) as a function of \( U_{10} \) there is always a possibility that the relationship between \( C_{DN} \) and \( U_{10} \) is contaminated by spurious ‘self-correlation’ because \( 1/U_{10}^2 \) is plotted against \( U_{10} \). However the increasing function of \( C_{DN} \) with \( U_{10} \) is free of ‘self-correlation’ because of the following fact: randomizing \( U_{10} \) (or randomizing both \( U_{10} \) and \( u^* \)) leads to decreasing trend of \( C_{DN,rand} \) with \( U_{10,rand} \). Subscript ‘rand’ means that \( U_{10} \) (or both \( U_{10} \) and \( u^* \)) is randomized in the calculation. Spurious correlation may also occur when \( C_{DN} \) is plotted against \( u^*/\sigma_p \), because \( u^* \) is plotted against \( u^* \). However, it has been shown in Paper I that the explained variance of the measurements is much larger than in the stochastic data set.
• The MOST is valid,
• The sea surface is dominated by wind-driven waves that are in equilibrium with wind,
• Neutral stability.

During pure wind-sea conditions, wind speed is usually high, so wave breaking is an ever-present phenomenon. In theory, more wave parameters than only \( u*/c_p \) are needed to describe the wave state. The WOWC model of MK2002 in Paper II indicates that for a fully developed sea at high winds, stress supported from waves of the equilibrium range of the wave spectrum contributes up to 50% of the total stress right at the surface. For younger waves, stress supported from waves at the spectral peak also becomes significant. In Paper II for pure wind-sea, the modeled stress is in excellent agreement with our measurements, but both deviate from that derived from a linear relationship of \( C_{DN} \) and \( U_{10} \). The higher values of \( C_{DN} \) at high winds might be an effect of wave breaking.

5.2 Mixed sea and following swell

Consider a simple wave field where long waves become more and more dominant, the local wind becomes gradually less important for the variation of drag. This is reflected in many experiments with open ocean conditions where, at light to moderate winds, the distribution of \( C_{DN} \) with \( U_{10} \) is dominated by scatter (e.g. review by Geerneart 1990; Paper I & II).

In Paper II at relatively light winds for swell cases, there is considerable discrepancy between the WOWC model and measurements. The model overestimates the stress. This reveals the fact that there is swell impact on the stress and the mechanism of wave impact on stress during swell conditions is different from that during pure wind-sea conditions.

When wind sea gives place to long waves, which do not have direct relation to wind speed, Paper I shows that \( C_{DN} \) varies both with \( u*/c_p \) and with \( E_1/E_2 \), so that

\[
C_{DN} = f(u*/c_p, E_1/E_2)
\]  

(20)

This is presented in Figure 8 in which \( C_D \) is plotted against \( u*/c_p \) in categories of \( E_1/E_2 \). As long waves become more and more dominant, i.e. \( E_1/E_2 \) increases, the dependence of \( C_D \) on \( u*/c_p \) becomes weaker and weaker. Eventually when \( E_1/E_2 \geq 4 \), \( C_D \) does not show any systematic dependence on

\[\text{In Paper I, } C_D \text{ is calculated with data of neutral stratification, so that stability correction is not needed, and for these data } C_D = C_{DN}.\]
Thus the function (20) is valid for pure wind-sea and mixed sea, but not for pure swell conditions.

Figure 8. Drag coefficient $C_D$ times $10^3$ plotted against $u^*/c_p$ in 5-bins of $E_1/E_2$.

Since the log-wind law is not valid during swell condition, $z_0$ calculated with Eq. (16.1) does not have the original physical meaning of roughness length and therefore it is in Paper I referred to as the apparent roughness length $z_0a$. The variation of $z_0a$ with $u^*/c_p$ and $E_1/E_2$ is similar to that of $C_{DN}$. Linear relations between $\ln(z_0a)$ and $\ln(u^*/c_p)$ may be obtained for pure wind-sea and mixed sea; see Figure 9a:

\[
\ln z_{0a} = 6.56 \ln(u^*/c_p) + 10.74 \quad E_1/E_2 < 0.2 \\
\ln z_{0a} = 6.17 \ln(u^*/c_p) + 10.98 \quad 0.2 \leq E_1/E_2 < 4
\]  

(21)

For swell ($\Diamond$ in Figure 9a), there is no statistical dependence of $z_{0a}$ on $u^*/c_p$.

One can say that Eq. (21) is a further extended Charnock’s formula, with $C$ and $D$ in (17.2) varying with $E_1/E_2$. Figure 9b shows the comparison between $z_0$ calculated from (21) and $z_0$ calculated from (17.1) and (17.2) with $C=1.59$ and $D=1.67$ (Drennen et al., 2002). The agreement is excellent for the pure wind-sea group. Discrepancy appears for mixed sea, suggesting that $C$ and $D$ need to be adjusted according to the value of $E_1/E_2$.

In both Paper I and Paper II, there are cases with surprisingly high values of $C_{DN}$ during pure following-swell conditions. There are two possible sources for this. One seems to be the history of the wind and wave fields, both of speed and direction. For a great deal of time, the sea surface is covered by waves in different developing stages. Wind that varies in both magnitude and direction drives waves from different directions with different forces. The interaction between wind sea, swell and atmosphere leads to complicated variation of $C_{DN}$. The other source can be detected from
the co-spectra of $uw$ and $vw$. The characteristic feature during swell is that there is low frequency stochastic variations of large amplitude, which could either be positive or negative. At the same time, during swell condition, the ‘ordinary’ stress in the high frequency region is of small magnitude, which makes the calculation of stress very sensitive to the cutoff frequency. Drennen et al. (1999a) found that the spectra and co-spectra during swell conditions are reduced compared to pure wind-sea runs. Although for a large amount of data the fluctuation in the low frequency range will lead to nothing but scatter.

Figure 9. (a) $\ln \left( \frac{z_0a}{\epsilon_u} \right)$ as function of $\ln \left( \frac{u^*}{c_p} \right)$ in 3-bin of $E_1/E_2$. (b) Comparison between $z_{0a}$ (estimated from Eq. (21)) and $z_{0ch}$ (estimated from Charnock’s formula with C and D from Drennan et al. 2002); for $E_1/E_2 \geq 4$, $z_0a$ is calculated from the dashed curve in (a).

5.3 Mixed sea and cross swell

Measurements have shown that quite often the crosswind component of the stress $\tau_{wx} = -\rho \bar{v} \bar{w}$ is not zero. This leads to a deviation of stress direction from wind direction:

$$\theta = \arctan \left( \frac{\bar{v} \bar{w}}{\bar{u} \bar{w}} \right) \quad (22)$$

Possible causes of this given in literature are the coastal jet (Zemba and Friese 1987), stratification (Geernaert 1988), and swell from directions different from that of the local wind (Geernaert et al. 1993; Rieder et al. 1994, 1996; Grachev et al. 2003 etc). The stratification impact could never cause a deviation larger than 5° according to Geernaert’s (1988) theory. The focus here is on the swell impact.

The impact of swell on the stress direction is comprehensively discussed in Grachev et al. (2003) for different wind and wave conditions. One key point emphasized by Grachev et al. (2003) is that $\tau_{wx}$ could be positive as
well as negative, depending on the wind and wave conditions. Accordingly, the swell-induced stress and the swell could be in the same or opposite directions, and the total momentum flux could either be downward or upward. The upward momentum flux (from water to air) has been observed by e.g. Smedman et al. (1994), Drennan et al. (1999a) and Grachev et al. (2001, 2003) during light winds and strong swell.

In the WBL, the total stress can be decomposed, so that one component is aligned with the wind ($\tau_1$) and one component is aligned with the swell ($\tau_2$), see Figure 10. In Figure 10a (case-a), $\tau_1$ and swell are in the same direction. In Figure 10b (case-b), $\tau_2$ and swell are in opposite direction. According to Grachev et al. (2003) case-a is supposed to be representative of ‘moderate to strong winds and weak swell’, and the stress direction is between wind and swell directions. This has also been observed by Geernaert et al. (1993) and Rieder et al. (1994; 1996). According to Grachev et al., the case-b is supposed to be representative of ‘light wind and strong background swell’.

![Figure 10](image)

**Figure 10.** Stress vector orientation, total stress $\tau$, stress align with wind $\tau_1$ and stress align with swell $\tau_2$. (a) Case-a, (b) Case-b.

According to Figure 10, the relationships between the total stress and the two stress components for case-a and –b can be written as:

\[
\tau = \left(\tau_1^2 + \tau_2^2 + 2\tau_1\tau_2 \cos \alpha\right)^{1/2}
\]

for case-a \hspace{1cm} (23.1)

\[
\tau = \left(\tau_1^2 + \tau_2^2 + 2\tau_1\tau_2 \cos(180 - \alpha)\right)^{1/2}
\]

for case-b \hspace{1cm} (23.2)

For pure wind sea, $\tau_2 = 0$, the total stress $\tau_{pws}$ equals

\[
\tau = \tau_{pws} = \tau_1
\]

(24.1)

For following swell/mixed sea, $\alpha \approx 0$, so that the total stress $\tau_1$ for case-a and –b, according to Eq. (23.1) and Eq. (23.2), can be expressed as:
\[ \tau = \tau_{fs} = \tau_1 + \tau_2 \quad \text{for case-a} \quad (24.2) \]
\[ \tau = \tau_{fs} = |\tau_1 - \tau_2| \quad \text{for case-b} \quad (24.3) \]

The total stress for cross swell/mixed sea, \( \tau_{cs} \), is identical to Eq. (23.1) and (23.2) for case-a and -b respectively.

For case-a, \( \tau \) decreases with increasing \( \alpha \), Eq.s (23.1), (24.1) and (24.2) imply
\[ \tau_{pws} < \tau_{fs} \quad \text{and} \quad \tau_{cs} < \tau_{fs} \quad (25.1) \]

For case-b, \( \tau \) increases with \( \alpha \), Eq.s (23.2), (24.1) and (24.3) suggest
\[ \tau_{fs} < \tau_{cs} < \tau_{pws} \quad (25.2) \]

The inequality (25.2) tells that during swell condition at relatively light winds the total stress is smaller than that of a pure wind sea, and the stress of wind-cross-wave cases is enhanced compared to wind-following-wave cases.

The enhanced drag for cross-swell condition at light to moderate winds has been observed in Paper II. The variation of the drag coefficient with wind speed is reproduced in Figure 11a. For \( U_{10}>10 \text{ m s}^{-1} \) (an approximation of case-a), the relation given by (25.1) can not be seen, probably because the magnitude of \( \tau_2 \) is too small or because there are too few swell data for \( U_{10}>12 \text{ m s}^{-1} \), or both. For \( U_{10}<10 \text{ m s}^{-1} \) (an approximation of case-b), the inequality (25.2) is well presented.

For case-b, Eq. (23.2) together with Figure 10b indicate that the total stress increases with \( \alpha \). In Figure 11b, the cross-swell data in subplot-a are selected and divided into groups of \( \alpha \), see the legend. Figure 11b clearly shows the increasing of drag with \( \alpha \) — the enhancement of drag by the wave-cross-wind effect. Enhanced drag for cross/counter swell conditions have also been observed by e.g. Donelan et al. (1997) and Drennan et al. (1999a).

5.4 Non-stationary wind field

The impact of flow acceleration/deceleration is expected to give larger/smaller drag coefficients for a given wind speed since flow acceleration is usually accompanied by developing young waves, as observed by e.g. Smith (1980), while flow deceleration is usually accompanied by mature sea and swell, as observed by Smedman et al. (1999). For the smaller drag observed during a wind decay, Drennan et al. (1999b) conclude that “it is unclear whether this difference is due primarily
to the presence of following swells, the decay of the wind field, or both”. Vickers and Mahrt (1997) show no significant difference between accelerating and decelerating runs. They conclude that “this is probably due to the fact that, for these data, flow acceleration only occasionally determines the wave state”. In other words, the wave-state is considered as the more direct cause of variation of the drag coefficient.

In Paper II, three cases of ‘non-stationary’ wind fields are selected, pure wind-sea, mixed sea/ following swell and mixed sea/ cross swell conditions respectively. Wind variation (direction and magnitude) during one hour is comparable for pure wind sea (acceleration) and cross-swell (deceleration) cases. The results show that for the pure wind sea case, $u_*$ varies in pace with $U_{10}$, giving slight increase of $C_D$ with $U_{10}$. For the two swell cases, $u_*$ remains more or less constant when the wind varies substantially, showing that the relation between $u_*$ and $U_{10}$ is weak, and $C_D$ fluctuates. The WOWC model gives accurate prediction of the drag of ‘non-stationary’ pure wind-sea case, but fails to capture the variation of drag during ‘non-stationary’ swell conditions. It seems that both the wind speed variation and the magnitude of acceleration/deceleration explain very little of the difference in the variation of $C_D$ for swell periods. The interaction between swell, wind waves and the atmosphere, the mechanisms of which are not yet clear, leads to the complicated variation of $C_D$.

Figure 11. $C_D$ as a function of $U_{10}$ (m s$^{-1}$). (a) Mean values of $C_D$ as a function of mean values of $U_{10}$ in $U_{10}$-bins of 2 m s$^{-1}$ for three wave conditions. Error-bars are the confidence intervals of the mean values of $C_D$ (see Paper II for the calculation). (b). Mean values of $C_D$ against $U_{10}$ for cross-swell in categories of $\alpha$. 

In Paper II, three cases of ‘non-stationary’ wind fields are selected, pure wind-sea, mixed sea/ following swell and mixed sea/ cross swell conditions respectively. Wind variation (direction and magnitude) during one hour is comparable for pure wind sea (acceleration) and cross-swell (deceleration) cases. The results show that for the pure wind sea case, $u_*$ varies in pace with $U_{10}$, giving slight increase of $C_D$ with $U_{10}$. For the two swell cases, $u_*$ remains more or less constant when the wind varies substantially, showing that the relation between $u_*$ and $U_{10}$ is weak, and $C_D$ fluctuates. The WOWC model gives accurate prediction of the drag of ‘non-stationary’ pure wind-sea case, but fails to capture the variation of drag during ‘non-stationary’ swell conditions. It seems that both the wind speed variation and the magnitude of acceleration/deceleration explain very little of the difference in the variation of $C_D$ for swell periods. The interaction between swell, wind waves and the atmosphere, the mechanisms of which are not yet clear, leads to the complicated variation of $C_D$. 

25
6 Sensible heat transfer

At a reference height of 10 m, the sensible heat transfer coefficient at neutrality, $C_{HN}$, is a function of $z_0$ and $z_{0T}$, according to Eq. (16.2). $z_{0T}$ can be obtained from Eq. (4.2) with input of $\theta_s$, $\theta_{10}$, $\bar{w} \theta$, $u^*$ and the stability function for temperature $\psi_h$. In Section 3.3 it has been shown that $\psi_h$ developed over land can be considered valid. $z_0$ is calculated from Eq. (4.1) with input of $U_{10}$, $u^*$ and the stability function for wind speed $\psi_m$. For mixed sea and swell conditions, new $\psi_m$ expressions, i.e. Eq.s (13) are used.

In the literature, the variation of $z_{0T}$ is quite confusing. Theoretical arguments (surface-renewal theory) and laboratory data suggest that $z_{0T}$ decreases with increasing wind speed (e.g. Brutsaert et al. 1975; Liu et al. 1979; Fairall et al. 1996b; Clayson et al. 1996; Zeng et al. 1998). Field measurements of $z_{0T}$ show large scatter with no clear trend (e.g. Large and Pond 1982; Dupuis et al. 1995; DeCosmo et al. 1996). Another often observed feature is that $z_{0T}$ values are larger for unstable conditions than for stable conditions (Smith 1980; Large and Pond 1982; Oost et al. 2000).

Field experiments show that $C_{HN}$ appears to be constant, being around $1.1 \cdot 10^{-3}$ during unstable conditions, and smaller during stable conditions. On the other hand, studies based on surface-renewal theory and laboratory data predict that $C_{HN}$ should decrease at high wind speed. Fairall et al. (1996b), Clayson et al. (1996) and Zeng et al. (1998) tested the predictions on data from the tropical ocean and claim good agreement with theory. However, in these studies there are no data with wind speed above 10 m s$^{-1}$.

In Paper III, for the entire unstable range $z_{0T}$ is found to be a constant on the order of $10^{-5}$ m. As $z/L$ changes from small negative to small positive values, $z_{0T}$ values drop dramatically many orders of magnitude. For $z/L>0.2$, $z_{0T}$ values seem to go back to the typical values for unstable conditions. No clear dependence of $z_{0T}$ on wind speed has been found. The unstable cases are in agreement with the prediction of Liu et al. (1979) for $U_{10}<8$ m s$^{-1}$, but deviate systematically with increasing wind speed.

In Paper III (its figure 7), around neutral stability $z_0$ shows a flat maximum, with high values extending well from unstable into stable conditions. Accordingly, there appears to be two striking features in the plot of $C_{HN}$ against stability $z/L$ given in Figure 12a: (i) a significant change of mean value for $C_{HN}$ at from slightly unstable to slightly stable condition; (ii)
a clear division of the unstable data in one rising and one descending branch as \( z/L \) approaches zero. The rising branch with high \( C_{HN} \)-values corresponds to high wind speed and relatively young waves \( (c_p/u^*<30) \), whereas the other branch corresponds to lower wind speed and mature sea \( (30<c_p/u^*<50) \). Figure 12b shows that the two ‘theoretical’ estimates of Fairall et al. (1996b) and Zeng et al. (1998) describe well our unstable data up to about 10 m s\(^{-1}\). With increasing wind speed our data deviate to an increasing extent relative to their estimates, giving \(~30\%\) higher values at 14 m s\(^{-1}\) and more than 50\% higher values at 16 m s\(^{-1}\). In the range of 14 –16 m s\(^{-1}\) as stability goes from slightly unstable to slightly stable, \( C_{HN} \) changes dramatically from about \( 1.6 \times 10^{-3} \) to \( 0.5 \times 10^{-3} \) (Figure 12c).

Andreas and DeCosmo (2002) reanalyzed the HEXOS data reported in DeCosmo et al. (1996). They assume that the surface-renewal theory of Liu et al. (1979) and of Zeng et al. (1998) is basically correct and derive spray-mediated contribution to the fluxes of latent and sensible heat. They find that at high winds the spray contribution is ‘about 10\% of the total flux’ and ‘20-30\% of the flux in 10 of the 14 highest wind speed runs’. In addition to wind speed, other factors of the spray-mediated sensible heat flux are sea surface temperature, relative humidity and air temperature close to the water surface.

The ‘spray effect’ would bring higher values of \( C_{HN} \) for unstable condition but smaller values of it for stable condition. It seems to be a reasonable assumption that the observed deviation of our data from the theoretical estimates at high winds is a result of ‘spray effect’. Stable cases from Nässkär illustrate that low-level jets can also influence the variation of \( z_0T \) and \( C_{HN} \). In Smedman et al. (1997) it is discussed that the sensible heat exchange was strongly suppressed of the presence of low-level jets.

![Figure 12](image-url)  

**Figure 12.** \( C_{HN} \) as a function of \( z/L \) in a limited range (a); as a function of \( U_{10N} \) for unstable and stable conditions (b) and (c). Legend shows the source of data.
7 Summary and conclusions

Long term semi-continuous meteorological measurements from an air-sea interaction station and wave data from a Wave-Rider Buoy are used to study the air-sea exchange of momentum and sensible heat in the marine atmospheric boundary layer. Based on the resemblance of our measurements to those made over the global oceans, this station can be relied on to give results representative of open ocean conditions in most cases.

The wave impact on the air-sea exchange of momentum is explored during neutral conditions for various wave states, namely, pure wind-sea, mixed-sea and swell (including wind following mixed-sea and swell, as well as wind cross mixed-sea and swell). A wind over wave coupled model of MK2002 is used as a tool.

It is found that during pure wind-sea conditions, the wave influence remains at a height lower than our first measuring level and the wind profile is logarithmic during neutral stratification. Roughness length $z_0$ resembles that of a solid rigid surface, it scales with surface elevation and increases with inverse wave age ($u*/c_p$). For these wind waves, which are in equilibrium with wind, the drag coefficient of neutral stability $C_{DN}$ can be estimated from the inverse wave age.

When wind waves are replaced by older waves, the wave influence can extend up to or even above the measuring levels, and the logarithmic wind law is not valid any more during neutral conditions. As long waves become more and more dominant, the slope of wind profile ($8-13$ m) decreases and sometimes can be negative, indicating a wind maximum below $8$ m. Accordingly, in addition to stratification ($z/L$), the normalized wind gradient $\phi_m$ is also a function of $E_1/E_2$, giving smaller values for longer waves.

The presence of swell changes the simple dependence of $C_{DN}$ on $u*/c_p$. As $E_1/E_2$ increases, the dependence of $C_{DN}$ on $u*/c_p$ becomes less and less significant. $C_{DN}$ is dependent on not only $u*/c_p$ but also $E_1/E_2$.

When the surface is dominated by pure swell ($E_1/E_2 > 4$), conventional parameters, such as $u*/c_p$, $U_{10}$ and $U_{10}/c_p$, all become irrelevant for $C_{DN}$. The interaction between swell, wind waves and the atmosphere, a mechanism of which is not yet clear, leads to a complicated variation of $C_{DN}$. Sudden turn of the wind could result in large fluctuation of $C_{DN}$ values. The wind-cross-swell effect is found to enhance the sea drag at light to moderate winds.
During swell conditions, the variation of $C_{DN}$ could also be related to mesoscale variability in the atmospheric boundary layer as revealed as the low frequency fluctuation in the stress co-spectra.

On the other hand, controlled by a total different mechanism, neither potential temperature profile nor the sensible heat exchange coefficient $C_{HN}$ demonstrates similar wave age dependence as wind speed profile and $C_{DN}$.

For unstable cases, the values of $C_{HN}$ were found to follow the predictions from the surface-renewal theory quite well up to a wind speed of about 10 m s$^{-1}$ but about 30% higher at 14 m s$^{-1}$ and about 50% higher at 16 m s$^{-1}$. At 15 m s$^{-1}$ the value of $C_{HN}$ is about $1.5 \cdot 10^{-3}$ for $z/L<0$ but $0.5 \cdot 10^{-3}$ for $z/L>0$. This is interpreted as an effect of spray, which becomes significant when wind speed is larger than about 10 m s$^{-1}$. Due to the spray-mediated sensible heat flux, the total sensible heat is expected to be larger during unstable conditions but smaller during stable conditions.

The average value of $C_{HN}$ for stable conditions is smaller than that for unstable conditions, partly due to the spray effect. It is also demonstrated that part of the strong suppression of the sensible heat flux can be explained as an effect caused by the presence of a wind maximum at a low height.
8 Acknowledgement

The Uppsala early spring comes for the fifth time just as I am writing the final words on this thesis.

First I would like to thank my supervisor, Prof. Ann-Sofi Smedman, for the opportunity to start my Ph D study here. Thank you, for years of ideas, discussions and encouragement. The cooperation with Prof. em. Ulf Högström has been joyful – His knowledge and enthusiasm for science has truly been an inspiration, something Ph D. students can look up to.

I will not forget Dr. Hans Bergström’s continuous help with using and analyzing data. Also, I am indebted to all those who are involved in obtaining these data. To name a few; the MIUU staff for collecting data at Östergarnsholm and Nässkär, Dr. Kimmo Kahma and Dr. Heidi Pettersson at FIMR for providing the wave measurements, and ‘Captain’ Ingvar Östman for his hospitality.

Fortunately, I have had the opportunity to work with Dr. Vladimir Makina very inspiring and educating cooperation. Thank you, Vladimir, for teaching me how a scientist faces and deals scientific problems.

I would acknowledge two projects that have supported my study: SFINCS (ENV4-CT97-0573) and AUTOFLUX (MAS3-CT97-0108).

The last five years would not be as colorful and easy without the colleagues and friends in Air and Water Science. Dr Anna Sjöblom and Dr Cecilia Johansson who are always around when I need help, the warm heart of Dr. Anna Rutgersson I will always feel, Babatunde, who needs 25 hours a day, but still has time for my interruptions – thank you all! Karin, thank you for sharing the same window and the same computer noise with me, and for the encouragement when I needed it. These years are full of warm memories of Birgitta, Mikael, Matthias, Erik, Elin, Linda…

I would also like to thank my friends Zouxiang, Ulrica, Xuchongyu, Åse…for introducing Uppsala to me and making it a very friendly place, and our friends from Västgötland for making me feel at home in Sweden.

Thank my families both in China and in Sweden for giving me love and strength, yet without the support and care from you, Sampe, nothing would have happened.
9 Reference:


Anisimova, E.P., G.E. Kononkova, V.V. Kusnetsov, A.S. orlov, G.I. Popov and A.A. Speranskaya, 1974: ‘Wind-wave generation and the wind velocity structure in the air above a wavy water surface’, Boundary−Layer Meteorology, 6, 5-11


